Greenhouse-Gas-Induced Climatic Change: A Critical Appraisal of Simulations and Observations

edited by

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Preface

This book is the culmination of a Workshop on Greenhouse-Gas-Induced Climatic Change: A Critical Appraisal of Simulations and Observations which was held at the University of Massachusetts, Amherst, during 8-12 May 1989. The objectives of the Workshop were to: (1) present and evaluate the current status of climate model simulations of greenhouse-gas-induced changes of both the equilibrium and nonequilibrium (transient) climates; (2) present and assess the current status of the observations of global and regional climates from the beginning of the industrial revolution to the present, circa 1850 to 1989; (3) present reconstructions of climatic change during the last millennium to determine the "natural variability" of climate on the intra-century time scale; (4) critically evaluate whether or not the climate has changed from circa 1850 to 1989; and (5) compare the observations with the model simulations to ascertain whether a greenhouse-gas-induced climatic change has occurred and, if not, to estimate when in the future such a climatic change will likely become detectable against the background of the "natural variability."

Acknowledgements

I thank Ray Bradley, Henry Diaz and Tom Karl for their guidance and assistance as members of the Workshop Organizing Committee. I particularly thank Ray Bradley for all his efforts as Local Organizer to make the Workshop a success. I also thank Deborah Salkaus of University Conference Services, University of Massachusetts, Amherst, for her help with the arrangements for the Workshop. I thank Tim Barnett, Mike MacCracken and Kevin Trenberth for their efforts as chairmen of the Working Groups whose reports appear herein. I express my gratitude to Susan McKinney for her efforts in the preparation of this book. The Workshop was sponsored by the Carbon Dioxide Research Program – Atmospheric and Climate Research Division, Office of Health and Environmental Research of the U.S. Department of Energy.

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Part 1:

Model Validation: How Good are the Models in Simulating the Present and Past Climates?
A Statistical Intercomparison of Temperature and Precipitation Predicted by Four General Circulation Models with Historical Data

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ABSTRACT. This study is a detailed intercomparison of the results produced by four general circulation models (GCMs) that have been used to estimate the climatic consequences of a doubling of the CO₂ concentration. Two variables, surface air temperature and precipitation, annually and seasonally averaged, are compared for both the current climate and for the predicted equilibrium changes after a doubling of the atmospheric CO₂ concentration. The major question considered here is: "How well do the predictions from different GCMs agree with each other and with historical climatology over different areal extents, from the global scale down to the range of only several gridpoints?"

Although the models often agree well when estimating averages over large areas, substantial disagreements become apparent as the spatial scale is reduced. At scales below continental, the correlations observed between different model predictions are often very poor. The implications of this work for investigation of climatic impacts on a regional scale are profound. For these two important variables, at least, the poor agreement between model simulations of the current climate on the regional scale calls into question the ability of these models to quantitatively estimate future climatic change on anything approaching the scale of a few (< 10) gridpoints, which is essential if these results are to be used in meaningful resource-assessment studies. A stronger cooperative effort among the different modelling groups will be necessary to assure that we are getting better agreement for the right reasons, a prerequisite for improving confidence in model projections.

1. INTRODUCTION

Recent studies with each of the four GCMs being used for CO₂ sensitivity studies in the United States have projected an equilibrium global-average annual surface warming of 3 to 4.5°C for a doubling of the CO₂ concentration. While agreement of global-scale averages is a necessary and encouraging first step, to address environmentally important
Issues one must examine how well current models represent “reality” over much more limited regions. To more quantitatively determine the degree to which model results are converging on these finer scales, model intercomparisons of regional and seasonal climatic change have been carried out. Because the separate modeling groups are all attempting to carry out the same numerical experiment, albeit with models that are slightly different, such convergence would be expected to be good, although not perfect.

In this study the results generated by four GCMs are intercompared. To avoid confusion in describing these results, they will be referred to using the commonly accepted acronyms for their institutional names: (1) CCM (Washington and Meehl, 1984), (2) GFDL (Manabe and Wetherald, 1987), (3) GISS (Hansen et al., 1984), and (4) OSU (Schlesinger and Zhao, 1989). Only two of the many fields produced by the GCMs are considered here, and only at the surface: (1) average seasonal surface air temperature (°C) and (2) precipitation (mm/day). For each of these fields results are calculated for a control climate, $X(1\times CO_2)$, and the predicted changes after a doubling of CO2, $\Delta X = X(2\times CO_2) - X(1\times CO_2)$ [for $X = T$ = temperature or $X = Pr$ = precipitation].

In the comparison with the current climate, the gridded (4° latitude x 5° longitude) global surface air temperature data set assembled by Oort (1983), covering the years 1958–1973, is used. A comparison of the Oort data with an earlier, but similar, compilation by Schutz and Gates (1971, 1972) gives an average absolute temperature difference of less than 0.7°C for all gridpoints in both the December–January–February (DJF) and June–July–August (JJA) seasons. To facilitate further, more detailed regional intercomparisons, the GCM results for surface air temperature for the CCM, GFDL and GISS models were interpolated using bicubic splines to the same 4°x5° grid used by the OSU model and Oort data.

For seasonal precipitation the data set of Jaeger (1976) is used. Because there are significant uncertainties about precipitation amounts over the oceans, the comparison emphasizes land points. However, even with this restriction, small-scale variations in precipitation are very large and probably only the qualitative characteristics of the modelled and observed distributions can be compared.

This study should not be construed as a “beauty contest” among models in which one model is in some sense “better than any other.” The fact that the DJF seasonally averaged surface air temperatures predicted by model X are closer to the historical observations than model Y over region Z does not mean that model X is “better” than model Y over region Z (let alone over region W or for JJA in region Z!) More details on the topics covered in this report may be found in Grotch (1988).

2. INTERCOMPARISON OF CONTROL (1×CO2) SURFACE AIR TEMPERATURE OVER DIFFERENT REGIONAL SCALES

It must be strongly emphasized at the outset that much of the GCM literature has focused on the use of large-scale averages as the primary measure of model agreement/disagreement. Although agreement of the average is clearly a necessary condition for model validation, even when averages agree perfectly, very much larger regional or pointwise differences can, and do, exist in practice. Thus, the spatial distributions of pointwise differences will be examined here as the regional extent is reduced from global to hemispheric to zonal to continental, and finally to small regions comprising only a few gridpoints.

Starting on the global scale, the seasonal surface air temperatures of the four GCMs show good agreement with Oort’s historical data for the averages: the average absolute temperature difference between the four GCMs and Oort’s global average is < 0.9°C for both DJF and JJA seasons. However, because ocean temperatures are established in
different ways in these models, a more sensitive test considers only land gridpoints. Here, the corresponding average absolute seasonal, global differences are higher: 1.7°C in DJF and 2.8°C in JJA.

Given model and historical data on a common grid, it is possible to examine pointwise differences between model predictions and historical data over different regional extents. Using the four GCMs and considering land gridpoints only, the average global area-weighted absolute differences [GCM-Oort] range from 3.1°C to 5.9°C for DJF and from 2.3°C to 5.1°C for JJA. With the second historical data set (Schutz and Gates, 1971, 1972) the corresponding seasonal-average absolute differences are noticeably smaller, 1.3°C and 1.1°C.

The latitudinal distributions of the pointwise differences (GCM-Oort) are displayed using boxplots (Tukey, 1977) in Fig. 1 for the DJF season. Each boxplot represents the quartiles of the 72-point differences at a given latitude. For these GCMs the pointwise temperature differences with respect to Oort are typically < 4°C over the latitude region 50°S to 50°N. However, even at these latitudes, but particularly at higher latitudes, individual point differences can exceed 10°C. The excellent agreement between the Oort and GISS data, particularly at lower latitudes, is almost certainly due to the setting of the model's implicit ocean circulation to assure a match of sea surface temperatures to their monthly varying climatological averages.

![Boxplots](image-url)

**Figure 1.** Boxplots displaying the zonal distributions for the pointwise seasonal surface temperature differences between the GCMs and the Oort historical data for DJF. Each boxplot is the distribution of the quartiles of these differences for the 72 gridpoints at each latitude. The median values at each latitude are connected. The temperature scaling is -25°C to +25°C, with horizontal lines every 5°C.
Figure 2 displays corresponding pointwise differences between the four GCMs and Oort for a 4°x5° grid of 170 points covering North America. For both seasons, more than one-half of the gridpoint temperatures are within 5°C of the Oort values (with the singular exception of the CCM model in JJA). However, in virtually every case, individual gridpoint differences exceed 10°C over this domain. When the area is reduced to the 43 land gridpoints covering the continental United States, three-quarters of the gridpoints have absolute pointwise differences—GCM-Oort—within 2.3°C to 4.5°C in DJF, and 4.1°C to 8.4°C in JJA.

Further subdividing the United States into west, central and east domains (with 15, 15 and 13 gridpoints, respectively) emphasizes the formidable difficulties faced in using available GCM results in regional studies. (Recall that the 4°x5° grid is the finest used by these four models, and the 15 western gridpoints cover 11 large states.) The resultant pairwise difference boxplots over the United States and these three regions for the JJA season are displayed in Fig. 3. It may be seen that over these limited regions, large differences exist with Oort’s data, especially in JJA when only the GISS model has all gridpoints within 10°C of the Oort values.

With data on a common grid it is possible to directly crossplot the predictions of any GCM with historical data over different regional extents to assess the accuracy of the model predictions. Figures 4A and 4B provide a scatter diagram comparison of the seasonal results (DJF, JJA) for surface air temperature predicted by, in this example, the

![Figure 2](image_url)
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Figure 3. Boxplots showing the distributions of unweighted JJA seasonal surface temperature differences (GCM-Oort) over three different areas: western USA, central USA, eastern USA (left to right). The temperature scaling is $-15^\circ C$ to $+15^\circ C$, with horizontal lines every $5^\circ C$.

Manabe-Wetherald (1987) GFDL model to the observed data for all land areas of the Northern Hemisphere from $2^\circ N$ to $62^\circ N$. For the four GCMs over this region, goodness-of-fit statistics for these intercomparisons are excellent, with cross correlations of GCM to historical data varying from 0.95–0.98 in DJF, and 0.80–0.91 in JJA. For this area, the spatial agreement of the seasonal range of temperature [e.g., the temperature difference, $T(JJA)-T(DJF)$], is also excellent.

However, the excellence of these fits can be largely attributed to the substantial latitudinal variation of temperature, which the GCMs simulate quite well. To more critically evaluate longitudinal variations, the zonal medians of the historical data at each latitude have been subtracted from all values, and the results for the GFDL model are presented in Figs. 4C and 4D. As may be seen here, when the climatological zonal means are subtracted from each gridpoint value, the correlations decrease. For the four models over the $2^\circ$–$62^\circ N$ land areas, the correlations range from 0.64 to 0.84 in DJF and from 0.55 to 0.69 for JJA. As the scale is reduced to the land area of North America, the correlations decrease markedly to 0.23–0.67 in DJF and 0.16–0.53 in JJA. Particularly during the NH summer, the models show poor skill in being able to reproduce longitudinal temperature distributions over these land areas.
3. INTERCOMPARISON OF CONTROL (1×CO₂) PRECIPITATION OVER DIFFERENT REGIONAL SCALES

Because of its obvious relevance to agriculture, precipitation is one of the most important variables predicted by the GCMs. Because precipitation is highly variable in both space and time, the resulting data sets are extremely noisy and highly variable spatially. These characteristics can cause serious distortions when attempts are made to fit or interpolate such fields, therefore, only the original gridded data and model results have been used. Similar to surface air temperature, the GCMs give global averages for seasonal precipitation that are in excellent agreement with historical data. For the entire globe, the four GCMs give area-weighted average seasonal precipitations which are, on average, only about 12–13% different from those recently given by Jaeger (1988). This is likely to be within the error with which precipitation can currently be estimated.

The zonal-median distribution of the predicted precipitation for the four GCMs and Jaeger's historical data are displayed in Fig. 5 for the two seasons, DJF and JJA. Although the forms of the zonal distributions are generally very similar, most showing
A STATISTICAL INTERCOMPARISON OF TEMPERATURE

Figure 5. Zonal distributions of the median zonal precipitation (mm/day) for the four GCMs and Jaeger's historical data for the two seasons DJF (left) and JJA (right). The two plots are scaled identically: 0 to 8.0 mm/day, the Jaeger data are differentiated using the heavy line and the four GCMs are labelled as: C = CCM, G = GFDL, GISS = I, OSU = O.

a distinct maximum at the equator and two symmetrically placed local maxima in the mid-latitudes, there are substantial quantitative differences between the predicted and measured median values. The absolute percentage difference between the GCMs and the historical data (area-weighted) varies from 38 to 76% in DJF and from 80 to 96% in JJA. The corresponding differences between the second historical data set of Schutz-Gates (monthly, not seasonal) are more modest: 15% in January and 38% in July.

More extensive side-by-side intercomparisons over smaller regional scales have shown very serious discrepancies between model predictions on subcontinental scales (Grotch, 1988, pp. 69-119).

4. SIMULATION OF A DOUBLED CO₂ CLIMATE: SURFACE AIR TEMPERATURE CHANGES

To examine the potential for regional assessments of the predicted effects of CO₂ doubling on surface temperature, ΔT, model predictions for seasonally averaged ΔT are compared over seven consecutively smaller regions using the spline-interpolated results on a common 4°×5° grid. The areas considered are:
10.

(1) the entire NH (land + water) (1584 gridpoints),
(2) only NH land gridpoints to 80° N (625 gridpoints),
(3) land gridpoints for North America (88 gridpoints),
(4) gridpoints for the United States (48 gridpoints),
(5) central United States (15 gridpoints),
(6) eastern United States (14 gridpoints), and,
(7) western United States (11 gridpoints).

Figure 6 presents boxplots showing the predicted DJF distributions for ΔT for the four GCMs over these seven areas. It is important to recall that the boxplot shows only the distribution of values in a specified region and it should be realized that, even if two distributions were to agree exactly, they could still be quite different in their spatial distribution. Note that although there is considerable overlap in these distributions over the largest regional extents, already at the scale of the United States the ΔT values for the CCM and the GFDL predictions show no overlap, whatever. The picture worsens for two of the regional United States' areas, where two of the four models show no overlap with the remaining two models in predicted ΔT. The JJA seasonal predictions are slightly different, with the GFDL values typically being the highest over all regions.

Consider the question: "If an estimate for ΔT is available at a particular gridpoint for model A, can this estimate be used to predict what model B would give for ΔT at this same gridpoint?" Examining only land gridpoints in the region 62°S to 62°N, the six pairwise intercomparisons among the four GCMs are shown for the predictions of JJA ΔT in Fig. 7. With one exception (GFDL vs. OSU), all yield cross correlations which are extremely poor, < 0.1. For the DJF season, the cross correlations are considerably higher, ranging from 0.4 to 0.7. Comparable crossplots using only gridpoints covering the land area of the United States yield similar results.

To return to the question just posed, how well can the results of these models be used to refine ΔT predictability over land, it appears that for DJF there is some measure of correlation between these results on a 4°x5° scale, but in JJA the results given by one model appear to be randomly correlated with the results of other models such that a particular choice of ΔT at a given location is no better than using the average of the points over the region. Further work (Grotch, 1988) has shown that poor JJA correlations are generally obtained for land gridpoints at all scales below zonal. These results are particularly provocative in that they seem to suggest that the detailed model predictions for ΔT over land in JJA may not yet be used reliably on gridpoint scales much below zonal. The implications of this suggestion require much further study.

5. SIMULATION OF A DOUBLED CO₂ CLIMATE: PRECIPITATION CHANGES

The four GCMs also predict changes in precipitation after a doubling of CO₂. The percent change in precipitation at each gridpoint will be used as a measure of this change: 
%ΔPr = [100 x (Pr(2xCO₂)–Pr(1xCO₂))/Pr(1x(CO₂)]

Annually, on a global basis, the four GCMs all agree reasonably well, predicting area-weighted average % precipitation changes of 9.4 to 14.8%. On a seasonal basis, however, the global changes are broader: 11.8–16.5% in DJF and 7.8–19.2% in JJA.
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Figure 6. Distributions of ΔT seasonally averaged for DJF for the four GCMs (CCM, GFDL, GISS, OSU, left to right) over seven spatially decreasing NH areas. For each of the upper four plots the vertical scale is −10°C to +15°C, with horizontal lines given at 0°C, 5°C and 10°C. For the lower three United States’ areas, the vertical scale in each plot is 0°C to 10°C, with horizontal reference lines shown every 2°C.
The distributions of percentage changes in annual precipitation predicted by the four GCMs are intercompared in Fig. 8 using boxplots for the following five geographical areas: (1) the entire globe, (2) the Northern Hemisphere, (3) the United States, (4) western Europe, and (5) Africa. Over these regions (with the exception of the United States) the GCMs predict positive changes in precipitation for more than 75% of the gridpoints. In Fig. 8 for both Europe and Africa, two of the models show no overlap whatever in the range of the central 50% box [75%–25%].

The zonal medians for the absolute changes in precipitation predicted by the four GCMs are given in Fig. 9 for the two seasons, DJF and JJA. It can be seen immediately that there are significant differences in these quantitative predictions even on the zonal scale.

Further calculations (Grotch, 1988) comparing the locations of the upper and lower quartiles for precipitation changes over different regions lead one to be very dubious about the spatial consistency of these model predictions for changes in precipitation.
6. THE DANGERS OF RELYING ON AVERAGES

Because the average is probably the "best" single value characterizing a distribution, the climate community has become beguiled by its routine application; for example, "the global-average temperature is expected to rise by 3±1.5°C due to a doubling of CO₂." While the average is clearly important, it is easy to lose sight of an obvious fact, namely, that there are an infinite number of distributions which will yield precisely the same average. The indiscriminate use of the average by the unwary can lead to misleading inferences as the following hypothetical example illustrates.

Let us assume that four GCMs yield for the average change in temperature after a doubling of CO₂ over North America 4.0, 4.1, 4.2 and -4.0°C. When confronted with such results many would say, "clearly the first three models show excellent agreement, whereas Model 4 is considerably different, even having the 'wrong' sign, and thus it is clearly 'suspect'." While, by definition, this statement is certainly true for the average value over the region, it may be completely untrue for, let us say, the average absolute difference between all point values or many other measures of "goodness of fit." In Fig. 10 we see four superposed distributions which yield the average values cited for the four models. In this case Model 4 is considerably closer to Model 1 throughout the region than...
7. CONCLUSIONS

After intercomparing the predictions of four GCMs for surface air temperature and precipitation with each other and with a variety of historical data sets over a broad range of spatial scales, the following conclusions can be made:

Figure 9. Zonal median changes in absolute precipitation (mm/day) predicted by the four GCMs for DJF (left) and JJA (right). The four models are identified as: C = CCM, G = GFDL, I = GISS, O = OSU.

is either Model 2 or 3. (On average, the absolute difference between Models 1 and 4 is about one-half that of either Models 1 and 2 or Models 1 and 3.) It is important to be particularly cautious when comparing spatial predictions using only the average value of the distribution (or for that matter, even higher moments). If the model results are to be used to infer behavior over smaller regions (which is often the case), the actual spatial distributions must be examined in detail. It should not be forgotten that the worst nightmare of modellers could also occur, namely that all of the models could agree perfectly in all details, and yet all could be wrong!
Figure 10. Superposed distributions of predicted change in temperature predicted by four different hypothetical models over the North American Region. Models 1–3 all give approximately the same average. Although Model 4 yields a very different average of opposite sign, it is, in fact much closer spatially to Model 1 than is either 2 or 3.

1) Although the predictions of the GCMs often agree well with each other and with historical data over large scales (global, hemispheric, zonal), when the model results are examined on successively smaller domains, eventually focusing on subcontinental regions containing only relatively few (5–20) gridpoints, significant differences arise.

2) Results reported on the near-zero cross correlations of the predicted change in surface air temperature due to a doubling of CO₂ give little assurance for confidence in quantitative model predictions on a fine scale.

3) The reliance of many investigators, both within and outside of the climate community, on the use of averages as the “bottom line” indicating goodness of agreement can lead to misleading conclusions. It is often too easy to forget that two models can have precisely the same average value for a variable over a region and yet exhibit very significant spatial differences for that variable over the region.

4) As stated at the outset, this study was not intended as a “beauty contest” among models. Many of the results found here suggest that there is no “best” model for all seasons, particularly for detailed regional or seasonal perturbation studies.
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Comparing and Contrasting Holocene and Eemian Warm Periods with Greenhouse-Gas-Induced Warming

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ABSTRACT. Periods of the past that are estimated to have been warmer than present are of great potential interest for comparison with simulations of future climates associated with greenhouse-gas-induced warming. Certain features of the climates of the mid-Holocene and Eemian periods, both interglacial maxima, are described. The simulated climatic responses to both types of forcing, in terms of land/ocean and latitudinal averages, are also compared. The zonal average and annual (or seasonal) average radiation fluxes associated with the different-from-present orbital conditions that existed for those interglacials are compared to the radiation flux associated with CO₂-induced warming. There are some similarities but also significant differences in the two types of radiation flux perturbations, and there are both similarities and differences in the simulated climatic responses.

1. Introduction

The rising concentrations of carbon dioxide and other greenhouse gases are projected to warm the global climate by a few degrees Celsius over the next century. The numerical models upon which these estimates are based, however, have many limitations, including
especially their approximations in representing physical processes and their coarse spatial resolution (e.g., Schlesinger and Mitchell, 1985, 1987; Groth, 1988; Groth and MacCracken, 1991). Therefore, results from these models cannot yet be used to make definitive assessments of potential ecological and societal impacts or to develop adaptation strategies in important areas such as agriculture and water resources.

The magnitude of the projected climatic changes associated with CO2-induced warming is large compared to climatic changes in recent historical times (for example, the past several centuries; see Webb and Wigley, 1985). Geologic records, on the other hand, provide examples of large climatic changes, although in many cases these changes have occurred slowly. For example, the global-average temperature, as estimated from a combination of boundary conditions (estimated sea-surface temperature from marine sediments; CLIMAP, 1981) and atmospheric general circulation model results (COHMAP, 1988), rose by several degrees Celsius between the last glacial maximum (∼ 18,000 yr BP) and the interglacial maximum (∼ 9,000 to 6,000 yr BP). This "global warming" was therefore of the same magnitude as the estimated warming associated with the projected CO2 rise (assuming no controls on emissions), but occurred over 10,000 years rather than a century.

Studies of past climates are useful for identifying mechanisms and processes of large climatic changes, including feedbacks, and for testing the accuracy of climate model simulations. Recent papers by Crowley (1989 and 1990), Gallimore and Kutzbach (1989), and Mitchell (1990), and a US/USSR report (MacCracken et al., 1990) have described, from various perspectives, the possible relations between warm climates of the past and CO2-induced climate warming of the future. In this paper we briefly summarize observations of the climate of the two most recent interglacials, and then compare these observed paleoclimates with simulations of interglacial climates and with simulations of the climates associated with CO2-induced warming.

2. ESTIMATES OF MID-HOLOCENE AND EEMIAN CLIMATES BASED UPON OBSERVATIONS

Extensive research is helping to document climatic changes during the Holocene (Velichko, 1984; COHMAP, 1988; MacCracken et al., 1990). There is considerable evidence that the mid-Holocene period, about 6,000 years ago, had a climate very different from the present. Intensified summer monsoons in parts of northern Africa and southern and eastern Asia brought moister conditions over many regions that are now dry (Kutzbach and Street-Perrott, 1985). There is evidence from vegetation and lake levels that the climate was warmer and drier in the continental interiors of North America and Eurasia. High northern latitudes were warmer. The data coverage, however, is insufficient to provide an accurate estimate of the change in global-average temperature. Webb and Wigley (1985) suggest "that the global mean temperature at 6,000 years B.P. was probably within 1°C of today's temperature" [p. 252]. They go on to say that "if this estimate is correct, then the data for moisture conditions at 6,000 years B.P. are impressive because they show that large changes in both precipitation and extent of deserts and grasslands can be associated with relatively small variations in the global mean temperature" [p. 252].

Paleoclimatic maps for the mid-latitude and polar regions of the Northern Hemisphere for the period 6,000 to 5,000 years ago have been constructed by Velichko et al., (1987) and, for a more limited area of the hemisphere, by the COHMAP group (COHMAP, 1988). Based on a wide variety of data sets for land and ocean conditions, the reconstructions of Velichko and many Soviet colleagues (see MacCracken et al., 1990) indicate that the climate was significantly warmer than present poleward of about 30°N, reaching 3°C or more above present conditions over northern continental and arctic regions (see Fig. 1). Equatorward of 30°N, they estimate slightly cooler conditions than present. Vinnikov and Lemeshko (1987) and Vinnikov et al. (1990) have
Figure 1. Soviet paleoclimatic reconstructions for the mid-Holocene (6,000 to 5,000 yr BP) showing departures from values characteristic of the second half of the nineteenth century for surface air temperature (°C) for January-February (top) and July-August (bottom). Maps are based on data from Klimanov (1982), Velichko (1984), and (except for January-February temperature) from Borzenkova and Zubakov (1984). Contours are dashed in regions where data are most sparse. (Reproduced from MacCracken et al., 1990.)
used estimates of paleotemperature and precipitation values to derive soil moisture maps for the mid-Holocene (Fig. 2). On continental scales, these results (Fig. 2) are in general accord with COHMAP results showing more soil moisture and wetter conditions in southern and eastern Asia and parts of northern Africa due to the strong summer monsoons and drier summer conditions over much of northern North America and northern Eurasia.

There has been less effort devoted to reconstructing the climate of the Eemian interglacial (also known as the Mikulino, Riss-Würm, or Sangamon interglacial), which dates to 125±5 thousand years ago. Data are especially sparse over North America (Heusser and King, 1988). Soviet scientists have also constructed temperature, precipitation and soil moisture anomaly maps for this period (see MacCracken et al., 1990). The Eemian patterns (not shown here) are similar to the mid-Holocene patterns (Figs. 1 and 2), but the magnitudes of the changes are larger and in some cases the boundaries separating warmer/cooler and wetter/drier conditions are shifted compared to their mid-Holocene counterparts. Some regions in the middle and high latitudes were estimated to be 4 to
6°C warmer than present in the Eemian, compared to 2 to 4°C warmer than present in the mid-Holocene (Fig. 1). Slightly cooler conditions (0 to 1°C below present) were estimated for the region south of about 40°N, perhaps a consequence of the intensified monsoonal precipitation or the decreased insolation caused by the differences in the Earth's orbit at that time, compared to present (see Section 4). Drier conditions were estimated for northern Eurasia north of 60°N, with wetter conditions to the south. At 6,000 yr BP, for comparison, this drier/wetter boundary was farther south, near 50°N (Fig. 2).

3. MODEL CALCULATIONS OF THE MID-HOLOCENE AND EEMIAN WARM PERIODS, AND COMPARISONS WITH OBSERVATIONS

Attempting to simulate past climatic changes offers a unique opportunity to gain insight into our ability to model future climatic conditions. Thus, although verifying that the climate models can represent the present climate (especially its seasonal variation) leads to some confidence in their accuracy, verifying that the models can simulate climates of the past provides the opportunity to test whether the models can simulate correctly the response to various changes in forcing. For example, observational studies have suggested that orbital variations are a primary forcing factor pacing glacial/interglacial and monsoonal climatic changes, but models have had an important role in helping us to understand the physical processes that transform "forcing" into "response" (COHMAP, 1988).

Comparisons of climate simulations with observations of the glacial maximum (about 18,000 yr BP) have been performed by many groups (e.g., Gates, 1976a,b; Manabe and Broccoli, 1985; Kutzbach and Guetter, 1986; Rind, 1987). More recently, the necessary observational data sets have been compiled so that comparisons can be made with simulations of past warm periods. In a number of papers, Kutzbach and other COHMAP researchers have reported simulations of the climate for each 3,000-year interval since the last glacial maximum (COHMAP, 1988), including the mid-Holocene warm period around 9,000 to 6,000 yr BP. They and others have also simulated the climate for the Eemian interglacial, taken to be about 126,000 years ago (Prell and Kutzbach, 1987; Royer et al., 1984).

Using a version of the NCAR Community Climate Model (CCM) similar to that used by Washington and Meehl (1984) to estimate the potential climatic effects of a doubled carbon dioxide concentration, the COHMAP group imposed the variations of solar insolation caused by the changes in orbital parameters. They first used an atmospheric GCM with prescribed ocean surface temperature. More recently, experiments have been run with atmospheric GCMs coupled to mixed-layer oceans. We will describe results from both kinds of models.

For the mid-Holocene, about 6,000 years ago, the change in insolation due to changes in the season of perihelion caused an enhancement (dimunition) of Northern Hemisphere summertime (wintertime) radiation by about 5%, with reverse changes in the Southern Hemisphere (less summer radiation). The increased axial tilt, compared to present, changed the latitudinal distribution of annual-average radiation (see Section 4). For the Eemian, about 126,000 years ago, the radiational changes were in the same direction as at 6,000 yr BP, but approximately twice as large (see Section 4). For both times, 6,000 and 126,000 yr BP, there was no net change in the global-average solar insolation, which is distinctly different from the situation for a carbon dioxide doubling, in which the net global radiative forcing of the surface-troposphere system is projected to increase by about 4 W/m², approximately equivalent to a 2% increase in the solar constant.
3.1. Results with atmospheric GCMs with prescribed ocean surface temperature

Figure 3 shows the July (northern summer) temperature differences, compared to present, simulated for 6,000 and 126,000 years ago. An important constraint in these particular simulations by Kutzbach and Guetter (1986) and Prell and Kutzbach (1987) is that the ocean surface temperatures were held fixed at their present seasonal values. Given that the thermal time constant of the upper ocean mixed layer is a few years, the large seasonal insolation variations associated with the orbital changes produce only a small change in the seasonal cycle of sea surface temperature. Because the annual integral of solar radiation differences is not zero at each latitude, we would expect some changes in ocean temperature; however, these changes are likely to be only a fraction of the land temperature response to the orbital forcing, except perhaps in high latitudes where sea-ice feedbacks could occur (see below).

The model results for July, 6,000 yr BP (Fig. 3, top), can be compared to the mid-Holocene reconstructions for summer, 6,000-5,000 yr BP, shown in Fig. 1. There is general agreement between the observations and simulations concerning a warming at high northern latitudes (2 to 3°C), with smaller changes south of 40°N. (The small warming in the Arctic is associated with changes in the surface energy budget over the prescribed (modern) sea ice.) The model simulates an intensified monsoon in July over parts of northern Africa and southern and eastern Asia, and drier conditions in summer and year-round in the mid-latitude continental interiors (Fig. 4, top) in general accord with the reconstructions (Fig. 2).

The simulation for July of 126,000 yr BP (Fig. 3, bottom) shows even warmer conditions in high latitudes (4 to 6°C warmer than present in some locations). The warming is about twice as great as at 6,000 yr BP (Fig. 3, top). This increased warming in the Eemian, compared to the mid-Holocene, is consistent with the increased solar radiation induced by the orbital changes at 126,000 yr BP, compared to 6,000 yr BP (see Section 4). The reconstructed climate also shows much warmer conditions (by 4 to 6°C) poleward of 40°N, but the reconstructed cooler conditions equatorward of 40°N are simulated only for south Asia. The simulated changes in moisture at 126,000 yr BP (Fig. 4, bottom) have the same general pattern as at 6,000 yr BP (Fig. 4, top).

3.2. Results with atmospheric GCMs coupled to mixed-layer ocean models

Various groups (Kutzbach and Gallimore, 1988; Mitchell et al., 1988) have now examined the response of coupled atmosphere/ocean models to orbital changes. These models employ atmospheric GCMs coupled to static mixed-layer ocean models, and include parameterizations that also allow sea-ice, snow cover and soil moisture to be simulated rather than prescribed. The models do not, however, allow for effects resulting from changes in ocean currents.

One set of experiments (Kutzbach and Gallimore, in prep.) used the model described by Kutzbach and Gallimore (1988) and Gallimore and Kutzbach (1989) to examine the climate's response to relatively extreme values of tilt (23.5 degrees plus and minus 1.5 degrees) and extremes of perihelion (June and January), with relatively large eccentricity (0.04). This choice of orbital parameters covers approximately the full range of orbital changes. We present here some examples of results for two of these extreme cases: (1) Case P, perihelion in June (eccentricity = 0.04), modern day tilt; and (2) Case P+T, perihelion in June (eccentricity = 0.04), 25 degree tilt. Orbital conditions for 126,000 years ago were between these two extreme cases; perihelion was in June and eccentricity was about 0.04, but the tilt was only about 0.5 degree greater than present. The radiation differences for Cases P and P+T, compared to present, bracket the differences for 126,000 yr BP. Therefore the magnitude of the simulated climatic response to 126,000 yr BP radiation conditions should be between the values shown for Cases P and P+T.
Figure 3. Paleoclimate simulations of land surface temperature differences (°C), compared to present, for July for 6,000 yr BP (top) and 126,000 yr BP (bottom). Experiments used an atmospheric GCM with surface temperature over oceans prescribed at modern seasonal values; so, temperature differences over oceans are zero. From Kutzbach and Guetter (1986) and Prell and Kutzbach (1987), respectively.
Figure 4. As in Fig. 3, but paleoclimate simulations of precipitation-minus-evaporation differences over land, compared to present, estimated annual averages for 6,000 yr BP (top) and 126,000 yr BP (bottom). Shading denotes decreased values (i.e., drier conditions). Clear areas denote increased values (i.e., wetter conditions). Annual estimates are weighted averages of January and July simulations. From Kutzbach and Guetter (1986) and Prell and Kutzbach (1987), respectively.
The changes in land surface temperature caused by these extreme orbital conditions, Cases P and P+T, are large. The simulated land surface temperatures in JJA are more than 4°C warmer than present in most regions north of 40°N (Fig. 5). As expected, the temperatures are higher in case P+T than in Case P. Ocean surface temperatures are slightly warmer in high latitudes and slightly cooler in middle and low latitudes, compared to present (Fig. 5). In DJF (not shown), it is cooler than present except along the western coasts of continents where advection dominates over radiation, and along the Arctic coasts where the warmer conditions are linked to less sea-ice cover.

At 58° and 70°N, two of the latitudes represented by the model grid, surface temperature (averaged over both land and ocean) is 2 to 4°C above present in JJA and 1–2°C below present in DJF; soil moisture is 10 to 20% less than present in JJA. Sea-ice cover at 81°N (the central Arctic) is 20–25% less than present (Case P) and 60–75% less than present (Case P+T) in summer and fall. These differences are qualitatively similar to, but larger in magnitude than, those reported for 9,000 yr BP simulations with the same coupled atmosphere/ocean model (Kutzbach and Gallimore, 1988; Gallimore and Kutzbach, 1989).

An experiment for 6,000 yr BP radiation conditions has been completed with a newer version of the NCAR CCM developed by Covey and Thompson (1989), one that is also coupled to a mixed-layer ocean (Kutzbach et al., 1990). North of 40°N, surface temperatures in JJA are 1 to 3°C higher than present and south of 40°N temperatures are slightly lower than present in many locations. These results (not shown) compare rather well with estimates from observations (Fig. 1). In DJF the simulated climate is slightly warmer in some regions and colder in others; few of the wintertime changes are statistically significant. Summertime (JJA) soil moisture is generally reduced in mid-latitudes, compared to present; there are wetter conditions in the monsoon lands of northern Africa and southern Asia. These changes in moisture show some similarity to the reconstructed values (Fig. 2) and to the previous simulations (Fig. 4). Sea-ice is more than 1 meter thinner in the central Arctic and there is less sea-ice cover along the Arctic coast in SON (10-40% reduction) and in the annual average (0-15% reduction).

4. COMPARISON OF CHANGES IN CLIMATIC FORCING AND RESPONSE ASSOCIATED WITH ORBITAL AND CARBON DIOXIDE PERTURBATIONS

In this section we compare the zonal-average and certain global-average responses of the climate models to orbital forcing and to increases in the carbon dioxide concentration, and we illustrate similar zonal-average differences, compared to present, from the climatic reconstructions.

First, we review characteristics of the zonal-average changes in radiation associated with CO2 increases and orbital changes. Manabe and Wetherald (1980) show the latitudinal distribution of the change in net annual-average downward radiation flux at the 200 mb level of the standard model atmosphere for a quadrupling of CO2 and, for comparison, for a 4% increase in the solar constant. These distributions are shown here (Fig. 6, top) along with the latitudinal distribution of the net annual-average change in insolation for 6,000 and 126,000 yr BP orbital conditions. The 4xCO2 forcing increases the downward (longwave) flux by about 6 W/m2 at the pole and about 10 W/m2 at the equator. The 6,000 yr BP orbital forcing increases the downward (solar) flux by about 4 W/m2 at the pole and decreases the flux by about 1 W/m2 at the equator. The cross-over point from increase to decrease is at about 40°. These annual-average changes due to orbital forcing are the result of the increased axial tilt at 6,000 yr BP and 126,000 yr BP, compared to present. The changed tilt also changes the latitudinal distribution of radiation.
Figure 5. Simulations of surface temperature differences (°C), compared to present, for JJA for extreme orbital conditions for Case P (left) and Case P+T (right). P indicates June perihelion with eccentricity equal to 0.04, modern axial tilt of 23.5 degrees. P+T indicates June perihelion and axial tilt of 25 degrees. Orbital conditions for 126,000 yr BP are intermediate between P and P+T. Experiments used an atmospheric GCM coupled to a mixed-layer ocean model, as described in Kutzbach and Gallimore (1988).

throughout the seasonal cycle (not shown). Changes in perihelion cause no change in annual-average radiation but cause large seasonal changes (Fig. 6, bottom).

From the perspective of these latitudinal distributions of the annual-average forcing differences, one could anticipate that warmer conditions in the polar regions would accompany both CO\textsubscript{2} increases and periods with increased axial tilt. However, whereas the CO\textsubscript{2} increases should cause warmer conditions at the equator as well, this would not be expected to be the case for the orbital forcing associated with increased tilt. The simulated zonal-average temperature changes generally agree with these anticipated changes. For the CO\textsubscript{2} doubling experiments (Fig. 7a), there is warming at all latitudes in both winter and summer. Due to feedbacks, the polar warming is larger than the tropical warming in winter. For the 6,000 yr BP orbital parameter experiment (Fig. 7b) there is warming in polar latitudes in both winter and summer, but greater in winter and with slight cooling equatorward of 20\degree-30\degree N. This temperature distribution is similar to that obtained by Kutzbach and Gallimore (1988) in an experiment with 9,000 yr BP orbital conditions. Soviet reconstructions for the mid-Holocene and Eemian (Fig. 7c) indicate strong warming in the polar regions and slight warming in the tropics in both seasons. One overall similarity, then, in both the CO\textsubscript{2}- and orbitally-induced simulations, and in the interglacial reconstructions, is the polar warming and the reduced north-south temperature gradient at the surface.

The simulated large warming at high latitudes is in part the result of feedbacks in the model that need further study and, if possible, further confirmation from observations. In the orbital forcing simulations, although there is reduced energy input to Northern Hemisphere high latitudes in winter, the fall and winter temperatures increase rather than decrease. This apparently occurs because the increased summertime heating is stored by the ocean and released during the fall and winter. As shown by Kutzbach and Gallimore
Figure 6. Comparisons of Northern Hemisphere latitudinal distributions of radiation flux differences (W/m²) associated with quadrupled atmospheric carbon dioxide (4×CO₂), four percent increase in solar constant (4% SC), and orbital configurations for 6,000 yr BP and 126,000 yr BP. Top: annual-average values. Bottom: seasonal values (orbital variations only). The differences for increased CO₂ refer to downward longwave flux at the 200 mb level of a standard model atmosphere. The other differences refer to downward shortwave flux at the top of the atmosphere. The annual average values for radiation flux differences at 126,000 yr BP and 6,000 yr BP (top) are nearly identical because the values for increased tilt at those times are nearly identical: 6,000 yr BP, 24.11°; 126,000 yr BP, 23.94°; modern, 23.44°. The differences associated with CO₂ and solar constant changes are redrafted from Manabe and Wetherald (1980).
Changes in Northern Hemisphere latitudinal distributions (land plus ocean) of surface temperature (°C) for DJF (top) and JJA (bottom) for (a): doubled CO2 simulations, (b): 6,000 yr BP orbital simulations, and (c): mid-Holocene (6,000 yr BP) and Eemian (126,000 yr BP) observations. (a) The simulated results for CO2 doubling are from four climate models: (1) Manabe and Wetherald (1987); (2) Schlesinger and Zhao (1989); (3) Hansen et al. (1984); and (4) Washington and Meehl (1984). The temperature differences are normalized to 1°C changes in global-average temperature; the global-average temperature increased by about 4°C (plus or minus one) in the four models. The normalized temperatures on the vertical axis should be multiplied by 4 (approximately) to read simulated temperature changes. Reproduced from MacCracken et al. (1990). (b) The simulated results are from experiments using the NCAR CCM (Kutzbach et al., in prep.). Because the simulated global-average temperature change was approximately zero, these results are not scaled (normalized) to 1°C changes as in parts (a) and (c).

(1988), sea ice forms later in fall and is thinner in the winter at 9,000 yr BP, compared to present. Because of the importance of summertime heating to the overall high-latitude climate, the changes in perihelion that produce more summer radiation also contribute to year-round decreases in sea ice, even though the annual-average radiation change is zero. In the case of carbon dioxide-induced changes, the year-round increase in infrared heating can contribute directly to the melting back of the sea ice in all seasons.

A major difference between the simulations with increased CO2 and the orbital simulations (and the observed warm periods) occurs in the low latitudes, where the simulated and observed paleoclimatic warm periods show virtually no temperature change (or slight cooling), whereas the response to the carbon dioxide increase is a warming of several degrees Celsius.
COMPARING HOLOCENE AND EEMIAN PERIODS WITH GREENHOUSE

Figure 7c. (c) Reconstructed Northern Hemisphere temperature anomalies, winter (top) and summer (bottom), for the mid-Holocene (1) and the Eemian interglacial (2) are based upon averages computed from the paleoclimatic maps described in MacCracken et al. (1990). The temperature differences are normalized, as in Figure 7a (but not 7b), to $1^\circ$C changes in global-average temperature; the Northern Hemisphere average temperature increase, as estimated in these reconstructions (see MacCracken et al., 1990) was $1^\circ$C for the mid-Holocene and $2^\circ$C for the Eemian. The normalized temperatures on the vertical axis should therefore be multiplied by two in the Eemian case (labelled 2) to read the reconstructed estimates of temperature change.

These above-mentioned changes in latitudinal forcing, while important, are overshadowed in the middle and tropical latitudes by the strong differential heating between land and ocean that arises in the case of the large seasonal changes in solar radiation due to orbital forcing compared to the more uniform year-round changes due to CO$_2$ forcing. As shown by Kutzbach and Guetter (1986), Mitchell et al. (1988) and others, a major effect of perihelion occurring in northern summer, compared to northern winter, is an increase in the seasonal temperature extremes over land—warmer summers and colder winters. Owing to its large heat capacity, the ocean’s response to these seasonal radiation changes is much smaller. As a result, continent-scale monsoon circulations are enhanced:
summitone land temperature is higher compared to present, land-ocean temperature
contrasts are enhanced, surface pressure falls over the land and rises over the ocean,
and enhanced ocean-to-land circulations develop. Thus, the enhanced precipitation in the
monsoon lands at 6,000 and 126,000 yr BP is caused primarily by circulation changes
and enhanced moisture convergence over land, whereas, in the case of CO₂ increases,
the tendency for increased precipitation is much more uniform over both land and ocean
and results from the generally increased moisture content of the warmer atmosphere.

The global-average radiation changes due to orbital forcing are near zero, whereas
the global-average increase in longwave radiation due to doubled CO₂ is approximately
4 W/m². The simulated global-average temperature in an orbital-change experiment for
9,000 yr BP was 0.1°C lower than in the control simulation (Kutzbach and Gallimore,
1988); the global-average temperature in the extreme orbital-change experiment, Case
P+T (above), was 0.25°C lower than in the control simulation (Kutzbach and Gallimore,
in prep.). Thus, in the model simulations, the simulated warmth of interglacial climates
is primarily associated with warmer conditions over land in northern summer and in high
latitudes year-round. Changes are very small (or even move toward lower temperatures)
elsewhere, leading to net cooler global-average conditions. These small simulated changes
in global-average temperature due to orbital changes are very different from the large
changes projected due to a doubling of the CO₂ concentration, where increases in global-
average temperature of from 3 to 5°C are reported (Houghton et al., 1990). Estimates of
increase in Northern Hemisphere average temperature for interglacial times, based upon
the Soviet reconstructions, are 1°C (mid-Holocene) and 2°C (Eemian); see MacCracken
et al. (1990). These estimates are smaller than the simulated global-average response to
CO₂ doubling, but larger than the simulated global-average response to orbital changes.
More work is needed to refine these estimates from both observations and models.

5. CONCLUSIONS

When the reconstructed climates of the mid-Holocene and Eemian warm periods are
compared to simulations of these climates (with changed orbital forcing) using climate
models similar to those being applied to simulate the climatic response to increased carbon
dioxide, the model-generated interglacial climates exhibit many features similar to those
of reconstructed interglacial climates, even on a sub-continental scale. This is encouraging,
but more work is needed with improved models and with improved reconstructions.

The model simulations for a carbon dioxide doubling have both similarities and
differences compared to simulations for the orbital conditions for 6,000 and 126,000 yr
BP. The similarities are perhaps most pronounced in the high latitudes, where warming
occurs in response to both kinds of forcing (and in both interglacial reconstructions).
As a whole, the simulations with increased CO₂ exhibit a broad, year-round warming of
both land and ocean. The changes in land/ocean temperature contrast, compared to the
control experiments, are relatively small. However, changes in land/ocean temperature
contrast are a key feature of the orbitally-induced changes in monsoon circulations. The
various CO₂ experiments do not agree in detail regarding the changes of soil moisture in
tropical and extratropical latitudes. In contrast, increased solar radiation at 6,000, 9,000
and 126,000 yr BP in northern summer produces significant enhancement of the seasonal
land/ocean thermal contrast and tropical monsoon intensity and produces distinctly wetter
conditions in parts of northern Africa and southern and eastern Asia and drier conditions
over northern continental interiors in middle latitudes, in comparison to present. At mid-
Holocene time, 6,000 and 9,000 yr BP, there is good confirmation of many of these
regional climatic responses to orbital forcing in the paleoclimatic record. At Eemian
time, about 126,000 yr BP, agreement is much less satisfactory and clearly more work is
needed both in the modeling and the climatic reconstruction efforts.

Our comparison of potential climatic effects of orbital changes and CO\textsubscript{2} changes does
not consider the transient response of the climate. For example, the transient response
to time-dependent increases in atmospheric CO\textsubscript{2} concentration may be different, at least
in some details, from the equilibrium response (e.g., Schneider and Thompson, 1981;
Manabe \textit{et al}., 1990). We have also not described here a number of other factors that
need to be considered. For example, the Eemian interglacial, lasting many millenia, may
have led to a significantly smaller Greenland ice sheet, compared to present (Koerner,
1989). If so, then this changed "boundary condition" would need to be accounted for in the
model studies. Other considerations include changes in the the atmospheric concentrations
of CO\textsubscript{2} and CH\textsubscript{4} at interglacial times (Lorius \textit{et al}., 1990).

In spite of difficulties associated with experimentation on past climates, such as lack
of knowledge of all possible factors influencing past climatic behavior, and in spite of
the difficulties involved in climate reconstruction, these studies continue to be important
for learning more about climate mechanisms and for testing of models. This is so even
though success in simulating past climates is not necessarily a sufficient condition for
success in simulating future climates.

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Utilization of Paleoclimate Results to Validate Projections of a Future Greenhouse Warming

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ABSTRACT. Paleoclimate data provide a rich source of information for testing projections of future greenhouse trends. This paper summarizes the present state-of-the-art as to assessments of two important climate problems. (1) Validation of Climate Models: The same climate models that have been used to make greenhouse forecasts have also been used for paleoclimate simulations. Comparisons of model results and observations indicate some impressive successes but also some cases where there are significant divergences between models and observations. However, special conditions associated with the impressive successes could lead to a false confidence in the models; disagreements are a topic of greater concern. It remains to be determined whether the disagreements are due to model limitations or uncertainties in geologic data. (2) Role of CO₂ as a Significant Climate Feedback: Paleoclimate studies indicate that the climate system is generally more sensitive than our ability to model it. Addition or subtraction of CO₂ leads to a closer agreement between models and observations. In this respect paleoclimate results in general support the conclusion that CO₂ is an important climate feedback, with the magnitude of the feedback approximately comparable to the sensitivity of present climate models. If the CO₂ projections are correct, comparison of the future warming with past warm periods indicate that there may be no geologic analogs for a future warming; the future greenhouse climate may represent a unique climate realization in earth history.

1. INTRODUCTION

A future greenhouse warming threatens to be a climate change of significant magnitude. In order to test climate projections, modellers have sometimes turned to the paleoclimate record for answers to a number of critical questions concerning greenhouse studies. In this paper I describe the state-of-the-art of paleoclimate studies with respect to two important climate problems: validation of climate models and assessment of the role of CO₂ as a significant climate feedback. Because of the scope of the problems being reviewed, all of the above questions cannot be answered in complete detail. The reader is referred to Crowley and North (1990) for a more detailed discussion of many of the observations and conclusions reviewed in this paper.
2. VALIDATION OF CLIMATE MODELS

The same climate models that have been used for making greenhouse projections have also been used for paleoclimate simulations. Comparison of model results with paleoclimate observations allows us to establish some level of confidence in the models. In this study I will focus on the time intervals most examined (the last 20,000 years), but include a few comments on the mid-Cretaceous (90-120 Ma [million years ago]) interval of inferred higher CO₂ levels.

2.1. Agreements

A detailed comparison of model simulations and observations for the last 20,000 years is summarized in Crowley and North (1990). The conclusions (Fig. 1) of this comparison are as follows. At the last glacial maximum (18,000 BP [Before Present]) there is overall a very good agreement between model simulations and observations in the high latitudes of the Northern Hemisphere. The giant Laurentide Ice Sheet exerted a very strong influence on regional circulation patterns over much of the area north of about 30°N. Regional comparisons of model simulations and observations are often very favorable (Kutzbach and Wright, 1985; Manabe and Broccoli, 1985a; COHMAP, 1988). One exception involves some indications that the simulations are not as cold in winter over North America as suggested by the observations.

Another outstanding agreement between model simulations and observations involves fluctuations of the African-Asian monsoon during the early part of the present interglacial period (~5,000-10,000 BP). Calculations indicate that increased summer insolation resulted in enhanced heating on land, which in turn caused increased precipitation

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<th>&quot;Paleo-Modeling Scorecard&quot;</th>
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<td><strong>Region and Time</strong></td>
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Figure 1. Status of agreement between model simulations and observations for various time and space slices over the last 20,000 years. Evaluation represents the author's judgment, but there is considerable support for each rating. [After Crowley, 1989.]
in the monsoon belt. Comparison of model-simulated precipitation-minus-evaporation (P-E) changes and observations indicate, in general, very good agreement (Kutzbach and Street-Perrott, 1985; COHMAP, 1988).

2.2. Disagreements

A less satisfactory agreement involves climate change in the high latitudes of the Southern Hemisphere at 18,000 BP. Initial calculations (Manabe and Broccoli, 1985b) suggested that the direct effect of the Northern Hemisphere ice sheets was quite small, as the temperature perturbation from the ice sheet radiates to space over a length scale of \( \sim 1500 \) km (cf. North, 1984). Inclusion of lower ice-age CO\(_2\) levels resulted in a much better agreement between model simulations and observations, especially with respect to sea ice changes around Antarctica (Manabe and Broccoli, 1985a). Comparison of CO\(_2\) and ice-core inferred temperature fluctuations over Antarctica further indicate a fairly good agreement (Genthon et al., 1987).

There are some problems with the apparently satisfactory convergence of models and data in the high latitudes of the Southern Hemisphere. Comparisons of CO\(_2\) changes with glacial-interglacial transitions indicates that CO\(_2\) lags Southern Hemisphere temperature changes (Fig. 2). This lag suggests that either CO\(_2\) is not an important feedback or that there are other agents contributing to climate change in this area. Possibilities include changes in atmospheric methane, interhemispheric deep-ocean heat transport, orbital forcing, or atmospheric dust levels (Crowley and Parkinson, 1988; Harvey, 1988; Chapellaz et al., 1990; cf. Petit et al., 1990). As an example of how these other mechanisms could modify the CO\(_2\)-temperature relation in Fig. 2, I have plotted the combined effect of CO\(_2\), interhemispheric deep-ocean heat transport and orbital forcing (weighted 3:1:1) in Fig. 3. The choice of weights is arbitrary, for the purpose at this stage is merely to illustrate a point, not rigorously determine relative importance. Notice that the CO\(_2\)-temperature offsets are not nearly so prominent in this comparison, despite the fact that CO\(_2\) still accounts for 60% of the variation in the synthetic curve. The effect of dust (not considered here) is probably relatively small, for absolute concentrations are almost an order of magnitude higher in the Northern Hemisphere than in the Southern Hemisphere (Thompson and Mosley-Thompson, 1981). Furthermore, the radiative effect of the dust may have been overestimated by almost an order of magnitude (Anderson and Charlson, 1990).

The apparently good agreement between model simulations and observations over Antarctica at 18,000 BP (Jouzel et al., 1989) may also be misleading, because the result is dependent on stipulation of a 500-m thicker ice sheet over Antarctica by CLIMAP (1981). Utilization of a near-dry adiabatic lapse rate of \( \sim 8^\circ\)C/km suggests that a 4°C temperature cooling—nearly one-half of the ice core range—is directly attributable to this elevation effect. However, more recent glaciologic studies over Antarctica (e.g., Denton, 1985) no longer support inferences of a thicker ice sheet. Thus, we can explain only about one-third of the observed temperature change over East Antarctica at 18,000 BP.

An even more vexing area of model-data disagreement involves climate change in the tropics at 18,000 BP (Webster and Streeter, 1978; Rind and Peteet, 1985). Because CLIMAP (1981) tropical SSTs are relatively unchanged, models do not show much differences between control and perturbed runs. However, observations consistently indicate that tropical snowlines were lowered by about 1 km; data less consistently suggest that lowland precipitation was decreased (see Crowley and North, 1990). Repeated efforts to validate the CLIMAP SST estimates with independent methods continue to support their main conclusion (Prell, 1985; Brassell et al., 1986; Broecker, 1986; Anderson et al., 1989).
Figure 2. Comparison of high-latitude Southern Hemisphere records for the present and last interglacial with the ice-core CO$_2$ record from Vostok, Antarctica. Both the Southern Ocean SST and ice-core temperature records indicate warmth greater than the present at the last interglacial. Note that CO$_2$ changes lag temperature in the Vostok ice-core. SST record from Martinson et al. (1987); ice-core deuterium record from Jouzel et al. (1987); CO$_2$ record from Barnola et al. (1987); ice-core chronology plotted according to revisions of Petit et al. (1990). [From Crowley and North, 1990.]

It is possible that some of the lowland P-E simulation/observation discrepancies may reflect processes not being incorporated into climate models. For example, lower CO$_2$ levels could generate a significant vegetation response, thereby giving
Figure 3. Example of how additional mechanisms can partially reconcile the Southern Hemisphere CO$_2$-temperature offsets in Fig. 2. Figure illustrates the combined effect due to CO$_2$, changes in North Atlantic Deep Water (NADW) production and changes in local orbital forcing over Antarctica (the latter from Short et al., 1990). Changes in NADW production affects interhemispheric exchange of heat (e.g., Hastenrath, 1980; Crowley and Parkinson, 1988); time variations in NADW based on equatorial Atlantic core analyzed in Mix and Fairbanks (1985). CO$_2$, NADW and tilt variations have been weighted at 3:1:1. These ratios give approximately the best fit to the Vostok temperature record (solid line). However, the physical justification of the relative weighting requires more examination. Thus, this figure is for illustrative purposes only.

a false impression of drier conditions. Biosphere feedback may also be important (e.g., Lean and Warrilow, 1989). However, the upland simulation/observation discrepancies are more difficult to explain. Geologists (e.g., van der Hammen, 1985) have sometimes suggested changes in lapse rate, but this suggestion is usually considered anathema to climate modelers (Webster and Streeten, 1978; Rind and Peteet, 1985). Furthermore, (very) scattered tropical lower-elevation (1-2 km) temperature estimates suggest decreases of 4-6°C in these regions (cf. Liu and Colinvaux, 1985; Colinvaux, 1989; Bonnefille et al., 1990), a result not consistent with stable SSTs offshore.

The above picture is further complicated by comparisons of upland and lowland temperature changes over Europe at 9,000 BP, in which observations suggest a larger temperature response than over lowlands (Porter and Orombelli, 1985; Huntley and Prentice, 1988). Geologic indications of an increased temperature sensitivity versus height are consistent with results from the 18,000 BP tropics and again diverge from model predictions (COHMAP, 1988). The 9,000 BP data provide additional support for the altered-lapse-rate scenario, and perhaps it is prudent to re-examine this process in climate models.
2.3. Summary of Model-Data Comparisons

Direct comparisons of models and data for the last 20,000 years yield a mixed picture. We have two very good agreements and two cases where there are significant discrepancies. Trying to determine an “average rating” is not satisfactory, for none of the test cases gives a middle-of-the-road response. A further consideration involves the fact that where models and data diverge, the discrepancy is not necessarily a condemnation of the model. Data are also open to reinterpretation and sometimes a model may not include certain processes that need to be considered.

Two other factors need to be considered when weighing the relative significance of the cases where there is a very good agreement between models and data. For the northern high latitudes at 18,000 BP, the very good agreement is impressive. However, it is not necessarily a justification for giving great credence to forecasts for greenhouse studies, because the Laurentide Ice Sheet was so huge that it essentially forced the atmosphere into a relatively narrow range of possible responses. The situation will be quite different for a future greenhouse perturbation, where there will be no massive localized forcing.

In the case of the second very good agreement between models and data (the monsoon at 9,000 BP), there is also reason for caution in extrapolating successes to greenhouse forecasts. Intercomparisons (Crowley et al., 1986) of linear and nonlinear model responses to orbital forcing (Fig. 4) and additional GCM sensitivity experiments (Oglesby and Park, 1989) yield the surprising result that in continental interiors the initial thermal response to orbital insolation forcing on seasonal time scales is very nearly linear (subsequent precipitation processes are more nonlinear, however). This result is also consistent with observations: if one were to compare the spectrum of observed temperature response in Eurasia to solar forcing at the top of the atmosphere, the amplitudes of the annual and semi-annual harmonics are nearly identical for forcing and response, and there is very little variance in higher harmonics of the temperature field. If the temperature response were highly nonlinear, one would expect significantly greater variance in the higher harmonics of this field.

The above results, which are not completely understood from a theoretical viewpoint, imply a significant limitation to extrapolating seasonal model behavior to changes in mean-annual forcing, such as occurs with greenhouse gases. For example, it has sometimes been stated that because climate models simulate the present annual cycle so well, and since this cycle is larger than glacial-interglacial changes, then we can use this success as one measure of confidence in greenhouse predictions (Schneider, 1989). The paleoclimate results suggest that the seasonal model example is sampling the system over the linear domain of its response. This result is a nice example of how paleoclimate studies have enhanced our understanding of how the climate system operates.

With the above considerations in mind I think we can say that at present we have overall only a fair agreement between models and data for the last 20,000 years. Going further back in time, modeling studies suggest that CO₂ is perhaps the most important forcing factor influencing long-term climate change, with mid-Cretaceous (90-120 Ma) levels perhaps 5-7 times present (Berger and Spitzy, 1988; Berner, 1990). Climate models would predict that such concentrations would produce tropical SSTs perhaps 2-4°C warmer than present and increased subtropical drought (Manabe and Bryan, 1985; Hansen et al., 1988; Schlesinger, 1989). Such changes are not observed in the record (Crowley and North, 1990; Horrell, 1990). It is possible that some of these discrepancies may reflect inadequacies in estimating paleo-SSTs (Horrell, 1990) or changes due to continental drift (Barron and Washington, 1985); nevertheless, there is a possibility of a very significant discrepancy for this time interval that warrants more attention.
Figure 4. Comparison of energy balance and general circulation model responses to increased summer insolation at 9000 BP. The 4°C contours indicate the difference between the model response for 9000 years ago and the present (the earlier time period is warmer). The energy balance model is from North et al. (1983); the low-resolution model and the NCAR CCM are from Kutzbach and Otto-Bliesner (1982) and Kutzbach and Guetter (1984), respectively. Note that all models have comparable sensitivity. Since the energy balance model is linear, this agreement suggests that the seasonal GCM response on land is also linear. [From Crowley et al., 1986.]

3. ROLE OF CO$_2$ AS A CLIMATE FEEDBACK

Critiques of greenhouse predictions seems to be almost a monthly occurrence. Given the importance of the problem, is there anything we can say about the role of CO$_2$ as a significant climate feedback based on observations and modeling of the geologic record? Despite the fact that our understanding of past changes is incomplete, I think it is fair to summarize the present state of knowledge by concluding that, although other factors are contributing to the long-term change of climate on 10$^3$-10$^6$ year time scales, modeling studies exploring the importance of these mechanisms generally yield smaller changes than observed in the geologic record. Changing CO$_2$ levels, supported by both observations and geochemical calculations (e.g., Berner et al., 1983; Barnola et al., 1987), result in a closer correspondence between models and data (Crowley and North, 1990).
However, some significant problems still remain. Thus, from a geologic perspective we can say that CO₂ appears to play a very important role in both warming and cooling climates, and although the precise sensitivity of climate change to CO₂ levels still remains to be more precisely determined, geologic studies suggest that the sensitivity is sufficiently large (approximately the same as the present generation of climate models) to justify concerns about future atmospheric perturbations due to greenhouse gas changes.

Geologic studies also provide us with other perspectives on the greenhouse problem. For example, the reaction of many greenhouse sceptics to the CO₂ predictions seems to assume that somehow there are any number of self-correcting features in the climate system that prevent large-scale climate change, that the system in effect supports the Gaia hypothesis. Yet the geologic record states the exact opposite — that the system changes, sometimes greatly, and that its changes are larger than we would expect from modeling studies; in other words the system may be more sensitive than our present ability to model it.

The above changes lead me to conclude that, from a geologic perspective, there is a likelihood that the greenhouse forecasts are approximately correct. Temperatures will probably rise in the future, perhaps reaching levels not achieved for tens of millions of years (Fig. 5). Yet it is not possible to cite these past time intervals as analogs for a future warming (Crowley, 1990), as future changes will be very rapid and involve a significant nonequilibrium component, whereas past warm periods represent long time-averaged equilibrium snapshots of warmer climates. Large changes in orography and ocean circulation also occurred in the past (e.g., Kutzbach et al., 1989; Maier-Reimer et al., 1990), thereby further making regional comparisons by analogy invalid. Thus, study of the geologic record yields the startling conclusion that the future greenhouse warming may be a unique climate realization in earth history.

![Figure 5. Schematic comparison of future greenhouse warming with past changes in temperature. Note that Pleistocene climate oscillations are idealized and that interglacial temperatures were probably not significantly warmer than present. Pre-Pleistocene changes are not well fixed in magnitude, but their relative warmth is approximately correct. Maximum warming in the Cretaceous based on estimates by Barron and colleagues (Thompson and Barron, 1981; Barron and Washington, 1985). Time intervals in between have been scaled accordingly. [From Crowley, 1989.]](image-url)
4. CONCLUSIONS

(1) Although there has been some success in modeling the climates of the last 20,000 years, there are significant areas of model/data disagreement. Where there is good agreement, it is felt that the model successes cannot at present be used as a gauge of the reliability of models for future greenhouse predictions. There may also be some significant discrepancies between model simulations and observations for the mid-Cretaceous time with inferred higher CO₂.

(2) Modeling studies suggest that past CO₂ changes have played a critical role in the long-term evolution of climate. From this perspective, geologic studies support the conclusion that the future greenhouse perturbation represents a significant climate change.

(3) If the climate does change in the future, it is likely to result in a climate state for which there may be no geologic analog, that is, it may be a unique climate realization in earth history.

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Part 2:

Climate Model Projections of Greenhouse-Gas-Induced Equilibrium Climatic Change: What Are the Expected Climatic Change?
The Equilibrium Response to Doubling Atmospheric CO₂

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ABSTRACT. The equilibrium response of climate to increased atmospheric carbon dioxide as simulated by general circulation models is assessed. Changes that are physically plausible are summarized, along with an indication of the confidence attributable to those changes. The main areas of uncertainty are highlighted.

1. INTRODUCTION

In the last year there has been an intensification of interest in the climatic effects of increases in greenhouse gases, partly because of extreme events, including the summer drought in the United States, and partly because of political initiatives such as the Toronto Conference (August 1988). Five centers, the Geophysical Fluid Dynamics Laboratory (GFDL), the Goddard Institute for Space Studies (GISS), the United Kingdom Meteorological Office (BMO), the National Center for Atmospheric Research (NCAR) and Oregon State University (OSU) have recently published details of one (or more) numerical studies of the climatic effects of doubling CO₂. At least four other centers (the Max Planck Institute, Hamburg; the Canadian Climate Centre; the Meteorological Research Institute, Japan and the Commonwealth Scientific and Industrial Research Organization (CSIRO) in cooperation with the Bureau of Meteorology in Australia) have run, or are about to carry out similar numerical studies, and institutes in Italy, the Soviet Union, France and the People's Republic of China have also expressed interest in such work. The Inter-Governmental Panel on Climatic Change (IPCC) is to produce a major assessment of the predictions of climate change to be presented at the World Climate Conference and to the United Nations in late 1990.

Given this rapid acceleration in scientific and political activity, the useful lifetime of a review such as this is limited. Furthermore, detailed reviews of the topic have been given recently by Dickinson (1986), Schlesinger and Mitchell (1987) and Mitchell (1989). Hence, in this paper I will summarize much of what has gone before, allowing the reader to refer to previous reviews or to the original papers for a more detailed explanation of...
results obtained to date, and I will attempt to highlight what I believe are the main areas of agreement and disagreement among different studies and indicate areas where I think further research is necessary.

2. EQUILIBRIUM EXPERIMENTS WITH MIXED-LAYER OCEANS

The thermal relaxation time of the deep ocean is of the order of centuries. Thus, equilibrium experiments with models including the deep ocean are prohibitively expensive. [Bryan (1984) has derived a method of accelerating the convergence to equilibrium in the absence of the seasonal cycle.] Most equilibrium experiments to date have used atmospheric models coupled to an oceanic mixed layer, generally of order 50 meters depth. In early experiments the advection of heat by the ocean was ignored; in later experiments this is allowed for by prescribing a heat convergence to represent non-local oceanic effects. This heat convergence is assumed not to change with doubling CO2. The horizontal resolution is generally about 550 km (880 km in the case of the GISS experiments), though experiments at higher resolution are now being carried out (Wetherald, personal communication).

2.1. Temperature

The global-mean warming simulated by atmosphere/mixed-layer ocean models varies from 1.9 to 5.2K (Fig. 1). Cess and Potter (1988) plotted the equilibrium global-mean temperature change obtained by doubling CO2 in five different studies against the global-mean temperature in the simulation of present-day climate. They noted that the largest warming occurred in the model with the coldest control climate, and vice versa. Although there are plausible explanations why this should be the case (the colder the control, the larger the sea ice extents and hence the greater the positive ice-albedo feedback; the warmer the control, the stronger the negative lapse rate feedback due to the increasing importance of moist processes), the rate of change of sensitivity with temperature suggested by the straight line fit in Fig. 1 is far greater than can be explained by such mechanisms. Mitchell and Warrilow (1987) modified the treatment of land surfaces in the Meteorological Office model, thereby increasing the global-mean temperature in the control simulations slightly, but without affecting the global-mean sensitivity. Cess et al. (1989) demonstrated that there are large differences in cloud feedback produced by at least three of the models represented in Fig. 1. Mitchell et al. (1989) obtained warmings which ranged from 5.2 to 1.9K with versions of the Meteorological Office model with increasingly sophisticated, but not necessarily more accurate, cloud parameterizations. The surface temperature of the control simulation only varied between 287.2 and 287.7K. In summary, the discrepancies in global-mean warming simulated by the different models is too large to be explained in terms of differences in the temperature of the simulated present-day climate. The biggest contribution must come from the differences in the parameterizations of physical processes, especially those associated with cloud.

The geographical and seasonal distribution of the warming shows some similarity from model to model. The warming is greatest around sea ice margins in winter (Fig. 2) and least over sea ice in summer. The larger winter warming is associated with changes in sea ice (Manabe and Stouffer, 1980; Ingram et al., 1989). There is little seasonal variation in the warming of the tropics. The warming is generally greater over land than over the open ocean at similar latitude. The annual-mean warming is greatest at high latitudes.

The main relative discrepancies occur in the magnitude of the warming: (1) in the tropics throughout the year, and (2) over the northern mid-latitude continents in summer. For example, both the GISS (Hansen et al., 1984) and GFDL (Manabe and Wetherald,
Figure 1. The simulated global-mean change in surface temperature due to doubling atmospheric carbon dioxide as a function of simulated control temperatures. GFDL - Manabe and Wetherald (1987); GISS - Hansen et al. (1984); NCAR - Washington and Meehl (1984); OSU - Schlesinger and Zhao (1989); BMO - Wilson and Mitchell (1987); M1, M2, M3 - Mitchell et al. (1989).

1987) models produce a global-mean warming of about 4K. However, the GISS model produces a larger warming in the tropics and a smaller warming at high latitudes (Fig. 3). The GISS model produces a much larger warming in the upper tropical troposphere, up to 8K compared with 4K in the case of the GFDL experiment (Fig. 4). This discrepancy may be related to the convective parameterization. The GISS results are qualitatively similar to those attained with the BMO model (Wilson and Mitchell, 1987) which uses a similar penetrative convection scheme. The GFDL results are more like those obtained with the NCAR model (Washington and Meehl, 1984) which uses an equivalent moist convective adjustment parameterization. However, an attempt to confirm that a penetrative scheme would produce a greater warming in the tropics has proved inconclusive (Cunnington and Mitchell, 1990).
Figure 2. Simulated change in surface air temperature due to doubling atmospheric CO₂ concentrations during northern summer (June to August). Contours every 2K, stippled where greater than 4K. (a) a GFDL model (Manabe and Wetherald, 1987); (b) an NCAR model (Washington and Meehl, 1984). (From Schlesinger and Mitchell, 1987.)

The GFDL and BMO models produce a relatively large warming over the northern mid-latitude continents in summer. This is associated with the drying of the land surface (Manabe et al., 1981). On the other hand, the NCAR (and to a lesser extent, the GISS) model produces a wetter land surface for much of this region in summer, so that enhanced evaporation occurs at the expense of a smaller surface warming (Fig. 2).
2.2. Precipitation

The range of intensification in the hydrological cycle varies from 4 to 15%. The size of the intensification is largely dependent on the size of the warming (Fig. 5). This is to be expected, given the increase in saturated vapor pressure with temperature. However, the rate of increase is typically less than half that given by the Clausius-Clapeyron relation (about 6%/K at 288K), and appears to be limited by energy budget considerations (Mitchell et al., 1987).

There is some agreement on patterns of the larger-scale features in the changes in precipitation. All models produce enhanced precipitation in high latitudes on doubling CO$_2$, associated with enhanced moisture convergence (Manabe and Wetherald, 1975; Mitchell, 1983) and in the tropics, and in mid-latitudes in winter (Fig. 6). Changes in the subtropics are generally small, with both increases and decreases.

There is, however, considerable disagreement on smaller scales, especially over the northern continents in summer and in the tropics. A number of GFDL and BMO simulations, along with that from OSU, suggest a reduction in precipitation over much of the northern mid-latitude continents in summer (particularly in the GFDL models) and an enhancement of the Asian southwest monsoon. In general, the precipitation in the intertropical convergence zone (ITCZ) is enhanced, and there is a tendency for a tropical precipitation maximum to shift into the summer hemisphere. This may be associated with a strong positive cloud feedback in the summer hemisphere (Wilson and Mitchell, 1987).
Figure 4. Change in zonally averaged atmospheric temperature due to doubling atmospheric CO\textsubscript{2} (June to August). Contours every 1K, areas of decrease stippled. (a) a GFDL model (Manabe and Wetherald, 1987); (b) a GISS model (Hansen et al., 1984). (From Schlesinger and Mitchell, 1987.)
In contrast, the GISS (Hansen et al., 1984) and particularly the NCAR (Washington and Meehl, 1984) models produce a widespread increase in precipitation over the northern mid-latitude continents in summer, and in the case of the GISS model, a tendency for the ITCZ to shift towards the winter hemisphere. As a result, there is little or no consensus on the equilibrium changes in precipitation in the tropics, or over the northern mid-latitudes in summer.

2.3. Soil Moisture

One of the major controversies arising from the studies made to date concerns whether or not the surface of the northern mid-latitude continents will become drier in summer. The bulk of the northern continents become wetter in winter in all simulations because of the enhanced winter precipitation, and the relatively small increase in evaporation found at the low winter temperatures. All the GFDL and BMO experiments and the OSU study produce a drying of the northern mid-latitude land surface in summer (Fig. 7) due to one or both of the following factors: (1) reduced precipitation and enhanced evaporation, the latter following earlier snow melt; and/or (2) reduced cloud cover and/or enhanced longwave radiation from the warmer, moister CO₂ enriched atmosphere. The NCAR and GISS models produce a wetter surface (Fig. 7). Mitchell and Warrilow (1987) changed the parameterization of runoff over frozen ground in a version of the BMO model, demonstrating that where the land surface is far from saturation at the time of spring snow melt
Figure 6. Simulated change in precipitation due to doubling atmospheric CO$_2$. Contours at $0$, $\pm 1$, $\pm 2$ mm/day. Areas of decrease are stippled. (a) December to February, (b) June to August.

in the control simulation the land surface will remain wetter for much of the spring and summer on doubling the CO$_2$ concentration. Meehl and Washington (1988) reported that the NCAR model produces a very dry land surface throughout the year in its simulation of present-day climate (possibly because of the use of a low surface albedo over
Figure 7. Simulated changes in soil moisture (June to August) due to doubling atmospheric CO$_2$ from 5 different models. GFDL - Manabe and Wetherald (1987); GISS - Hansen et al. (1984); NCAR - Washington and Meehl (1984); OSU - Schlesinger and Zhao (1989); BMO - Wilson and Mitchell (1987). Areas of decrease are stippled. Note that the contour interval varies. (From Schlesinger, 1989.)

land). Hence, the increase in summer soil moisture found in that model is consistent with Mitchell and Warrilow's results. The GISS model uses a parameterization over frozen ground (Hansen et al., 1983) which is similar to that used in the alternative version of the Meteorological Office model. The NCAR results are dubious, given that the simulated soil moisture and runoff in the control simulation are excessively small. The GISS model uses a very coarse horizontal grid (880 km) and so may not adequately represent the weak disturbances associated with summer rainfall. Simulations with GFDL and BMO models at higher (250 km) horizontal resolution consistently produce a drying of the mid-latitude surface in summer (Manabe et al., 1981; Mitchell and Lupton, 1984; Wetherald, 1989 personal communication). Nevertheless, a 440-km-grid version of the GISS model with prescribed changes in sea temperature failed to produce a substantial summer drying on doubling atmospheric CO$_2$ (Rind, 1987).

In summary, most, but not all, models produce a drying of the northern extratropical continents in summer. In those models producing a drying, the mechanisms are well understood and are very plausible.
3. EQUILIBRIUM STUDIES WITH DYNAMICAL OCEAN-ATMOSPHERE MODELS

The use of dynamical ocean/atmosphere models in equilibrium $2 \times CO_2$ simulations has been limited, partly because of the difficulty in bringing the deep ocean to equilibrium. Manabe and Bryan (1985) carried out a series of equilibrium simulations using the low-resolution model with idealized geography and annually averaged insulation. The main characteristics of the atmospheric warming obtained on enhancing $CO_2$ (Fig. 8) are similar to those using a mixed-layer ocean model, namely, a stratospheric cooling, and maximum warming at upper tropospheric levels in the tropics and in high latitudes at the surface. The warming at high latitudes is propagated into the deep ocean at all latitudes. Manabe and Bryan (1985) note that in the warmer climates the thermohaline circulation changes little, as the larger warming at high latitudes is compensated by the smaller coefficient of thermal expansion at lower temperatures. Thus, the changes in density in high and low latitudes are similar.

No equilibrium studies have been carried out to date with a seasonal cycle and a realistic geography, so as yet it is not possible to say whether or not changes in ocean circulation would substantially alter the patterns of climate change determined using mixed-layer ocean models. Manabe and Stouffer (1988) have demonstrated that coupled models can produce more than one equilibrium, though there is no evidence to suggest that more than one equilibrium may exist for enhanced $CO_2$ simulation.

4. SUMMARY OF RESULTS DEDUCED FROM EQUILIBRIUM $2 \times CO_2$ EXPERIMENTS

The following changes are virtually certain to occur on enhancing atmospheric $CO_2$:

* the stratosphere will cool
* the troposphere will warm
* the global water cycle will be enhanced

The following changes are simulated by all comparable models and will probably occur:

* the warming will be largest in high latitudes in winter
* the annual-mean warming is smallest in the tropics
* the warming in the tropics varies little with season
* precipitation increases in mid- to high latitudes in winter
* soil moisture and runoff increase in mid- to high latitudes of the northern mid-continents in winter
* seasonal sea ice melts earlier and forms later

The following changes are common to some, but not all, models and hence are less certain:

* seasonal snow cover accumulates later and melts earlier
* soil moisture amounts decrease in the northern mid-latitude continents in summer

The global-mean warming at equilibrium varies from 2 to 5K and the enhancement of the water cycle ranges from 4 to 15%. Much of the discrepancy is due to differences in simulated cloud feedback. In general, there is better agreement for simulated changes in temperature than in the hydrological cycle.
Figure 8. Height-latitude diagram of zonally averaged equilibrium changes in temperature due to quadrupling CO$_2$ in an idealized coupled ocean-atmosphere model. (From Spelman and Manabe, 1984.) Contours at $-8K$, $-4K$, $-1K$ and then every $1K$ upwards. Extra contours at $3.5K$ and $6.5K$. Areas of decrease lightly stippled, and warming greater than $7K$ is heavily stippled.

5. PRIORITIES FOR FUTURE RESEARCH TO IMPROVE ATMOSPHERE MODELS

The equilibrium studies carried out so far indicate several areas with either the need or potential for improvement. First, there is considerable uncertainty in the estimates of cloud-climate feedbacks. Climate models are just beginning to include cloud water content as an explicit variable which will enable an estimate of the feedbacks due to change in cloud radiative properties. Further feedbacks may occur due to changes in cloud droplet distribution. Very little is known about the radiative properties of ice clouds and how they may be represented in large-scale models. Secondly, there are large discrepancies in the simulated warming in the tropics which appear to depend on the choice of convection scheme. Third, there is still much to be done on including the effects of vegetation in climate models. It is likely that improvements will have little effect on the global climate sensitivity, but may have pronounced effects at and near the surface, and hence are important for climate impact studies. Fourth, it is evident that increased resolution from the 15 spectral waves or $5^\circ$ latitude by $8^\circ$ longitude typical of current models will produce a substantial improvement in the simulation of current climate, particularly hydrology. However, with the current generation of computers, it is unlikely that one can make accurate prediction of the regional effects without resorting to other techniques such as
nesting high-resolution, limited-area models over regions of specific interest within global climate models. Finally, current simulations of equilibrium climate change have not taken into account possible changes in ocean circulation. Although such changes are unlikely to affect global sensitivity significantly, they may have substantial local effects.

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THE EQUILIBRIUM RESPONSE TO DOUBLING CO₂


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Computer Experiments with a Coarse-Grid Hydrodynamic Climate Model

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ABSTRACT. A climate model is developed on the basis of the two-level Mintz-Arakawa general circulation model of the atmosphere and a bulk model of the upper layer of the ocean. A detailed model of the spectral transport of shortwave and longwave radiation is used to investigate the radiative effects of greenhouse gases. The radiative fluxes are calculated at the boundaries of five layers, each with a pressure thickness of about 200 mb. The results of the climate sensitivity calculations for mean-annual and perpetual seasonal regimes are discussed.

1. MODEL DESCRIPTION

The CCAS (Computer Center of the Academy of Sciences) climate model is used to investigate the climatic effects of anthropogenic changes of the optical properties of the atmosphere due to increasing CO₂ content and aerosol pollution, and to calculate the sensitivity to changes of land surface albedo and humidity.

The climatic changes are studied with the aid of a joint atmospheric general circulation model, based on the two-layer Mintz-Arakawa model (Gates et al., 1971), and a bulk thermodynamic model of the upper layer of the ocean. The finite-difference approximation of the governing equations is carried out on a coarse spatial grid with a resolution of 12° latitude by 15° longitude. Realistic geography and relief are used. In the vertical direction the troposphere is divided into two layers of equal mass. A convective adjustment procedure is used to calculate cloudiness and precipitation. The
The climatic effects of increases in CO$_2$ and aerosol are calculated for mean-annual insolation. The ocean temperature is calculated, but sea ice and snow cover are fixed (Aleksandrov et al., 1983; Aleksandrov and Stenchikov, 1985). Radiative transfer is calculated by means of the parameterization of Katayama (1972). The effects of changes in land surface properties are calculated for perpetual July with the ocean temperatures fixed.

2. RESULTS

The global-mean heating of the surface air due to a CO$_2$ doubling is 1.4°C. The largest temperature changes are located in the middle and high latitudes. The temperature increase reaches 5.5°C at the northern coastline of North America, 4.8°C near the Barents sea, and 5.8°C over Alaska. The air is warmed by 4.6°C and 4.3°C in the equatorial regions of the Pacific Ocean and Arabian Sea, respectively. The regional character of the effect is shown by the existence of cooling in the central and South Atlantic, and near the coastline of Antarctica. The low global-average sensitivity of the simulated climate change is connected with the absence of sea ice and snow cover feedbacks (Fig. 1).

For the calculation of the climatic change induced by aerosol pollution of the atmosphere, only the effects of the aerosol on the scattering and absorption of solar radiation are taken into account. The optical properties of the aerosol are calculated by means of the simplified model developed by Coakley et al. (1983).

The global-mean temperature decrease due to aerosol pollution is −1.76°C. The Northern Hemisphere is cooled by −1.9°C and the Southern Hemisphere by −1.6°C.
The maximum temperature change of $-6.9^\circ$C occurs near the coastline of Antarctica. In the middle latitudes of the Southern Hemisphere, the air is 3–4°C colder than in the normal case. The cooling is $-4.3^\circ$C in northern Canada, $-3.4^\circ$C at the western coastline of North America, $-4.2^\circ$C in Southeast Asia, and $-4.6^\circ$C in northern Siberia. In spite of the global cooling, the air over the North Atlantic Ocean and northern South America is heated by $1.3^\circ$C (Fig. 2).

The numerical experiments with the changed surface properties were carried out in collaboration with A. Krenke and D. Turkov from the Institute of Geography, Moscow. The present July regime of the atmospheric circulation was calculated with fixed climatic distributions of the ocean temperature, albedo, and ground wetness (Arhipov et al., 1987; Stenchikov and Turkov, 1988).

The observed surface albedo depends on the type of vegetation and the incidence angle of the solar radiation to the surface (Kondratiev et al., 1983). For example, the albedo of glaciers in Antarctica is 0.9, while the albedo of sea ice is 0.6–0.7. The ocean has a minimum albedo in tropical regions of about 0.07. The albedo of land has maximum values in grasslands and deserts, where it varies from 0.2 to 0.28. Comparison of the results of perpetual July calculations with the observed albedo and Katayama's albedo (Katayama, 1972) parameterization (globally observed albedo is 3% greater than Katayama's) shows that the most significant temperature changes occur over the Northern Hemisphere land. The land surface temperature decreases by 3.7°C while the global-mean temperature decreases by 0.9°C. In North Africa, the temperature drops by 6–8°C, in Central Asia by 12°C, and in Europe by 4–6°C. The air temperature over the oceans in equatorial and subtropical regions increases by 1–2°C. The largest temperature changes in the disturbed case are in the deserts where the changes in albedo are maximum. Temperature changes are simulated in the entire troposphere. The summer hemisphere is more sensitive to changes in the surface albedo. The global surface air temperature sensitivity reaches 0.3°C per 1% albedo change which is lower than the estimate by Budyko (1980). The cooling of the air over the land leads to a shift in the convergence zone to the south and a decrease in the monsoon precipitation (Fig. 3).

The sensitivities for a 1% change of the surface albedo are: surface air temperature, $-0.3^\circ$C; lower troposphere temperature, $-0.5^\circ$C; upper troposphere temperature, $-0.9^\circ$C; and precipitation, $-0.1$ mm/day.

For the investigation of the sensitivity of the climate to the evaporation from land, the surface wetness parameter was changed from its July climatic value, which was obtained on the basis of observations (Atlas World Water Balance, 1974), to the maximum possible value of unity. Globally the surface wetness parameter increases by 0.15. The results of the calculations show that the surface air temperature decreases by $3.6^\circ$C as a result of the evaporation increase. But the changes in the upper and lower troposphere are not as large. Temperature decreases are simulated mainly in the Northern Hemisphere and reach 9–10°C in the 30–40°N latitude band where the changes of surface wetness parameter were the largest (from 0 to 1 in deserts). The areas with a large decrease of the surface air temperature are located in the tropical and subtropical continental regions. In the central regions of Europe and Asia, and in North America, the temperature decreases by more than 17°C, and in North Africa by more than 13°C (Fig. 4). The increase of the evaporation leads to an increase of precipitation in many regions of the Earth, and globally by 0.5 mm/day. The geographical distribution of the precipitation change shows that the maximum increase of precipitation occurs in North America and Europe, and also in the northern region of the Indian Ocean and the Gulf of Guinea. It is obvious that changes in the evaporation rate cause the most significant intensification of the hydrological processes in these regions. It is very interesting that the increase of evaporation in the Northern Hemisphere deserts in the southwest regions of North America, North Africa, and Central Asia does not lead to a substantial increase of precipitation in these regions (Fig. 5).
Figure 2. The surface air temperature change as a result of natural aerosol pollution of the atmosphere.

Figure 3. The surface air temperature change as a result of 3% global surface albedo increasing.
Figure 4. The surface air temperature change as a result of 15% global surface moisture increasing.

Figure 5. The precipitation change as a result of 15% global surface moisture increasing.
The sensitivities for a 1% change of the surface wetness parameter are: surface air temperature, $-0.24^\circ C$; lower troposphere temperature, $-0.13^\circ C$; upper troposphere temperature, $+0.02^\circ C$; and precipitation $+0.03 \text{ mm/day}$.

In addition, the effects of simultaneous regional changes of albedo and surface wetness for perpetual July and January regimes have been investigated. These changes may be connected with the characteristics of the snow cover, deforestation processes or transport of water to the deserts. Realistic scenarios in the case of deforestation lead to an increase of albedo and a decrease of evaporation. The effects of these changes on the surface temperature have different signs and compensate each other. The effects are not very large and have a regional character.

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Climate Variability and Climate Change

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ABSTRACT. Changes of variability with climate change are likely to have a substantial impact on vegetation and society, rivaling the importance of changes in the mean values themselves. A variety of paleoclimate and future climate simulations performed with the GISS global climate model is used to assess how the variabilities of temperature and precipitation are altered as climate warms or cools. In general, as climate warms, temperature variability decreases due to reductions in the latitudinal temperature gradient and precipitation variability increases together with the intensity of the hydrologic cycle. If future climate projections are accurate, the reduction in temperature variability will be minimized by the rapid change in mean temperatures, but the hydrologic variability will be amplified by increased evapotranspiration. Greater hydrologic variability would appear to pose a potentially severe problem for the next century.

1. INTRODUCTION

For assessments of the impact of climate change on society and the interpretation of paleoclimate indicators, understanding changes in climate variability is of first-order importance. It is the extremes in temperature that are responsible for killing trees in winter, or causing health problems in summer. It is the extreme hydrological conditions (floods/droughts) that affect water availability and lead to property damage or fires. While all of these effects are related to the mean conditions about which variability is occurring, the magnitude of the variability will amplify or diminish the importance of the mean-condition changes.

Climate variability occurs on all time scales and results from a variety of processes. In addition to the variability of climate-change forcing mechanisms such as solar insolation, volcanic aerosols or atmospheric composition, internal processes produce a spectrum of variations. These include turbulent fluctuations on the order of an hour, weather fluctuations of a few days, atmospheric internal dynamics and interaction with the land surface for weeks to months, ocean interactions from months to years, and cryospheric processes.
of up to 10,000 years (Mitchell, 1976). The variability on the shorter time scales adds to the spectrum at larger periods, and a significant part of the total variance is likely generated by internal stochastic processes.

The change of variability with mean climate can be expected to follow general governing principles. All models agree that as climate changes the effect should be largest at high latitudes, especially in winter. This "high-latitude amplification" of climate change results from high latitudes' having greater atmospheric stability so that warming is preferentially kept near the surface there. Furthermore, in the warmer climate, snow and sea ice are diminished at high latitudes, reducing the surface albedo and allowing more of the sun's energy to be absorbed.

The expected greater high-latitude temperature sensitivity to climate change thus indicates that in a warmer climate temperatures will warm more at high latitudes than elsewhere. This would reduce the latitudinal temperature gradient, which has two effects: it reduces the temperature contrast of air masses that invade from low and high latitudes, and it reduces the intensity of extratropical storms which advect different air masses into a region. Both effects reduce the variability in temperature that occurs; in the extreme case, if all the air masses were the same, or advection was nil, there would be little if any temperature variability at a particular site, except that associated with cloud cover/radiative changes. Observations for the last century, however, have failed to verify unambiguously the relationship between warmer climates and reduced variability (Angell and Korshover, 1978; Barnett, 1978; Diaz and Quayle, 1980; Ratcliffe et al., 1978), or found no apparent connection between trends of temperature and trends of variability (van Loon and Williams, 1978).

Weaker eddies should also lead to smaller variations in rainfall. However, variability in precipitation will in addition be affected by the intensity of the hydrologic cycle. A warmer climate will have the capacity for greater moisture loading of the atmosphere, more evaporation, and thus more precipitation. At mid- and high latitudes, the warmer atmosphere could produce much more rain than possible currently, with the potential for greater variability in absolute amount. At low latitudes, rainfall intensities will be strongly affected by the sea surface temperature gradients, currently a major uncertainty in climate change assessments (Rind, 1987a). Thus the overall effect of a warmer climate on hydrologic variability is difficult to foresee, and no trend in precipitation variability has been found associated with the historical temperature trend (van Loon and Williams, 1978).

2. MODEL EXPERIMENTS

Temperature changes over the last century have been relatively small, and their change does not represent movement to a new equilibrium climate. To investigate the dependence of variability on climate state, we use the Goddard Institute for Space Studies general circulation model (GCM) simulations of a variety of warm and cold climates. The climates include the last ice age, 18,000 years before the present (B.P.) (Hansen et al., 1984; Rind and Peteet, 1985; Rind, 1986, 1987b); an alternate simulation of the last ice age with 2°C colder sea surface temperatures (Rind and Peteet, 1985; Rind, 1986); the Allerod, 11,000 B.P. (Rind et al., 1986); the Younger Dryas, 11–10,000 B.P. (Rind et al., 1986); the doubled CO₂ climate (Hansen et al., 1984; Rind, 1986); and a simulation of the warm Cretaceous, approximately 65 million B.P. (Rind, 1986). The interannual variability from these different climates is compared with simulations of the current climate control run.

The model simulations are generally for five years, with a one-year spin-up. An exception is the doubled CO₂ climate and its separate control run, for which the last ten years of a 35-year run are used. The boundary conditions for the past climates and
the simulation of the future climate contain many uncertainties. Nevertheless, the results can be used to investigate the response of model variability to conditions of changed latitudinal temperature gradients and mean temperature. A comparison of the interannual temperature variability from the current climate with observations shows that the values are comparable, although slightly too large over land in summer, and too small over regions of dynamical ocean events (El Niño) (Hansen and Lebedeff, 1987; Rind et al., 1989).

3. RESULTS

3.1. Warm and Cold Climates

The results of changes in variability and other parameters are presented in Table 1, given as percentage change from the model's current climate values. The values are averaged over 108 gridpoints for each month, about half of which are independent, as the correlation between observed temperatures 1000 km apart (one gridbox) drops to 50% (Hansen and Lebedeff, 1987). To determine the significance of the changes in standard deviations, we use a 100-year control run, broken up into 20 five-year periods, to calculate the standard deviation of the standard deviations. Values at each gridpoint are then compared using the t test. We also compare the magnitudes of the standard deviations for each climate experiment and control run over land, determining the number of gridpoints in which the altered climate value is greater or smaller than the current value, and evaluating the probability of such results using the binomial distribution. In general, variability changes on the order of 15% or higher are significant at the 5% level, while changes of 10% are of marginal significance. While some of the changes are therefore not statistically significant, a basic trend from warm to cold climates is apparent.

At mid-latitudes, the colder climates have greater latitudinal temperature gradients, greater eddy energy, and greater interannual temperature variability. The changes are most significant during winter, early spring and late fall when the temperature-gradient and eddy-energy changes are most noticeable. The difference between the January ice age temperature standard deviations and those for the current climate control run are shown in Fig. 1; increased variability occurs for most locations at mid- to high latitudes in the winter hemisphere. The difference between the doubled CO$_2$ climate standard deviations and today's (Fig. 2) illustrates the general reduction in the warmer climate winter hemisphere. Percentage changes are smaller at low latitudes and in the summer hemisphere, where the changes in latitudinal temperature gradients are muted, and eddy energy less important. The Cretaceous simulation has an altered continental distribution, which provides for greater continentality and an increase in temperature variability over what would have been expected from the decreased latitudinal temperature gradient.

The variability in precipitation tends to track the variation in precipitation itself, although variations in eddy energy also are important (at low latitudes signified to some extent by the change in temperature variability). For example, exceptions to the general trend include the ice age climate, which has reduced precipitation but increased precipitation variability and increased eddy energy, and the Cretaceous at mid-latitudes in summer, which has decreased eddy energy and precipitation variability, despite increased precipitation. In the altered ice age experiment, with colder sea surface temperatures, the greater reduction in precipitation is sufficient to decrease precipitation variability during the warmer months. The general increases in precipitation variability for the doubled CO$_2$ run are shown for July in Fig. 3, reflecting the overall increase in hydrologic activity in the warmer climate.
Table 1. Percentage changes relative to current climate.

<table>
<thead>
<tr>
<th>CLIMATE EXPERIMENT</th>
<th>ICE AGE</th>
<th>ICE AGE</th>
<th>Y. DRYAS</th>
<th>ALLEROD</th>
<th>2CO₂</th>
<th>CRETAC</th>
</tr>
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<td></td>
<td>ALT 18K</td>
<td>REG 18K</td>
<td>11-10K</td>
<td>11K</td>
<td></td>
<td>65 M.Y.</td>
</tr>
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<td>MID-LATITUDES (31-55°N)</td>
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</tr>
<tr>
<td>Annual</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Δ Surf Temp (°C)</td>
<td>-7.2</td>
<td>-5.3</td>
<td>-2.0</td>
<td>-0.4</td>
<td>4.0</td>
<td>7.8</td>
</tr>
<tr>
<td>Δ(ΔT 27-67°N)(%)</td>
<td>34.5</td>
<td>34.5</td>
<td>15.6</td>
<td>11.3</td>
<td>-6.3</td>
<td>-23.3</td>
</tr>
<tr>
<td>Δ Eddy K. E. (%)</td>
<td>15.2</td>
<td>16.5</td>
<td>11.2</td>
<td>6.0</td>
<td>-8.0</td>
<td>-20.5</td>
</tr>
<tr>
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<td>21.0</td>
<td>25.6</td>
<td>11.6</td>
<td>0.0</td>
<td>-7.2</td>
<td>-11.6</td>
</tr>
<tr>
<td>Δ Precip. (%)</td>
<td>-15.7</td>
<td>-10.4</td>
<td>-6.0</td>
<td>-2.2</td>
<td>8.0</td>
<td>20.4</td>
</tr>
<tr>
<td>Δ P St. Dev. (%)</td>
<td>-10.3</td>
<td>6.9</td>
<td>2.3</td>
<td>4.6</td>
<td>5.3</td>
<td>24.1</td>
</tr>
<tr>
<td>Δ(Sum T/Win T) (%)</td>
<td>36.6</td>
<td>21.1</td>
<td>31.2</td>
<td>25.8</td>
<td>-8.8</td>
<td>-0.1</td>
</tr>
<tr>
<td>January</td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Δ Surf Temp (°C)</td>
<td>-10.2</td>
<td>-7.1</td>
<td>-4.4</td>
<td>-2.3</td>
<td>4.5</td>
<td>8.5</td>
</tr>
<tr>
<td>Δ (ΔT 27-67°N)(%)</td>
<td>35.5</td>
<td>32.9</td>
<td>11.7</td>
<td>10.6</td>
<td>-11.8</td>
<td>-27.9</td>
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<tr>
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<td>15.2</td>
<td>13.7</td>
<td>8.8</td>
<td>7.4</td>
<td>-5</td>
<td>-20.6</td>
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<tr>
<td>Δ T St. Dev. (%)</td>
<td>50.0</td>
<td>48.7</td>
<td>54.1</td>
<td>14.9</td>
<td>-8.0</td>
<td>1.3</td>
</tr>
<tr>
<td>Δ Precip. (%)</td>
<td>-15.7</td>
<td>-9.6</td>
<td>-7.5</td>
<td>-4.4</td>
<td>10.5</td>
<td>19.5</td>
</tr>
<tr>
<td>Δ P St. Dev. (%)</td>
<td>2.2</td>
<td>0.0</td>
<td>-4.4</td>
<td>-7.6</td>
<td>14.7</td>
<td>9.7</td>
</tr>
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<td>July</td>
<td></td>
<td></td>
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<td></td>
<td></td>
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<tr>
<td>Δ Surf Temp (°C)</td>
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<td>-0.3</td>
<td>1.5</td>
<td>3.6</td>
<td>6.9</td>
</tr>
<tr>
<td>Δ (ΔT 27-67°N)(%)</td>
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<td>33.9</td>
<td>18.1</td>
<td>12.9</td>
<td>4.8</td>
<td>-14.6</td>
</tr>
<tr>
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<td>19.1</td>
<td>29.2</td>
<td>22.5</td>
<td>15.7</td>
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<td>-16.9</td>
</tr>
<tr>
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<td>8.2</td>
<td>20.6</td>
<td>4.1</td>
<td>3.1</td>
<td>0.0</td>
<td>-2.1</td>
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<tr>
<td>Δ Precip. (%)</td>
<td>-22.0</td>
<td>-18.6</td>
<td>-11.2</td>
<td>-4.3</td>
<td>8.7</td>
<td>29.1</td>
</tr>
<tr>
<td>Δ P St. Dev. (%)</td>
<td>-23.3</td>
<td>-1.2</td>
<td>-17.4</td>
<td>1.2</td>
<td>12.0</td>
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<td>LOW/SUBTROPICS(0-30°N)</td>
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<td></td>
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<td></td>
</tr>
<tr>
<td>Δ Surf Temp (°C)</td>
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<td>-2.0</td>
<td>-0.3</td>
<td>-0.1</td>
<td>3.6</td>
<td>2.1</td>
</tr>
<tr>
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<td>-8.7</td>
<td>2.4</td>
<td>-4.4</td>
<td>0.0</td>
<td>2.6</td>
<td>-26.1</td>
</tr>
<tr>
<td>Δ Precip. (%)</td>
<td>-7.7</td>
<td>-2.0</td>
<td>3.7</td>
<td>2.5</td>
<td>5.0</td>
<td>-5.0</td>
</tr>
<tr>
<td>Δ P St. Dev. (%)</td>
<td>-11.9</td>
<td>9.5</td>
<td>9.5</td>
<td>4.8</td>
<td>1.9</td>
<td>-19.1</td>
</tr>
<tr>
<td>January</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Δ Surf Temp (°C)</td>
<td>-4.0</td>
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<td>-0.9</td>
<td>-0.6</td>
<td>3.9</td>
<td>3.0</td>
</tr>
<tr>
<td>Δ T St. Dev. (%)</td>
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<td>30.0</td>
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</tr>
<tr>
<td>Δ Precip. (%)</td>
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<td>8.2</td>
<td>5.7</td>
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<td>-8.0</td>
</tr>
<tr>
<td>Δ P St. Dev. (%)</td>
<td>10.1</td>
<td>5.5</td>
<td>13.8</td>
<td>20.2</td>
<td>5.9</td>
<td>-6.4</td>
</tr>
<tr>
<td>July</td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Δ Surf Temp (°C)</td>
<td>-4.4</td>
<td>-2.5</td>
<td>0.3</td>
<td>0.4</td>
<td>3.3</td>
<td>1.1</td>
</tr>
<tr>
<td>Δ T St. Dev. (%)</td>
<td>-17.0</td>
<td>10.3</td>
<td>-5.2</td>
<td>0.0</td>
<td>7.1</td>
<td>-5.2</td>
</tr>
<tr>
<td>Δ Precip. (%)</td>
<td>-15.6</td>
<td>-7.9</td>
<td>-0.7</td>
<td>-1.4</td>
<td>16.9</td>
<td>-5.4</td>
</tr>
<tr>
<td>Δ P St. Dev. (%)</td>
<td>-12.8</td>
<td>3.4</td>
<td>-17.9</td>
<td>-10.3</td>
<td>5.3</td>
<td>2.6</td>
</tr>
</tbody>
</table>
The results for the changes in interannual variability with climate are in general consistent with expectations, with temperature variability decreasing as climate warms and precipitation variability increasing.

How do changes in interannual variability relate to changes in variability on other time scales? Also shown in Table 1 is the change in seasonal variability indicated by the ratio of summer to winter temperatures. Seasonal variability is larger in the colder climates, associated with the tendency for surface temperature changes to be greater in winter, although here an additional effect occurs due to the increased seasonal oscillation in radiation during the Allerod/Younger Dryas time frame.

3.2. Future Climate

A more detailed study was conducted of the changes in variability which arise in the GISS doubled CO₂ climate model simulation (Hansen et al., 1984) and the GISS transient climate change simulation (Hansen et al., 1988) on several shorter time scales, (day-to-day (daily) changes and changes in the amplitude of the diurnal cycle) in addition to the interannual variations. A full discussion of the results from this study is presented in Rind et al. (1989). Here we simply summarize the results and put them in a broader perspective.
Temperature variability on all three time scales is generally reduced as the climate warms, although this is not true for all locations and all seasons. The reduction on the interannual and daily time scales is due to the same factor, the "high-latitude amplification" of climate change reducing the latitudinal temperature gradient. Since the latitudinal temperature contrast is not eliminated, only weakened, the reduction in variability does not occur everywhere, being experienced in about 2/3 of the model's extratropical gridboxes. The reductions are generally on the order of 10-15%, and are greatest in winter, the season for which the normal latitudinal temperature contrast is most severely reduced.

The diurnal temperature cycle is also reduced; for this time scale the effect is greatest in summer, and is associated with the increased greenhouse capacity of the atmosphere. With more CO₂ and water vapor, the loss of energy to space is inhibited, which limits nighttime cooling but does not noticeably impact daytime warming (as water vapor and CO₂ are poor absorbers of solar energy). Thus the day/night temperature contrast is reduced, again on the order of 10%, in most locations. The effect is greatest in summer when radiative processes become more important (relative to advection) and the model's cloud cover changes do not interfere.

In contrast to the reduced temperature variability, hydrologic variability tends to increase in the model as climate warms, on both the interannual and daily time scales. The hydrologic cycle (evaporation/precipitation) intensifies by about 10% in the doubled CO₂ climate. The change in variability is strongly connected with the change in mean conditions, being of the same sign 3/4 of the time. Overall, 2/3 of the locations show increased hydrologic variability, with changes again on the order of 10%. During summer,
Figure 3. Change in the interannual standard deviation of precipitation during July between the doubled CO₂ experiment and its current climate control run.

Increased rainfall amounts are associated with greater penetrating convection, as the warmer and moister atmosphere has greater potential instability.

The relationship of changes in the mean to changes in the variability can be generalized to include other climate characteristics in addition to precipitation. Presented in Table 2 are the changes that arise in selected dynamic or thermodynamic mean values, along with the changes in variability between the doubled CO₂ climate and its control run. The variability changes have the same sign as the changes in the mean, and the magnitudes are often of the same order.

<table>
<thead>
<tr>
<th>Quantity</th>
<th>Location</th>
<th>Change in mean (%)</th>
<th>Change in St. Dev. (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eddy K. E.</td>
<td>N. Hem.</td>
<td>-7</td>
<td>-10</td>
</tr>
<tr>
<td>N. Tr. Sens. Ht. by Eddies</td>
<td>N. Hem.</td>
<td>-8</td>
<td>-13</td>
</tr>
<tr>
<td>Total N. Tran. St. Energy</td>
<td>N. Hem.</td>
<td>6</td>
<td>11</td>
</tr>
<tr>
<td>Barocline EKE Gen.</td>
<td>43°N, 600 mb</td>
<td>-7</td>
<td>-29</td>
</tr>
<tr>
<td>Jet Stream</td>
<td>31°N, 200 mb</td>
<td>14</td>
<td>3</td>
</tr>
<tr>
<td>Specific Humidity</td>
<td>N. Hem.</td>
<td>33</td>
<td>27</td>
</tr>
<tr>
<td>Low Level Cloudiness</td>
<td>Global</td>
<td>-6</td>
<td>-22</td>
</tr>
</tbody>
</table>
4. DISCUSSION

The results of these studies are: as climate warms, temperature variability decreases and hydrologic variability increases. What will this mean from a practical standpoint for the coming century, assuming the greenhouse-gas-induced warming comes to pass? The temperature variability reduction is likely to have little direct influence, since the mean temperatures are expected to be changing significantly from decade to decade. So, while the variability about the mean may undergo some reduction, the fact that the mean itself is changing rapidly indicates that the temperatures experienced will be reaching new levels anyway.

The hydrologic variability increase is likely to be an underestimate of the actual effect. The above study looked specifically at precipitation variability. However, the effect on water resources also involves evaporation, which together with precipitation and runoff determine the available soil moisture. Evaporation increases as the climate warms, and potential evaporation (the atmosphere's demand for moisture) increases even more. The greater evaporation will reduce available moisture. In those locations where precipitation increases, the effect may be ameliorated, but where precipitation decreases, the effect will be amplified. For example, the global soil moisture variability in July increases by 29%, while the precipitation variability increases only 11%. Overall, then, the likelihood of droughts should increase substantially.

This effect is quantified with the use of a drought index, which measures the difference between the atmospheric supply and demand for moisture (Rind et al., 1990). The results provide a dire prediction for the likelihood of future droughts: for Northern Hemispher summer, drought intensities, which occur just 1% of the time in the current climate control run, occur 40% of the time by 2060, the date for the equivalent doubled CO₂ warming in that particular trace gas accumulation scenario. The drought increase is driven by the large increase in potential evaporation which accompanies a warming of approximately 4°C during summer. This effect has not been properly appreciated in prior discussions of associated greenhouse effects and is underestimated in GCMs (Rind et al., 1990).

What effects have been left out that could contribute to changes in variability? The model does not include tropical storms, and recent analysis (Emanuel, 1987) indicates that the warmer ocean temperatures are likely to lead to an increase in the potential destructive power of hurricanes of 40–50%. This would augment the increased hydrologic variability discussed above.

Currently, some percentage of interannual variability appears to be related to changes in tropical ocean temperatures associated with El Niño events. While the model cannot make any specific forecast of how ocean dynamics will change, it is worth noting that the increased radiative forcing of the doubled CO₂ climate works to equalize temperatures in the eastern and western Pacific; the warming climate attempts to warm the cooler eastern Pacific more, for the heat put into the warmer western Pacific is largely used for increasing evaporation. It is this temperature contrast between the east and west Pacific that produces the most dramatic El Niño effects. Were El Niños really to decrease with time, this would reduce interannual variability.

Other ocean circulation changes may have contributed to climate variability in the past. Specifically suggested in this regard are possible changes in North Atlantic Deep Water production (Broecker et al., 1985). Again, the model cannot comment upon this possibility; however, the largest changes in this effect seem to be associated with ice sheet growth or ice melting effects, and it is not clear how the warming climate would force the system in this regard, although changes in ocean transports cannot be ruled out.
Finally, there has recently been some attention given to the possibility that significant interannual variability is related to solar cycle effects, as mediated by the phase of the quasibiennial oscillation of zonal winds in the tropical lower stratosphere (van Loon and Labitzke, 1985). While this process is not well understood, and may ultimately turn out to be a statistical fluke, if it is real it implies that the climate system is much more sensitive to small-amplitude perturbations than heretofore believed. The impact of the solar cycle on the received radiation is only on the order of 0.1%, and the doubled CO₂ climate is equivalent to a solar radiation increase of about 2%. The possibility thus exists that the real system may show a much larger change in variability than predicted by current models.

ACKNOWLEDGEMENTS

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Broecker, W. S., D. M. Peteet and D. Rind, 1985: Does the ocean-atmosphere system have more than one stable mode of operation? Nature, 315, 21–25.


Characteristics of Coupled Atmosphere-Ocean CO₂ Sensitivity Experiments with Different Ocean Formulations

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Boulder, CO 80307
U.S.A.

ABSTRACT. The Community Climate Model at the National Center for Atmospheric Research has been coupled to a simple mixed-layer ocean model and to a coarse-grid ocean general circulation model (OGCM). This paper compares the responses of simulated climate to increases of atmospheric carbon dioxide (CO₂) in these two coupled models. Three types of simulations were run: (1) control runs with both ocean models, with CO₂ held constant at present-day concentrations, (2) instantaneous doubling of atmospheric CO₂ (from 330 to 660 ppm) with both ocean models, and (3) a gradually increasing (transient) CO₂ concentration starting at 330 ppm and increasing linearly at one percent per year, with the OGCM. The mixed-layer and OGCM cases exhibit increases of 3.5°C and 1.6°C, respectively, in globally averaged surface air temperature for the instantaneous doubling cases. The transient-forcing case warms 0.7°C by the end of 30 years. The mixed-layer ocean yields warmer-than-observed tropical temperatures and colder-than-observed temperatures in the higher latitudes. The coarse-grid OGCM simulates lower-than-observed sea surface temperatures (SSTs) in the tropics and higher-than-observed SSTs and reduced sea-ice extent at higher latitudes. Sensitivity in the OGCM after 30 years is much lower than in simulations with the same atmosphere coupled to a 50-m slab-ocean mixed layer. The OGCM simulates a weaker thermohaline circulation with doubled CO₂ as the high-latitude ocean-surface layer warms and freshens and the westerly wind stress decreases. Convective overturning in the OGCM decreases substantially with CO₂ warming. Geographical distributions of surface air temperature change in the transient case show regional climate anomalies different from those in the instantaneous CO₂ doubling OGCM case, particularly in the North Atlantic and northern Europe. These two sets of experiments demonstrate that different ocean models and types of CO₂ forcing in the climate system result in different CO₂ climate responses. Earlier studies with energy-balance climate models confirm that instantaneous CO₂ doubling simulations respond differently than transient simulations.
I. INTRODUCTION

This paper discusses coupled atmosphere-ocean experiments with a simple mixed-layer ocean model and a dynamical ocean model developed at the National Center for Atmospheric Research (NCAR). The focus is on demonstrating the effects on the climate system of sea ice albedo feedback, horizontal ocean heat transport, thermohaline circulation, and convective overturning.

The earliest simulations of coupled atmosphere-ocean models included a simple ocean component and assumed a saturated surface based upon a surface energy balance. This type of ocean can be thought of as a shallow swamp, having no seasonal heat storage and no vertical or horizontal heat transport (e.g., Manabe and Wetherald, 1975; Washington and Meehl, 1983, 1986). Because the ocean is a wet surface with no seasonal ocean heat storage, such a model requires annually averaged solar forcing.

The next generation of simple coupled atmosphere-ocean models made use of a shallow mixed layer with a thickness approximately equal to 50 m. This coupling allowed for an annual cycle and a crude approximation of seasonal heat storage but no horizontal or vertical heat transport (e.g., Washington and Meehl, 1984; Manabe and Stouffer, 1980), and thus could better be compared with observations. Hansen et al. (1984, 1988) and Wilson and Mitchell (1987) have used models that, in addition to the mixed-layer ocean, prescribed seasonal horizontal ocean heat transport. This type of formulation necessarily produces sea surface temperatures (SSTs) that are close to observed values compared to mixed-layer formulations without specified ocean heat transport. However, since the specified heat transport cannot change as ocean temperature and sea ice respond in climate-change sensitivity experiments, an important and perhaps crucial feedback mechanism is not included.

A number of carbon dioxide (CO$_2$) climate sensitivity studies with a simple mixed-layer ocean formulation have been conducted at NCAR: (1) intercomparison of the observed, computed, and simple mixed-layer sea surface temperatures (SSTs), as well as the surface-energy balance (Meehl and Washington, 1985); (2) changes of tropical circulation in the model (Meehl and Washington, 1986); (3) changes in atmospheric circulation as indicated by persistent height anomalies or “blocking” (Bates and Meehl, 1986); (4) assessment of ice-albedo feedback in the model (Dickinson et al., 1987); (5) tropical-midlatitude interactions (Meehl, 1988); and (6) soil-moisture sensitivity (Meehl and Washington, 1988).

A more recent type of model used at NCAR for studying the sensitivity of climate to increased CO$_2$ is an atmospheric model coupled to an ocean general circulation model (OGCM). This type of coupled model needs extremely long computer runs, even with coarse-resolution oceans, due to the long adjustment time of the global ocean model. Examples of CO$_2$ experiments with OGCMs coupled to atmospheric models include Bryan et al. (1982), Bryan and Spelman (1985) and Bryan et al. (1988), using a Geophysical Fluid Dynamics Laboratory (GFDL) model with idealized geographical domain (a sector of the globe); and Schlesinger et al. (1985) and Schlesinger and Jiang (1988), using an Oregon State University model with a realistic geographical domain. In several of these studies, CO$_2$ was increased instantaneously to investigate the transient response of the system. The GFDL coupled models followed the coupling method of Manabe (1969) and Bryan (1969) which includes the seasonal cycle. Some recent results of NCAR CO$_2$ studies with an OGCM coupled to an atmospheric model (Washington and Meehl, 1989) are shown here with a system that includes the seasonal cycle. An assessment of errors introduced from the respective components of the NCAR coupled models (mixed layer and OGCM) has been made for the Indian and Pacific sectors of the tropics (Meehl, 1989).
The CO$_2$ experiments in this paper use the atmospheric model described in Washington and Meehl (1983, 1984). Section 2 describes the atmospheric, oceanic, and sea ice models, along with the method of coupling the atmospheric and oceanic models, and how the experiments were conducted. Section 3 compares simulations with the observed atmosphere, ocean and sea ice distributions. Section 4 discusses the differences between a control simulation with 330 ppm and instantaneous doubling of CO$_2$ concentrations; zonal and some geographical distributions are also shown. Section 4 also describes a transient experiment with a gradually increasing (one percent per year) concentration of CO$_2$. Section 5 summarizes the differences in climate sensitivity between the simple mixed-layer ocean model and an OGCM in terms of a CO$_2$ increase. Also included are suggestions for improvements in the various aspects of the coupled model system.

2. DESCRIPTION OF ATMOSPHERIC, OCEANIC AND SEA ICE COMPONENTS

Figure 1, a schematic of the components of the coupled ocean-atmosphere sea ice model, shows atmospheric and oceanic levels at the left and physical processes at the middle and right. (See Washington and Meehl, 1984, for mixed-layer details, and Washington and Meehl, 1989, for OGCM details.) The atmospheric component of the interactive climate model, a version of the NCAR Community Climate Model (CCM), has nine levels and uses the spectral transform method for computing horizontal nonlinear transport terms with a truncation wavenumber of rhomboidal 15. The horizontal resolution is approximately 4.5° in latitude and 7.5° in longitude. The radiation-cloudiness scheme is as described in Ramanathan et al. (1983), where the cirrus formulation assumes an emissivity of unity. The blackbody assumption precludes the possibility of cloud-emissivity climate feedbacks. The surface-hydrology cycle uses model-derived rates of precipitation, evaporation and sublimation to simulate the changes of soil moisture and snow cover, as described in Washington and Meehl (1984) and Meehl and Washington (1988), whose treatment is similar to that of Manabe (1969).

Figure 1. Schematic of vertical layer structure of coupled atmosphere-ocean model depicts various physical processes and interactions.
The simple mixed-layer ocean has a thickness of 50 m with no transports or diffusion into the sides or bottom of the layer. The heat fluxes into and out of the top of the ocean's mixed layer are the sum of radiative, latent and sensible heat fluxes. As noted in the introduction, several modeling groups (e.g., Hansen et al., 1984; Wilson and Mitchell, 1987) have used a horizontal ocean heat flux which also compensates for atmospheric model deficiencies so as to keep the model climate close to the observed. It is unclear how this affects the model sensitivity to climate change, however, since such flux correction limits the ocean feedback mechanisms.

The OGCM uses primitive equations for a hydrostatic Boussinesq ocean with a rigid lid and follows a numerical method developed by Bryan (1969) and Semtner (1974, 1987). Large-scale bottom topography has been included. This OGCM was used in an earlier asynchronously coupled prototype model (Washington et al., 1980). Various sensitivities of the OGCM have been analyzed by Meehl et al. (1982). If adjacent vertical layers become statically unstable, then convective adjustment or overturning is invoked, thus returning the respective layers to a neutral state by uniformly mixing temperature and salinity in the two layers. Table 1 summarizes the major characteristics of the OGCM.

### Table 1. World ocean model characteristics.

<table>
<thead>
<tr>
<th>1) Vertical coordinate</th>
<th>depth</th>
</tr>
</thead>
<tbody>
<tr>
<td>2) Horizontal domain:</td>
<td>global (latitude and longitude)</td>
</tr>
<tr>
<td>3) Horizontal resolution</td>
<td>$5^\circ$</td>
</tr>
<tr>
<td>4) Horizontal approximaton of derivatives</td>
<td>second-order</td>
</tr>
<tr>
<td>5) Vertical resolution</td>
<td>$\Delta z_1 = 50$ m, $\Delta z_2 = 450$ m, $\Delta z_3 = 1500$ m, $\Delta z_4 = 2000$ m</td>
</tr>
<tr>
<td>6) Time step</td>
<td>30 minutes</td>
</tr>
<tr>
<td>7) Topography</td>
<td>included</td>
</tr>
<tr>
<td>8) Convective adjustment</td>
<td>included</td>
</tr>
<tr>
<td>9) Salinity flux at surface</td>
<td>included</td>
</tr>
<tr>
<td>10) Computed sea ice</td>
<td>included</td>
</tr>
</tbody>
</table>

Small-scale features — mid-ocean eddies and concentrated aspects of the western boundary currents — are not resolvable by the $5^\circ$ latitude-longitude horizontal grid. The parameterization of subgrid-scale processes is discussed in Washington and Meehl (1989). Sea ice forms when ocean temperature becomes equal to the approximate freezing point for sea water. The amount of growth and melting of sea ice depends upon the energy balance at the top of sea ice, as described in the zero-layer model of Semtner (1976).

Atmospheric and oceanic models essentially are coupled synchronously and communicate with each other once every model day. For the mixed-layer ocean model, the sum of surface-energy fluxes is used at the top of the layer, and for the OGCM the atmospheric model provides wind stress, precipitation minus evaporation and the sum of the surface-energy fluxes to the top of the ocean. The ocean models furnish surface temperature and sea ice data to the atmospheric model (interactions shown schematically in Fig. 2 for the OGCM). Several modeling groups studying CO$_2$ sensitivity have applied constraints to the coupled system that “adjust” the model to remove or minimize climate drift or bias. We chose not to apply such a constraint because of its possible effect on the simulated climate change.
SYNCHRONOUS COUPLING METHOD

ATMOSPHERIC GCM

OCEAN SURFACE TEMPERATURE AND SEA ICE HELD FIXED FOR ONE DAY

WIND STRESS, PRECIPITATION MINUS EVAPORATION, HEAT FLUX INTO OCEAN HELD FIXED FOR ONE DAY

OCEAN GCM AND SEA ICE PREDICTION

Figure 2. Schematic of synchronous coupling method for atmospheric and oceanic sea ice models.

The experiments with a mixed-layer ocean were started from the end of the annual-mean solar-forcing experiment, as described in Washington and Meehl (1984). The doubled CO₂ (2xCO₂) and control (1xCO₂) experiments were run through 11 years, with the last three years of the experiment used in the time average of the results that follow.

Figure 3 describes how the OGCM experiments were conducted with 1xCO₂, 2xCO₂ and slowly increasing CO₂ (or transient forcing). The ocean model started from the end of the uncoupled model experiment with observed atmospheric forcing (approximately equal to 50 years), where the first year was forced by a Newtonian-type observed zonally averaged temperature and salinity distribution (Meehl et al., 1982). This forcing decreased the time required to bring the overall zonal temperature and salinity patterns to roughly that of the observed. The atmospheric part of the coupled mode began from the last year of Washington and Meehl's (1984) climate simulation, which used a simple mixed-layer ocean. The coupled model system was run for 16 years with several changes in ocean diffusion parameters. The model was then run for an additional 30 years for the 1xCO₂, 2xCO₂ and transient experiments.

Although the overall system is far from complete equilibrium, the top parts of the ocean model appear to change little from year to year. Figure 4 shows the time evolution of globally averaged ocean surface temperature and ocean surface temperature differences from the coupled model with 1xCO₂, 2xCO₂ and transient forcing (slowly increasing) CO₂ OGCM experiments. The globally averaged ocean temperature in the top model
Figure 3. Description of running of coupled experiments. At the left, the atmospheric and oceanic models were run separately and then coupled for 16 years. At the right, the control (1xCO2) was extended for 30 years, and an instantaneous 2xCO2 experiment and a transient CO2 experiment were run for 30 years, with CO2 increasing one percent per year.

layer with 1xCO2 exhibits a very slow cooling trend or climate drift (approximately equal to 0.02°C per year), whereas the 2xCO2 experiment shows little globally averaged drift. The climate drift of the coupled model is substantially smaller than the increased CO2 warming signal. The transient experiment depicts a small warming by year 30. In the 2xCO2 experiment, the upper part of the ocean warms quickly and shows a smaller secular change in the 15- to 30-year range. The globally averaged SST difference (Fig. 4) between the 2xCO2 and 1xCO2 experiments is about 1.2°C and 1.6°C for globally averaged surface air temperature at 30 years, which is larger than the value found at 20–25 years by Bryan et al. (1982) and Schlesinger et al. (1985). The transient experiment shows a much smaller difference of about 0.7°C after 30 years for surface air temperature increase and about 0.3°C to 0.4°C for globally averaged SST differences. See studies with simpler energy-balance climate models (e.g., Schneider and Thompson, 1981; Thompson and Schneider, 1982) where equilibrium and transient behavior are explored.

The depth-versus-latitude pattern of climate drift in the ocean for the three cases is similar, with small-amplitude changes occurring in the surface layer, cooling of the second layer (less than 0.05°C per year), a warming of the third layer (less than 0.03°C per year), and a slight warming of the bottom layer (less than 0.005°C per year). Most of the climate drift is removed by subtracting the control case averaged over the same time period.

3. SIMULATION OF OBSERVED CLIMATE

Given the present state of coupled model simulations, many of the errors in ocean surface temperatures can be attributed to a limited knowledge of cloudiness, radiation and surface transfers of sensible heat, momentum, salinity and moisture. Nevertheless, the basic
aspects of the climate system can be simulated. Figure 5 compares the latitude-versus-pressure distribution of zonally averaged temperature from the two coupled models with the observed data of Newell et al. (1972) for December-January-February (DJF) and June-July-August (JJA). The mixed-layer ocean model shows vividly the effects of lack of ocean dynamics on model climate since the surface temperatures are too warm in the tropics and too cold in the polar regions. Conversely, in the OGCM experiment, temperatures near the surface in the tropics are cooler by up to 5°C in the zonal average than the
Figure 5. Zonal-mean temperatures as a function of latitude and height (K). Computed values are three-month seasonal averages taken over the last three (mixed layer) and five (OGCM) years of the respective control cases. Observed values are long-term, three-month seasonal means from Newell et al. (1972): (a) mixed layer for DJF; (b) OGCM for DJF; (c) observed for DJF; (d) mixed layer for JJA; (e) OGCM for JJA; (f) observed for JJA.
observed temperatures in the tropics, and in the higher-latitude Southern Hemisphere (SH) ocean temperatures are substantially warmer than the observed in both seasons. The OGCM yields cooler tropical and subtropical ocean temperatures than the mixed-layer ocean version, principally because equatorial upwelling and substantial poleward heat transport are not present in the mixed-layer model. Overall, the tropospheric and stratospheric simulations are similar in both ocean models, with one important exception — the overall tropospheric latitudinal temperature gradient in the OGCM is less, partly due to the effects of the ocean poleward heat transport (heat from low latitudes is exported to high latitudes).

Through the horizontal temperature gradient, which can be related to the vertical wind shear via the thermal wind relationship, the zonal wind maxima are smaller in mid- and high latitudes in the simulation with the OGCM. Figure 6 shows the zonally averaged zonal wind component as a function of latitude and pressure for the two seasons. During DJF, some features with the OGCM simulation are closer to the observed of Newell et al. (1972) and van Loon et al. (1971) than to the model with the mixed-layer ocean (Washington and Meehl, 1984). In the DJF simulation, there is a continuous distribution of easterlies throughout the mid-troposphere in the OGCM and the midlatitude westerly jet streams are weaker, a result of the warmer-than-observed high-latitude temperature distribution.

The surface air temperature geographical distributions for both coupled models are shown in Fig. 7. In the Atlantic and Pacific Oceans, the Northern Hemisphere (NH) winter shows more northward-shifted zonal temperature gradients with the OGCM than the experiment with the mixed-layer ocean. Temperatures are substantially warmer over the North Atlantic and near the edge of Antarctica. Over the tropical ocean, surface temperatures in the present model are more than 5°C too low in places. Continental temperatures over North America, Asia, and Africa are similar in the two simulations. The simulation of less-than-observed sea ice distribution, noted by Washington and Meehl (1989) for the OGCM, is caused by substantially warmer-than-observed surface air temperatures in the SH, a result of large horizontal ocean heat transport. At high southern latitudes, this poleward heat transport is mainly attributed to horizontal diffusion which must be kept at a large value to maintain computational stability in the coarse-grid ocean model (Washington and Meehl, 1989).

The geographical surface salinity distribution from the OGCM is compared with annual-mean observations from Levitus (1982) in Washington and Meehl (1989). In the coupled model, because river runoff is not considered, there is no fresh-water input to the oceans from the land areas. In spite of the simplifications and limitations of the model, most of its major features are simulated. This can be explained partly by the fact that salinity fluxes at the top of the ocean contribute little to the total salt content of the oceans over the time scale of these experiments (50–60 years). The salinity distribution is still heavily influenced by the initial distribution. The subtropical maxima of salinity in the North Pacific and North Atlantic are simulated, including the maxima in the North Atlantic, while the Antarctic circumpolar and Arctic Ocean minima are also simulated.

The latitude-depth distribution of the zonally averaged temperature for DJF from the model is shown in Fig. 8 and compared with the observed of Levitus (1982). JJA is not shown because it is similar, except near the poles. Even though the general pattern is simulated, the permanent thermocline is not as sharp as the observed, which can be partly explained by the coarse vertical resolution. In the equatorial region, surface temperatures are too cold and the simulated surface temperatures near 60°S are too warm. The latitude-depth distribution of salinity (Fig. 9) shows that the zonal-mean surface features are reproduced with relatively higher salinity in the subtropics and lower salinities at higher latitudes and in the equatorial tropics. Beneath the surface layer are discrepancies mainly...
Figure 6. Mean zonal (east-west) wind component $u$ as a function of latitude and height (m s$^{-1}$). Computed values are compiled as in Fig. 5. Observed SH tropospheric values are geostrophic winds for (c) January and (f) July, from van Loon et al. (1971). All other observed values are three-month seasonal means from station data (Newell et al., 1972): (a) mixed layer for DJF; (b) OGCM for DJF; (c) observed for DJF (SH, January); (d) mixed layer for JJA; (e) OGCM for JJA; (f) observed for JJA (SH, July).
Figure 7. Geographical distribution of surface air temperature (degrees K; $\sigma = 0.991$). Computed values are compiled as in Fig. 5. Sea ice values are stippled for mixed layer and observed. (a) mixed layer for DJF; (b) OGCM for DJF; (c) observed for DJF; (d) mixed layer for JJA; (e) OGCM for JJA; (f) observed for JJA.
Figure 7. Continued.
Figure 8. Zonally averaged (a) computed temperature (DJF, °C); (b) observed annual-mean temperature (Levitus, 1982; °C). Contours every 2°C. Computed JJA patterns are qualitatively similar.

involving the relatively lower salinity values in the SH mid- and high latitudes. The tongue of highly saline water at 35°N is due to inclusion of the Mediterranean Sea in the observed which is not included in the model.
Figure 9. Zonally averaged (a) computed salinity (o/oo) for DJF; (b) observed annual-mean salinity (Levitus, 1982). Computed JJA patterns are qualitatively similar. Note the tongue of high salinity at 35°N in observed caused by the Mediterranean Sea which is not included in the model ocean configuration.

4. DIFFERENCES IN 2×CO₂, TRANSIENT CO₂ AND 1×CO₂ EXPERIMENTS

Most climate simulations of greenhouse warming have been with an instantaneous doubling of CO₂. This section describes such experiments with the NCAR coupled models as well as an additional experiment of increasing the CO₂ concentration by one percent per year.
The associated zonal-mean temperature changes in the atmosphere for a doubling of CO₂ are shown in Fig. 10 for both types of ocean models. In the NH of the OGCM, the reduction of sea ice during DJF is associated with a maximum warming of 5.0°C at the surface near 75°N, largely a result of the ice-albedo feedback. This feedback process, which shows largest change in winter, is caused by increased solar absorption during the summer months (Washington and Meehl, 1984). Similar ice-albedo effects are not evident in the SH (Fig. 10d) because of the near absence of sea ice around Antarctica in the OGCM. Both winter and summer temperature differences with the mixed-layer ocean show a warming of about 2–4°C throughout the troposphere, with strong upper tropospheric warming in the tropics (Fig. 10a,c). At most latitudes, stratospheric cooling increases above about 18 km to at least 5–6°C at the top of the model. Warming at the surface is roughly 2–3°C, except at high latitudes in the winter hemispheres. During DJF, warming increases to more than 7°C near 65°N, and a warming of about 13°C occurs near 60°S during JJA (Fig. 10a,c). Both temperature maxima are near the sea ice margin. The vertical pattern of atmospheric temperature change in Fig. 10 in the mixed-layer version is similar to that of the OGCM.

Figure 10. Zonal-mean atmospheric temperature differences (°C), 2×CO₂ minus control (1×CO₂): (a) mixed layer for DJF; (b) OGCM for DJF; (c) mixed layer for JJA; (d) OGCM for JJA.
Figures 11a and b depict latitude-depth plot of zonal-mean differences, 2x CO$_2$ minus control, for (a) temperature and (b) salinity for DJF. The ocean warming has penetrated into the third model layer with largest concentration near the surface (1.7°C near 60°N and 1.3°C around 50°S). Schlesinger and Jiang (1988) found similar maxima in their coupled model simulation. In addition, the minima of warming in the ocean in the tropics and high latitudes are quite different from those in the ocean mixed-layer model results. The zonal-mean salinity differences (Fig. 11b) also show large changes in the surface layer of the ocean where salinity decreases in the midlatitudes and high latitudes. The OGCM shows salinity increases in the subtropics and a decrease near the equator. These changes in salinity indicate alterations of precipitation minus evaporation, as well as changes of vertical mixing. Warmer SSTs caused by CO$_2$ doubling produce increased evaporation at most latitudes. Precipitation increases correspond to areas of salinity decreases, and vice versa for precipitation decreases. Washington and Meehl (1989) discuss the effects on salinity distributions of lack of river runoff.

Geographical difference patterns of surface air temperature change, 2x CO$_2$ minus control, for the two ocean types are shown in Fig. 12 for DJF and JJA. The temperature differences are much larger in the mixed-layer ocean model, with the largest values over the edges of the sea ice, whereas the differences with the OGCM are smaller and not obviously associated with sea ice position. As noted earlier, because the control simulation with the OGCM has much less sea ice, especially in the SH, there should be less sea ice albedo feedback on temperature. Although the difference satisfies a significance test (Chervin, 1981; Washington and Meehl, 1984, 1989), the results are sensitive to the control climate. The transient simulation exhibits a different geographical pattern for reasons other than dissimilar ocean models, as will be discussed later.

In Fig. 12, the large warming in areas of sea ice reduction is statistically significant during DJF. During JJA, significant warming is not as uniform over the NH. The CO$_2$-caused warming in the SH is smaller — about 1–2°C. The small areas of cooling are not statistically significant. With the earlier set of experiments, Washington and Meehl (1983) demonstrated that cooler SSTs in the tropics contributed to reduced atmospheric moisture amounts and less sensitivity. Other groups have found similar dependence of sensitivity on the control temperature distributions.

It is noteworthy that, after 30 years, this model produces less CO$_2$ warming than the simple mixed-layer ocean version. Lower tropical OGCM SSTs are associated with reduced atmospheric moisture, and the warmer high-latitude SSTs result in less-extensive sea ice. Both factors contribute to less feedback and lower sensitivity.

Ice albedo as a feedback is important in that more extensive sea ice leads to greater model sensitivity (Washington and Meehl, 1986). Estimates for the simple mixed-layer ocean version show that ice-albedo feedback contributes more than a degree to the 3.5°C global warming (Dickinson et al., 1987).

Changes in soil moisture for the two ocean types are qualitatively similar (Fig. 13). During winter over the NH midlatitude continents, there are mostly increases of soil moisture, and during the summer season in these regions there are areas of both increase and decrease. The OGCM allows for circulation changes in the ocean, a distinct advantage over the mixed-layer ocean. From a comparison of the zonally averaged wind stress over the ocean for the control, the 2x CO$_2$ cases and the observed of Hellerman and Rosenstein (1983), Washington and Meehl (1989) found that the model agreed well with the observed. Washington and Meehl (1989) also show that changes in wind stress have a profound effect on the tropical ocean temperatures, such that a reduction in wind stress can result in a warming of the equatorial waters by several degrees. This is due to the close association in the model of wind stress, upwelling and SST. That is, reduced wind stress results in weakened upwelling and higher SSTs. In the midlatitudes, the doubled
CO2 experiment exhibits a noticeably reduced westerly wind stress of 0.03 N m\(^{-2}\) and a warming of the ocean surface layer at those latitudes.

The changes in the zonal-mean mass circulation in the ocean caused by doubling CO2 are large (Washington and Meehl, 1989). Figure 14a shows the global meridional overturning from the control experiment during DJF (Washington and Meehl, 1989), and Fig. 14b shows the difference in circulation between the 2xCO2 and 1xCO2 experiments. Unfortunately, there are no reliable observations of this quantity with which to compare the model results. Negative values indicate counterclockwise circulation and positive values are clockwise. In the control, near the surface, are shallow wind-driven cells, with
deeper overturning near 40–50° latitude below the wind-driven surface cells in both hemispheres. A weak upward mass circulation is seen near 60°S.

The difference of meridional streamfunction (Fig. 14b) has positive values with anomalous clockwise circulation in the NH and the converse in the SH with anomalous counterclockwise circulation. Because of the reduction of the westerly wind stress in the 2xCO₂ experiment, the shallow, wind-driven midlatitude cells weaken in both hemispheres. The deep-ocean midlatitude cells also weaken. The warming and freshening of the surface layer are especially evident near 50°N where sinking is weaker in the deep ocean. Washington and Meehl (1989) examine this process more closely for the North Atlantic.

Bryan and Spelman (1985) showed, in a coupled ocean-atmosphere sector model, a similar partial collapse of the ocean thermohaline circulation with a quadrupling of CO₂ after 25 years of simulation due to a warming and freshening of the upper ocean from the effects of increased CO₂. Bryan (1986) further demonstrated with specified atmospheric
forcing that changes of the thermohaline circulation in a basin-ocean model are associated with changes in the surface salinity forcing. Moreover, these salinity perturbations can cause rapid mass circulation changes. The Washington and Meehl (1989) results with a coupled global ocean-atmosphere model with realistic geography and bottom topography lend support to other model results. These experiments indicate that warming and freshening of the high-latitude surface layer make the oceans more stable and weaken the thermohaline circulation on relatively short time scales (tens of years). The weakening of the westerlies in the 2×CO₂ experiment is associated with decreased overturning between the upper layers of the model.

Unique to the OGCM is the possibility of convective overturning. When this occurs in the OGCM, there are rapid vertical fluxes of heat and salinity. The convective mechanism of the ocean model devised by Bryan (1969) and used in the OGCM specifies that, if the densities in two adjacent layers are statically unstable, then the temperature and salinities in the two layers are uniformly mixed, yielding a neutral stratification. The process is repeated for the next pair of layers. Depending upon the preexisting condition, this can mean an upward flux of heat or downward flux of salt or, under certain conditions, both. Killworth (1982) has shown the geographical locations of deep convection in the observed ocean and Reid (1965) for locations of intermediate waters. Because
the model does not simulate all of the observed temperature and salinity distributions, particularly at the higher latitudes in the North Atlantic, errors exist in some of the locations. The control ($1 \times CO_2$) and the $2 \times CO_2 - 1 \times CO_2$ difference of convective overturning between levels 1 and 2 are shown in Fig. 15 for the late winter season in each hemisphere—Feb–Mar–Apr for NH, and Aug–Sept–Oct for the SH. Convective overturning, less frequent between layers 2 and 3, rarely occurs between layers 3 and 4. If convective overturning occurs every time during the year, the value is unity in Fig. 15; if it occurs half the time, the value is 0.5.

Convection occurs mainly in the fall and winter of both hemispheres, suggesting that in the model it is predominantly temperature-driven rather than salinity-driven. Important exceptions occur in the Mediterranean, Red and Caspian Seas where convection takes place year-round. In these regions the large evaporation causes a salinity-driven overturning of the first two layers.
The $2\times CO_2 - 1\times CO_2$ difference (Fig. 15) shows a substantial decrease in convection, particularly in the higher latitudes, indicating an increased stratification of the model, as expected with a CO$_2$ warming.

Another major difference between the mixed-layer ocean model and the OGCM is the presence of ocean-heat transport in the latter. This quantity consists of: (1) the meridional overturning component; (2) the gyre component, such as the midlatitude gyres in the North Pacific and North Atlantic; and (3) subgrid-scale lateral diffusion. Figure 16a shows the components from the control case, and the total values can be compared with observed latitudinal estimates from Oort and Vonder Haar (1976), Trenberth (1979), Hastenrath (1980), and Carissimo et al. (1985) in Fig. 16b. In low and subtropical latitudes, the largest horizontal heat transport is by meridional overturning, and in the middle and high latitudes the largest contributor is diffusion transport, especially in the SH. Employing a smaller horizontal diffusivity for heat yields more heat transport by meridional and gyre components (Meehl et al., 1982), but the total transport remains the same. Cox (1985) has shown similar compensations with higher-resolution models in which ocean eddies are resolved. Increasing ocean resolution, therefore, decreases lateral diffusion heat transport and increases the gyre component. Figure 16c depicts the total annual-mean horizontal ocean heat transport for $2\times CO_2$ and $1\times CO_2$. Transport decreases at the midlatitudes,
large part as a result of a decrease in diffusion caused by a slackening of the north-south oceanic temperature gradient. The small increase of poleward transport near 10°S is associated with a somewhat stronger meridional circulation in the model's second layer.

The annual cycles of ocean heat storage among the observed, the mixed-layer ocean model and the OGCM are remarkably similar, thus establishing that a simple mixed
Figure 15. (a) convective adjustment between top two layers in the ocean model for the winter seasons in both hemispheres; a value of 1.0 means convective adjustment occurs every time step; (b) same as part (a) except for the difference 2×CO₂ minus control case.

layer can account for heat storage in the upper ocean at least to the first order (Meehl and Washington, 1985; Washington and Meehl, 1989).

A great deal of interest has been generated in the differences between experiments with instantaneous increase of CO₂ and the effects of a gradual increase of CO₂ with time. Washington and Meehl (1989) extensively discuss such experiments. The present paper concentrates on how the type of ocean model may affect the result. At the end of the transient experiment (years 26–30), the temperature change in the ocean appears similar in some respects to the temperature change in the early part (years 6–10) of the 2×CO₂ experiment where most of the warming is at the top of the ocean (Fig. 17). Small changes, evident in the transient experiment compared to the CO₂ doubling case, are mostly near the surface, with maxima of warming near 45°S and 20°N. Poleward of 30°N, cooling takes place in the zonal mean with a maximum at 65°N and is reflected in the atmospheric
zonal temperature difference (Fig. 18). Apparently, the cooling in the zonal-mean ocean temperatures at the high latitudes in the NH results in a cooling of the atmosphere zonal mean through the troposphere. As discussed in Washington and Meehl (1989), the
circulation changes in the NH of the transient case are associated with the southward advance of the sea ice limit in conjunction with southward advection of cold air and the cooling of SSTs into the North Atlantic region.

The response of the transient case is different because the changes in forcing are much slower and the climate system has a chance to evolve more gradually than the $2\times$ CO$_2$ case. Although some patterns of changes are similar, the transient time evolution allows the ocean to adjust to changes in atmospheric circulation. This strongly suggests
that the $2\times CO_2$ instantaneous experiments may not be reasonable analogues to a gradual increase of $CO_2$. The transient experiment will be discussed further in the next section.

5. DISCUSSION AND CONCLUSIONS

Despite the shortcomings and errors in simulated ocean temperature of the coupled atmosphere-OGCM, it simulates many features of the present climate system. The zonal-
mean patterns of temperatures and winds in the atmosphere are reproduced, as are the
ocean zonal-mean temperatures, salinities, heat transport and heat storage. The most
important errors are colder-than-observed tropical SSTs and warmer-than-observed high-
latitude temperatures. There is very limited extent of sea ice in the Bering Sea and around Antarctica. From the additional parameter studies discussed in Washington and Meehl (1989), the following are important factors that contribute to the SST distribution computed by the coupled model: (1) decreased wind-stress forcing results in less equa-
torial upwelling and warmer tropical SSTs, (2) a decrease in the depth of the second
layer results in warmer water brought to the surface and higher SSTs in the tropics; (3) a
shallower second layer at extratropical latitudes results in cooler SSTs at those latitudes
following similar reasoning; and (4) a decrease in the horizontal heat diffusion results
in less heat diffused out of the low and midlatitudes, warmer tropical SSTs and cooler
high-latitude SSTs.

The coupled model has lower sensitivity to increased CO₂ after 30 years of integra-
tion compared to the coupled model with a 50-m slab-ocean mixed layer (Washington
and Meehl 1984). The mixed-layer model has warmer-than-observed tropical SSTs and
cooler-than-observed high latitude, causing over-extensive sea ice. As a result of warm
tropical SSTs and over-extensive sea ice, the mixed-layer model is very sensitive to CO₂-
induced warming (3.5°C globally averaged surface air temperature change for a doubling
of CO₂). Conversely, the OGCM has cool tropical SSTs, reduced sea ice extent, and a
sensitivity of 1.6°C after 30 years of integration, suggesting that the sensitivity of the real
climate system to a doubling of CO₂ is somewhere between the two sets of experiments.

Patterns of temperature change in the atmosphere are similar for the two types of
ocean with CO₂ instantaneously doubled. Warming occurs in the troposphere with the
largest surface air temperature change taking place where sea ice has retreated poleward.
Soil moisture changes are also similar in both sets of experiments.

CO₂ warming in the ocean is confined mostly to the upper portion, especially in
the surface layer near 60°N and 50°S. Through increases in precipitation, weakened
westerly wind stress, and reduced overturning, salinity amounts decrease at high latitudes
of each hemisphere. Salinity also increases in the subtropics. The resultant warming and
freshening of the high-latitude ocean surface layer stabilize the ocean and cause a weaker
thermohaline circulation.

The transient experiment shows the greatest changes to ocean temperatures in the
ocean surface layer, and most of the troposphere warms while the stratosphere cools.
Important differences from the 2×CO₂ instantaneous experiment, such as cooling in the
North Atlantic and, to a lesser extent, the northwest Pacific, suggest that patterns of
climatic change from an instantaneous CO₂ doubling experiment may not be accurate
analogues to climate change from a transient or slowly increasing CO₂ amount.

Analysis of observations (e.g., Karoly, 1989) indicates a secular cooling trend in the
lower troposphere over regions of the North Atlantic and North Pacific, with warming
over the rest of the NH similar to that noted in the transient case (Fig. 19). Over the
North Atlantic, the more recent cold years in that region (1960s through 1980s) have been
characterized by a sea level pressure anomaly pattern similar to that noted in the transient
case for the coupled model (Fig. 20). In particular, the centers of positive SLP anomalies
over the North Atlantic and negative SLP anomalies over the subtropical Atlantic and
northern Europe are similar between the transient case and the more recent observations
(Fig. 20).

The correspondence of recent observed temperature and circulation anomalies with
those simulated during the latter period of the coupled model transient case raises two
intriguing possibilities: (1) both the actual and the simulated coupled system display cli-
mate anomaly regimes on decadal time scales that function in similar ways and the
Figure 19. (a) trend in °C/century of the least-squares fit to the time series of annual temperature anomalies in the lower troposphere (700 mb) derived from radiosonde data. Stippled areas indicate cooling trend (negative values); positive values indicate warming trend (after Karoly, 1989); (b) temperature difference for DJF, transient (years 26–30) minus control for the lower troposphere of the coupled model (σ level = 0.811). Stippled areas indicate cooling at mid- and high latitudes greater than 0.5°.
correspondence is coincidental; or (2) the coupled model is actually simulating the real transient response to the observed slow increase of CO$_2$ presently taking place. Clearly, further inspection of the observed climate regimes in these regions, as well as additional analysis of climatic regimes in the coupled model, is warranted. In addition, because of model shortcomings and failure to reproduce all of the regional aspects of climate, one should be cautious in using the climate changes exhibited by these experiments. Increased resolution and improved physical processes are necessary in the next generation of coupled atmosphere-ocean general circulation models.
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The Southern Oscillation in a Coupled GCM: Implications for Climate Sensitivity and Climate Change

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ABSTRACT. Results are presented from a global coupled ocean-atmosphere general circulation climate model developed at the National Center for Atmospheric Research. The atmospheric part of the coupled model is a global spectral (R15, 4.5° latitude by 7.5° longitude, 9 layers in the vertical) general circulation model. The ocean is a coarse-grid (5° latitude by 5° longitude, 4 layers in the vertical) global general circulation model. The coupled model includes a simple thermodynamic sea-ice model. Due mainly to inherent limitations in the ocean model, the coupled model simulates sea surface temperatures that are too low in the tropics and too high in the extratropics in the mean. In spite of these limitations, the coupled model simulates active interannual variability of the global climate system involving signals in the tropical Pacific that resemble, in some respects, the observed Southern Oscillation. These signals in the tropics are associated with teleconnections to the extratropics of both hemispheres. The implications of this model-simulated interannual variability of the coupled system relating to climate sensitivity and climate change due to an increase of atmospheric carbon dioxide are discussed.

1. INTRODUCTION

To understand climate change resulting from an increase of carbon dioxide (CO₂) and other trace gases, models must simulate the processes observed in the climate system. Numerous observational studies (as discussed by other papers in this volume) have shown that the Southern Oscillation (SO) and attendant El Niño (warm events) and cold events play an important role in global climate variability. El Niño-Southern Oscillation (ENSO) phenomena involve the coupled ocean-atmosphere system and display global teleconnections. Therefore, to reproduce these processes a model must simulate the mechanisms involved with the dynamically coupled ocean-atmosphere system. In the past few years,
advances in computer power and climate modeling capability have enabled several modeling groups to undertake extended integrations with global coupled ocean-atmosphere climate models. At least three of the coupled general circulation models (GCMs) (Sperber et al., 1987; Meehl, 1989a, 1990; Philander et al., 1989) show that Southern Oscillation (SO)-like variability is present in some form in the coupled simulations. For the first time, global coupled models are indicating that aspects of the SO may be inherent in the dynamically coupled climate system and can be studied with such models. The implication, then, is that changes in the processes associated with the SO in the coupled models can be examined in a simulated climate with increased CO$_2$.

As shown by Meehl (1990) and discussed by Sperber et al. (1987) and Philander et al. (1989), however, the current generation of global coupled GCMs is probably portraying only a subset of all the processes that take part in the observed SO. As such, the models are a reflection of the current level of understanding of the observed ENSO, and many of these mechanisms are not well understood.

The purpose of this paper is to examine the NCAR global coupled GCM simulation (the control experiment with present amount of CO$_2$ as described by Washington and Meehl, 1989) to: (1) determine if ENSO phenomena are present, (2) determine how ENSO phenomena are represented by the coupled GCM compared with observations, and (3) elucidate the mechanisms in the model that are producing ENSO phenomena and decide if analogues exist in the observed climate system. An important goal of this work is to ascertain how such ENSO phenomena change in the integration of the model with increased CO$_2$. These analyses have just begun and will not be discussed here.

2. THE NCAR GLOBAL COUPLED GCM

The NCAR global coupled ocean-atmosphere GCM was illustrated schematically earlier in this volume by Washington and Meehl (their Fig. 1). The atmospheric version of the NCAR community climate model (CCM) has been used specifically for coupling to various ocean formulations (e.g., Washington and Meehl, 1983, 1984, 1989). It is a global, spectral GCM and includes rhomboidal 15 (R15) resolution (about 4.5° longitude by 7.5° latitude), nine layers in the vertical, computed clouds, and parameterized land-surface processes. Meehl and Washington (1988) have analyzed in detail the sensitivity of land-surface processes in this model.

The ocean model has a coarse resolution (5° latitude by 5° longitude, four layers in the vertical) and includes realistic geography, bottom topography consistent with resolution, and a simple thermodynamic sea-ice formulation. Meehl et al. (1982) have tested this ocean model with a variety of parameter changes.

The strategy of spinup and experimental design was discussed earlier in this volume by Washington and Meehl. (See their Fig. 2 for coupling method and their Fig. 3 for a schematic of the experimental design.) Because the model was not originally designed to study long-term time series, only the last ten years of the control run (with present amounts of CO$_2$) were available for the present analysis. The NCAR coupled model does not contain flux correction or flux adjustment (Latif et al., 1988; Manabe and Stouffer, 1988). Therefore, simulated sea surface temperatures (SSTs) in the tropics are cooler than observed. This cool water from the active upwelling, coarse vertical resolution and simplified geography in the model extends across the equatorial Pacific Ocean to the Indian Ocean with warmer water to the north and south in the tropics (see Washington and Meehl, 1989, Fig. 10). Meehl (1989b) analyzed the source of this coupled-model error for the tropical Indian and Pacific regions and concluded that the inherent limitations of the ocean model in resolution and geography contribute most to these errors. A slight drift of about $-0.02^\circ$C/year is evident in the surface layer of the ocean. For the ten-year period under consideration here, this drift is small compared to the signals of SST variability with amplitude on the order of 1°C.
3. REPRESENTATION OF EI NIÑO-SOUTHERN OSCILLATION IN THE COUPLED MODEL

In spite of the inherent limitations and errors in the simulation mentioned above, the interannual variability is pronounced in the tropical Pacific in the NCAR coupled model. As a first step in analyzing these fluctuations, area averages of sea-level pressure (SLP) anomalies (seasonal mean SLP minus the ten-year seasonal means in the model) for an area in the eastern Pacific (“Niño3”; 90°W to 150°W, 10°N to 10°S) and an area over the Indonesian region (110°E to 155°E, 10°N to 10°S) are plotted for years 21 to 30 in the coupled model (Fig. 1b). This can be compared to a similar plot of observed SLP anomalies for Tahiti and Darwin (Fig. 1a). The familiar fluctuation of atmospheric mass, characteristic of the SO (as indicated by the opposite sign of SLP anomalies between the eastern and western Pacific), is evident in both model and observations, with the amplitude of the oscillations in the model less than the observed.

Figure 2 shows a time-longitude plot of SST anomalies in the model from 50°E in the Indian Ocean to 80°W in the Pacific, averaged from 10°N to 10°S. For this ten-year period in the model, SST anomalies generally appear in the eastern Pacific and become established progressively farther west. Maximum amplitudes are roughly ±1°C. The magnitude of the SST anomalies in the coupled model is about 30% less than the observed, composite warm-event SST anomalies shown, for example, by Rasmusson and Carpenter (1982). The establishment of SST anomalies greater than ±0.5°C between about 110°W and the Dateline in the model occurs near the seasonal transition between March–April–May (MAM) and June–July–August (JJA). The horizontal lines in Fig. 2 denote this seasonal transition. For example, northern spring seasonal transitions from warm to cold SST anomalies in this area occur in years 24 and 27, while similar seasonal timing for transitions from cold to warm occurs in years 22 and 26. SST anomaly events with a duration of one year are present in the model near 110°W and the Dateline (e.g., the warm event starting in MAM of year 26 and ending after MAM of year 27). Two-year warm and cold events also occur (e.g., a warm event lasting from MAM year 22 to MAM year 24 and a cold event lasting from MAM year 24 to MAM year 26) and are analogous to single-year and multiyear events observed in the tropical Pacific (van Loon, 1984).

Warm-water events in the model eastern Pacific (Fig. 2) correspond with low SLP in the Niño3 region, high SLP over Indonesia (Fig. 1), and a negative SOI (Southern Oscillation Index, not shown), and vice versa for cold events. This is similar to the association between Tahiti and Darwin SLP and warm and cold events observed in the Pacific. Like the observed ENSO phenomenon, the model SO involves processes of the dynamically coupled ocean-atmosphere system. Time-longitude plots similar to the SST plot in Fig. 2 for SLP, precipitation and surface wind stress show a close association in the model in time and space among positive SST anomalies, negative SLP anomalies, positive precipitation anomalies, and westerly wind-stress anomalies to the west (i.e., wind-stress convergence in the areas of warmest SSTs), and vice versa for negative SST anomalies (see Meehl, 1990, Fig. 5).

4. COMPOSITE ENSO PHENOMENA IN THE COUPLED MODEL

To study ENSO in the coupled model, composites are formed for warm and cold events in the model. The year of initiation of a warm event (year zero) is defined as a year with SOI (calculated from gridpoints nearest Tahiti and Darwin in the model) less than −0.5 and SST in the Niño3 area greater than 0.25°C. Year zero for a cold event in the model is defined by SOI greater than 0.5 and SST in Niño3 less than −0.25°C. Two warm-event
Figure 1. (a) SLP anomalies (five-month running mean) at Tahiti (solid line) and Darwin (dashed line) from data supplied by the Climate Analysis Center (CAC). (b) Seasonal SLP anomalies (three-season running mean) from an area in the tropical eastern Pacific ("Niño3": 90°W to 150°W, 10°N to 10°S) and an area over Indonesia in the far western Pacific ("Indonesia": 110°E to 155°W, 10°N to 13°S).

(years 22 and 26) and two cold-event initiations (years 24 and 27) are defined from years 21–30 in the coupled model run. Examination of the seasonal composite anomalies shows that, as in the observed events (Meehl, 1987; Kiladis and van Loon, 1988), the warm- and cold-event patterns are near-mirror images of each other. To emphasize the composite patterns, differences of warm-event seasons minus cold-event seasons are computed. The sign of the difference, a transition from a cold phase to a warm phase, will be discussed in that context. The designations of warmer, colder, higher, lower refer to relative anomalies in the model.

Negative SST differences characteristic of the waning stages of a cold event (Fig. 3a) cover most of the central and western tropical Pacific during December-January-February (DJF) (subscript refers to the year of the initiation of a warm event). Small-amplitude warming is just starting in the far eastern Pacific. Positive SLP anomalies (Fig. 3b) overlie the cool water in the tropical central and western Pacific. In the far eastern Pacific, the small-amplitude positive SSTs are associated with negative SLP anomalies that extend across most of tropical South America. This anomalous SLP gradient is associated with
Figure 2. Time-longitude plot of seasonal mean SST differences from the long-term seasonal means for years 21–30 in the coupled model, 50°E to 80°W, averaged from 10°N to 10°S. Vertical line near center is Dateline; horizontal lines demarcate seasonal boundary between northern spring (MAM) and northern summer (JJA). Stippling greater than +0.5°C, hatching less than −0.5°C.
ageostrophic westerly wind-stress anomalies directed from high to low pressure in Fig. 3c along the equator between about 100°W and 150°W. Similar relationships exist between SST and low-level winds in the observed systems as well (Lindzen and Nigam, 1987), particularly in the eastern Pacific (Gutzler and Wood, 1989). The westerly wind-stress anomalies in the model (weakened easterly Trade Winds) along the equator in that region result in a decrease in upwelling in the ocean (Fig. 3c). Most areas under the influence of westerly anomaly wind stress and suppressed upwelling lie just to the west of the warm SST anomaly.

When warmer water appears in the equatorial eastern Pacific as a result of the weakened easterlies and reduced upwelling in the model, lower pressure overlies this warm water similar to that noted in the observed system in this region (Lindzen and Nigam, 1987). The SLP gradient between the positive SLP anomalies in the central equatorial Pacific and the negative SLP anomalies in the eastern Pacific is established west of the positive SST anomalies, the westerly wind-stress anomalies set up by this surface pressure gradient extend farther west, upwelling is weakened, and SSTs become higher farther west during MAMo (Fig. 4).

This set of coupled processes involving the atmosphere and the ocean continues to establish the positive SSTs farther west in the model until, by JJAn, the positive SSTs almost reach the Dateline and positive SLP extends all the way to India (Fig. 5). There, the positive SLP is associated with a suppression of the Indian Monsoon, as evidenced by the weakened southwesterly flow (easterly wind-stress anomalies) in the Arabian Sea. This is analogous to the observed relationship between warm events in the Pacific and weak Indian Monsoons (Rasmusson and Carpenter, 1983).

The mechanism in the model, then, for the movement of the coupled anomalies from east to west is the establishment of an SLP gradient (consistent with the SST anomaly pattern) that is associated with westerly anomaly wind stresses and suppressed ocean upwelling to the west of the positive SST anomalies. Decreased upwelling means that warmer SSTs form to the west in the model and the entire set of coupled anomalies continues to the west. These mechanisms are associated with thermally direct effects throughout the tropical troposphere (not shown).

The SO-type interannual variability in the tropics involves near-global changes in atmospheric circulation and SST patterns in the coupled model as well. Figure 6 shows composite global patterns of SST (a) and SLP (b) anomalies corresponding to the waning stages of a cold phase in the tropical Pacific (DJF) as a warm phase is about to begin. The analogous anomalies from the observed system appear in Fig. 6c (SST anomalies from December 1988 indicative of general SST anomaly patterns for that time of year during a cold event) and Fig. 6d (composite SLP anomalies for DJF before the inception of observed warm events when low SSTs dominate the central equatorial Pacific) after van Loon (1986). For the SST anomaly patterns, both computed and observed show low SSTs in the subtropical northwest and southwest Pacific, low SSTs south of the Aleutians, and a band of low SSTs near 50–55°S at most longitudes. The corresponding SLP anomaly patterns for computed and observed show high SLP over the areas of cool water in the tropical Pacific just east of the Dateline, low SLP to the north and south in the subtropical Pacific, high SLP near the Gulf of Alaska, low SLP over southern North America, and a band of high SLP near 45°S with low SLP to the south. This latter feature implies a strengthening of the north-south pressure gradient in the southern mid-latitudes, an increase of the mean westerly low-level winds (not shown), and lower SSTs at 50–55°S in both model and observations.

For the season at the peak of a warm event (DJF), both model and observations (Fig. 7) show high SSTs in the equatorial Pacific, low SSTs in the subtropics to the north and southwest, and a band of warm SSTs at most longitudes south of about 45°S. Model and
observed composite SLP anomaly patterns for this season indicate low SLP in the tropical Pacific east of the Dateline, high SLP to the northwest and southwest, a center of low SLP in the north Pacific, low SLP in the Indian and Pacific Oceans near 40°S (model, Fig. 7b) and 45°S (observations, Fig. 7d) with positive SLP anomalies around 55°S (model, Fig. 7b) and 60°S (observations, Fig. 7d). This indicates a weakening of the SLP gradient, reduced westerlies and warmer water near 50° in both model and observations. Clearly, the SST anomaly patterns in the tropical Pacific are associated with global atmospheric and oceanic anomalies in both model and observations.

Returning to the seasonal evolution of composite anomaly patterns in the tropics, Figs. 8a and b show warm-minus-cold-event, composite, time-longitude plots for SST anomalies and u-component, wind-stress anomalies. As before, the sign of the anomalies
is a warm event. Both are averaged from 10°N to 10°S. Cool water just to the east of the Dateline early in the cycle (JJA\(_{t-1}\) through MAM\(_{t}\)) gives way to the large-amplitude continuous positive anomalies (stippled areas) in the eastern Pacific in MAM\(_{t}\) (Fig. 8a). These positive SST anomalies with narrow longitudinal extent move west as warm water is established farther west. This band of warm water is associated with the zero line separating westerly and easterly wind-stress anomalies in Fig. 8b (i.e., \(u\)-wind-stress convergences). The establishment of easterly anomaly wind stresses in the far eastern Pacific in JJA\(_{t}\) (Fig. 8b) is associated with a reduction in amplitude of the positive SST anomalies there. By SON\(_{t}\), the sign of the SST anomalies is negative around 90°W in association with the strong wind stresses near that longitude.

Comparable time-longitude plots of composite observed warm events (1957, 1965, 1972 events from Rasmusson et al., 1986) for SST and surface \(u\)-wind anomalies are shown in Figs. 8c and d (composite SST and wind anomalies for six warm events—1951, 1953, 1957, 1965, 1969, 1972—are given by Rasmusson and Carpenter (1982), Fig. 22, and show a similar pattern). The seasonal timing between model and observed is comparable in the eastern Pacific, with the appearance of large-amplitude positive SST anomalies in northern spring of year zero. These positive anomalies then become established farther

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**Figure 4.** As in Fig. 3, except for MAM\(_{t}\).
west to the Dateline by the following northern winter (JAN+1). The pattern of low-level
\(u\)-wind (observed) and \(u\)-wind-stress (model) anomalies is also similar in the eastern
Pacific, with the zero line dividing westerly anomaly and easterly anomalies (indicative
of \(u\)-wind-stress convergence) coinciding with the establishment in the eastern Pacific (and
movement westward) of large-amplitude SST anomalies beginning in northern spring of
year zero. (For comparison, largest \(u\)-component, westerly surface wind anomalies from
the coupled model are about 2.0 m sec\(^{-1}\).)

Two interesting differences between the composite model and observed warm events
emerge, however. First, observed positive SSTs and westerly wind anomalies in the west-
ern Pacific at the beginning of year zero migrate eastward (Figs. 8c,d) and meet the
anomaly signals from the east in the central Pacific around northern summer of year
zero. The coupled model shows no comparable features in the western Pacific. Second,
the warmest water in the composite events in the far eastern Pacific in the observations
(Fig. 8c) occurs in northern fall of year zero at a time when easterly anomaly winds are
established there (Fig. 8d). In the model, easterly anomaly wind stress appears with
similar seasonal timing in the far eastern Pacific as noted earlier, but the SSTs immedi-
ately cool under the influence of these easterly anomaly wind stresses in the model. This
confines the positive SSTs in the model warm events to a narrow longitudinal band
Figure 6. (a) Global SST differences for the warm-minus-cold-event composites in the coupled model for DJF₀, as discussed in the text. Sign of differences is representative of a cold episode in the central equatorial Pacific prior to the beginning of a warm episode. Stippling indicates cooler water. (b) Same as (a), except for SLP. Stippling indicates higher pressure.

(Fig. 8a) compared to the wide extent (80°W to the Dateline) of the observed positive anomalies (Fig. 8c).

Several factors explain the lack of simulation of processes in the western Pacific in the coupled model. Primarily, the coupled model does not simulate the western Pacific warm pool or any of the processes thought important to maintain it there. The model’s failure to maintain warm SST anomalies late in year zero of an event in the eastern Pacific is probably related to the lack of vertical resolution and inability to simulate properly some of the processes involving mixed-layer dynamics. To better illustrate this point,
Fig. 6 (cont.). (c) Observed global "blend" SST anomalies from data supplied by the CAC for December 1988. Stippling indicates cooler water. This period is representative of a cold episode in the central equatorial Pacific. (d) Mean SLP anomalies for the northern winter season prior to the inception of composite warm events. Stippling indicates higher pressure. Period is representative of a cold phase in the central equatorial Pacific. Event composites are differenced from mean SLP that does not contain those event years (van Loon, 1986).

Fig. 9 is a schematic of coupled processes in the eastern Pacific in the present model and those thought to be occurring in the observed system (e.g., Wyrtki, 1975). In the model, easterly wind stress in the equatorial eastern Pacific causes Ekman divergence in the surface layer, and the cool water of the second layer is brought to the surface (Fig. 9a). This cool water is associated with high SLP and weak convection, precipitation
Figure 7. As in Fig. 6, except for computed and observed SST and SLP anomalies representative of a warm event in the central equatorial Pacific.

and upper-level divergence. Westerly anomaly wind stress causes weakened upwelling and warming of the surface layer (Fig. 9b). This is associated with low SLP and active convection, precipitation and upper-level divergence. With the reestablishment of easterly wind stress (Fig. 9c), upwelling of the cool second-layer water resumes, SST is lower, and the associated SLP is again higher with weakened convection, precipitation and upper-level divergence as in part (a).

For the observed system, strong easterly wind stress causes vigorous upwelling of cool water from below the thermocline (Fig. 9d). SSTs are low, SLP is high, with weak convection, precipitation and upper-level divergence. As anomalous westerly wind stresses appear, upwelling is reduced, the surface layer warms and the thermocline deepens.
During this period, the arrival of downwelling Kelvin waves generated by westerly anomaly winds in the western Pacific several months earlier could contribute to deepening the thermocline. Finally, when the easterly wind stress is reestablished in October, upwelling resumes, but its vertical extent is confined to the warm layer above the deepened thermocline. Therefore, warm water is brought to the surface and the SSTs remain high.

The coarse vertical resolution of the ocean model's upper ocean makes the SSTs very sensitive to upwelling and the associated wind-stress forcing from the atmosphere. A weakening of the wind stress from the atmospheric model to 0.25 of its original value results in a reduction in equatorial upwelling by an order of magnitude and an increase in tropical SSTs by about 4-5°C (Washington and Meehl, 1989). However, the timing of the phenomena in the eastern Pacific and the westward movement of the anomalies there
Figure 8. Time-longitude plots showing the evolution of composite warm events: (a) Warm-minus-cold-event composite SST differences (°C) from the coupled model (10°N–10°S). Stippled areas greater than +0.75°C, hatching less than –0.75°C. (b) Warm-minus-cold-event composite $u$-component wind-stress differences (10°N–10°S). Stippling greater than +0.5 dynes cm$^{-2}$ (+0.05 N m$^{-2}$). (c) Observed composite SST anomalies along the equator for 1957, 1965 and 1972 events minus the long-term mean (after Rasmussen et al., 1986). Stippling indicates positive SST anomalies. (d) Same as (c), except for $u$-component surface wind anomalies. Stippling indicates positive (westerly) wind anomalies.

Point to some interesting correspondence between the model and observed. Both appear to be linked to the seasonal cycle in the eastern Pacific.
It is likely that the coarse resolution of the ocean model either distorts or does not adequately resolve internal ocean waves. The westward mode-1 Rossby-wave speed in the ocean model (deduced from mean vertical temperature profiles in the equatorial Pacific) is about 1.0 m s\(^{-1}\). This is close to the theoretical phase speed of about 0.9 m s\(^{-1}\). The mean westward phase speed of SST anomalies in the coupled model (warm and cold anomalies deduced from Fig. 2) is 0.20 m s\(^{-1}\), much slower than the calculated mode-1 westward Rossby-wave phase speed, but about twice as fast as the mean westward surface current speed in the model of about 0.10 m s\(^{-1}\). As mentioned earlier, no well-defined western boundary in the equatorial western Pacific exists in the ocean model because of the coarse model grid and simplified continental outlines. Wave reflection and returning eastward Kelvin waves, therefore, are not factors in the model. Westward advection of the SST anomalies evidently does not control the westward propagation. However, the coupled propagation mechanisms outlined in Figs. 3–5 that have to do with the westward movement of anomalies also play similar roles in the long-term mean seasonal cycle as well, as discussed by Meehl (1990) for the model and Horel (1982) for the observations.
5. SUMMARY

Results from the coarse-grid global coupled model (global, spectral R15, atmospheric GCM coupled to a global, 5° latitude-longitude, four-layer ocean GCM) suggest that the model simulates one aspect of the observed system having to do with interannual variability in the tropical eastern Pacific manifested by a modulation of the long-term mean seasonal cycle there. The coupled processes identified in the model appear to have analogues in the observed system that provide one possible set of mechanisms to explain the transition and seasonal dependence of the SO-type signals involving the coupled system in the eastern Pacific, as outlined by Meehl (1987).

The principal conclusions are:

1. The coarse-grid, global, coupled ocean-atmosphere model displays an inherent and active interannual variability in the tropical Pacific that resembles some aspects of observed ENSO phenomena.

2. Sets of coupled anomalies in the model (SST, SLP, wind stress, convection/precipitation) appear in the far eastern Pacific during northern spring and move west.

3. The movement of the anomalies cannot be explained by westward Rossby-wave propagation (too fast) or mean westward advection (too slow). Instead, the coupled anomalies depend on a kind of coupled propagation whereby SLP gradients, set up at the surface and associated with the precondition of SSTs in the tropical Pacific and the land-sea contrast in the eastern Pacific, produce ageostrophic near-equatorial, westerly wind-stress anomalies and suppressed upwelling to the west. This causes SSTs to warm to the west and the whole system of coupled anomalies is pulled to the west. Horel (1982) has suggested this type of coupled interaction to explain the long-term mean seasonal cycle in the eastern Pacific.

Because a transition from warm to cold phase or back again in the model (and the observations) depends only on the precondition of the SSTs in the tropical Pacific and the passage of the forcing from the annual cycle in northern spring, the system should have an inherent biennial tendency as observed (Meehl, 1987; Kiladis and van Loon, 1988). That is, if there is cool water in the equatorial Pacific in northern spring, the system should make a transition to warm water there, and vice versa if the system starts out with warm SSTs in the tropical Pacific. Yet, as in the real ocean-atmosphere system, multiyear events disrupt the biennial cycle. The cause of these multiyear events is under investigation and appears to involve heat storage in the upper ocean.

Since the observed SO accounts for the largest known source of tropical interannual variability and also is associated with global atmospheric teleconnections, a climate model that does not simulate at least some aspects of the SO cannot be expected to capture these essential elements involved with global climate. The frequency of droughts in the tropics or modulation of monsoon rainfall, as noted for the earlier slab model by Meehl and Washington (1986), must depend critically on the model's ability to simulate the largest-known influence on these events—the SO. Any discussion of changes in tropical circulation or climate, or mid-latitude weather and associated interannual variability affected by the SO, must be qualified by whether or not the SO is reproduced in the model. Even in the present generation of coupled models, as noted in this paper, only certain aspects of the SO are present. Any attempt to study changes in these features in an increased-CO₂ integration must take into account exactly which mechanisms in the model produce such phenomena and which elements are not simulated by the model. All have important
implications concerning what can and cannot be said with any degree of certainty about changes of global climate with increased CO₂.

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Climatic Response to a Gradual Increase of Atmospheric Carbon Dioxide*

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ABSTRACT. The transient response of a coupled ocean-atmosphere model to an increase of carbon dioxide has been the subject of several studies (Bryan et al., 1982; Spelman and Manabe, 1984; Bryan and Spelman, 1985; Schlesinger and Jiang, 1988; Schlesinger et al., 1985; Bryan et al., 1988; Manabe et al., 1990; Washington and Meehl, 1989). The models used in these studies explicitly incorporate the effect of heat transport by ocean currents and are different from the model used by Hansen et al. (1988). Here we evaluate the climatic influence of increasing atmospheric carbon dioxide using a coupled model recently developed at the NOAA Geophysical Fluid Dynamics Laboratory. The model response exhibits a marked and unexpected interhemispheric asymmetry. In the circumpolar ocean of the Southern Hemisphere, a region of deep vertical mixing, the increase of surface air temperature is very slow. In the Northern Hemisphere of the model, the rise of surface air temperature is faster and increases with latitude, with the exception of the northern North Atlantic, where it is relatively slow because of the weakening of the thermohaline circulation.

1. MODEL AND SIMULATIONS

The model we use here consists of general circulation models (GCMs) of the atmosphere and oceans and a simple model of the continental surface that involves the budgets of heat and water. It is a global model and includes realistic geography. The atmospheric component of the model used here is very similar to the model used in by Wetherald and Manabe (1988). The seasonally varying insolation

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at the top of the model atmosphere is taken into account in calculating the solar radiation. Whenever the relative humidity at the gridpoint exceeds a critical value (99%), the gridbox is filled with cloud; otherwise, it is free of cloud. The atmospheric GCM uses a spectral transform method in which the horizontal distributions of predicted variables are represented by spherical harmonics (15 associated Legendre functions for each of 15 Fourier components) and gridpoint values. For vertical-finite-difference calculations, nine unequally spaced levels are chosen. The ocean GCM uses a full finite-difference technique and has a regular grid system with 4.5° × 3.75° (latitude × longitude) spacing and 12 vertical levels. This model is very similar to the model of Bryan and Lewis (1979), except that the model used here incorporated diffusion on isopycnal surfaces. The atmospheric and oceanic components of the model interact with each other continuously through the exchange of heat, water and momentum fluxes.

Two 100-year integrations of the coupled model were performed. In the first integration the atmospheric concentration of carbon dioxide increased by the rate of 1% per year (compounded). The CO2 concentration remains unchanged in the second integration (the control). We have evaluated the influence of gradually increasing atmospheric carbon dioxide upon the coupled system from the difference between the first and second (control) integrations. If the CO2 increase of 1% per year begins in AD 1958, the doubling of CO2 concentration would be achieved around AD 2030. Such an increase of CO2 is larger than its projected increase, but is approximately equal to the present rate of increase of the combined thermal forcing of all greenhouse gases other than water vapor (Ramanathan et al., 1985).

The initial conditions for both integrations have realistic seasonal and geographical distributions of surface temperature, surface salinity and sea ice, with which both the atmospheric and oceanic components of the model are nearly in equilibrium. This quasi-equilibrium condition was obtained by separate time integrations of these two components of the model using observed surface conditions. The convergence of the oceanic component towards equilibrium was accelerated by the method described by Bryan (1984).

To avoid the systematic climatic drift caused by the bias of the model during the two 100-year integrations, the oceanic surface fluxes of heat and water are adjusted seasonally and geographically, but not interannually, using the method described by Manabe and Stouffer (1988). Identical adjustments are applied to both the first and the second integrations defined above. By contrast to the recent studies (Schlesinger and Jiang, 1988; Washington and Meehl, 1989), no systematic trends are evident during the control integration, indicating that the initial condition is close to stable equilibrium and the flux adjustments are working as expected.

2. RESULTS

Figure 1 shows the latitude-time distribution of the CO2-induced change of zonally averaged, decadal-mean surface air temperature during the 100-year experiment. In the Northern Hemisphere the warming of the zonal-mean surface air increases with latitude due partly to the poleward retreat of both snow cover and sea ice which have high surface albedos. The relatively large warming in high northern latitudes is essentially confined to the lower troposphere because of the stable stratification. This is indicated in Fig. 2 which shows the difference in zonal-mean temperature between the two integrations for the seventh decade of the experiment. The ocean warming is essentially confined to the upper layer, except near Antarctica. It also increases with latitude in the Northern Hemisphere, but becomes small near the North Pole which is covered by sea ice.

The meridional profile of the near-surface warming in the Northern Hemisphere is qualitatively similar to the difference between two equilibrium climates of a model with the normal and the above-normal concentrations of atmospheric CO2. (See, for example, Manabe and Stouffer, 1980 and Manabe and Bryan, 1985 — studies that were conducted using an atmosphere/mixed-layer ocean model and an atmosphere-ocean GCM, respectively.)
Over the northern North Atlantic the increase of surface air temperature is significantly smaller than the zonal average. This is evident in Fig. 3 which shows the geographical distribution of the difference in decadal-mean surface air temperature between the two integrations for the seventh decade of the experiment. Because of the increase in runoff and precipitation over evaporation, surface salinity decreases and the stability of the upper ocean layer increases in the northern North Atlantic. Thus, the thermohaline circulation of the North Atlantic is weakened by about 25% by the seventh decade of the experiment. The weakening is shown in Fig. 4 which illustrates the meridional circulation zonally averaged over the entire oceanic segment of a latitude circle. (Note the reduction from 16 to 12 Sv in the Northern Hemisphere.) Because of the weakening of the thermohaline circulation, the northward advection of warm and saline subtropical surface water is reduced, thereby reducing surface salinity further and making the increasing of sea surface temperature relatively small over the northern North Atlantic Ocean as indicated in Fig. 3. The reduction of surface salinity stabilizes the near-surface layer and provides a feedback that further weakens the thermohaline circulation in the northern North Atlantic.
Figure 2. Zonal-mean difference in temperature (°C) of the ocean-atmosphere model between the two integrations. The difference represents the decadal average over years 61-70. The atmospheric data have been interpolated to isobaric surfaces.

In an earlier study, Bryan and Spelman (1985) investigated the influence of an abrupt quadrupling of atmospheric CO$_2$ on the climate of an atmosphere-ocean model with a sector computational domain bounded by the equator and two meridional boundaries. They found that the thermohaline circulation of the model collapsed completely towards the end of the third decade of the experiment. Washington and Meehl (1989) noted that the response of their global model to a linear increase of atmospheric CO$_2$ was a weakening of the thermohaline circulation in the Atlantic. Manabe and Stouffer (1988) found that their global model has two stable equilibria, one with and one without the thermohaline circulation in the Atlantic Ocean. Although we do not observe a complete collapse of the thermohaline circulation here, it is a possibility that deserves further assessment (Broecker, 1987).

In the Southern Hemisphere the CO$_2$-induced warming of surface air decreases with increasing latitude in sharp contrast to the situation in the Northern Hemisphere and to the results of equilibrium experiments. The rate of the warming is particularly slow in the Antarctic circumpolar ocean. A similar feature is noted and analyzed in studies by Bryan...
et al. (1988) and Manabe et al. (1990) that explored the effect of an abrupt doubling of the atmospheric CO2 on coupled models with "sector" and global computational domain, respectively. The latter model is very similar to the present model except that it has annual-mean insolation. Because the fraction of the area covered by oceans in the Southern Hemisphere is much larger than the corresponding fraction in the Northern Hemisphere, the effective thermal inertia of the Southern Hemisphere tends to be larger than in the Northern Hemisphere, thereby accounting for the slower rise of surface air temperature in the former.

Based upon a detailed study of the heat budget by Bryan et al. (1988) and Manabe et al. (1990), there is another important mechanism responsible for the interhemispheric difference of effective thermal inertia particularly in high latitudes. From 45°-65°S the strong surface westerlies induce an Ekman drift current towards the equator. Because of the absence of a continental barrier in the upper oceanic layer, the zonal pressure gradient is small at the latitudes of the Drake Passage and the southward geostrophic return flow is weak beneath the surface mixed layer. Instead, as indicated in Fig. 4, the surface drift currents are compensated by a deep downwelling north of the Drake Passage gap, southward geostrophic flow at the base of the gap, and deep upwelling south of the circumpolar current. The existence of such an upwelling region has been inferred from water mass analysis (Deacon, 1937; Sverdrup, 1942). A deep wind-driven cell is also simulated by a recent high-resolution model of Semtner and Chervin (1988). In addition to the wind-driven deep cell of meridional circulation described above, we note that (Fig. 4) there is another cell circulating in the reverse direction in the immediate vicinity of Antarctica where deep convective overturning prevails, forming the Antarctic bottom water. These deep cells of meridional circulation, together with associated convective overturning and other subgrid-scale mixing processes, enhance the vertical mixing of heat over a thick ocean layer, thereby increasing the effective thermal inertia of the
Figure 4. Streamlines of zonal-mean meridional oceanic circulation (in Sverdrups) averaged over the seventh decade (that is, years 61–70) of the experiment. Top: the control integration with constant CO$_2$ concentration; Bottom: The integration in which the atmospheric CO$_2$ concentration is increased with the rate of 1% per year.
coupled ocean-atmosphere system. This large effective thermal inertia is responsible for the very slow rise of the near-surface temperature in the circumpolar ocean of the model. It is likely that a similar effective vertical mixing is also realized in the actual circumpolar ocean which is characterized by relatively small vertical variations of density and other trace substances such as phosphate and $^{14}$C.

3. CONCLUSION

The marked interhemispheric asymmetry of our result differs from the study of Schlesinger et al. (1985) in which an almost symmetric response was obtained. Recent results of Washington and Meehl (1989) indicate smaller warming of the Southern Hemisphere than in the Northern Hemisphere. However, the asymmetry is much less pronounced than in the present results. This discrepancy may result from the fact that the wind-driven cell in the circumpolar ocean obtained in the earlier studies is much weaker and shallower than the cell simulated by the present model. Obviously, a careful comparison of results is needed to determine the underlying cause of this discrepancy.

Further evaluation of the mechanisms involved in the effect of ocean circulation on the CO$_2$-induced climate change is in progress. The general applicability of the present results will be evaluated by conducting further simulations with different rates of CO$_2$ change and with models of higher computational resolution.

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Preliminary Assessment of the Performance of A Global Coupled Atmosphere-Ocean Model

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ABSTRACT. A low-resolution version of the ECMWF global atmosphere model has been coupled to a global ocean model developed at the Max Planck Institute in Hamburg. The atmosphere model is driven by the sea surface temperature and the ice thickness calculated by the ocean model, which, in return, is driven by the wind stress, the heat flux and the fresh-water flux diagnosed by the atmosphere model. Even though each model reaches stationarity when integrated on its own, the coupling of both creates problems, since the fields calculated by each model are not consistent with the ones the other model has to have in order to stay stationary, because some of the fluxes are not balanced. In the coupled experiment the combined ocean-atmosphere system drifts towards a colder state. To counteract this problem, a flux correction has been applied which balances the mean biases of each model. This method almost eliminates the climate drift of the coupled model. Problems still arise over ice covered regions.

1. INTRODUCTION

The confidence in model simulations of the greenhouse effect of increased carbon dioxide or other trace gases is limited, since most of the models suffer from severe simplifications. Mitchell et al. (1987) obtained a rise in the atmospheric temperature, which they claim to be comparable to the greenhouse effect, but only after raising the sea surface temperature, their lower boundary condition, by 2°C. Other global models use a mixed-layer ocean and thereby neglect the redistribution of heat by currents (Washington and Meehl, 1983, 1984; Hansen et al., 1988) or restrict the ocean and atmosphere to a fraction of the globe (Spelman and Manabe, 1984; Bryan et al., 1988), thereby neglecting the interoceanic heat exchange which might be crucial for climate catastrophes such as the ice ages (Broecker et al., 1985). The only published global coupled atmosphere-ocean model (Schlesinger et al., 1985) has a very coarse vertical resolution (2 atmospheric levels). Its ability to represent the atmospheric circulation is therefore restricted. In this paper the development of a comprehensive global, coupled atmosphere-ocean model from
"state of the art" atmosphere and ocean models will be described. Earlier stages of this development work have been described by Cubasch (1989a,b).

2. THE COUPLED MODEL

2.1. The Atmosphere Model

The atmosphere model has been developed as a medium-range forecasting model at the European Centre for Medium Range Forecasts (ECMWF) (Louis, 1986). Its resolution has been decreased for climate studies to a spectral resolution of T21 and 16 vertical levels. Its physical parameterization includes turbulent vertical diffusion, shallow convection, large-scale condensation, cumulus convection, an interactive radiation scheme, and a full hydrological cycle. In addition to the operational ECMWF model, the temperature over sea ice is calculated via a balance equation, and the deep soil temperature and deep soil wetness are prescribed by their climatological value every fourth day of the simulation.

2.2. The Ocean Model

The ocean model has been developed by Maier-Reimer et al. (1982). Its fully implicit formulation allows a time step of over 30 days which makes it very suitable for climate experiments. In the coupled version it runs with a horizontal resolution of approximately 4° x 4° and 10 vertical levels. It includes a thermodynamic sea ice model and a realistic bottom topography.

2.3. The Coupling

The ocean model provides to the atmosphere model the sea surface temperature and temperature over sea ice, and the atmosphere model, in turn, calculates the total heat flux (i.e., latent, sensible, shortwave and longwave radiative fluxes), the wind stress and the fresh water flux for the ocean model (Fig. 1). Since in the ECMWF model the continental runoff is estimated only as a global mean, it has not yet been incorporated into the calculation of the freshwater flux. As each of the models individually driven, even by realistic boundary forcing, tends to generate its own climate which is different from observation, their coupling creates a nonlinear feedback that forces some coupled models to drift far from the observed state (Han et al., 1985). Sausen et al. (1988) proposed a flux correction to compensate for the individual model errors which should stabilize the coupled model. This method has been tested in some of the experiments. The flux correction for temperature creates subfreezing temperatures in some regions without ice cover. In these regions a basic ice cover of 0.1 m has been assumed.

2.4. The Initial Data

The initial data for the atmospheric part of the coupled model have been taken after integrating the atmosphere model alone (i.e., with climatological sea surface temperature after Alexander and Mobley, 1976) for one year to let it obtain its own climate. From that date onwards the integration has been continued for another five years as a control experiment. This uncoupled run has been used to establish the flux-correction fields. The coupled runs start at the same date as the control experiment at a January condition. The initial data for the ocean have been derived from a 10,000-plus-year integration with the ocean model, but at a slightly different resolution (Maier Reimer et al., 1982). This data
set has been interpolated to the resolution of the coupled model. From this data set a 1500-year integration has been performed until all imbalances from the interpolation had faded away and a stationary state had been obtained. The ocean model in its uncoupled mode has been driven by the wind stress and a "Newtonian coupling" to salinity and temperature.

The wind stress has been taken from a publication by Hellermann and Rosenstein (1983). Since the observed wind stress is stronger than the one simulated by the atmosphere model (Fig. 2), particularly in the Southern Hemisphere and in the tropics, the coupling of this quantity creates some problems: a simulated ocean circulation initially forced by the strong winds of the observation reacts with a backlash when suddenly driven by the winds generated by the atmosphere model. There exist two possibilities to compensate this effect: (a) introduce again a bias correction, which will be added to the atmosphere model wind stress field to make it look in strength like the observed wind field or, (b) drive the ocean model with the atmosphere wind stress field derived from the control experiment until it reaches stationarity and then couple the models directly. A bias correction in the Southern Hemisphere and in the tropics has the same magnitude as the model-simulated forcing field.

To test the suitability of the model-generated wind stress to drive the ocean model circulation, a 1000-year run with the ocean model driven by this wind stress has been performed. The oceanic surface currents simulated by the ocean model forced by the model-simulated wind stress are generally weaker than those forced by the observed wind stress (Fig. 3). The circumpolar current around Antarctica is halved, but still is reasonable, indicating that it is not only forced by the wind stress, but also maintained by the thermohaline structure of the ocean. The equatorial upwelling in the central Pacific has also become weaker (Fig. 4), but also more detailed structures emerge in the
Indonesian region in the ocean model forced by the model-simulated wind stress. To generate a realistic temperature structure in the ocean, the model has been driven by a "Newtonian coupling" to the "equivalent" temperature by a variable coupling coefficient as defined by Oberhuber (1988). Figure 5 displays the temperature field as observed and as simulated by the ocean model driven by both the observed and modelled wind stresses. Considerable differences of both simulations from observation appear in the Gulf Stream and Kurishio region, where the gradient of the simulated sea surface temperatures is not strong enough. This is possibly connected with the too-coarse resolution of the ocean model. Large temperature differences also appear on the ice limits. The observations also contain temperatures over ice that can be quite low, while the ocean model never simulates temperatures lower than -1.9°C, a threshold at which it forms ice. In the tropical Pacific the temperature is still about 2°C colder than observed for the experiment driven by the observed wind stress. This error decreases in the run driven by the simulated wind stress, since in this case the equatorial upwelling of cold water is smaller. The oceanic fields of the model driven by the modelled wind stress are reasonable enough to use as an initial condition for the coupled experiment, thereby eliminating the need for a wind stress flux correction.

It is possible to estimate, from the Newtonian coupling, the heat flux that would have been necessary to maintain the simulated circulation. The atmosphere model simulates a heat flux comparable to observation (Esbensen and Kushnir, 1981), while the magnitude of the ocean model's heat flux is about 30% larger in the Kuroshio and Gulf Stream regions, and about twice as large in the oceans of the Southern Hemisphere. In this hemisphere the flux correction is as large as the quantity it is correcting. In the monthly global-mean heat flux over the ocean (Fig. 6), one finds a clear annual cycle in the model simulations and in the observations. None of the heat fluxes is totally balanced within a year (not even in the observations). The heat flux simulated by the atmosphere model has a constant negative bias of about 7.3 W/m², which means it extracts heat from the ocean which, in an uncorrected run, will cool the ocean if no compensating feedbacks occur.
The fresh water flux can be derived from the Newtonian coupling to the salinity observed by Levitus (1982). Figure 7 shows the fresh-water flux calculated by the ocean model and the one obtained by the atmosphere model control experiment. The global water budget is balanced in the ocean model with a loss of 24 mm/year, and in the atmosphere model with a loss of 46 mm/year which accounts at least partially for the continental runoff that has not been included in this calculation. The largest differences between the ocean model and the atmosphere model fresh-water flux appear in the Gulf Stream region, where the pattern of maximum evaporation is closely linked to the Gulf Stream itself, which, as mentioned before, is displaced in the ocean simulation; in the tropics where the cumulus convection is underestimated in the atmosphere simulation; and along the Antarctic coast where the melting and freezing of ice is not treated properly in the ocean model. It is interesting to see a lot of comparatively small-scale structures in the ocean simulation, which makes it difficult to apply a flux correction directly, since it would perpetuate these patterns locally. A spatial filter has therefore been applied to the flux correction for this quantity. Compared to observation, both model simulations underestimate the amplitude of the fresh-water flux, particularly in the Indonesian region.
In the first coupled experiments the ice thickness has been calculated by the ocean model, which was then used by the atmosphere model to calculate a temperature over ice via a balance equation. This method created problems with the flux correction, because it was not clear what to correct the ice thickness with, since in the atmospheric control run the temperature over ice had been prescribed. Additionally, the flux-corrected sea surface temperature might indicate ice coverage where the ocean does not simulate one, thus what basic ice thickness has to be assumed there? These inconsistencies in the treatment of the ice resulted in a complete melting of the Arctic ice. It was therefore decided to recalculate the temperature over ice from the heat flux due to the Newtonian coupling and the simulated ice thickness, and to flux correct that value in a similar fashion as the sea surface temperature. Figure 8 shows the recalculated temperature over the ice-covered regions which agrees well with observation, but is too high near the North Pole.

Figure 4. The vertical velocity in the top layer of the ocean for January. Top: ocean model driven by observed wind stress; bottom: ocean model driven by simulated wind stress. (From Cubasch, 1989b)
Figure 5. The sea surface temperature (°C) for January. Top: observed (after Alexander and Mobley, 1976), middle: ocean model driven by observed wind stress, bottom: ocean model driven by simulated wind stress. Contour interval 2°C. (From Cubasch, 1989b)
3. RESULTS

After reformulating the treatment of the ice-covered regions, one 5-year integration has been run so far.

3.1. The Oceanic Fields

The run shows in the global-mean ocean surface temperature hardly any drift (Fig. 9), even though some low-frequency variations can be found whose impact and origin have yet to be analyzed. A clear annual cycle is visible, caused by the unbalanced distribution of the landmasses in the hemispheres. Even though the global average performs almost perfectly, regional problems can be detected: the coupled model simulates too high temperatures near Indonesia and in the tropical Atlantic, but is too cold in the eastern Pacific (Fig. 10). Also, the Arctic is slightly too warm, caused by a too-far-north-reaching Gulf Stream which causes the eastern North Atlantic and the North Cape to be ice free even in January (Fig. 11). Even though the ice coverage seems to be generally realistically simulated, a number of points appear with very large ice thicknesses. These points could be traced back to an interaction of the time step and the formulation of the coupling of the ice model. While in reality the formation of a large ice thickness would almost immediately inhibit any heat flux through the ice, in the model it is kept constant for 30 days, which is translated into further ice growth. The resulting ice cover will never melt. Since the flux correction does not know anything about that thick ice point, which was not there in the uncoupled run, it can demand a further large heat flux though the ice, resulting in even more ice growth. The current run had to be stopped when single ice points reached
Figure 7. The fresh-water flux for January; top: simulated by the atmosphere model, middle: simulated by the ocean model, and bottom: observed after Oberhuber (1988). Unit: mm/month. (From Cubasch, 1989b)
Figure 8. The January sea surface temperature (°C) simulated over ice. (From Cubasch, 1989b)

Figure 9. The global-mean sea surface temperature simulated by the coupled model as function of integration time. (From Cubasch, 1989b)
Figure 10. The ocean surface temperature for January after five years of integration time for the coupled experiment. (From Cubasch, 1989b)

Figure 11. The ice coverage in January for the uncoupled and the coupled ocean model. (From Cubasch, 1989b)
the thickness of the top level of the ocean model, which resulted in a numerical failure of the calculation of the advection in the ocean model.

Figure 12 shows the heat transport within the ocean model before and after the coupling. The heat transport in the Southern Hemisphere has hardly been altered, but has been reduced in the Northern Hemisphere in all three oceans.

3.2. The Atmospheric Fields

The global-mean temperature of the atmosphere has hardly any drift in the coupled experiment (Fig. 13). A clear annual signal can be detected, together with appreciable short-term variability. It is interesting to note that the annual cycle flattens out as the integration progresses. In the first year its amplitude amounts to about 3°C, while in the fifth year it is only about 2°C. The causes for this reduction in variability are not yet clear. It might be caused by the diminished zonal heat exchange in the ocean, by a damping effect of the flux correction, or might also be connected with the problems in the ice model.

Figure 12. The oceanic heat transport in January, year 5. Top: uncoupled, bottom: coupled ocean model.
4. SUMMARY

A multiyear integration with a coupled global ocean-atmosphere model shows that by application of the flux-correction method the mean state of the ocean and the atmosphere can be stabilized.

After 5 years of integration time the sea surface temperature simulated by the coupled model is still as realistic as the one simulated by the uncoupled ocean model. The heat transport, however, has become weaker in the Northern Hemisphere. The ice model simulated the ice coverage fairly well, but has problems in simulating the right thickness. The global-mean atmospheric temperature has hardly any climate drift, but the amplitude of the annual cycle diminishes steadily.

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Part 3:

Observations of Climate
Circa 1850 to the Present:
Has the Climate Changed?
Marine and Land Temperature Data Sets: A Comparison and a Look at Recent Trends

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ABSTRACT. Comparisons are made among the various data sets of marine and land temperatures. Emphasis in the analyses is placed on the first intercomparison of the two marine data sets, the United Kingdom Meteorological Office (UKMO) and the Comprehensive Ocean-Atmosphere Data Set (COADS). The results of the analyses show that the two data sets are not the same, as some authors have assumed. Important differences are noted prior to 1940, with hemispheric averages differing by up to 0.2°C for some decades during the nineteenth century. Patterns of regional temperature change over the two major periods of global warming this century, 1920–39 and 1967–86, are shown.

1. INTRODUCTION

Temperature data from land and marine areas often form the basis for studies of climatic variations on local, regional and hemispheric scales, and the global-mean temperature series is a fundamental measure of the state of the climate system. The global-mean series, however, masks all the regional and seasonal variations that it encompasses. This paper seeks to set out the guidelines necessary for studies on these smaller scales.

Foremost, it is necessary that all the temperature data used in any analysis be homogeneous. Several studies have discussed strategies for assessing the homogeneity of land-based data (see, e.g., Bradley and Jones, 1985). Few, however, have considered the homogeneity of marine data. Here we consider this aspect in detail and show examples of the first comparison of the United Kingdom Meteorological Office (UKMO) and the Comprehensive Ocean-Atmosphere Data Set (COADS) marine data bases. The combined analysis of both marine and land data is vital if we are to understand, and in time explain, the course of regional temperature variations over the last 140 years.
The paper concludes by comparing the regional temperature trends over the two major periods of global warming this century, namely, 1920–39 and 1967–86.

2. LAND DATA SETS

For data over the land areas, three global-scale data sets of temperature anomalies are available: Jones et al. (1986a,b) and Jones (1988), hereafter J; Hansen and Lebedeff (1987, 1988), hereafter HL; and Vinnikov et al. (1987), hereafter V. These data sets are either gridded on a regular latitude/longitude grid (J, V) or "boxed," that is, condensed to a set of latitude/longitude boxes (HL). For simplicity, we will refer to all three data sets as "gridded." They are all based on the same basic set of station temperature data, but there are important differences. For example, J have added many additional stations to the data set, especially in the early years of the record, and they have carried out quite detailed station-by-station assessments of homogeneity (see below) and corrected many of the individual station time series (see Jones et al., 1985, 1986c for details). The data sets have been compiled independently, except for the V results for the Southern Hemisphere which are based on the J data set, gridded and averaged using different methods from J.

Hemispheric means based on the three data sets are shown in Figs. 1 and 2. These hemispheric means have been compared by a number of authors (e.g., Grotch, 1987; Wigley et al., 1989). Correlations between the various time series are extremely high, but the three versions disagree on the amount of warming that has taken place since about 1880. For example, in the HL Northern Hemisphere estimate, the warming over 1880 to 1987 is about 0.2°C greater than in both the J and V series (Wigley et al., 1989). The extra warming in HL is believed to be due either to coverage differences and/or to the fact that they did not carry detailed station-by-station intercomparisons to ensure the homogeneity of the individual station data. The good agreement between V and J suggests that V was also careful in the basic station data used, but this is not clear in their published material.

Unless corrected, most of the potential inhomogeneities in station temperature time series are likely to affect a gridded data base at the gridpoint level, with the magnitude of the effect depending on how many stations go into a gridpoint or gridbox series. However, on larger spatial scales, many types of inhomogeneity will tend to have cancelling effects, and so have little influence on the homogeneity of hemispheric or global means. For example, changes in site or in methods used to calculate monthly means are equally likely to warm or cool the measured surface air temperature. When the gridded data are averaged to annual hemispheric estimates, such local effects are likely to cancel out. Other types of inhomogeneity, however, are less benign. The prime example is the inhomogeneity caused by the growth of cities around sites during the twentieth century. In this case the effects at individual stations will all be of the same sign, a warming of the city site relative to its rural environs.

Studies of the urbanization effect, on individual cities, have found that individual night minimum temperatures can have a warm bias of up to 20°C (see Oke, 1979). However, these types of study are not relevant to the monthly time scale and monthly means for which the effect is considerably reduced by temporal averaging. Furthermore, for gridded data, additional dilution of the effect occurs due to the mix of rural and urban stations used in the gridding and the averaging of gridpoint data over a wide area.

For monthly mean gridded data, the magnitude of the effect can be assessed by comparing regional time series from the gridpoint networks with regional analyses based principally on rural stations. The best available data set for such a comparison is the Historical Climate Network (HCN) for the contiguous United States, produced by Quinlan et al. (1987). This data set is composed mainly of rural stations and has also been assessed.
for station inhomogeneities (Karl and Williams, 1987). Urban sites have been corrected for urban warming using population data (Karl et al., 1988). Comparisons between estimates of the average contiguous U.S. temperature based on HCN and that from the J and HL data sets show a surplus warming in the gridded data sets of 0.1 and 0.4°C, respectively, over the 1901-84 period (Karl and Jones, 1989; Jones et al., 1989).

Generalizing this result to other parts of the world is not a simple task. The degree of urban warming depends on such factors as the growth of urban population and per capita energy consumption. The effect on large-area averages depends on the mix of urban and non-urban stations, and on how well individual records have been homogenized—which, in turn, depends on station density for intercomparisons. Few regions of the world can have experienced growth rates comparable to North America over the present century, but this may not be the case over the last twenty years in some developing countries. Studies in some other industrialized regions of the world tend to indicate that North America is more affected by urban warming (Jones et al., 1990).
Figure 2. Comparison of the J, HL and V estimates for the Southern Hemisphere. The reference periods used by the three analyses are 1951–70, 1951–80 and 1951–1975, respectively.

3. MARINE DATA SETS

Two global compilations of marine data have been assembled. These are the United Kingdom Meteorological Office (UKMO) marine data bank (Shearman, 1983; Bottomley et al., 1990) and the Comprehensive Ocean-Atmosphere Data Set (COADS) (Slutz et al., 1985; Woodruff et al., 1987). The analysed part of the UKMO data set contains only observations of sea surface temperatures (SST) and nighttime marine air temperatures (NMAT), while COADS contains the complete set of shipboard observations. There is considerable overlap between the two data sets, but the COADS set contains 40% more raw individual SST observations than UKMO (mainly since 1950) and does not distinguish between nighttime and daytime marine air temperatures (Woodruff et al., 1987; Bottomley et al., 1990). The UKMO data set for SST contains extra analysed data supplied by the Massachusetts Institute of Technology (Folland et al., 1984; Bottomley et al., 1990) for the period 1949–1979. These data are mainly in the Pacific. In addition, the
UKMO analysed data sets of SST and NMAT currently use telecommunicated observations beginning in January 1982, but the UKMO marine data bank contains no telecommunicated observations.

It is generally agreed that both types of marine data require some form of adjustment or correction, particularly before 1945. Although this has been known for some time (Saur, 1963; Tabata, 1978; Barnett, 1984, 1985), some authors have, nevertheless, failed to realize the primary importance of the inhomogeneities (e.g., Paltridge and Woodruff, 1981; Oort et al., 1987). Corrections are necessary because of changing methods of measuring SST, changes to ships, including the change from sail to steam and trends in ship sizes, and changes to shipping routes. Before considering how adjustments to the marine data should be made we will first compare the two data sets.

The basic raw UKMO and COADS data sets are made up of the mean SST or MAT value for each available “square” of the ocean for each pentad (UKMO) or month (COADS) since the beginning of recording during the 1850s. The only other difference between the two data sets at this stage, apart from the greater number of raw SST observations in COADS and the rejection of daytime MAT observations in UKMO, is that the box size for UKMO is 1° by 1° while for COADS it is 2° by 2°. Because the gridbox sizes are small, the coverage is “spotty” in some regions and the number of boxes is too large for studies of global and hemispheric temperatures. The basic data have therefore been combined to form larger boxes, 5°x5° for UKMO (Bottomley et al., 1990) and 4°x10° (SST) and 5°x5° (MAT) for COADS (Wright and Jones, 1986). [For the COADS data, the MAT observations have been gridded, as in Jones et al. (1986a,b) rather than simply averaged into boxes.]

Both data sets use very similar techniques to compress the data onto the more manageable grids. First, all small-box values are reduced to anomalies from a common reference period (1951–80 for UKMO; 1950–79 for COADS). Small boxes with too few observations to form a reference-period mean or individual monthly anomaly values are ignored. Only a small number of observations are required to form a monthly value because of the high temporal autocorrelation between successive observations (Parker, 1984). Generally, only 3 measurements were considered sufficient (see Wright and Jones, 1986) or three 5-day means for UKMO (Bottomley et al., 1990). With the COADS data, some effort was made to ensure that, if the number of observations was small, they were reasonably well distributed throughout the month, based on the mean day of the month for observations (see Wright and Jones, 1986). For the UKMO data, the use of three 5-day means and other quality control procedures (Bottomley et al., 1990) largely achieves the same end. The small-box values are then averaged to the larger units, with distance weighting in the COADS MAT case (cf. Jones et al., 1986a,b). In data-sparse regions, only one small square was considered sufficient to produce a large-box value. These procedures are valid only if anomaly values are used. Use of anomalies is an important aspect of the analyses, since it allows greater coverage than would otherwise be possible.

The area of ocean covered by the two data sets is similar up to 1960, although UKMO SST coverage is better than COADS SST over the Southern Hemisphere between 1910 and 1980. Since 1961, however, the area covered by the COADS data set in the Northern Hemisphere has been consistently greater. Coverage tends to increase for both marine variables in both data sets in a linear fashion from the 1860s to the 1980s if the marked reductions in coverage during the early 1900s and the two world-war periods are ignored. Despite the relatively poor coverage during the nineteenth century in both hemispheres, studies using the “frozen grid” method of Jones et al. (1986a,b) show that there should be no systematic biases in either hemispheric estimates (Parker, 1987).

Direct comparison of the uncorrected hemispheric estimates from UKMO and COADS using MAT and SST reveals that there are important differences between the
two data sets (Fig. 3, MAT, and Fig. 4, SST). The scale of these differences has not been generally realized. Indeed, they are somewhat surprising since there is so much common data.

For MAT (NH), UKMO is consistently warmer than COADS by between 0.10-0.15°C up to 1940. Since 1950, UKMO is warmer up to 1960, then generally cooler. In the Southern Hemisphere, the difference is near zero during the twentieth century. Between 1870 and 1900, UKMO is about 0.2°C cooler, particularly during the late 1880s. More detailed comparisons using 10° zonal means show similar patterns to those in each
Figure 4. Differences of hemispheric means of uncorrected sea surface temperatures (SST) between UKMO and COADS [UKMO (SST) minus COADS (SST)]. The reference periods used by UKMO and COADS are 1951–80 and 1950–79, respectively.

Hemisphere, except for the most-poleward zones (70-60°N, 60-50°N and 40-50°S) where differences, although highly variable, do not show the consistent variations of other zones. A further seasonal breakdown shows that most of the features are common to all seasons. These differences cannot be related to the use of only NMAT data in UKMO and both day and night MAT in COADS, since both are analysed in anomaly terms. Differences may arise if the bias in either daytime or nighttime MAT observations varied with time (e.g., due to changing ship characteristics or observation practices), or if the mix of day and night temperatures in COADS varied with time, and/or if the UKMO NMAT data were contaminated with varying amounts of daytime data.
For SST (Fig. 4) there is a greater similarity between the corresponding hemispheric-mean time series. Up to 1880, differences are small, although there is a cooling in UKMO relative to COADS in the Southern Hemisphere. Between 1880 and the 1960s, UKMO is warmer than COADS by between 0.05 and 0.10°C, nearer the higher value in Northern Hemisphere and the lower value in the Southern Hemisphere. After 1960 there is a sharp drop in the difference to near zero. Analysis of zonal-mean differences shows that these features are apparent in all but the most-poleward zones, but are strongest in the mid-latitude zones between 20-50°N and 20-40°S. There appears to be no immediate causes for these differences, although the sharp jump at 1960 may be related to the inclusion in UKMO of Massachusetts Institute of Technology (MIT) data in the Pacific during the early 1960s.

While it may not be immediately apparent what the exact causes of the differences between the UKMO and COADS data sets are, the differences mean that any corrections for instrumental-practice changes should be different between the two data sets. The fact that corrections are necessary to either one or both measures of marine temperature can be seen by plotting the difference in hemispheric MAT and SST values for the COADS data set (Fig. 5). If instrumentation had remained constant, there should be little difference between annual anomaly values of MAT and SST, with no change on time scales of 5 years and upwards. Figure 5 shows MAT significantly, and erroneously, higher than SST prior to 1940. From the 1850s to the 1890s, the air is relatively warm by about 0.6°C in the NH and 0.45°C in the SH. This drops to around 0.4°C (NH) and 0.3°C (SH) from 1900 to 1940.

Two different approaches to the correction of marine data series have been made. We will refer to these as the \textit{a priori} and \textit{a posteriori} approaches.

In the \textit{a priori} approach, exemplified by Folland \textit{et al.} (1984), Bottomley \textit{et al.} (1990) and Folland and Parker (1989b), the adjustments are made partly on theoretical grounds. Folland and Parker (1989b) refer to the approach as an \textit{a priori/a posteriori} mix. In order to avoid confusion with the \textit{a posteriori} approach discussed later, we refer to it here as \textit{a priori}. In this approach, it is assumed that the causes of inhomogeneities in the data are understood, and correction factors derived based on a model of the causal factor(s). For SST, Folland and Hsiung (1986) have assessed, with the aid of a model, the amount of cooling that will take place while a canvas bucket is being hauled onto and left to stand on deck before reading. The evaporative cooling that takes place depends on temperature, the air-sea temperature difference, relative humidity and windspeed. The correction factor so derived varies according to month and to the region of the ocean.

The implied seasonal and spatial variability of the correction factor provides a way of verifying that the model is producing reasonable results. Because of the changing mix of bucket and intake measurements with time, the observed seasonal cycle of SST for the period prior to 1940, when observations were mainly made using canvas buckets, should have a slightly different amplitude compared with more recent (mainly intake) measurements. This difference in annual cycles should also vary spatially. In practice, these differences will show up as spurious annual cycles in the anomaly data, since the reference period spans an interval of virtually constant measurement techniques. After correction, both the temporal and spatial variability of the amplitude of the annual cycle should be substantially reduced.

Climatological values are used for most of the parameters in the bucket model. The only unknown is the length of time between taking the sample and the temperature measurement. Folland and Parker (1989a,b) estimate this by finding the time that best minimizes the spurious annual cycles. This turns out to be 3-6 minutes, which agrees with some recommendations (see, e.g., WMO, 1954). Furthermore, as anticipated, the same inferred exposure time applies to all regions of the world's oceans. Figure 6 shows the
average hemispheric correction between 1856 and 1941. The slight variations from year to year result from slightly changing coverage in each year. The average correction is slightly larger in the Northern Hemisphere because of a greater proportion of high-latitude ocean where the corrections required are greater.

In the \textit{a posteriori} approach, exemplified by Jones \textit{et al.} (1986d), adjustments are made so that the mean hemispheric and regional estimates of MAT and SST are in accord with adjacent land-based data on decadal and longer time scales. For this approach to be valid, the land-based data must be homogeneous. In the previous section and elsewhere (Jones \textit{et al.}, 1985; Wigley and Jones, 1988; Jones \textit{et al.}, 1989) we have detailed the methods used to ensure this homogeneity.
Figure 6. Average hemispheric correction between 1856 and 1941 using the Folland and Parker (1989a,b) bucket model.

In Jones et al. (1986d), 15 different regions of the world were chosen where both MAT (COADS) measurements and adjoining coastal and island data from the land-based data exist. Both types of measurement purport to measure the same quantity. In addition to these 15 region-by-region comparisons, the two hemispheric-mean MAT series (for the NH and SH) were compared with the appropriate means of the coastal and island temperatures from the land data set. The similarity of all 17 annual-difference time series implied that the MAT data were affected by consistent inhomogeneities, which were similar in both hemispheres. Corrections to the MAT data are derived from the hemispheric differences.

The same approach is used to correct the SST data. Hemispheric means of the uncorrected SST data and the corrected MAT data are compared. These show long periods of roughly constant differences which are similar in both hemispheres. The differences are presumed to relate to changes in SST instrumentation, and correction factors for SST are based on these differences. The basic character of the SST corrections derived this way is similar to that for the a priori corrections, with one important difference. Based on the a posteriori approach, the corrections required in the nineteenth century differ from those required between 1900 and 1940. We presume that this reflects the change over from (relatively well-insulated) wooden buckets to uninsulated canvas buckets, a presumption which is supported by historical evidence on instrumentation. Although one could account for this using the a priori method, this has not yet been done. It would be quite difficult to do because the quality of the nineteenth-century data is insufficient to pin down an exposure time with any confidence. This point is discussed further below.
Here we repeat part of the \textit{a posteriori} correction exercise with the UKMO NMAT and SST data. In Jones \textit{et al.} (1986d) we compared both the differences between NMAT and coastal land data for both 15 regions where the two data sets overlap and for the hemispheric averages. Here, simply for comparative purposes we use the hemispheric differences. The differences between hemispheric NMAT and the coastal and island hemispheric estimates are shown in Fig. 7. When this figure is compared with the corresponding result using the COADS data (Jones \textit{et al.}, 1986d, Fig. 2), there are marked differences in the Southern Hemisphere during the nineteenth century. Such differences were to be expected, however, given the difference between the COADS MAT and UKMO NMAT data shown in Fig. 3. Using UKMO NMAT there is a clear discrepancy between the implied corrections necessary for the NMAT data in the two hemispheres. This was not evident for the COADS MAT data used by Jones \textit{et al.} (1986d).

The corrections which must be applied to the NMAT data are $+0.08^\circ C$ over 1903-41, $-0.59^\circ C$ ($-0.32^\circ C$ in SH) over 1874-89 and $-0.44^\circ C$ ($-0.36^\circ C$ in SH) over 1861-73. During the war period of 1942-45 there is again a difference between the hemispheres of $-0.81^\circ C$ in NH and $-0.56^\circ C$ in SH. This, however, also occurs with the COADS data and can be readily explained (see, e.g., Jones \textit{et al.}, 1986d). Over the period 1890 and 1902, a gradual change between the two correction levels is necessary. Although the corrections are similar to the COADS corrections derived by Jones \textit{et al.} (1986d), there are differences (as noted above, and in addition) which reflect the differences inherent in Fig. 3. However, the times of transition between different correction factors are the same, even though these have been estimated independently.

Possible reasons for the MAT corrections have been given by Jones \textit{et al.} (1986d). The corrections required during the nineteenth century might imply that the screens were poorly exposed or that they were not used at all. With the UKMO data it can also be argued that the NMAT data are contaminated by a higher percentage of daytime MAT observations than during the twentieth century. The difference between the two implied hemispheric corrections for NMAT arises partly from the fact that NMAT has been shown to be erroneously warm in the Mediterranean Sea and North Indian Ocean between 1876 and 1893 (Folland and Parker, 1989b).

Once the NMAT data have been corrected, it is a simple matter to correct the SST by comparison with the corrected NMAT. The implied SST corrections necessary to ensure compatibility with the NMAT series are: 1903-41, $+0.45^\circ C$ ($+0.40^\circ C$ in the SH) and 1856-89, $+0.10^\circ C$. No correction is necessary for the data after 1941. As with NMAT, gradual changes take place between 1890 and 1902. A value between 0.40 and 0.45°C between 1903 and 1941 is consistent with the change from uninsulated bucket measurements to engine-intake readings. The value lies in the range 0.3-0.7°C, estimated for such a correction independently by James and Fox (1972), Barnett (1984) and Ramage (1984).

The corrections derived by the two methods, \textit{a priori} and \textit{a posteriori}, are compared in Figure 8. The implied corrections during the twentieth century are extremely similar. Both methods even agree in finding slightly lower average correction factors in the Southern Hemisphere compared to the Northern Hemisphere. The major difference between the methods occurs during the nineteenth century and is of the order 0.2°C between 1856 and 1889. The difference hinges entirely on the type of bucket used during the last century. Folland and Parker (1986; 1989a,b) have assumed that canvas buckets, which were undoubtedly in general use during the first half of the twentieth century, were also used in the nineteenth century. At the first international meeting in Brussels which set up the current system of marine data collection, it was recommended that wooden buckets should be used (Quetelet, 1853). It is also clear that wooden buckets were in use around this time and probably for some time subsequently (Maury, 1855). This is also supported
Figure 7. Difference between hemispheric estimates made using NMAT and those made using coastal and island temperature data from the Jones et al. (1986a,b) land-based data set.

by Krummel (1907), who refers to the canvas bucket as being a recent innovation. It is thus likely that there was a mix of bucket types in use during the nineteenth century, with a transition to a preponderance of canvas buckets probably occurring over the last few decades of the nineteenth century. Both wooden and tin buckets (which might also have been used) are considerably better insulated than their canvas counterparts and would not be affected as much by evaporative cooling.
Figure 8. Comparison between the corrections implied by the \textit{a priori} (Fig. 6) and \textit{a posteriori} approaches.

4. REGIONAL TEMPERATURE CHANGES

Figures 1 and 2 show time series of hemispheric-average temperatures based on the three data sets discussed earlier J, HL and V. Clearly, not all regions show the same fluctuations, and there are important spatial differences. Here we will show spatial patterns of the linear trend in annual-mean temperature over the two major warming periods in the global-mean temperature record, 1967–86 and 1920–39 (Figs. 9 and 10, respectively).

The maps combine land-based estimates from the J data set and marine data from COADS (MAT). The choice of MAT rather than SST is somewhat arbitrary. There is excellent agreement between anomaly patterns of MAT and SST, as has been shown by
Figure 9. Change in temperature (°C) over 1967–86 accounted for by a linear trend fitted to each annual gridpoint time series. Calculations over the oceanic areas are made using COADS MAT data. Shading indicates areas with no data.
Figure 10. Change in temperature (°C) over 1920–39 accounted for by a linear trend fitted to each annual gridpoint time series. Calculations over the oceanic areas are made using COADS MAT data. Shading indicates areas with no data.

other workers (Cayan, 1980; Barnett, 1984; Jones et al., 1988), and our own comparisons confirm this. Similar trend maps to those presented here have been calculated using
land (J) and marine (UKMO SST) data by Jones et al. (1988) for the 1947–86 period. Trend maps based only on land data have also been produced by van Loon and Williams (1976a,b; 1977), Jones and Kelly (1983), Mo and van Loon (1985) and Hansen and Lebedeff (1987).

In the last section we saw that marine data are considered usable without correction only over the past 40 years. Prior to this, spatially and seasonally varying corrections are required and, furthermore, coverage is particularly poor during 1940–45. This does not mean, however, that we need to confine ourselves to periods since 1946. Provided we choose periods between the major changes in marine instrumentation and observation practices we can produce meaningful results. The period 1920–39, between the two world wars, is one in which a correction to the marine data is required, but over which the correction is thought to have remained relatively constant (Jones et al., 1986d; Bottomley et al., 1990). Another period which could be chosen is the 1861–1889 period. Periods which would not be appropriate are 1889 to 1905 and any period encompassing 1940–45.

Over the Northern Hemisphere between 1967–86 (Fig. 9), cooling is confined mainly to two regions, the northern halves of the North Pacific and North Atlantic Oceans. The cooling areas extend over the adjacent coasts to cover Japan, northwestern Europe, Iceland and the maritime states in Canada. Elsewhere, warming dominates and is strongest over the Soviet Union (particularly Siberia), western North America (particularly Alaska and the Yukon), and parts of North Africa. Warming is also evident over the eastern equatorial Pacific, the central Atlantic and the Indian Oceans.

Coverage over the Southern Hemisphere is poorer than in the Northern Hemisphere, with inadequate data over much of the 45–60°S zone and in the southeastern Pacific except near the South American coast. Warming tends to dominate. There are only three small areas of cooling, the largest of which is centred on the Amazon Basin. Warming is strongest, and the most significant statistically, over the eastern equatorial Pacific Ocean and over other mid-latitude (20–40°S) zones of the Atlantic and Indian Oceans.

Figure 10 shows patterns of temperature change over the period 1920–39. The most notable feature is the markedly poorer coverage during the period, particularly over the Southern Hemisphere where, apart from southern South America, there are no data south of 40°S. Coverage is also poorer over the North Pacific Ocean, except for the shipping route from North America to Australia via Hawaii.

The pattern of temperature change over the Northern Hemisphere between 1920 and 1939 shows the strongest warming over the entire North Atlantic basin extending over Europe, North Africa and the United States. Warming is strongest over Greenland, Sahelian Africa and the United States. The warming during this period over the North Atlantic basin is in striking contrast to the 1967–86 picture. Cooling affects two main areas, central Asia from western Siberia to southern China, and central northern Canada from Hudson Bay to southern Alaska. Over the Southern Hemisphere, warming tends to dominate over the tropical oceans. Cooling is found only over the southern half of South America.

5. CONCLUSIONS

Comparison of the MAT and SST data in the two main marine data banks, UKMO and COADS, has shown that the two data sets are not the same, as might have been believed from their large common data source. Although the differences (0.1–0.2°C in terms of hemispheric averages) may appear small, they are particularly important in the context of the overall global warming trend over the last 100 years (0.5°C).

The necessity of applying some form of correction to both types of marine data, MAT and SST, was illustrated by simply plotting the differences between the hemispheric
means of these two variables (Fig. 5). The two major approaches that have been used to quantify the degree of correction required were compared. The \textit{a priori} approach makes assumptions about the main sources of inhomogeneities in MAT and SST data and makes appropriate corrections using physical reasoning and models. The \textit{a posteriori} approach, on the other hand, makes corrections to ensure conformity with the land data over common regions. The homogeneity of the land data has been previously assessed using independent methods.

Comparison of the correction factors for SST data on a hemispheric basis shows that they are in excellent agreement during the early twentieth century (1900–41). During the nineteenth century, the two methods differ, with the \textit{a priori} approach implying, on average, an additional 0.2°C correction and producing warmer corrected data. The reason for this difference appears to be related to the types of bucket used during the nineteenth century. The \textit{a priori} calculations carried out by Folland and Parker (1989a,b) and Bottomley \textit{et al.} (1990) assume that the predominant bucket type in the nineteenth century was as in the 1900–41 period, viz. uninsulated canvas buckets. More likely, a mix of wooden, tin and canvas buckets was used, making a smaller hemispheric correction necessary. Although, in application, the \textit{a priori} approach has produced results that are probably too warm in the nineteenth century, the method is clearly superior in that it is essential in order to derive regional and seasonally specific correction factors.

In the final section of this paper, the land and marine data have been used to derive and compare the spatial patterns of temperature change over two periods, 1920–39 and 1967–86. These periods cover the two main global warming episodes this century. Although warming dominates, the patterns of temperature change over the Northern Hemisphere in both periods show two large areas of cooling. These are not the same regions, however, and the cooling areas are considerably larger during the 1967–86 period. The contrast in trends over the North Atlantic is particularly striking.

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1. INTRODUCTION

Our perception of recent climatic changes, and therefore our interpretation of them in terms of the increasing atmospheric "greenhouse" effect and other, natural, forcings, may be affected by false trends in the observational record. In this paper we present fields and trends of marine temperatures, corrected for instrumental biases, and compare them with corresponding land air temperatures as quality-controlled and analysed by Jones (1988) and sources referenced therein.
2. CORRECTION OF MARINE DATA

2.1. Sea Surface Temperature

Systematic large-scale biases in sea surface temperature (SST) data have occurred because of widespread changes in instrumentation. Earlier this century, and probably since the 1870s (Toynbee, 1874; Deutscher Seeewarte, 1875; Meteorological Office, 1876), sea water was usually collected by observers using uninsulated canvas buckets, and occasionally in metal or leather buckets. In the mid-nineteenth century wooden buckets were more general (Maury, 1858). Around the end of 1941 a sudden widespread change is believed to have taken place, in response to wartime conditions, so that engine-intake measurements became predominant for the remainder of the Second World War. Since then, engine-intake measurements have retained a prominent role, but some insulated and uninsulated buckets are known to have been used, the latter especially before 1960. The effects of the major change around 1941 have been twofold: firstly, SSTs were lower relative to corrected marine air temperatures (Sec. 2.2) until 1941 (Fig. 1); secondly, SSTs until 1941 showed stronger annual cycles in the extratropics because of enhanced heat losses from uninsulated buckets in winter.

![Figure 1. Seasonal global anomalies (with respect to 1951–80) of uncorrected sea surface temperature (solid line) and corrected night marine air temperature (dashed line), 1856–1988. Season are Jan–Mar, etc. Corrections applied to night marine air temperature: Up to 1890: – 0.15°C; 1891–1910: linear rise to zero; 1911–March 1940: zero; April 1940–Dec. 1941: –0.1°C; 1942–Sept. 1945: – 0.9°C in Pacific north of 25°N and 20–25°N, west of 165°W, – 0.6°C in Atlantic north of 20°N, and – 0.5°C elsewhere; October 1945 onwards: zero. In addition, bad marine air temperature anomalies in 1876–93 in the Mediterranean and the North Indian Ocean were replaced by sea surface temperature anomalies.](image-url)
We have corrected the SSTs from the beginning of our data set (1856) up to 1941 using a set of versions of a mathematical model of the physics of an uninsulated canvas bucket. The model takes account of convective transfer of both sensible and latent heat, and of short and long wave radiative heat transfer. It assumes an environment for the bucket derived from a 1951–80 monthly mean climatology, modified by the influence of the ship. Faster ships' speeds, and therefore stronger ventilation, are assumed on steamships than on sailing ships, with linear interpolation of ships' speeds between 1890 and 1910. The nonlinearity of the relationship between wind speed and forced convective heat transfer and mass transfer (evaporation) is taken account of. The time allowed for heat and mass transfer in the model is tuned by a statistical assumption that the augmented annual cycles of SST in the uncorrected data should be reduced to a mean amplitude as consistent as possible with that observed in 1951–80. In this way the influence of the uncertainty regarding the duration of exposure of the bucket on deck has been largely avoided. Exposure times required in the model for optimal adjustment of the annual cycles are broadly consistent with documented instructions to observers. They vary with the size of the bucket used, essentially with the thermal inertia. However, the calculated corrections hardly vary with the size of the buckets, because although small buckets cool quickly and large ones cool slowly, the annual cycle is minimized only after the same amount of cooling has been accounted for.

Corrections applied to data for June and December for 1911–41 are shown in Figs. 2a and b, respectively. Note the large corrections applied in winter over the Gulf Stream and Kuroshio where the water is warm but the air relatively cold and dry, with frequent strong winds, so that uninsulated buckets cool rapidly.

Figure 3 shows distributions of differences between May-to-July and November-to-January anomalies of uncorrected SST data for extratropical 5° latitude x 5° longitude areas for 1911–40, with superimposed distributions using corrected data. The tendency to positive values in the extratropical Northern Hemisphere shows the augmented annual cycles in the uncorrected data, as does the tendency to negative values in the extratropical Southern Hemisphere. Application of the corrections in Fig. 2 strongly reduces these spurious inter-seasonal differences.

Further details of the method used to correct SST are given by Folland and Parker (1990). We are also developing a model of a wooden bucket for application to the nineteenth-century part of the SST record; see Sec. 3 below.

2.2. Night Marine Air Temperature

Marine air temperatures were restricted to nighttime values to avoid the potentially variable effects of solar heating of the decks. Corrections applied are described in the caption to Fig. 1. A small adjustment was applied to allow for progressive increases in the elevation of ships' decks, and adjustments were applied to compensate for procedural changes made temporarily during the Second World War (Folland and Parker, 1990; Folland et al., 1984).

3. COMPARISON OF CORRECTED SST AND NIGHT MARINE AIR TEMPERATURES

Figure 4 illustrates the good agreement between hemispheric trends of corrected SST and night marine air temperatures (NMAT). In view of the independence of the observational techniques, and of the correction procedures for SST and NMAT, except in 1876–1893 (see the caption to Fig. 1), the agreement in Fig. 4 provides good evidence of the validity of the results. Corrections to SST have also been computed using techniques
Figure 2a. Provisional corrections to uninsulated bucket sea surface temperature data for June, 1911–41. Hatched: > 0.5°C. Stippled: < 0.0°C.

Figure 2b. Provisional corrections to uninsulated bucket sea surface temperature data for December, 1911–41. Hatched: > 0.5°C. Stippled: < 0.0°C.
similar to those described above and in Folland and Parker (1990), but with a model of a wooden bucket. The corrections are less positive (or, in mid-latitude summer, more negative) than those derived for canvas buckets. The reasons include a greater sensitivity to solar radiation because, in the wooden-bucket model, this was allowed to affect the open top of the bucket. Use of these corrections causes SST anomalies to be relatively lower than NMAT anomalies, weakening the agreement in Fig. 4. The dotted curves in Fig. 4 show SST trends assuming a linear transition from wooden to canvas buckets from 1856 to 1910 and a linear transition from sailing ships to steamships from 1876 to 1910. Because the evidence from the literature (e.g., Toynbee, 1874) suggests that canvas buckets were general by the mid 1870s, the dotted curves represent lower limits for the estimated SST anomalies. On the other hand, some of the 19th Century NMAT data appear to be unreliable (Folland and Parker, 1990), though marine screens were in use from 1868 or before (Meteorological Office, 1868).

A further indication of the reliability of the corrected SST (with the canvas-bucket model) and NMAT since 1900 has been gained through the creation of global covariance eigenvectors representing the principal geographical patterns of their natural variability. Figures 5 through 7 show the first 3 (unrotated) eigenvectors of annual anomalies of SST and NMAT, and annual time series of their coefficients. To depict the large-scale
patterns, the analyses were made on 10° latitude x 20° longitude resolution. Note the similarity of the SST and NMAT patterns, and the high correlations between the series of coefficients for each of the first 3 eigenvectors. These represent, respectively, worldwide climatic trends of the same sign nearly everywhere (compare Fig. 4 with the bottom panel of Fig. 5), the worldwide effects of the El Niño-Southern Oscillation (Fig. 6), and a mainly interhemispheric contrast of anomalies (Fig. 7) which has been linked to drought in sub-Saharan Africa (Folland et al., 1986). The recent relative coldness in the mid-latitude North Atlantic (Fig. 7 with positive coefficients) requires further study in view of coupled atmosphere-ocean model simulations of a weaker thermohaline circulation with enhanced atmospheric carbon dioxide (Washington and Meehl, 1991). The fourth and higher-order eigenvector patterns of SST and NMAT anomalies reflect (after rotation to isolate regional patterns more effectively) regional phenomena such as the atmospheric North Atlantic Oscillation (Barnston and Livezey, 1987) and its impact on SST.

Figure 4. Smoothed and corrected (assuming canvas buckets) SST (solid) and NMAT (dashed) anomalies (with respect to 1951-80) for the Northern and Southern Hemispheres, 1856 to 1988. The data are plotted against the end-date of a 10-year triangular smoothing filter. The dotted curves show SST trends assuming some wooden buckets and some sailing ships until 1910 as described in the text.
Figure 5. The first covariance eigenvector of annual anomalies of: (a) SST and (b) night marine air temperature (NMAT), defined for 1901–80 using data with 10° latitude × 20° longitude resolution. The SST data have been corrected using canvas-bucket model as in the text, and the NMAT data have also been corrected as in the text. The units are arbitrary. The time series of the coefficients are in (c), with SST solid and NMAT dashed.
Figure 6. As Fig. 5, but for the second eigenvector.

4. COMPARISON OF CORRECTED SST WITH LAND AIR TEMPERATURES

Comparisons have been made between coastal and island land air temperatures from gridpoint data provided by P. D. Jones (University of East Anglia) and colocated 5°-area SSTs with canvas-bucket mode; corrections as described above. All data are anomalies referenced to 1951–80.
Figure 7. As Fig. 5, but for the third eigenvector.

Ten-year running mean differences for the globe (Fig. 8) show relative warmth of land air temperature of up to 0.2°C in 1900–1940. However, agreement is good in the mid- and late 19th century. The numbers of 5° latitude × 5° longitude areas used are shown in Fig. 9. Similar results (not shown) were obtained in each 3-month season for the globe and in each hemisphere for the year as a whole. The relative warmth of the land air temperatures in the early 20th century is also evident in the tropics and in a
variety of tropical or subtropical sectors (Fig. 10), as well as in Europe (Fig. 11). The sample size was small in southern Asia and Australia until after 1900, so the early parts of the series for these regions in Fig. 10 are less reliable.

A reason for the relative warmth of land air temperature in the tropics in the early 20th century is likely to have been the siting of thermometers at that time in wire cages in open-sided, thatched-roofed sheds. These yielded higher observed temperatures than do Stevenson-screen exposures, by about 0.4°C on average in experiments carried out at Agra (India) by Field (1920). Use of the shed siting became general in India in the 1870s (Chambers, 1876, 1877, 1878, 1879). As a consequence of Field's results, Stevenson screens became standard in India by the 1930s. Indian data published in "World Weather Records" (Smithsonian Institution, 1947) up to 1940 were corrected so as to be consistent with the earlier sitings, but it is not apparent that the corrections were carried forward after 1940. "World Weather Records" reports similar procedures in Ceylon. Hunt (1925) still recommends the thatched sheds for Australian tropical stations, but Marriott (1924) recommends a transition to Stevenson screens in tropical countries. Marriott (1902) had recommended the thatched sheds for the tropics, as did Ravenstein et al. (1892) and Meteorological Office (1907) for tropical Africa, but there is no reference to the thatched shed in Marriott (1892). It is thus not clear whether thatched sheds had been general in Africa from the establishment of the stations. A memorandum issued by the Meteorological Office (1935) to observers in tropical Africa recommended transition to Stevenson screens on the basis of Field's results. Some stations in Africa were, however, according to "World Weather Records," already using Stevenson screens in the 1920s and before. Information on thermometer exposure in South America has not been found, except for a few citations by Rotch (1894) and in "World Weather Records" indicating, for example, the use of (North American type) Stevenson screens in Argentina from the late 19th century and the use of a "large louvred pavilion" in Rio de Janeiro from 1870 until at least 1920. The tropics will have contributed little to the global comparison before 1900 because of the relatively small proportion of 5° areas there with data for comparison. The greater relative warmth of land air temperature in the late 19th century in Southern Asia (Fig. 10) is consistent with the evidence that thatched sheds were in general use in India at that time.

The comparisons for Europe show relative warmth over land from about 1910 to 1950 (Fig. 11), but the comparisons for the Mediterranean area, shown in the same diagram, indicate negligible relative changes over land and sea. Much of the relative warmth before 1950 over Europe results from the winter season (Fig. 12), suggesting that the enhanced westerly circulation over the North Atlantic in winter in the early 20th century was responsible, that is, the effect is real and not an instrumental artifact. When European winters are warm (cold) the coastal waters also become warm (cold) but to a lesser extent: the interannual standard deviation of winter air temperature over land greatly exceeds that of the coastal water temperature. This even applies to small islands where, for example, low air temperatures can occur in cold air masses as the atmospheric boundary layer rapidly stabilizes over land. A minimum as low as -7°C has been recorded on the Scilly Islands (off S.W. England) in severe northeasterly airflow.

The atmospheric circulation changes affecting Europe on interdecadal time scales have not affected the Mediterranean region in the same way, partly because the temperatures there are less sensitive to the zonal wind component, and partly because there have not been systematic changes in the meridional wind component (to which the Mediterranean region is sensitive) on these time scales.
Figure 8. 10-year running mean differences (°C), land air temperature anomalies minus SST anomalies, for coastal and island locations around the globe.

Figure 9. Number of locations used to construct Fig. 8.
Figure 10. 10-year running mean differences (°C), land air temperature anomalies minus SST anomalies for coastal and island locations. Solid line, Tropics (20°N–20°S); dashed line, S. Asia (0°–30°N, 60°–130°E; 15°–30°N, 40°–60°E; 25°–30°N, 35°–40°E); dotted line, Australasia (0°–50°S, 100°–180°E).

Figure 11. As in Fig. 10, but solid line, Europe (45°–65°N, 25°W–60°E); dashed line, Mediterranean (30°–45°N, 25°W–60°E).
Fig. 12. 10-year running mean differences (°C), land air temperature anomalies minus SST anomalies, for coastal and island locations for Europe (45°–65°N, 25°W–60°E). Solid line, Jan–Mar; dashed line, Apr–Jun; dotted line, Jul–Sept; starred line, Oct–Dec.

Fig. 13. As in Fig. 12, but for N. America (30°–50°N).
North America between 30°–50°N, approximately representing the USA and southern Canada, also shows relative warmth over land in the early 20th century, but this is only marginally statistically significant. It appears to originate largely from the winter season (Fig. 13) and thus may have the same real causes as for Europe. For example, enhanced westerly flow would give enhanced positive (negative) anomalies on the west (east) coasts, with greater magnitude in the air over land than in the coastal waters. Results before 1880 are unreliable owing to sparse data.

Figures 14a to e show combined fields of SST and land air temperature anomalies for 15-year periods from 1871-85 until 1931-45. SST data are very sparse over the Pacific for the earlier periods and over the Southern Ocean throughout, but coverage is good over the Atlantic and Indian oceans (see also Fig. 2 of Folland and Parker, 1990). The blended fields exclude monthly 5°-area SST anomalies that are outside the range ± 6°C, or more than 2°C different from the average of surrounding 5°-area SST anomalies. The markedly greater coldness (relative to modern climatology) over land than over the oceans in the late 19th century reported, for example, by Jones et al. (1986) appears to arise from inner-continental and high-latitude regions, especially in winter (Fig. 14f). The reversal of signs over western Greenland, southwestern Canada and parts of Siberia between 1886-1900 (Fig. 14b) and 1931-45 (Fig. 14e) suggests that a natural fluctuation of atmospheric circulation might have caused the differences in oceanic and continental trends. Eigenvector analysis of the global surface temperature fields covering land and ocean is planned as a means of investigating this possibility. On the other hand, north-facing exposures of thermometers were common in the 19th century in Canada (Kingston, 1878) and in Russia (Wild, 1879), so that the possibility of instrumentally induced trends needs to be considered, in addition to the influence of urbanization.

Figures 15a and b are latitude-time sections of decadal zonal-mean anomalies of corrected SST and of land air temperature, respectively. Both show greater interdecadal variability at higher latitudes, suggesting that any greenhouse-gas-induced signal may be more difficult to detect at high latitudes than in the tropics. Comparison of these two sections shows that significant land-sea differences at a given latitude are not restricted to the data-sparse 19th century, and are thus likely to be real. Differences are small between 1951 and 1980 because that is the common reference period. A noteworthy recent example of relative marine coldness is the mid-latitude North Pacific (Fig. 7 with positive coefficients; Fig. 20(b) of Folland and Parker, 1990; Jones et al., 1991).

5. SATELLITE DATA AND THE FUTURE

The continuing paucity of marine data over the Southern Ocean and some parts of the tropical Pacific emphasizes the need for satellite data to ensure complete global coverage of information on surface temperature. However, biases also need to be removed from satellite data. Such biases may arise from changes in instrumentation or processing algorithms, or from changes in atmospheric transparency due, for example, to volcanic eruptions. Techniques for blending satellite and ships' SSTs in such a manner as to minimize biases are well-developed (Reynolds, 1988).

The need for regional calibration of satellite data with surface-based data, as is effectively done by Reynolds (1988), is demonstrated by the 1°C per decade global warming trend found by Strong (1989) on the basis of satellite SST for 1982–1988 calibrated using buoys which were mainly in the Southern Hemisphere. This warming is almost absent in the blended analyses (Reynolds et al., 1989) and Strong's data appear to contain biases resulting from the eruption of El Chichón in Mexico in 1982 and possibly
Figure 14. SST and land air temperature anomalies (with respect to 1951–80). SST data are with canvas-bucket corrections as in the text. Land air temperatures were provided by R. D. Jones (University of East Anglia). Contours are every $1/2^\circ$C (negative, dashed; zero, heavy solid; positive, light solid). (a) 1871–85 annual; (b) 1886–1900 annual; (c) 1901–15 annual.
Figure 14.  (continued). (d) 1916–30 annual; (e) 1931–45 annual; (f) 1886–1900 October through March.

from more recent instrumental problems. A comparison of satellite, blended and ship data with expendable bathy-thermograph (XBT) data for 1982–1984 (Fig. 16) indicates that the ship and blended data in the Northern Hemisphere were unbiased relative to the XBT data, whereas the satellite data contained serious time-varying biases and more variability
Figure 15a. Zonally averaged decadal-mean SST anomalies (°C) (with respect to 1951–80) updated every 5 years for 1861–70 to 1976–85. Values are calculated for every 5° of latitude. At least 10% of the ocean of a zone had to be covered for a value to be plotted. Canvas-bucket corrections applied as in text.

than the ship and blended data. The fact that the blended data compare so well with the XBT data indicates that satellite data can be used to help monitor climatic trends so long as these data are continually corrected for biases.
Figure 15b. Zonally averaged decadal mean land air temperature anomalies (°C) (with respect to 1951-80), updated every 5 years for 1861-70 to 1971-80 then 1976-84. Values are calculated for every 5° of latitude. At least 10% of the land of a zone had to be covered for a value to be plotted. Data were provided by P. D. Jones, University of East Anglia.
Figure 16. Mean differences between Meteorological Office (UK) ship, Climate Analysis Center (USA) \textit{in situ}, satellite and blended analyses and quality controlled expendable bathythermograph data for the region between 0° and 60°N for January 1982 to December 1984.

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Wild, H., 1879: Thermometer installations for the determination of the true temperature of the air. Imp. Akad. Nauk., Met. Sbornik, VI, No. 9, St. Petersburg. 18 pp. + 5 plates. [In German, but English translation available at the Meteorological Office Library, Bracknell, UK.]
Climate Trends, the U.S. Drought of 1988, and Access to Data

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ABSTRACT. Selected climate trends are presented, together with information about selected data sets, and where a scientist can find more data for climate research. One source of plots of climatic indices is referenced. Plans for re-analysis of the atmosphere for 35 years are summarized. The greenhouse effect and the U.S. drought of 1988 are discussed.

1. INTRODUCTION

It is important that good climate-trends data be available for such variables as surface temperature, pressure patterns, ocean surface temperature, precipitation, snow, glaciers (extent and volume), sea ice (extent and thickness), tropospheric temperature, stratospheric temperature, ozone, methane and carbon dioxide. Fortunately, a lot of good research has already been done to determine trends, but it is not complete. Many data sets must be prepared, and the relationships among variables and the teleconnections over the globe have to be clarified further.

Individual variables, such as sea ice extent or glacier indices, cannot be understood independently of the environment that influences them. For example, we need to know how the temperature, winds and radiation over the ice caps and Arctic Ocean have changed with time. We need to know how ocean surface winds and currents have changed. Our knowledge of the overall day-to-day atmospheric changes and monthly summaries is still not adequate. To improve our information base, it should be possible to make new analyses of the atmosphere from 1950 or 1957 onward. This initiative will be discussed later.

A selection of the climate-trends research projects and data sets will now be summarized.
2. OBSERVED WARMING SINCE 1880

Over the continental U.S. there has been only a small temperature trend from 1901-84 (+0.16°C). The trends over the U.S. show a temperature increase from 1901-40 and a decrease after that (Karl and Jones, 1989). These authors used a U.S. data set with monthly data from 1219 stations having long periods of record prepared by the National Climatic Data Center (NCDC). (A list of acronyms is given in the appendix.) These are usually rural stations that have had few changes for long periods of time. Karl and Jones (1989) picked a large subset of these stations, with the fewest changes, to determine that the U.S. had only this minor temperature trend.

There is a remaining urban bias in the studies of global warming from 1880-1990, but it should not be greater than about 0.1°C. Probably the best number for the amount of real global warming in the last century is about 0.4°C (Kerr, 1989). Based on the various studies, I agree with this number.

3. ATMOSPHERIC TEMPERATURE FOR 1979-88 BASED ON SATELLITE DATA

A series of NOAA satellites that started making observations in October 1978 have had several microwave channels that measure temperature for different layers in the atmosphere. NASA scientists (Spencer and Christy, 1990) used Channel Two, which has most weight between about 1 and 10 km height in the troposphere. The results for the 10-year period show almost no trend, only excursions of about ±1.0°C in monthly temperatures. NOAA-6 and -7 satellites, in this series, overlapped for two years. There is only a 0.01°C difference between the monthly global temperatures derived from each satellite. The offset between satellites is usually about 0.1°C; the biggest offset was 0.3°C. The offsets are removed by the processing methods. The authors think that the precision of their monthly global anomalies is about 0.01°C (Spencer and Christy, 1990). Jones and Wigley (1990) commented on the NASA paper to point out that the satellite data are not inconsistent with surface data. Both are important.

3.1. Correlation of the Tropospheric Data with Surface Temperatures

Spencer and Christy (1990) correlated their monthly tropospheric temperatures (from microwave data) over the U.S. with surface observations from Tom Karl of NCDC. The correlation is 0.9. Over ocean areas the surface air temperature is coupled more closely with the ocean temperature than with the mid-tropospheric temperature. Therefore, one expects the correlation to be worse. Over the Caribbean the correlation is 0.6. Over the Western Pacific and the ocean near Australia the correlation is near zero. They have used rawinsonde (raob) data to calculate the expected correlation between mid-tropospheric temperature and surface temperature. Their results with satellite data are consistent with results from raob data.

3.2. Comparison with Raob Temperature Trends

Folland et al. (1990) have compared the global, annual mid-tropospheric microwave temperatures from NASA (obtained for 1979-88, ten numbers), with similar world mid-tropospheric temperatures (850-300 mb thickness temperature) from Jim Angell’s sample of world raob (63 stations, also ten numbers for ten years; Angell, 1988). The correlation between the two sets of annual means is 0.975, a very high correlation. The RMS difference between the annual global means from the two series (after averages are removed) is only 0.02°C. This gives us more confidence that the global temperature trends from 1958, based on raob data, are probably correct.
3.3. Trends of River Discharge Data

Fifty major rivers of the world were used by Probst and Tardy (1987) to study long trends in river discharge. Only one figure from their paper is presented here, Fig. 1.

3.4. A Sea Water Temperature Trend That Wasn’t

In early 1989 it was reported that satellites had detected a significant warming (about 0.1°C each year) in the world’s sea surface temperatures (SST) from 1982–88 (Strong, 1989). In Science News (April 22, 1989), Richard Reynolds of the NOAA Climate Analysis Center was quoted as saying that he was “flabbergasted” by the reported warming in the

![Figure 1](image.png)

**Figure 1.** Comparison between the total runoff fluctuations for the different continents and for the whole world. Five-year moving averages were calculated on standardized data. [From Probst and Tardy, 1987.]
satellite data. "I think the whole thing is an error." Reynolds et al. (1989) showed that in 1982–83 the satellite data were nearly $1.0^\circ$C colder than analyses based on a blend of in-situ and satellite data. The methods of analysis-blending remove any biases in the satellite data. The blended data show very little temperature rise from 1982–88. The satellite data were affected by the eruption of El Chichón in 1982. I was told that the satellite SST algorithms could detect dust contamination only if the effect was greater than about $2^\circ$C. In any case, a six- or seven-year period is a very short period to use for climate trends. Also see Robock, (1989) for a letter and reply on this subject. Groups that use the basic Spot SST data from satellites need to be aware of this information.

3.5. Plots of Climatic Indices

A useful compilation of a number of climatic indices is in Hattemer-Frey et al. (1986). The authors gathered figures from many publications and provide background information about the data. A partial listing of the plotted data is presented in Table 1.

Table 1. Climatic data compiled by Hattemer-Frey et al. (1986).

<table>
<thead>
<tr>
<th>Quantity</th>
<th>Years</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tropospheric and stratospheric temperatures</td>
<td>1958–1981</td>
</tr>
<tr>
<td>Central England temperature</td>
<td>1670–1975</td>
</tr>
<tr>
<td>England and Wales precipitation</td>
<td>1766–1980</td>
</tr>
<tr>
<td>Beijing precipitation</td>
<td>1482–1960</td>
</tr>
<tr>
<td>Arctic region air temperature</td>
<td>1880–1980</td>
</tr>
<tr>
<td>Indian monsoon rainfall index</td>
<td>1871–1978</td>
</tr>
<tr>
<td>Rainfall series from sub-Sahara</td>
<td>1901–1980</td>
</tr>
<tr>
<td>Contiguous U.S. area average temperature</td>
<td>1895–1984</td>
</tr>
<tr>
<td>New Zealand temperature and precipitation</td>
<td>1853–1975</td>
</tr>
<tr>
<td>Changes in levels of three E. African lakes</td>
<td>1880–1974</td>
</tr>
<tr>
<td>Temperature, Sea Level Pressure,</td>
<td></td>
</tr>
<tr>
<td>Southern Oscillation Indices (tree rings)</td>
<td>1602–1961</td>
</tr>
<tr>
<td>Southern Oscillation Indices</td>
<td>1942–1972</td>
</tr>
<tr>
<td>Global sea level, tide gages</td>
<td>1880–1980</td>
</tr>
<tr>
<td>Dust veil index</td>
<td>1500–1983</td>
</tr>
<tr>
<td>Volcanic activity from Greenland ice cores</td>
<td>533–1972</td>
</tr>
<tr>
<td>Sunspot numbers</td>
<td>1875–1981</td>
</tr>
</tbody>
</table>

4. IS THE OBSERVED WARMING CAUSED BY THE GREENHOUSE EFFECT?

If we assume that the global surface warming over the previous century is about $0.4^\circ$C, the next question is whether it is caused by the greenhouse effect. Many historical and paleoclimatic studies have been made that give a considerable amount of evidence for the climate over the last 1000 years and much longer. These studies are based on data such as tree-ring growth, harvests of grapes and other crops, pollen deposits, and various historical documents. It appears that during the past five or ten centuries the climate showed periods of warming and cooling that are not unlike what we have measured in the past century. This climate variability happened without a greenhouse effect caused by the activities of mankind. Even long runs of climate models (with no change in greenhouse gases) show a natural variability. Therefore, on the basis of observations, the present warming cannot be said to be caused by the greenhouse effect.
5. THE GREENHOUSE EFFECT

In spite of the relatively small amount of warming that has been observed, there is good reason to believe that the basic physics behind the greenhouse effect is correct, and that a strong increase in greenhouse gases would cause a warming. However, there is still much uncertainty about the climate feedback processes in climate models, especially the cloud feedbacks. There is, therefore, an uncertainty in the amount of warming that should be expected from a given change in greenhouse gases.

Simple one-dimensional models have been run to estimate the magnitude of the effect of increased greenhouse gases. GISS found that the earth’s surface temperatures would increase by 1.2°C (caused by primary radiative changes) due to a doubling of CO₂. The associated increase in water vapor (by 33%) produced an additional greenhouse effect of 1.6°C. Changes in cloud amount (a decrease) added 0.8°C. With changes in sea ice and snow cover, the earth’s surface was darker so that more radiation was absorbed; this added another 0.4°C to the warming, giving a total of 4.0°C for a CO₂ doubling. This is a large change and no one believes that the clouds are right, as yet. Also, note that the water vapor feedback, a large 1.6°C, is associated with the total temperature change of 4°C. If the overall temperature changes were less, this feedback would also be reduced.

We will now present some results from climate model runs to show the climate effect of increasing the energy from the sun, or of increasing the greenhouse gases. The cloud response will also be shown. If the sun’s energy intensity is increased by 2%, the effect on the simulated earth surface temperature is the same as for a doubling of CO₂ (about 3 to 4.5°C in most present models).

Table 2 shows the response of clouds and planetary radiation in three climate models. Note that the model response to a doubling of CO₂ is to have fewer clouds and a darker planet that absorbs more of the sun’s energy. This accounts for part of the warming. At first this response seems surprising since the models give more precipitation which could mean more clouds. However, the low clouds are reduced in the warmer model climate. It is clear that these are all complicated processes and it will take some years of research to narrow our uncertainty.

This table shows the response of clouds, planetary albedo and other quantities to a doubling of CO₂. Please note that the planetary albedo drops significantly in the 2xCO₂ runs compared to the control run. This lets more energy into atmospheric system and accounts for part of the surface warming. The earth’s present albedo, based on measurements from satellites is between 28 and 30%.

The United Kingdom Meteorological Office (UKMO) climate model run (completed June 1986) had a warming of 5.2°C when CO₂ is doubled. In the standard model, cloud cover is a function of relative humidity (RH). Another version of the model was constructed with an explicit cloud-water variable (CW); in this case, a balance is maintained between cloud water content and water vapor. Ice particles fell out quickly (probably too fast) with a fixed speed of 1 ms⁻¹. On replacing the RH cloud scheme with the CW scheme, the global annual average surface warming (for double CO₂) changed from 5.2°C to 2.7°C (Mitchell et al., 1989). With the RH scheme, cloud reductions (for 2xCO₂) occur throughout the mid-latitude troposphere. With CW, the main difference is near the freezing level; in the warmer doubled-CO₂ simulation, ice cloud which is depleted rapidly by precipitation, is replaced by water cloud which is depleted slowly. Therefore the amount of clouds increases, which gives a smaller temperature increase.

Since 1 ms⁻¹ may be too large a fall rate for ice particles, an experiment was run in which these rates were reduced to about 0.2 ms⁻¹ for high clouds and to 0.7 ms⁻¹ near the freezing level. The resulting global warming was 3.2°C compared to 2.7°C above. Another experiment was run in which cloud radiative properties depended on cloud water content.
Table 2. Planetary albedo and clouds in three model runs. The model response for clouds, planetary albedo and other quantities is given for a doubling of CO₂.

**GFDL Run with Q-flux** (completed Feb 1988). Total global cloud cover (area weighted) and total global planetary albedo are given from the R15 Q-flux* run (resolution 4.4°latitude x 7.5°longitude).

<table>
<thead>
<tr>
<th>Total Global Clouds</th>
<th>Planetary Albedo</th>
<th>Temp./Precip. Change</th>
</tr>
</thead>
<tbody>
<tr>
<td>1xCO₂ Run</td>
<td>50.93%</td>
<td>33.83%</td>
</tr>
<tr>
<td>2xCO₂ Run</td>
<td>50.59%</td>
<td>32.66%</td>
</tr>
<tr>
<td>Change</td>
<td>-0.34%</td>
<td>-1.17%</td>
</tr>
</tbody>
</table>

**GISS Model Run** (dated 1982). This simulation has relatively low resolution (8°latitude x 10°longitude), but it already (in 1982) employed a Q-flux procedure and had a diurnal cycle.

<table>
<thead>
<tr>
<th>Planetary Albedo</th>
<th>Outgoing IR</th>
<th>Temp./Precip. Change</th>
</tr>
</thead>
<tbody>
<tr>
<td>1xCO₂ Run</td>
<td>30.24%</td>
<td>233.0 W/m²</td>
</tr>
<tr>
<td>2xCO₂ Run</td>
<td>28.80%</td>
<td>237.5 W/m²</td>
</tr>
<tr>
<td>Change</td>
<td>-1.44%</td>
<td>+4.5 W/m²</td>
</tr>
</tbody>
</table>

**Canadian Model Run** (completed November 1989). Model Resolution 3.75°latitude/longitude

<table>
<thead>
<tr>
<th>Total Global Clouds</th>
<th>Planetary Albedo</th>
<th>Temp./Precip. Change</th>
</tr>
</thead>
<tbody>
<tr>
<td>1xCO₂ Run</td>
<td>51.82%</td>
<td>-</td>
</tr>
<tr>
<td>2xCO₂ Run</td>
<td>50.68%</td>
<td>-</td>
</tr>
<tr>
<td>Change</td>
<td>-1.14%</td>
<td>-</td>
</tr>
</tbody>
</table>

*R15 indicates a spectral model with rhomboidal truncation. The effective spatial resolution is as given. The Q-flux procedure is used to insure that sea surface temperature and sea ice are simulated correctly in the present climate. A heat flux (Q-flux) is calculated and then used to accomplish this goal.

For double CO₂, the global temperature increased by only 1.9°C compared to 2.7°C in the related experiment above. These reductions in CO₂ warming caused by changing the cloud algorithm are very large. The magnitude of the change surprised everyone when it happened. A consensus seems to be developing that the temperature change to expect from a doubling of CO₂ is probably about 2.5°C. The main model runs now all show a greater temperature change than this.


The U.S. experienced a major drought in the summer of 1988 that impacted agriculture, river shipping and other sectors of society. The way much of the news came across to
the public was that the drought was definitely caused by the greenhouse effect, and that society should expect a continuation of climate about as bad. There are real reasons to worry about a decrease in soil moisture with greenhouse warming. However, the situation was hyped beyond reason. It was a golden opportunity to give the public a feeling for the natural variability of climate, how this drought compared with past droughts, the history of past climates, and how all this fits in with the output of climate models. In this case, even reality was an exciting story. I'm afraid that much of the opportunity was lost. However, it was gratifying to see that many scientists mobilized to put the drought into a reasonable perspective. For example, the National Climatic Data Center (Karl and Heim, 1989) prepared a 100-year chart for the portion of the U.S. (48-state) area covered by severe wetness or drought (Fig. 2). The Climate Analysis Center (CAC) prepared many charts, including a 70-year series of precipitation departure for the North Central U.S. The better news stories carried information that helped provide balanced information for discerning readers. Finally, the drought was over, Thank Heavens!

At NCAR, a selection of research clippings about climate and drought was collected that included technical articles and some of the better popular news stories (Jenne 1988a, 1989a).

A colleague was in Mexico in 1989. There was a bad drought in Mexico then. I understand that Mexico then had similar publicity to the U.S. news of 1988.

7. DATA FOR STUDIES OF CLIMATE TRENDS

In the following sections, we will discuss selected data sets that can be used for climate-trend studies. Since space is not available to provide comprehensive information, we will survey a few data sets and give sources of additional information.
7.1. Daily Surface Data

For various climate studies one needs daily maximum-minimum (max.-min.) temperature and precipitation for many stations. Because of the variability of precipitation, there should be one precipitation measurement about each 70 km. The U.S. archives of daily cooperative-station data, with many stations back to 1900, are now accessible. However, similar archives from other countries have not been drawn together.

Daily precipitation observations are taken at about 140,000 stations in the world. In the U.S., for example, there are about 9,000 daily stations and 3,000 that report hourly precipitation.

The NOAA Climate Analysis Center (CAC) prepares an archive of daily max.-min. temperature and precipitation from GTS data (global telecommunication), about 7000 stations. It is available back only to 1979. In principle, it could be extended back in time to 1967, using the GTS source data.

The Air Force gathered about 107 million surface synoptic observations, worldwide, for varying periods between 1930 and 1965. These usually include daily rainfall. They also include elements that are needed such as the clouds, winds, pressure and humidity. They are on tape at Asheville and at NCAR. There are corresponding data sets of daily data that have variables such as max.-min. temperature and precipitation. It should be noted, however, that the task of gathering adequate global data sets of these types of data is far from done.

7.2. Hydrological Data

About 14,000 stations report streamflow in the U.S. The daily data are on magnetic tapes. WMO (1977) contains information about the global station density of hydrological data including precipitation, stream flow and evaporation.

UNESCO has published monthly river-discharge data for a number of the world's rivers. NCAR key-entered much of these data. The USGS and Australia did similar work. The data sets need to be combined.

![Figure 3. Meteorological daily analyses for the Southern Hemisphere. SLP means sea level pressure. Automated analyses at many levels start in 1972. The data are all digital, except for the one marked “paper.” Archive details and data are at NCAR.](image)
7.3. Analyses for Northern Hemisphere

Daily sea level pressure analyses starting in 1899 are available from NCAR on tape. The Arctic sea level pressure was too high for many years. The analysts had few observations and believed that an intense polar high pressure cell must exist. NCAR has hemispheric upper-air analyses starting about 1946, at 700 and 500 mb. Other levels start in 1950 (300 mb) and 1963 (many levels).

7.4. Analyses for Southern Hemisphere

The availability of daily meteorological analyses for the Southern Hemisphere is summarized in Fig. 3. Most of these are on tape at NCAR.

8. COMMENTS ABOUT DATA FOR ANALYSES IN SOUTHERN HEMISPHERE

From about 45°S and northward, one can accept the analyses for the whole 1951–57 period. To describe conditions further south, there were lots of ship reports for the summer whaling season (November–March) for November 1955 and later. van Loon says that the summer (November–March) analyses from November 1955 on were of equal quality with IGY analyses. The sector for South America, Antarctic Peninsula and Falklands had enough observed data for the whole period. There were very few Antarctic stations before 1956. The IGY analyses started June 1957.

The whalers were in the Atlantic and Indian Ocean areas for summers prior to November 1955, but the Pacific Ocean did not have good ship observations until the November 1955 summer.

8.1. Monthly Gridded Pressure Data, Antarctic Region

Jones and Wigley (1988) gathered monthly pressure data for a total of 29 Antarctic stations for 1957–85. This number included six sub-Antarctic islands from 47-55°S. They gathered the data directly from individual country archives because other sources had missing data and some errors. They used periods with both station data and analyses to develop a principal-component regression technique that they used to prepare sea level analyses for each season for the period 1957–85.

9. NEW ANALYSES OF PAST ATMOSPHERIC CONDITIONS

The existing daily atmospheric analyses for the period 1950–present are not of sufficient quality to permit the type of climate studies we need. There are several reasons for this: (1) there were many changes of analysis procedures, (2) much of the data did not reach the operational centers, and (3) the methods used were not good enough. For example, it is only since about 1986-88 that the ocean surface-wind analyses have been good enough to drive ocean models.

As early as 1985 there were discussions about the possibility of making re-analyses that would help achieve the goals of the TOGA (Tropical Ocean, Global Atmosphere) experiment. Some efforts to argue for a re-analysis project can be traced to as early as about 1980. From 1987–90 there has been an increasing interest in making daily re-analyses of the global atmosphere. The data assimilation methods used to analyze the state of the atmosphere have shown major advances during the 1985–90 period.
forecast is a part of these methods; therefore, the analyses improve when the forecast model is improved and when the methods improve. There has been a large advance in the capability of forecast models.

Bengtsson and Shukla (1988) published a paper that helped to start the movement toward planning for re-analyses. In early 1989 a small workshop was held to consider the initiative of making re-analyses (Kinter and Shukla, 1989).

There has been a clear interest in analyzing the whole period 1979–90 starting with two years in the 1980s as a pilot project. Now there are plans to get ready to do the whole period 1958-90. There were very few Antarctic observations prior to 1956. Therefore, if we go back further in time, data will not be available to help the analysis in some regions.

Do we have the data inputs needed to re-analyze the atmosphere? There are many source data sets that can be combined to produce data sets that will be significantly better than the data that any center used for the operational analyses that are now available. Such projects to prepare data are starting for surface land data, rawinsonde data, aircraft data, and satellite cloud-wind observations.

There are several existing data sets of global surface land observations and rawinsonde data. Data are available from telecom (real-time), from delayed, high-quality archives in national data centers and from previous data collection projects. The plan is to combine data from several sources and have initial data sets ready to start long-period re-analyses by November 1992. It will take a 10-year effort to keep improving the availability and quality of data, especially for the earlier years. The results of these data-preparation efforts will be valuable for many purposes beyond re-analyses.

For the necessary surface marine data, a project (COADS) was started in 1982 to prepare the best data set of world observations from 1854 on. This has involved cooperation between NOAA/ERL, NCAR and NCDC/Asheville (Woodruff et al., 1987). Canada is helping to prepare the drifting-buoy data. We obtain a significant number of additional ship reports by waiting until delayed reports become available. During the 1980s the number of unique ship reports available is about $1.20 \times 10^6$ per year in real time, $1.86 \times 10^6$ after a delay of one year, and $2.22 \times 10^6$ after five years.

What data would be available from a re-analysis effort? The normal variables (temperature, pressure, wind, moisture) for the atmosphere would be available. In addition, diagnostic terms such as precipitation, clouds, surface radiation and total surface energy budget would also be available.

The new surface winds and the surface flux data would be used to drive ocean models for the whole period. Also, the boundary layer conditions over the ocean and ice caps will be saved.

A paper is available that summarizes the status of data sets (and years of coverage) that can be combined to provide inputs for re-analyses (Jenne, 1988). It also gives a history of re-analyses done for the FGGE period (1979) to help scale the effort. Many additional details are available at NCAR.

9.1. Output of Snow, Temperature and Radiation from Analyses

In the recent years the assimilation schemes used to analyze the atmosphere are good enough to produce many diagnostic terms from the associated forecast model. The variables include precipitation, snowfall, temperature, winds, clouds, radiation, surface stress, surface heat fluxes and planetary radiation. NCAR has a list of the variables that are available.

These terms are being saved from some operational forecast models (NMC and ECMWF), in addition to the normal pressure, temperature and wind.
9.2. Satellite Sounder Data

Several sets of global satellite sounder data are available. The sounder data have channels (such as window IR) that can be used for other purposes besides deriving atmospheric temperature soundings.

- NIMBUS SIRS, April 1969 – April 1971
- NOAA VTPR, November 1972 – 28 February 1979 (8 IR channels)
- NOAA TOVS, 29 October 1978 – Present (channels in VIS, IR, microwave, and for stratosphere)

10. DATA FROM CLIMATE MODELS (1xCO₂ and 2xCO₂)

NCAR has data from several of the world's climate models. These include five different model experiments for the present climate (1xCO₂) and for a doubling of CO₂ (2xCO₂). There are also two transient runs from GISS, one to the year 2062. EPA sponsored this effort (starting in 1987) to prepare data in a common format to support assessment studies of climate changes on crops, forests and rivers. The primary data sets contain 10 to 20 variables (e.g., precipitation, temperature and surface radiation) to support these studies. There are data for 10-year monthly means of 1x and 2xCO₂ climates. For some runs there are data for each individual model month. For one GFDL run, NCAR has daily output for a sample of three years in each of the 1xCO₂ and 2xCO₂ runs. The resolution of most climate models is now about 500 km, so they cannot show the details of climate changes caused by local topography. The model resolution is improving.

11. SNOW COVER AND SEA ICE DATA

There are data sets of weekly sea ice and land snow cover. Many land stations report snowfall, but these data are still often hard to obtain, worldwide. A workshop "Snow Watch 85: Workshop on CO₂/Snow Interactions" produced a report "Snow Watch 85," that has over 20 papers that describe snow-cover data sets and make comparisons. It includes a discussion of weekly sea ice charts available from 1972. Weekly charts of Northern Hemisphere snow and sea-ice boundaries have been prepared starting 1966.

Satellites that observed passive microwave data from which sea ice and other variables can be estimated were: NIMBUS (ESMR instrument), December 72–May 77; NIMBUS-7 (SMMR), October 78–August 87; DMSP (SSMI instrument), 9 July 87–present. The status of the ice products evolves with time. The data are primarily located at NSSDC (Goddard; Olsen, 1990), JPL, and the Snow and Ice Data Center (WDC-A) in Boulder, Colorado. Sea-ice products on CD-ROM are becoming available.

For the north polar area, NCAR has a tape from John Walsh that has tenths of ice coverage on a 1° (60 nautical mile) resolution grid. The period is for each month (1953–1988, inclusive).

12. USE OF CD-ROM AND DAT TECHNOLOGIES TO DISTRIBUTE DATA

The first CD-ROMs with data for the geosciences were produced in mid–1987. Now a flood of them is being prepared. A CD-ROM holds about 650 MB, compared to 125 MB on a high-density, half-inch tape. The access time to any part of the disk is about 0.5 second. A reader costs about $700. The 4 mm digital audio tape drives (DAT) can now
be purchased for about $2,200. The price may come down to $1,000 by 1992. These small tapes hold 1300 MB of data.

When a CD-ROM is received, software to access the data on a PC is also provided. Some simple browse displays are often included.

13. SOURCES OF INFORMATION ABOUT AVAILABLE DATA

Selected sources of information about data that are available are presented in the sections that follow. Some indication of the types of information that can be found within each of these major sources is also included.

13.1. National Climatic Data Center and USAF/ETAC

This NOAA center (NCDC) at Asheville, North Carolina gathers the climate observations for the U.S. It also helps WMO by gathering selected world data such as monthly surface and upper-air data, CO₂ flask observations, and atmospheric turbidity for the world. There are many data sets. The most comprehensive summary of these data is in the data chapter (Jenne and McKee, 1985) of the Handbook of Applied Meteorology. Also see the INFOCLIMA listings (see below).

The Air Force Data Center (USAF–ETAC) is co-located with NCDC. It has done a fine job of gathering selected worldwide observations. There are many sets of digitized observations that were not prepared elsewhere, especially international data for periods prior to 1965.

13.2. National Center for Atmospheric Research (NCAR)

The relatively small data center within NCAR has a large archive of over 350 data sets (over 16 trillion bits). The data are from many sources including NMC, NCDC, various countries, ECMWF, USAF, Navy and research laboratories. The catalog "Data Availability at NCAR," (Jenne, 1989) describes the data sets within 24 categories of data. (The data categories include analyses, rawinsondes, ocean data, stratospheric data sets, paleoclimate, clouds, climate models and data received from the USSR). The publication includes references to catalogs at other centers.

13.3. DOE Carbon Dioxide Information Center

This center at Oak Ridge, Tennessee has a number of data sets relating to the carbon cycle and to climate. These include carbon dioxide measurements, fossil fuel emissions, and the role of oceans (tracers and coral growth). Many sets of biospheric data are included such as carbon in vegetation, FAO land use, and changes in soils and carbon in rivers (U.S.-DOE, 1989).

A number of climate and paleoclimate data series are also available (northern hemisphere temperature, 1851–1900; central England temperature, 1659–1983; and worldwide cloud cover). Some of the paleoclimate series include: CLIMAP data 18,000 years ago, and tree-ring data bank and dryness/wetness indices in China for the past 500 years.

13.4. INFOCLIMA (Catalogue of Climate System data sets), 1989 edition, WMO/ TD-No. 293, Geneva

The WMO (World Meteorological Organization) in Geneva includes divisions for both meteorology and hydrology. This data catalog (WMO, 1989) has 507 pages (2.5 cm
thick). It includes individual observations and summaries held at various data centers. It lists data centers worldwide. It has data set descriptions in the categories:

- Upper-air data (54 pages)
- Surface climatological data (140 pages)
- Radiation data at the surface (36 pages)
- Maritime and ocean data (50 pages)
- Cryosphere data (14 pages)
- Atmospheric composition data (18 pages)
- Hydrological data (42 pages)
- Historical and proxy data (28 pages)

A total of 920 data sets are listed. Data sets that cover global and regional areas are handled separately from those that include only national data. The catalog may be made available on computer floppy disks in 1990.

13.5. U.S. National Online Data Catalog

An effort was organized by NASA, starting about 1987, to provide a central point where the user community could do an online computer search to help locate data sets for climate and other disciplines. Various government agencies and research laboratories contribute information about their data sets. This central catalog has descriptions of about 1000 data sets and listings of various data sources. Contact NSSDC, code 633, Goddard, Greenbelt, MD 20771.

APPENDIX. ACRONYMS AND ABBREVIATIONS

CAC Climate Analysis Center (NC A)
CD-ROM Compact Disk; Read Only Memory (holds 650MB of data)
DMSP Defense Meteorological Satellite Program (A series of orbiting satellites also has this name.)
ERL Environmental Research Laboratories (NOAA)
ESMR Electrically scanning microwave radiometer (provides sea-ice concentration, snow cover, etc.)
ETAC Environmental Technical Applications Center (USAF)
FAO Food and Agriculture Organization
FGGE First GARP Global Experiment (in 1979)
GARP Global Atmospheric Research Programme
GFDL Geophysical Fluid Dynamics Laboratory (NOAA)
GISS Goddard Institute for Space Studies (NASA)
IR Infrared
JPL Jet Propulsion Laboratory (NASA)
NASA National Aeronautics and Space Administration
NCAR National Center for Atmospheric Research
NCDC National Climatic Data Center (Asheville)
NIMBUS A series of NASA remote sensing satellites
NMC National Meteorological Center
NOAA National Oceanic and Atmospheric Administration
NOTOS A Greek word for the south wind. Also the name of a journal in South Africa
NSSDC National Space Science Data Center
RAOB: Station data for temperature, humidity and winds aloft obtained from rawinsonde balloons
RMS: Root mean square
SLP: Sea level pressure
SMMR: Scanning Multichannel Microwave Radiometer (provides sea-surface temperature, marine wind speed, sea ice concentration, snow cover, water-vapor content, etc.)
SSMI: Special Sensor Microwave Imager (provides data similar to SMMR)
SST: Sea surface temperature
TOVS: TIROS Operational Vertical Sounder (data from a sensor on a satellite series that started October 1978)
UKMO: United Kingdom Meteorological Office
UNESCO: United Nations Educational, Scientific and Cultural Organization
USAF: United States Air Force
USGS: United States Geological Survey
WDC: World Data Center
WMO: World Meteorological Organization

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Jenne, R. L., 1989: Data Availability at NCAR. NCAR, P.O. Box 3000, Boulder, CO 80307. 45 pp. (This is updated each 15-20 months.)


WMO, 1977: Statistical Information on Activities in Operational Hydrology. WMO No. 464, Geneva. (Includes numbers of observing stations by country for rainfall, evaporation, stream flow, etc.).

Regional Greenhouse Climate Effects*

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ABSTRACT. We discuss the impact of an increasing greenhouse effect on three aspects of regional climate: droughts, storms and temperature. A continuation of current growth rates of greenhouse gases causes an increase in the frequency and severity of droughts in our climate model simulations, with the greatest impacts in broad regions of the subtropics and middle latitudes. But the greenhouse effect enhances both ends of the hydrologic cycle in the model, that is, there is an increased frequency of extreme wet situations, as well as increased drought. Model results are shown to imply that increased greenhouse warming will lead to more intense thunderstorms, that is, deeper thunderstorms with greater rainfall. Emanuel has shown that the model results also imply that the greenhouse warming leads to more destructive tropical cyclones. We present updated records of observed temperatures and show that the observations and model results, averaged over the globe and over the United States, are generally consistent.

The impacts of simulated climate changes on droughts, storms and temperature provide no evidence that there will be regional "winners" if greenhouse gases continue to increase rapidly.

*This report, except for the first two paragraphs of the Introduction, is a condensed version of a paper published in the proceedings of the Second North American Conference on Preparing for Climate Change, Climate Institute, Washington, D.C., 1989.
1. INTRODUCTION

In testimony to the United States Senate on June 23, 1988, one of us (Hansen, 1988) made the following three assertions: (1) it can be stated with about 99% confidence that the earth is getting warmer; the 1980s are the warmest decade in the period of instrumental record, the four warmest years are all in the 1980s, and 1988 is likely to be the warmest year in the record; (2) it can be stated with a high degree of confidence that there is a cause and effect relationship between the greenhouse mechanism and the global warming trend; and (3) in the Goddard Institute for Space Studies (GISS) climate simulations, by the late 1980s and the 1990s, there is an increased tendency toward droughts in the United States.

The first two assertions, about global warming and its relation to the greenhouse effect, are being debated, as evidenced by other papers in this volume. If global warming proceeds near the rate projected in the GISS climate simulations (Hansen et al., 1988), both of these issues may become moot sometime in the 1990s, as global climate change becomes apparent to the man-in-the-street. In that event, the focus of attention will undoubtedly be on regional climate change. In this paper we attempt to identify regional climate changes which are likely to accompany an increasing greenhouse effect. Most of these results were presented at the Second North American Conference on Preparing for Climate Change, held in Washington, D.C., December 6–8, 1989.

The conventional wisdom is that it is not yet possible to obtain reliable conclusions about regional climate impacts, principally for two reasons. First, the representations of atmospheric and surface processes in current climate models are highly simplified, and the models show a very wide range in their predictions for climate change at any particular region. Second, none of the existing climate models simulates the ocean realistically, and changes in ocean currents could alter regional climate.

These are valid concerns, especially if attention is focused on climate change at some specific time and locale. However, an increasing greenhouse effect undoubtedly implies some broad changes in the nature of regional climate, which it may be possible to investigate with existing modeling capabilities. In Section 2 we consider changes of a drought index, defined as the difference between atmospheric supply of moisture and atmospheric demand for moisture. In Section 3 we consider the impact of greenhouse warming on atmospheric stability as it affects convective storm intensity on scales from thunderstorms to tropical cyclones. In Section 4 we discuss possible impacts of rising temperature itself, and we illustrate global and United States temperature trends. In Section 5 we discuss the 1988 North American and Asian heat waves and their possible relation with the greenhouse effect. Finally, in Section 6, we summarize our conclusions.

2. DROUGHT

We define a drought index, D, which is a normalized measure of the difference between atmospheric supply of moisture and atmospheric demand for moisture, as follows:

\[ D(\text{current month}) = 0.9D(\text{previous month}) + \frac{d}{\sigma} \]

where

\[ d = (\text{precipitation} - \text{potential evaporation})_{\text{actual}} \]

\[ - (\text{precipitation} - \text{potential evaporation})_{\text{climatology}} \]
and

\[ \sigma = \text{standard deviation of } d, \]

with

\[ D(\text{first month}) = \frac{d}{\sigma}. \]

Potential evaporation is the evaporation which occurs if water is available. The ratio \( \frac{d}{\sigma} \) is a dimensionless measure of the precipitation deficit (or excess, if positive) in the current month. The drought index, \( D \), includes a memory of precipitation deficit over preceding months. Note that the drought index continues to yield negative values after evaporation ceases due to lack of available water. It thus provides an indication of stress on vegetation. It is also a relevant measure of reservoir water balance, since there is normally water available for evaporation from a reservoir.

The drought index we have defined is similar to, but simpler than, the Palmer Drought Index (Alley, 1984). The Palmer Drought Index has many locally defined parameters that would make it impractical for us to obtain global results. However, we have calculated the Palmer Drought Index for the United States and have verified that the results obtained using it show characteristics of future drought intensification similar to those illustrated below. The factor 0.9 in the definition of \( D \), which implies a time scale for recovery (i.e., a memory) of 9–10 months, is the same as in the Palmer Drought Index. We have tested recovery times as short as 1–3 months, verifying that the results discussed below are qualitatively unchanged.

We have computed the drought index \( D \) for some of our computer climate simulations published elsewhere (Hansen et al., 1988). These simulations were carried out with our global climate model (GCM) which has a global sensitivity 4.2°C for doubled CO\(_2\). Ocean heat transports were assumed to remain the same in the next few decades as estimates for the recent past. Other characteristics and qualifications for these climate simulations have been documented (Hansen et al., 1988, 1983).

The drought index obtained for our trace-gas scenario A is shown in Fig. 1 for June–July–August (Northern Hemisphere summer) of four specific years. The temperature anomalies for the same years, relative to the 100-year control run, are shown for comparison in Fig. 2. Scenario A has rapid growth of trace-gas emissions, for example, 1.5% per year for CO\(_2\) and 3% per year for CFC's; we describe this scenario as “business as usual,” because it may be realistic if there are no controls on trace-gas emissions. The shading scheme in Fig. 1 divides the drought index into categories according to the percent of time that a given drought index occurs in the 100-year control-run of the climate model; the control run had 1958 atmospheric composition. The drought index is defined locally, that is, relative to the control-run climate at each location. Thus, for example, dry conditions in a rainforest indicate only that it is dry relative to the mean for that location in the control run.

The exact patterns of the drought index and temperature are of course not intended to be forecasts for individual years because the climate patterns fluctuate almost chaotically on a year-to-year basis. However, it is meaningful to search for overall trends in the results. In the 1990s there is a tendency for more extensive dry conditions than in the control run. By the 2020s there is no mistaking the great intensification of drought at almost all middle-latitude and low-latitude land areas. There is also an intensification of wet regions, especially at high latitudes and in the intertropical convergence zone, the latter being the region where the low-latitude trade winds of the two hemispheres collide.
A summary of the drought intensification with time is shown in Fig. 3 for scenario A, averaged over all land areas except Antarctica. In this figure we define the degrees of dryness intensification which occur 1%, 5% and 16% of the time in the control run as extreme drought, drought, and dry, respectively. Drought conditions, which occur 5% of the time in the control run, have increased to 10% in the 1990s. In scenario A drought conditions continue to increase rapidly, to about 25% in the 2020s and about 45% in the 2050s.

What processes in the model lead to such rapid drought intensification? Of course the principal factor is the higher surface air temperature which increases the potential evaporation. More detailed analysis is hampered by the fact that droughts occur in different places in different years and are interspersed with wet periods. Therefore, we have sorted model diagnostics according to drought index, which allows us to examine how the drought characteristics change without concern as to where the droughts occur. We find that the regions with more negative drought index ("dry" regions) tend to warm more than the "wet" regions as the greenhouse warming increases. Several characteristic differences between the dry and wet regions in the control run, specifically, reduced rainfall, fewer low clouds and less spring soil moisture in the dry regions, all tend to be further enhanced as the greenhouse warming increases. No doubt low antecedent soil moisture is one factor which helps determine the location of droughts, in part through positive feedbacks such as reduced evaporation and cloud cover. But many other factors, such as atmospheric longwave patterns and ocean temperature distributions, can influence
the location and timing of droughts. We emphasize that, even as droughts intensify with a growing greenhouse effect, all of the droughts continue to be "natural," in the sense that their location and timing can be related to antecedent land, atmosphere and ocean conditions.

The qualitative picture that emerges is an intensification of both dry and wet extreme conditions as global temperature increases. A similar result is found for rainfall variability by itself (Rind et al., 1989), but the effect is stronger for the drought index because of the effect of warmer temperatures. The fundamental mechanism is increased heating of the surface. In dry regions, where little water is available for evaporation, the increased heating goes mainly into increasing the air temperature, which reduces low-level cloud and thus causes further heating. But over the oceans and land regions which happen to be wet, the added greenhouse heating increases evaporation rates, leading to more intense storms, as discussed below, and to increased rainfall and floods.

Although the increasing drought frequency that we obtain may seem extreme, the changes of the drought index would be even larger if we used the climate parameters obtained by the GFDL (Geophysical Fluid Dynamics Laboratory) model of Manabe and Wetherald (1987) because they obtain greater temperature increase and precipitation reduction at middle latitudes than we obtain with our model. The qualitative changes that we obtain in regions of increased drought, for example, decreased low-cloud cover and reduced spring soil moisture, are similar to the results that Manabe and Wetherald obtained for doubled CO₂ in North America and Asia, where their model developed strong drought conditions.
Figure 3. Drought occurrence as a function of time in scenario A. Results are averaged over all gridboxes that are more than 90% land, except that Antarctica is excluded.

The intensification of both dry and wet extreme conditions is a plausible consequence of increased surface heating and evaporation. The sense of this result is unlikely to depend upon precise simulation of regional climate patterns or on possible changes in ocean circulation. The magnitude of the effect does depend on regional climate feedbacks, such as decrease of low clouds with increasing drought intensity; this cloud feedback should be analyzed on the basis of global cloud observations. The results may also change somewhat as we improve the realism of the model, for example, by increasing the model's resolution (Rind, 1988a,b) and improving the representations of ground hydrology and moist convection, which affect precipitation patterns. But, it seems unlikely that such uncertainties will modify the sense of our result, that is, the intensification of both dry and wet extreme conditions.

3. STORMS

Storms are generally not resolved by the coarse horizontal resolution of present global climate models. However, models can provide many climate diagnostic parameters which indicate how storm intensity is likely to change with increased greenhouse warming. Some of the specific quantities we have looked at are described below.
3.1. Moist Static Energy

Moist static energy, the sum of sensible heat, latent heat and geopotential energy, is a useful indicator of the likelihood and penetration depth of moist convection. High values of moist static energy near the surface, relative to the air above, and high relative humidity favor deep penetrating convection (e.g., typical thunderstorms). Figure 4 shows the changes in the global-mean vertical profile of moist static energy which occur in our transient scenario A and doubled CO$_2$ simulations. As the greenhouse effect grows, the maximum increase of moist static energy occurs near the surface, because of the higher absolute humidity associated with increased evaporation, and in the upper troposphere, because of the peak in greenhouse warming there. However, the surface moist static energy increase is 2 kJ/kg greater than that at higher altitudes, which represents a 20% enhancement of the lower tropospheric gradient of moist static energy compared with current climate. Relative humidity changes at low levels are negative but very small in the climate simulations. The implication from these changes is that the warmer climate is prone to deeper, more penetrating convective events. A similar conclusion follows from the results obtained by Wetherald and Manabe (1988) in a GCM with a completely different cumulus parameterization. The tendency of deep convective cloud top heights to increase with increasing sea surface temperature in the tropical Pacific in the current climate (Fu et al., 1990) is also consistent with this conclusion.

3.2. Mass Flux by Moist Convection

In our climate simulations the vertical mass exchange due to deep moist convection increases and the mass flux due to shallow convection decreases, consistent with the altered thermodynamic state discussed above. The global-average depth of penetration by moist convection increases 20 mb of atmospheric pressure, from Δp = 395 mb to Δp = 415 mb, in the doubled CO$_2$ experiment, and to Δp = 405 mb by the 2050s in scenario A. The increases are largest near the equator and at middle latitudes. The height of these pressure surfaces increases by several hundred meters as a result of the warming.

3.3. Precipitation

The increased precipitation in the model as the climate warms is almost entirely in the form of moist (penetrating) convection. Changes in large-scale (stratiform) rainfall are small, in fact slightly negative, especially at middle latitudes; the latter characteristic is probably a result of reduced synoptic-scale wave activity (Rind, 1986). Atmospheric heating by moist convection, which is related to precipitation, increases 17% in the doubled CO$_2$ climate and by 10% by the 2050s in scenario A. These increases are caused by the higher absolute humidity (and latent heat content) of the warmer atmosphere and by the deeper penetration of moist convection which allows a greater percentage of the latent heat to be released in condensation. Thus thunderstorms are more intense in the model, in the sense that they have higher cloud tops and produce more rainfall for a given mass flux.

Emanuel (1987) used a simple Carnot-cycle model to estimate the effect of greenhouse warming on the maximum intensity of tropical cyclones, based on the sea surface temperature changes in the doubled CO$_2$ experiment of the GISS model. Figure 5 shows the resulting minimum sustainable surface pressure that he obtained. With today's climate the minimum sustainable surface pressure is about 880 mb, but this decreases to about 800 mb for the doubled CO$_2$ climate. The corresponding maximum wind speed increases from about 175 mph to 220 mph. Since the kinetic energy increases with the
square of the wind speed, Emanuel estimates that the destructive potential of hurricanes could increase by 40–50% with doubled CO₂.

These quantitative results were obtained for the ocean temperature warmings in the GISS model, which are as large as 4°C. Some other GCMs yield ocean warmings of only 2°C for doubled CO₂, which would imply wind speed increases half as large as those indicated here. Also, note that the maximum potential velocities are obtained in only a small percentage of hurricanes, and that the existing analyses do not permit prediction of the change in storm frequency. Nevertheless, it is obvious that the impact of greenhouse warming on tropical storms would have important implications for the Caribbean, Mexico and parts of the United States coast. And in addition to increased hurricane strength, it seems likely that higher ocean temperatures will lead to an expansion of the region that hurricanes frequent. For example, if the greenhouse warming continues unabated, hurricanes could become common along the entire east coast of the United States in the next century.

The picture that emerges from these diagnostics is an increased intensity of storms which are driven by latent heat of vaporization, both ordinary thunderstorms and mesoscale tropical storms. The basis for this is the increased evaporation and higher
temperatures at low levels in the atmosphere, which yield more moist static energy at low levels and greater vertical penetration of moist convection. The magnitudes and detailed spatial and temporal patterns of changes in storms are sensitive to uncertainties in the parameterization of moist convection and other feedback processes in the model. For example, we cannot determine changes in the updraft speed or frequency of storms with the current version of the model, nor can we predict the nature of changes in special categories of storms which depend on local wind shear and mingling of different air masses (e.g., squall lines and tornadoes). However, the general nature of those changes we have described is determined largely by the Clausius-Clapeyron equation and is thus a straightforward consequence of fundamental moist thermodynamics.

4. TEMPERATURE

We examined elsewhere (Hansen et al., 1987, 1988) temperature changes forecast by our global climate model for increasing greenhouse gases. In those papers we stressed the importance of a possible increase in the frequency of temperatures above some critical level. For example, we computed the average number of days per year that the simulated temperature exceeds certain limits in specific United States cities. Such quantities have an extremely large year-to-year variability, so they certainly will not increase smoothly as the world becomes warmer. However, it is meaningful to estimate how the probability of such extreme temperatures may change. For example, our climate model suggests that the probability of a hot summer in most of the United States may increase to 60–70%
by the middle 1990s, as compared to 33% in the period 1950–1979 (Fig. 6, Cayan et al., 1986).

The simulated warming in our model is somewhat larger in Canada and smaller in Mexico and the Caribbean, as compared to the United States. But natural climate variability (fluctuations from year to year) also increases with increasing latitude, as illustrated elsewhere (Hansen and Lebedeff, 1987). As a result, to a first approximation, the probability of a warm season relative to the local climatology is predicted to increase at a similar rate in these different regions (Hansen et al., 1987).

The impact on the biosphere of increasing temperature will be dramatic at all latitudes, if the results computed for scenario A (rapid growth of trace gases, Hansen et al., 1987) are realistic. The poleward shift of isotherms by 50 to 75 km per decade in that scenario is faster than most plants and trees are thought to be capable of naturally migrating (Davis, 1989), and thus the warming could cause a decline of many species in North American forests. The productivity of crops that are sensitive to a run of consecutive hot days could suffer also, as indicated by calculations published elsewhere (Hansen et al., 1987; Mearns et al., 1984). One impact in the Caribbean could be on coral reefs, since many coral populations are unable to survive if water temperatures rise above 30°C (Roberts, 1987). The expected increase in storm intensities, discussed in the section above, would be particularly important in the Caribbean.

Observations of current global temperature change are of special interest, because of the search for a long-term warming trend attributable to the greenhouse effect. A preliminary update of the global temperature analysis of Hansen and Lebedeff (1987), which uses MCDW (Monthly Climatic Data of the World) data available from NCAR, is shown in Fig. 6. The last six months of 1988 are based on NOAA near-real-time data, adjusted for reporting biases as described elsewhere (Hansen and Lebedeff, 1988). The use of these adjusted near-real-time data for six months affects the global temperature for the full year by at most a few hundredths of a degree. Note that Fig. 6 has not been corrected for “urban” effects. As discussed by Hansen and Lebedeff (1988) and below, approximately 0.1–0.2°C of the global warming in the past century in the MCDW data is estimated to result from urban growth effects.

Figure 6 indicates that 1988, based on the MCDW and NOAA data, was equal to the warmest year in the history of instrumental records. Jones et al. (private communication) have recently reported that their analysis, based on land and ship data, shows 1988 as the warmest year on record. The annual warmth for the globe in 1988 occurred despite rapid cooling at low latitudes between May and December, which was associated with an unusually strong negative phase of the El Niño cycle (Ropelewski, 1988). It will be interesting to see whether this cooling of tropical surface air propagates to high latitudes and dominates global temperature trends over coming years, since some scientists have expressed the expectation that this negative El Niño will slow down the greenhouse warming by 30 to 35 years. Global temperature change does have some correlation with El Niños (Fig. 3a of Hansen and Lebedeff, 1988), especially the past two El Niño events, but the correspondence is far from overwhelming. It is a classical confrontation, like the tortoise versus the hare: how long will it take a “small” global climate forcing to overcome the effects of a large negative El Niño fluctuation? If some recent climate simulations (Hansen et al., 1988) are realistic, it will be at most only a few years before the global temperature records are raised further.

The global temperature record in Fig. 6 is the average for all MCDW stations, urban and rural. We illustrated elsewhere (Hansen and Lebedeff, 1987) that if all stations associated with urban areas of population 100,000 or greater (about one third of the MCDW stations) are eliminated from this record, the global warming is reduced by 0.1°C. Based on studies of how the urban warming varies with population, we estimated
Figure 6. Global surface air temperature change estimated from meteorological station data. Uncertainty bars (Hansen and Lebedeff, 1988) account only for the incomplete spatial coverage of the stations. The error bar for 1988 is larger than that indicated for 1987 because of poorer station coverage and approximations in the near-real-time data used for the last several months of the year. Note also that no correction has been made in this figure for urban warming, which is estimated as 0.1–0.2°C for the century.

that there may be an additional urban effect of about 0.1°C due to smaller cities. Jones et al. (1986) used an independent labor-intensive procedure to analyze urban effects, comparing nearby stations on a one-by-one basis to try to identify and partially correct for urban bias. Their results and ours yield a net global warming of approximately 0.5°C in the past century, which is the magnitude of warming that we and others have used for empirical studies of the greenhouse effect and climate sensitivity (Hansen et al., 1984).

The temperature trend in the United States can be examined in detail, based on comprehensive studies by Karl et al., (1988a,b). They employ the recently completed Historical Climatology Network (HCN) data for 1219 stations, data which were meticulously scrutinized for biases resulting from such factors as station moves, time-of-observation changes and instrumental changes. By comparing urban temperature records with those of nearby rural stations, Karl et al. (1988a) obtained empirical relations between population and urban warming. The dependence of urban warming on population that they found is generally consistent with earlier studies, and thus does not modify our estimate of 0.5°C global warming in the past century.
Most of the stations in the Historical Climatology Network are located in sparsely populated regions, over 70% in areas with 1980 population below 10,000. Thus the urban warming which Karl et al. (1988a,b) found in the raw HCN data was small, amounting to 0.06°C in this century. Here we use the urban-adjusted HCN record, that is, after removal of this urban warming to: (1) estimate the urban warming in the MCDW data for the United States, (2) examine the United States temperature record for evidence of a trend, and (3) compare with the results of our global climate model simulations. The HCN record has been updated to include 1985, 1986 and 1987.

Figure 7a compares the urban-adjusted HCN data for the contiguous United States with uncorrected data of Hansen and Lebedeff (1987) based on MDCW stations. In this comparison the HCN data have been used to calculate a U.S. area-average by equally weighting each HCN station within 23 regions defined by Karl et al. (1988a,b). The MCDW data have been averaged by using the interpolation procedure described by Hansen and Lebedeff (1987). This comparison suggests that there is an urban warming bias of about 0.13–0.14°C/century in the MCDW data for the United States under the assumption that urban effects are the cause of the difference in the temperature trends. [The linear trend of the urban-adjusted HCN data of Karl et al. (1988a,b) for the interval 1901–1987 is 0.26°C/century; the Hansen and Lebedeff (1987) published data for the contiguous United States have a trend 0.39°C/century. If the Hansen and Lebedeff analysis is repeated using stations only within United States borders, the trend is 0.40°C/century.] Karl and Jones (1989) estimated that the urban warming in the Hansen and Lebedeff data for the contiguous United States was close to 0.4°C/century; however, the assumed Hansen and Lebedeff temperature trend (provided by Hansen and Lebedeff) was incorrect, an error having been made in the integration over the contiguous United States. The correct comparison is that shown in Fig. 7a. However, with different area-averaging methods we have found differences between the MCDW and HCN trends as great as 0.26°C/century; we are preparing a comprehensive description of the tests we have carried out. Incomplete corrections for changes in time of daily observations (Karl et al. 1986) also affect the comparison of HCN and MCDW data. Overall, our results indicate that urban warming for the United States is probably not larger than our estimate of 0.1–0.2°C for the full globe, despite the fact that per capita energy use in the United States is high and United States cities have experienced large vertical growth, relative to cities in the remainder of the world. Oke (1981) and others have shown that, except perhaps for population and energy use, the most important factor in urban warming is the reduction of the skyview factor by vertical growth of the cities. Although indications are that urban effects do not qualitatively modify estimated global temperature trends, it is clearly important to carry out much more comprehensive analyses of the urban effects as needed to provide optimum estimates of unbiased temperature change.

Figure 7b shows the HCN data for the contiguous United States (48 states) for the period 1901–1987. The linear trend of HCN data is 0.26°C/century. [The change from the trend reported by Karl and Jones (1989) (0.16°C/84 years or 0.19°C/century) is due to the addition of data for 1985–87.] If ΔT were a linear function of time and the deviations from the straight line were normally distributed, it could be stated with 90% confidence that the slope of the temperature trend is positive, that is, that there is a warming trend. But this confidence interval should not be taken literally because of inherent errors in these assumptions.

Figure 7c shows the Climatic Division (CD) data for the contiguous United States for 1901–1987, recently reported by Hanson et al. (1989). The CD data are from 6,000 stations, including second-order and cooperating stations. The small cooling bias on the CD data relative to the HCN data, about 0.1°C, may be a result of incomplete correction for a changing mix of stations in the CD data (Cayan et al., 1986). In any case, the HCN
Figure 7. (a) Comparison of urban-adjusted Historical Climatology Network (HCN) temperatures for the contiguous United States and uncorrected Monthly Climatic Data of the World (MCDW) temperatures for the same region; both curves are 5-year running means. (b) Annual urban-adjusted HCN temperatures (Karl et al., 1988a,b) for the contiguous United States with linear trend. Data of Karl et al. (1988a,b) are updated with results for 1985, 1986 and 1987. (c) Annual Climatic Division (CD) temperatures (Hanson et al., 1989) for the contiguous United States with linear trend. (d) Temperature for the full United States (50 states). Contiguous United States data are from urban-adjusted HCN record. Alaska and Hawaii are based on MCDW stations associated with population centers of less than 5,000.

data, which have undergone elaborate station-by-station scrutiny, are the most reliable record of temperatures in the United States. Figure 7d shows temperatures for the entire United States (50 states) for the period 1901–1987. This is the area-weighted average of the HCN data for 48 states and MCDW data for Alaska and Hawaii. The Alaska and Hawaii data are based on MCDW records only for stations associated with population centers of less than 5,000. The linear trend for United States temperatures is $0.33^\circ$C/century. The mathematical confidence (as defined above) that the slope of the temperature trend is positive is 97%. (Stated differently, the
warming trend is significantly different from zero trend at the 5% significance level for a 1-tailed test of positive slope.)

The temperature trend in the United States is positive, consistent with long-term warming. As expected, since the United States covers only 1.5% of the globe, the variability is too great to permit United States temperatures to be used as proof of a long-term climate change. But, contrary to recent reports in the popular press, the United States temperatures do not provide a basis for questioning the reality of the global warming trend.

Finally, the HCN data are compared in Fig. 8 with the surface air temperatures obtained for the United States in the three transient climate simulations with the GISS model (Hansen et al., 1988). There is warming at the end of the 30-year period in both the model and observations, but it is small compared to the natural variability. The model results indicate a clear tendency toward warming beginning in the late 1980s, but it is too early to determine whether the observations bear this out.

We conclude that there is no basic inconsistency between the model results and the observations of United States temperatures. If the model predictions for the 1990s prove to be realistic, it implies a substantial climate change, to a warmth at least comparable to that of the 1930s. That mean level of warmth remains somewhat smaller than maximum interannual fluctuations, so not every season would be warmer than normal. But it would represent a sufficient "loading" of the climate "dice" to be clearly noticeable.

Figure 8. Surface air temperature for the contiguous United States as simulated by a global climate model for three trace-gas scenarios, (Hansen et al., 1988) compared with the urban-adjusted HCN (Karl et al., 1988a,b) temperature record.
The summer of 1988 was warm and dry in much of the United States and Asia. The global context of recent temperature anomalies is given by Figs. 10 and 11 of Hansen et al. (1989) which, respectively, show the temperature anomalies for the past four years and the four seasons of 1988. In the summer (June–July–August) of 1988, the mean temperature anomaly for the contiguous United States was about +1°C, but it was about +3°C in the hottest region near the United States–Canada border.

Is it possible that greenhouse warming played a significant role in the heat waves and droughts of 1988? Trenberth et al. (1988) state “Any greenhouse gas effects may have slightly exacerbated these overall conditions during the 1988 drought, but they almost certainly were not a fundamental cause.” Manabe (1988), although he has been the principal proponent of the likelihood of increased drought with doubled carbon dioxide, stated that, in view of the fact that global warming of 0.5°C in the past century was so small compared to the 4°C warming in his doubled carbon dioxide experiment, the greenhouse effect had “only a minor role, at most, compared to natural variability.”

That rationale is reasonable. The natural variability of precipitation is even greater than the variability of temperature. On the other hand, the results illustrated in Fig. 3 suggest that the greenhouse mechanism may be beginning to compete effectively with natural variability at about the present time, that is, there begins to be a noticeable increase in the frequency of drought in the model. And the increasing greenhouse effect certainly does not have to cause changes which exceed the range of natural variability in order to have important consequences.

It should also be realized that some of the excursions of “natural variability” may be associated with global or large-scale climate forcings. The great droughts of the 1930s came at a time of global warmth. That degree of global warmth could have arisen from purely internal fluctuations of the climate system. Our GCM produces similar long-term variations in the absence of greenhouse forcing increases (Hansen et al., 1988). On the other hand, it is also possible that the warmth of the 1930s was related to global forcings, such as the net effect of the near absence of volcanic eruptions in the period 1910–1940, the growth of greenhouse gases in that period, and perhaps other factors such as change of solar irradiance. In any event, the Northern Hemisphere was very warm in the 1930s, and the present Northern Hemisphere temperature seems to be approaching a similar degree of warmth. This is another reason to be concerned about a possible relationship between large-scale warming and drought.

We have examined the summer climate changes in our transient scenario A for the period 1986–1995 in comparison with the 100-year control run. For all gridboxes which are more than 90% land between the equator and 55°N (the region where the drought index increases most, see Fig. 1), by the period 1986–1995 the mean summer warming is 0.7°C. For the same region and decade the temperature in the driest regions (the 10% of the gridboxes with most negative drought index) has increased about 1°C relative to the driest regions in the control run, to a level 3°C warmer than the control-run climatology. Thus the average calculated warming is similar in magnitude to the mean warming in the contiguous United States in the summer of 1988, and the calculated warming in dry regions is comparable to the observed warming in the 1988 drought region.

Although these model results provide no information on the pattern of drought in 1988, they suggest the possibility that the increasing greenhouse effect could already play a significant role in summer heat. We emphasize that the timing and geographical distribution of any specific drought is determined primarily by short-term meteorological fluctuations and antecedent land, ocean and atmospheric conditions. The occurrence or intensity of a single drought cannot be used to identify the role of the greenhouse effect...
in the drought. At the same time, determination of meteorological factors involved in any
drought cannot be used to disprove the role of the greenhouse effect; such a meteorological
analysis provides no information as to whether the greenhouse effect is increasing the
frequency and severity of droughts.

We are analyzing the results of our transient climate simulations in greater detail,
but there are severe limitations on the information that can be extracted from the model
with its present resolution and representation of physical processes. Thus our main effort
now is in constructing the next version of the climate model, which will include higher
resolution and improved representations of ground hydrology, moist convection and other
processes. This should make it possible to do a more thorough analysis of the greenhouse
role in summer heat waves and drought.

6. SUMMARY

Our climate simulations indicate that an increasing greenhouse effect causes an intensifi-
cation of the extremes of the hydrologic cycle: (1) greater frequency (or areal coverage)
and intensity of drought, and (2) more intense wet and stormy conditions. These general
conclusions are not per se dependent on the accuracy of the climate model for specific
regions, because the analysis avoids the need to predict exactly where the changes occur.

We find no evidence of regional climate "winners" with an increasing greenhouse
effect. Droughts increase in the model at essentially all low-latitude and middle-latitude
land areas, where almost all the world's population is located. Although annual precipi-
tation increases at most locations in response to the global warming, the added rainfall
occurs in intense (moist convective) events, not as gentle large-scale rainfall, implying the
likelihood of an increased frequency of flooding. Furthermore, our model results imply
that an increased greenhouse effect will lead to more intense thunderstorms and tropical
cyclones. Temperature increases may be considered to be beneficial in some regions, but
the predicted rates of temperature change are much greater than those to which the bio-
sphere has adapted in the past. And a principal anticipated impact of higher temperature
is a rising sea level, as a result of thermal expansion of the oceans and melting of land
ice. None of these major impacts of a rapidly increasing greenhouse effect appears to
produce a substantial number of "winners."

The rapid increase of drought intensity in our climate model, which begins at about
the present time, is of particular concern. Is it possible that regional droughts could
become substantially more frequent, in effect a near-term, severe local manifestation of
the greenhouse effect analogous to the Antarctic "hole" of ozone depletion? Recent events
provide little guidance. A single drought cannot be used either to prove or disprove a
role of greenhouse warming. Concurrent and antecedent meteorological factors, such
as jet stream, soil moisture and ocean temperature distributions provide "causes" of
every specific drought pattern and timing, but they are irrelevant to the issue of whether
greenhouse warming increases the frequency and coverage of drought.

Empirical verification of the major effects of greenhouse warming may require ob-
servations over 10 or 20 years. Analyses would be aided if global models were improved
so as to provide more specific predictions of regional climate effects. As far as droughts
are concerned, we can only state at this time that our model suggests that greenhouse
warming may have its biggest impacts in certain regions in the subtropics and middle
latitudes, such as, in the Northern Hemisphere: United States/Mexico/southern Canada,
southern Europe/Mediterranean region, middle latitudes and lower latitudes of Asia, and
the African Sahel; and, in the Southern Hemisphere: Australia, the southern quarter of
Africa, and parts of Brazil and Argentina. But in analyzing observed drought trends it will
be important to account for other anthropogenic effects, such as destruction of vegetation
cover in semi-arid regions like the Sahel, which could be as important or more important in disrupting regional climate.

Current climate models are inadequate for detailed, reliable predictions of greenhouse climate impacts in any specific region. Improved investigation of regional climate impacts will depend upon: (1) studies based on models with higher spatial resolution, and, especially, more realistic representation of key aspects of the "physics," such as moist convection, clouds, ground hydrology, and vegetation effects; and (2) global observations which permit analysis of key feedback parameters, such as cloud cover, soil moisture levels, vegetation cover and atmospheric water vapor profiles. Of course, prediction of long-range climate change will also require improved knowledge of many global factors (Hansen et al., 1988), the principal ones being global climate sensitivity, the rate of heat uptake and transport by the ocean, and future trends of various climate forcings.

But there appears to be little chance that the uncertainties in climate simulations can qualitatively change our principal conclusion, that a growing greenhouse effect will increase the frequency and severity of the extremes of the hydrologic cycle: droughts, on the one hand, and extreme wetness and storms, on the other. If global climate sensitivity proves to be near the lower end of the range that is considered plausible, the magnitude of the simulated effects would be reduced and the time when the impacts clearly exceed natural variability probably would be delayed, but the impacts would not be reduced to negligible proportions. Similarly, although there are major uncertainties about ocean circulation and mixing which could modify regional climate distributions, they would not remove the mechanisms that cause increased hydrologic extremes. The potential for sudden changes in ocean circulation, which cannot be modeled presently, must be recognized, but any such lurches in ocean circulation would only increase regional climate dislocations.

Given our present knowledge of the climate system, and the uncertainties accompanying any climate predictions, we believe that it is appropriate to encourage those steps which would reduce the rate of growth of the greenhouse gases and which would make good policy independent of the climate change issue. Specific examples are: phase out chlorofluorocarbons (which have been implicated in the destruction of stratospheric ozone, and which represent 25% of current increases in greenhouse climate forcing), encourage energy efficiency (improving balance of payments and energy independence), and discourage deforestation (preserving natural resources for sustainable use and the habitat of invaluable biological species).

The opinions expressed in this paper are those of the authors and are not meant to represent policy of NASA or NOAA.

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ABSTRACT. Based on a network of 63 radiosonde stations distributed fairly evenly around the world, the global tropospheric temperature in 1988 was the maximum observed since the beginning of the record in 1958, 0.02°C warmer than 1983, 0.06°C warmer than 1987, and about 0.16°C warmer than 1980 and 1981. The global tropospheric temperature is indicated to have increased by a significant 0.2°C between 1958-72 and 1974-88, but with most of the warming in the Southern Hemisphere and the north temperate zone even cooling slightly. Between these two intervals there was cooling in all climatic zones in the tropopause layer, the cooling of 0.2-0.3°C being significant in both hemispheres. The low stratosphere cooled by more than 1.5°C following the 0.5°C warming occasioned by the El Chichón volcanic eruption, with most of the cooling in the Southern Hemisphere and, in particular, in the south polar zone (Antarctic “ozone hole” phenomenon). Emphasized is the strong influence of El Niño on global tropospheric temperatures about two seasons later, and because of the El Niño in 1987, the need for caution in relating the record warmth of 1988 to any greenhouse effect. Discussed is the extent to which these tropospheric and stratospheric temperature changes support the presumption that a greenhouse effect is already being observed.

1. INTRODUCTION

Tropospheric and stratospheric temperatures are especially relevant to the problem of “greenhouse” warming because present-day General Circulation Models (GCMs) can readily provide estimates of the temperature changes to be expected in these layers due to increases in CO₂ and other trace gases. Furthermore, the “fingerprint” technique for detecting anthropogenic climate change depends in part on the ability to discriminate between tropospheric warming and stratospheric cooling as a function of climatic zone and season (Epstein, 1982; Barnett and Schlesinger, 1987). Attempts to define observationally the differences in temperature change with height have been made by Parker (1985), Oehlert (1986), Barnett (1986), Karoly (1987) and Sellers and Liu (1988).
With the acquisition of temperature data for 1988, 31 years of data have become available for most of the stations in the 63-station radiosonde network (see Fig. 1) used by Angell and Korshover (1983) to estimate temperature variations in the troposphere and low stratosphere. The record was begun in 1958 because the International Geophysical Year (IGY) resulted in an increase in the quantity and quality of radiosonde data at that time, and because the change in radiosonde observation time in 1957 (from 0300 and 1500 UT to 0000 and 1200 UT) made it difficult to extend a consistent radiosonde record further back in time. This paper presents the variation in tropospheric and low-stratospheric temperatures obtained from the 63-station network, and compares this variation with surface and low-stratospheric temperature variations obtained by others.

2. PROCEDURES

At each of the 63 radiosonde sites, mean monthly geopotential heights were obtained for the 850, 300, 100, 50 and 30 mb mandatory pressure surfaces using both daily teletype data collected at the National Meteorological Center of the National Weather Service, NOAA, and data published in "Monthly Climatic Data for the World" by the National Environmental Satellite, Data, and Information Service of NOAA. According to the hydrostatic equation, the difference in height of two pressure surfaces (thickness) is proportional to the mean temperature of the layer between the pressure surfaces. For the purpose of this analysis, the mean monthly heights have been averaged by season (December–January–February, etc.), and mean seasonal temperatures for the troposphere, tropopause layer and low stratosphere have been determined from the mean seasonal thickness between 850 and 300 mb, 300 and 100 mb, and 100 and 50 mb surfaces, respectively. A mean seasonal surface temperature was also obtained at the 63 sites. At each radiosonde station, seasonal deviations from long-term seasonal means were then evaluated for those seasons in which all 3 months of data were available. (A month was accepted even if there was only one observation in that month, an unusual occurrence.) Every effort has been made to ensure that the times of radiosonde observation are consistent through the record. A drawback to the use of the thickness procedure, particularly for the 100–50 mb layer, is that if there are differing data amounts at the two bounding surfaces, the thickness of the monthly mean height may not equal the monthly mean of the daily thickness values (Trenberth, 1989). This is a minor problem for the 850–300 mb and 300–100 mb layers since almost all soundings attain a pressure of 100 mb. However, because of this problem, 100–30 mb thicknesses, though tabulated, have rarely been used, though they are used here for comparison purposes in Section 6.

The station deviations have been averaged (with equal weighting) to obtain seasonal temperature deviations for north and south polar (60–90°), north and south temperate (30–60°), north and south subtropic (10–30°), and equatorial (10°S–10°N) climatic zones. Table 1 presents the number of seasons of missing data for the stations within these zones, expressed as a percentage of the total number of station-seasons. For example, in the north polar zone there are 8 stations with 31 years of data each, yielding a total of 992 seasonal values. Of the possible 992 seasonal values, 7 were missing, resulting in a percentage-missing value of 0.7%. In the equatorial and south subtropic zones, more than 10% of the seasonal data at the surface and in 850–300 mb and 300–100 mb layers were missing, and in the south temperate and south polar zones more than 25% of the seasonal data in the 100–50 mb layer were missing. The average seasonal temperature deviations for climatic zones were obtained by omitting stations with missing seasonal data. A 1, 2, 2, 1 weighting of the seasonal temperature deviations for polar, temperate, subtropical and equatorial zones, respectively (roughly proportional to their areal extent), defines the seasonal temperature deviation for the hemisphere, and the average of the seasonal deviations for the two hemispheres defines the seasonal temperature deviations for the world.
Figure 1. Distribution of the 63 radiosonde stations on which the global temperature analysis is based. (Courtesy of K. Trenberth.)

Table 1. Percent of missing seasons for stations within climatic zones. The record extends from 1958 to 1988 at the earth's surface and in 850–300 mb and 300–100 mb layers, and from 1970 to 1988 in the 100–50 mb layer.

<table>
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<td>0.7</td>
<td>0.7</td>
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<tr>
<td>North temperate</td>
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<td>0.3</td>
<td>4.5</td>
</tr>
<tr>
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<td>0.5</td>
<td>3.5</td>
<td>8.3</td>
</tr>
<tr>
<td>Equator</td>
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<td>13.4</td>
<td>17.9</td>
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<tr>
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<td>12.4</td>
<td>12.4</td>
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</tr>
<tr>
<td>South polar</td>
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<td>5.2</td>
<td>5.2</td>
<td>10.4</td>
<td>25.1</td>
</tr>
</tbody>
</table>

For presentation of the temperature variations in some diagrams, the seasonal deviations have been smoothed by applying a 1-2-1 weighting twice to successive seasonal deviations (binomial weighting), with the first two and last two deviations twice averaged. All statistical evaluations, however, are based on unsmoothed seasonal or annual data.
3. TEMPERATURE VARIATION AS A FUNCTION OF HEIGHT

Figure 2 presents smoothed estimates of the variations in global temperature at the earth’s surface and in 850–300 mb (troposphere), 300–100 mb (tropopause), and 100–50 mb (low stratosphere) layers. The traces begin in the boreal winter (DJF) of 1958 and extend through the winter of 1989, except for the 100–50 mb trace which begins in the winter of 1970 owing to the absence of Soviet Union 50 mb data in “Monthly Climatic Data for the World” prior to that time. The vertical arrows denote the volcanic eruptions of Agung (Bali) in 1963 and El Chichón (Mexico) in 1982.

Points of interest in Fig. 2 include:

1) The similarity of the trends in surface and tropospheric data, that is, there is no evidence of an urban heat-island effect (Karl et al., 1988) in these data.
2) The evidence for cooling (until recently) in the tropopause layer, a layer which models do not show as cooling due to a greenhouse effect (Manabe and Wetherald, 1980).
3) The pronounced cooling in the low stratosphere following the warming occasioned by the eruption of El Chichón (Labitzke et al., 1983).
4) The lack of evidence for appreciable surface or tropospheric cooling following the El Chichón eruption (Angell, 1988a).
5) The “turning over” of the surface and tropospheric temperature traces following the record or near-record warmth observed in 1987 and early 1988.

Surface temperatures from this analysis will not be considered further in this paper, both because the surface temperature variations obtained by others should be more reliable and representative owing to the far greater number of stations used in their analysis (Folland et al., 1984; Jones, 1988; Jones et al., 1986a,b, 1988; Hansen and Lebedeff, 1987, 1988; Vinnikov et al., 1987) and because, given the sparse network of 63 stations used here, the layer-mean temperatures should be more representative than temperatures at a level or the surface.

4. COMPARISON OF TROPOSPHERIC AND SURFACE TEMPERATURE VARIATIONS

Figure 3 presents the variation in year-average global tropospheric temperature between 1958 and 1988, as estimated from the 63-station radiosonde network. The annual average is based not on the calendar year, but an average of the 4 seasons. For example, December of 1987 is included in the average for the year 1988. With this proviso, the warmest year of the 31-year record turns out to be 1988, 0.02°C warmer than 1983, 0.06°C warmer than 1987, and about 0.16°C warmer than 1980 and 1981. The 1988 value would have been closer to the 1987 value if calendar years had been used (December of 1988 considerably cooler than December of 1987, but December of 1986 and 1987 comparable).

Figure 4 compares the global tropospheric temperature variations shown in Fig. 3, with the annual variations in global surface temperature obtained by Hansen and Lebedeff (1988), Jones et al. (1988), and Vinnikov et al. (1987 plus update) between 1958 and 1987 and based on the calendar year. The agreement between tropospheric temperature variations and surface temperature variations is encouraging (see the relatively high correlations in Table 2) because the sampling network and the method of analysis are very different in the two instances. Some discrepancies in interannual variations are
Figure 2. Variation in global temperature at the earth's surface and in 850–300 mb (troposphere), 300–100 mb (tropopause) and 100–50 mb (low stratosphere) layers through the boreal winter (DJF) of 1989. A 1-2-1 weighting has been applied twice to successive seasonal deviations from long-term seasonal means, with the first two and last two deviations twice averaged. Abscissa tick marks are in the summer (JJA), and note the twofold difference in ordinate scale.

Apparent, however. For example, 1980 is indicated to be slightly warmer than 1981 in the troposphere whereas all three surface analyses have 1981 warmer than 1980. Also, 1965 is indicated to be cooler than 1964 in the troposphere, whereas all three surface analyses have 1964 cooler than 1965. In general, though, the year-to-year changes in surface and tropospheric temperature have been similar, increasing one's confidence in the representativeness of the indicated global temperature variations.
Figure 3. Variation in global year-average temperature in the tropospheric 850–300 mb layer, 1958–88. Because the annual average is determined from an average of the 4 seasons, the value for December of the previous year is included in a given yearly value.

Table 2. Based on the data of Fig. 4, the correlations among year-average global tropospheric (850–300 mb) temperature deviations, 1958–87, obtained by Angell and surface temperature deviations obtained by Hansen and Lebedeff (1988), Jones et al. (1988), and Vinnikov et al. (1987).

<table>
<thead>
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<th>Correlation</th>
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<td>Angell-Jones</td>
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<td>0.99</td>
</tr>
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<td>Jones-Vinnikov</td>
<td>0.88</td>
</tr>
</tbody>
</table>

Figure 4 also shows that the overall temperature increase during 1958–87 is somewhat smaller in the Jones et al. (1988) analysis than in the other analyses, perhaps in part because this is the only data set that explicitly includes marine temperatures. Thus, based on a comparison of the average temperature during the first half (1958–72) and last half (1973–87) of the record, the tropospheric temperature increased by $0.17 \pm 0.14^\circ$C between the two intervals, whereas based on the analyses of Hansen and Lebedeff (1988),
Figure 4. Comparison of the variation in global year-average temperature in the tropospheric 850–300 mb layer, 1958–87, with the variation in surface temperature obtained by Hansen and Lebedeff (1988), Jones et al. (1988), and Vinnikov et al. (1987).

Jones et al. (1988), and Vinnikov et al. (1987) the surface temperature increased by 0.16 ± 0.10°C, 0.09 ± 0.09°C, and 0.17 ± 0.10°C, respectively. Here the 95% confidence limits are obtained from twice the square root of the sum of the squares of the standard deviations of the mean for each interval (Brooks and Carruthers, 1953). Note the larger confidence limit for the tropospheric temperature (indicative of a larger standard deviation), the result partly of the sparseness of the radiosonde network used, but also partly the result of the greater influence of El Niño on tropospheric temperature than on surface temperature (see Section 5). Furthermore, it is apparent from Fig. 4 that the similarity in temperature change between 1958–72 and 1973–87 in tropospheric, Hansen et al. and Vinnikov et al.
analyses results from the larger temperature decrease in the early part of the tropospheric record being compensated by a larger temperature increase during the remainder of the record. Thus, not only are the interannual temperature variations indicated to be larger in the troposphere than at the surface, but so are the trends.

5. TROPOSPHERIC TEMPERATURE AND EL NIÑO

Figure 5 shows the relation during the last 31 years between sea surface temperature (SST) in the eastern equatorial Pacific (0°-10°S, 180°-80°W), obtained from the NOAA publication “Oceanographic Monthly Summary,” and 850–300 mb tropospheric temperature in the tropics (30°S–30°N) and world. Unlike the tropospheric air temperature, this SST is not indicated to have increased during the last three decades. Even so, based on unsmoothed seasonal deviations from the mean, the correlation between this SST and tropospheric temperature in the tropics is 0.45, 0.65, 0.67 and 0.53 at lags of 0, 1, 2 and 3 seasons, respectively. Thus, on the average, the tropospheric temperature in the tropics has lagged this SST by about 1 1/2 seasons (see also Newell and Weare, 1976; Navato et al., 1981; Pan and Oort, 1983; Angell, 1988c). Using Fisher’s Z test (Hoel, 1947) and taking into account the considerable serial correlation in the data (Quenouille, 1952), a zero-lag correlation of 0.32 would be significant at the 5% level, so that the correlation between this equatorial SST and tropical tropospheric temperature is impressive, with significance at about the 0.1% level. Note that the warmth of the tropical troposphere in 1982–83 is not as pronounced as might be expected from the very warm SST (El Niño) at this time, presumably because of the tropospheric cooling brought about by the eruption of El Chichón in the spring of 1982 (Angell, 1988a). Note also from Fig. 2 that there is a more obvious relation between El Niño and 850–300 mb temperature than between El Niño and surface temperature.

Because the tropics as defined here embrace half the earth's surface, one would expect a relation between global tropospheric temperature and tropical tropospheric temperature. Indeed, as shown in Fig. 5, the relation is a close one (zero-lag correlation of 0.90 based on the unsmoothed seasonal deviations), pointing-up the fact that the interannual variations in global tropospheric temperature have mainly reflected the interannual variations in tropical tropospheric temperature. The correlations between equatorial SST and global tropospheric temperature are 0.34, 0.54, 0.60 and 0.51 at lags of 0, 1, 2 and 3 seasons, respectively, where a zero-lag correlation of 0.33 would be significant at the 5% level. Accordingly, more than one-third of the variance in global tropospheric temperature can be explained by the SST in eastern equatorial Pacific two seasons earlier, that is, from knowledge of this SST one can estimate with some skill the global tropospheric temperature two seasons in advance.

On the basis of this relation, a relatively warm global tropospheric temperature would have been expected in early 1988 due to the El Niño occurring in 1987. The global troposphere would have been even warmer in 1988 if it were not for the sudden demise of this El Niño in the spring of 1988 and its replacement in the summer and autumn of 1988 and winter of 1989 by the most pronounced cool episode (La Niña) of the last 30 years (a lower SST in this region was observed only in 1892 and 1916 based on COADS data going back to 1865). As a consequence of La Niña, the global tropospheric temperature should be considerably lower in 1989 and 1990 than it was in 1988, as evidenced already by the "turning over" of the surface and tropospheric temperature traces in Figs. 2 and 5. From the point of view of the possibility of a greenhouse effect now, it will be interesting to see how much lower the global tropospheric temperature becomes in the next year or two.
6. COMPARISON OF LOW-STRATOSPHERIC TEMPERATURE VARIATIONS

While the tropospheric temperature variations estimated from the 63-station radiosonde network can be compared with surface temperature variations in order to judge their representativeness (see Fig. 4), there are few long-term data sets (there are several short-term satellite data sets) with which the 63-station, low-stratospheric temperature variations can be compared. One such comparison is with the temperature variations derived from Northern Hemisphere stratospheric analyses carried out on a routine basis by the Free University of Berlin (e.g., Labitzke et al., 1986). Figure 6 shows the comparison of the respective temperature variations in the 100–30 mb layer of the north temperate zone between 1970 and 1985, where the Labitzke temperature variations are based on
gridpoint values of temperature estimated from mean-monthly analyses and the Angell temperature variations are based on the 12 radiosonde stations in this zone (see Fig. 1). In general, there is fair agreement between the two traces except around 1982 when the Angell analysis indicates the low stratosphere to be warmer than does the Labitzke analysis. Part of this discrepancy is due to a temporary fault at one of the 12 radiosonde stations, resulting in too high a temperature for the north temperate zone. This illustrates the problem with a sparse network, in that if one station is in error, the error may go undetected and affect the zonal average. However, if many stations are used, the error will probably be detected and the station discarded, but with a smaller impact in any event because of the greater number of stations.

7. STRATOSPHERIC TEMPERATURE AND THE QBO

Just as the interannual temperature variations in the tropical troposphere are dominated by the interannual SST variations in eastern equatorial Pacific, so the interannual temperature variations in the equatorial stratosphere are dominated by the quasi-biennial oscillation or QBO (e.g., Naujokat, 1986; Angell, 1986). However, unlike the case in the troposphere, in the low stratosphere the variation does not show up in the global average (see Fig. 2) because the oscillations in polar and equatorial zones are basically out of phase. Thus, Fig. 7 shows that since 1970 there has been a quite precise out-of-phase relation between 100–50 mb temperature variations in equatorial and north polar zones. The out-of-phase relation is not indicated to be as good in the case of the south polar zone (Antarctica), but this may partly reflect data problems in this region. Even so, it is this tendency for a quasi-biennial variation in stratospheric temperature above Antarctica that is related to the tendency for a quasi-biennial variation in the depth of the "ozone hole" in this region (Komhyr et al., 1988; Angell, 1988b).
A striking feature of Fig. 7 is the abrupt increase in 100-50 mb south polar temperature at the end of the record (following a decade of general temperature decline), resulting in the warmest 100–50 mb temperature since 1970. In the austral spring of 1988 the south polar temperature in this layer was about 19°C higher than in the austral spring of 1987, a phenomenal change even though some warming was expected due to the QBO effect. Associated with this warming was an equally striking increase in total ozone between the two years, so that the “ozone hole” phenomenon (Farman et al., 1985; Stolarski et al., 1986) was not nearly so pronounced in 1988 as in 1987 (W. Komhyr, personal communication). The recent increase in 100–50 mb global temperature (Fig. 2) resulted mainly from this anomalous increase in south polar temperature (recall that the south polar zone enters into the global average with a weight of 1 in 12).

Unrepresentative trends may be obtained through the use of linear regression if extreme values occur at the ends of the record (Trenberth, 1989). Because of the warm tropospheric temperatures in 1988 associated with the 1987 El Niño, it was decided here to estimate temperature trends from the change in mean temperature between the first half and last half of the record. The 95% confidence limits for these changes have been estimated from twice the square root of the sum of the squares of the standard deviations of the mean for each half of the record (Brooks and Carruthers, 1953, p. 47), where the standard deviation of the mean is defined as the standard deviation divided by the square root of the number of cases.

Figure 8 presents, for climatic zones, hemispheres and the world, the temperature changes so obtained between 1958–72 and 1974–88 for 850–300 mb and 300–100 mb layers, as well as the 95% confidence limits for these changes (horizontal bars). Owing to the time intervals involved, the temperature changes so obtained can be considered as changes over 16 years. There has been a significant increase in 850–300 mb global temperature, but the 0.2°C increase has been due almost entirely to significant, and consistent, temperature increases of 0.3–0.4°C in Southern Hemisphere climatic zones, as well as the equatorial zone. The abrupt transition from no temperature change in north subtropics to an 0.4°C increase in the equatorial zone is unsettling since it is difficult to imagine that climatic-zone temperature trends could be partitioned in such a striking manner. Note that in the north temperate zone there is indicated to have been an 0.1°C temperature decrease between the two intervals, though the decrease is far from significant.

Unlike the case in the 850–300 mb tropospheric layer, where the warming between 1958–72 and 1974–88 is significant in the Southern Hemisphere but not the Northern Hemisphere, in the 300–100 mb layer bracketing the tropopause, the cooling has been significant in both hemispheres, as well as for the world as a whole. This is so despite the warming in this layer (see Fig. 2) associated with the 1987 El Niño. (Because the 300–100 mb layer is in the troposphere of the tropics, it warms following an El Niño in the same way as the 850–300 mb layer does.) While there has been a cooling of the 300–100 mb layer in all climatic zones, the cooling has not been uniform. In north subtropics, south temperate and south polar zones there has been a significant cooling of about 0.4°C between the two intervals (the cooling in the south polar zone associated with the Antarctic “ozone hole” phenomenon), whereas in north temperate and south subtropic zones there has been only an 0.1°C cooling. Note the extent of the 95% confidence limits in the north polar zone, reflecting the tremendous interannual temperature variability in the 300–100 mb layer of this zone.

Figure 9 shows the tropospheric temperature changes between 1958–72 and 1974–88 by season. Only in the south temperate zone do the seasonal changes differ appreciably (much greater warming in the austral autumn than in the spring). The relative seasonal changes in temperature have been similar in north polar and north temperate zones, with the greatest warming in summer and the greatest cooling in autumn. There is no evidence of the greater wintertime warming predicted by greenhouse models. With the exception of the south temperate zone, also in the Southern Hemisphere there has been a tendency for a greater warming in summer (DJF) than in winter. In the north subtropics there has been essentially no temperature change in any season, and in the equatorial zone the warming has been slightly greater in DJF and MAM than in JJA and SON. The length of the horizontal bars in Fig. 9 suggests that the detection of meaningful differences in seasonal temperature changes may require a relatively long record.

Figure 10 presents the temperature changes between 1970–78 and 1980–88 in the 100–
50 mb layer of the low stratosphere (recall that Soviet Union stratospheric data were
not available to us prior to 1970). Owing to the time intervals involved, the temperature
changes so obtained can be considered as decadal changes. Shown for comparison are the
temperature changes between these two intervals also for the 850–300 mb troposphere
layer and the 300–100 mb tropopause layer.

Over this time period there has been a significant cooling of the 100–50 mb layer of
the low stratosphere in both the Northern Hemisphere (see also Labitzke et al., 1986) and
the Southern Hemisphere, but with the decadal temperature decrease approximately twice
as great in the Southern Hemisphere (−1.0°C) as in the Northern Hemisphere (−0.4°C).
Note that, because of the tremendous interannual temperature variability in the south
polar zone (see Fig. 7), the 100-50 mb temperature decrease of −1.0°C in this zone is
not significant. The cooling is not significant in the north polar zone either, and in the
north temperate zone there is no evidence of cooling. One of the controversial aspects of
this analysis is the finding of appreciable cooling in the low stratosphere of the tropics.
Figure 9. Temperature changes between 1958–72 and 1974–88 in 850–300 mb layers of climatic zones by season, where DJF is December–January–February, etc. Within each climatic zone, from top to bottom the temperature changes are for DJF, MAM, JJA, and SON. Otherwise, see Fig. 8 caption.

Unlike the case with the intervals 1958–72 and 1974–88 (Fig. 8), between 1970–88 and 1980–88 the 850–300 mb temperature increased significantly, and nearly equally, in both hemispheres. The temperature increase is significant, or nearly significant, in all climatic zones except the north polar zone where there is no evidence at all of tropospheric
CHANGES IN TROPOSPHERIC AND STRATOSPHERIC TEMPERATURES

Figure 10. Temperature changes between 1970–78 and 1980–88 in 850–300 mb, 300–100 mb and 100–50 mb layers of climatic zones, hemispheres and the world. Otherwise, see Fig. 8 caption.

warming (again not supporting the greenhouse models). The variation of 300–100 mb temperature change with climatic zone has been the same as the variation of 100–50 mb temperature change with climatic zone, though with a greater temperature decrease in the 100–50 mb layer (except in the north temperate zone where there has been a slight warming of both layers). This suggests that the relative variation of the 300–100 mb temperature change with climatic zone in Fig. 8 may also reflect the relative variation of the 100–50 mb temperature change with climatic zone over the longer period.

10. CONCLUSION

Based on a 63-station radiosonde network, the global tropospheric temperature increased significantly (at the 5% level) between the first half and last half of the 1958–88 period. However, the variation of this warming with climatic zone and season is not the same as that obtained from models of the greenhouse effect, at least the non-transient models. For example, the observed tropospheric warming has been mostly in the Southern Hemisphere, with the warming for the north polar region small and not significant, and with slight cooling indicated for the north temperate zone. Other disagreements with the models include the observation that, in general, the tropospheric warming has been greater in summer than in winter in both hemispheres (though not significantly so), and the observation that between 1970–78 and 1980–88 there is no evidence at all of tropospheric warming in the north polar zone.
On the basis of this analysis, all climatic zones have cooled in the 300–100 mb layer bracketing the tropopause. Models of the greenhouse effect do not indicate a cooling this low in the atmosphere. The pronounced cooling of the global low-stratosphere following the warming occasioned by the El Chichón eruption is certainly intriguing, but this cooling is much too abrupt to be related directly to a greenhouse effect. Furthermore, this temperature decrease in the low stratosphere has been mostly in the Southern Hemisphere (in particular in the south polar zone is association with the Antarctic "ozone hole"), and in the north temperate zone there is even evidence for a slight warming of this layer during the last 18 years.

In summary, either the observations from this sparse network are not representative, the greenhouse influences on temperature are not yet out of the noise level, or the changes in temperature induced by the greenhouse effect are more complicated than suggested by the non-transient models. It must be emphasized that, because the temperature record ends in 1988, a year with a global tropospheric temperature strongly influenced by the 1987 El Niño, the record warmth of 1988 (as well as the near-record warmth of 1987) should be used with great care as a harbinger of the greenhouse effect. The second warmest year in this record, 1983, was also an El Niño year. It will be informative to see how much the global troposphere cools in response to the presently occurring La Niña. This may allow us to make more reasoned judgments concerning whether or not the greenhouse effect is already being observed.

REFERENCES


CHANGES IN TROPOSPHERIC AND STRATOSPHERIC TEMPERATURES

Representativeness of a 63-Station Network for Depicting Climate Changes

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ABSTRACT. Errors arising from the use of a sparse 63-station network in monitoring global and regional temperature changes are estimated by comparing results from a complete global data set from European Centre for Medium Range Forecasts analyses with those from the 63 gridpoints nearest the stations. For nine years of seasonal means, the correlations between the station network and the true values are generally quite high, but root-mean-square errors are also large outside of the tropics and are of the same order as the signal being sought. For the most part, this occurs because the variance of the 63-station network is too large so that although the overall low-frequency fluctuations are reasonably well depicted, their amplitude is too large. However, any missing data could greatly exacerbate errors arising from spatial sampling.

1. INTRODUCTION

With the intense interest in global temperature trends that has been sustained in the past decade or so, several analyses have been made of the available station data. The primary analyses of the free atmosphere have been performed by J. K. Angell, either alone or with J. Korshover, and there have been updates every year or two (Angell and Korshover, 1975; 1978a,b; 1983; 1984; and Angell, 1986; 1988). The temperature variations have been computed from a relatively sparse network of 63 radiosonde stations (Fig. 1) to estimate global and regional temperatures. The purpose of this paper is to estimate the errors arising solely from the spatial sampling using this network. Additional errors occur because of missing data.

To provide a test of Angell’s network and procedures, we have used nine years of global European Centre for Medium Range Weather Forecasts (ECMWF) 12-hourly analyses as a baseline, with regional and global means computed exactly. Then the regional and global means have been recomputed using the procedures of Angell and Korshover (1983), with each of the 63 stations represented by the nearest single gridpoint on a T42 grid (about 2.8° resolution).
CO-LOCATIONS OF RADIOSONDE STATIONS ON T42 GRID

Figure 1. Location of “stations” on a T42 grid used to represent the 63 Angell and Korshover (1983) stations.

2. PROCEDURES

The raw data used in this study were nine years (1979–1987) of monthly mean ECMWF analyses on a T42 grid with a resolution of approximately 2.8°×2.8° (Trenberth and Olson, 1988). Because data for December 1979 are missing, December 1978 data were used as replacement data to make a complete time series.

Time series the 63 “station” data were extracted from these analyses following the distribution of upper-air stations used by Angell and Korshover (1983) (henceforth AK) by picking the nearest grid point to each station (Fig. 1). The AK stations have changed slightly with time and Fig. 1 does not depict the latest array. Following AK, values chosen for the study were the 1000 mb temperature and 1000, 850, 300, and 100 mb height fields, which were used to form layer thicknesses and then were averaged by zone for each month. AK do not use a 1000–850 thickness temperature. Seven latitudinal zones were defined and the means of those zones are simply the means of all the stations in those zones without any area weighting: South Polar (SP): 90°-60°S; South Temperate (ST): 60°-30°; South Subtropics (SS): 30°-10°S; Equatorial (EQ): 10°-10°; North Subtropics (NS): 10°-30°N; North Temperate (NT): 30°-60°N; and North Polar (NP): 60°-90°N. In addition, following AK, tropical, hemispheric and global area averages were computed as Tropical (TR): 30°S-30°N = (NS + EQ + SS)/3; Southern Hemisphere (SH): 90°S-0° = (1/6 SP + 1/3 ST + 1/3 SS + 1/6 EQ); Northern Hemisphere (NH): 0°-90°N = (1/6 NP + 1/3 NT + 1/3 NS +1/6 EQ); and Global (GL): 90°S-90°N = (NH + SH)/2.

Clearly the AK methodology is not exact, either in forming zonal means or in computing regional averages from those zones, but the effect of this should be fairly systematic. Because all their analyses are based upon departures from the long-term means, most of the systematic component is effectively removed. Nevertheless, because
climate trends also have distinctive patterns associated with them, the possibility remains for biases to exist in the AK results.

The control data set against which to compare this time series of regionally averaged station data was derived from the T42 gridded analyses themselves. Exact regional averages were determined for the seven zones and other areas defined above using latitudinal weights determined from Gaussian quadrature.

The following procedures were then performed on each time series:

1. Compute thickness temperatures for the 1000–850, 850–300, 300–100, and 1000–100 mb layers using the hydrostatic relation, \( T = -\frac{g\Delta Z}{R \ln p} \), where \( g = 9.8 \text{ ms}^{-1} \), \( R = 287.04 \text{ J/kg K} \), \( \Delta Z \) is the layer geopotential height thickness in gpm, and \( T \) is in degrees K.

2. Determine the annual cycle of the thickness temperatures and the 1000 mb temperature by computing the grand monthly mean for all months.

3. Produce monthly and seasonal time series of the anomalies (deviations from the annual cycle and seasonal climatology, respectively).

4. Compute root-mean-square (rms) differences and correlations between the two unsmoothed monthly time series of anomalies for each region. Do the same for the seasonal anomalies.

5. When plotting graphs of the anomaly time series, a 1-2-1 smoother (1-1 on the ends) was applied twice to seasonal values, as in AK.

3. RESULTS

3.1. Annual cycle

To show the effects of the inexact computational procedures and bias in the location of stations within several zones, we first present several mean annual cycles of temperature. Figure 2 shows 850–300 mb thickness temperatures and Fig. 3 shows results for the 300–100 mb layer. These provide indications of those regions where biases in trends might arise because of inadequate or biased spatial sampling.

In Fig. 2 the AK values are biased low in the SP region, likely due to the more poleward location of the SP stations within the region. The tropical stations (not shown) reveal a weak warm bias, while the temperate zones have strong biases (by as much as 3°-4°C) of opposite sign: a warm bias enhanced in summer in the ST and a cold bias enhanced in winter in the NT region. These biases appear to arise because of the northward displacement in the centroid of the stations relative to the center of both temperate zones. Globally, the station results (not shown) reveal a 1°C cold bias consistently throughout the year in the lower troposphere. In the lower and middle troposphere of the SH, as a whole, there is virtually no bias between the station and gridded results. In the upper troposphere/lower stratosphere (300–100 mb) (Fig. 3), the AK data for the SP zone switch from a weak warm bias in summer to a >2°C cold bias in winter. In the ST region, the two time series peculiarly reveal quite different annual variations. The "true" results peak in January/February while the AK data peak in September/October. Also for the 300–100 mb layer, and for NT (not shown), the AK values generally exhibit a warm bias, especially in summer, but there is virtually no bias
in the global average at these levels. The results in Fig. 3 apparently arise from the bias in station location combined with the seasonal variation in tropopause height.
3.2. Anomalies

Graphs of the time series of the smoothed seasonal anomalies (Figs. 4-7) show the largest variations to be in the polar regions of the lower troposphere. Part of these variations are spurious (Trenberth and Olson, 1988), especially in the 1000 mb temperature (not shown), and arise from changes in analysis procedures. Nevertheless, this is still a valid test of whether the AK stations can depict the changes, whether spurious or real. For this reason we have preferred to present the 1000–850 thickness as representing the lower atmosphere. Correlations (Table 1) and rms differences (Table 2) measure the discrepancies in the time series due to spatial sampling from the AK station network for both monthly and seasonal series. Standard deviations of the two seasonal anomaly time series are shown in Table 3. With the exception of the 1000 mb temperature, the station and “true” time series
are virtually identical in the tropics. Outside the tropics, amplitudes of the AK data variations are generally larger than those of the "true" values. In particular, for NT the true standard deviation is 0.44, 0.23 and 0.26°C for the 1000–850, 850–300 and 300–100 mb layers versus 0.73, 0.42 and 0.36°C for the AK series, respectively. Thus in the northern temperate zone AK variations are 40 to 80% larger in amplitude.

Table 1. RMS differences between time series of regionally averaged, mean AK-station and gridded (T42) thickness temperature anomalies (non-smoothed time series) for: (a) monthly and (b) seasonal values. Units are °C.

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Table 2. Correlation between time series of regionally averaged, mean AK-station and gridded (T42) thickness temperature anomalies (non-smoothed time series) for: (a) monthly and (b) seasonal values.

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Correlations between the regional averages of the AK-station and gridded anomalies (Tables 1 and 2) are strongest in the lower troposphere and lower stratosphere, and are weaker in the mid-troposphere (850-300 mb) in all regions in general. Correlations in midlatitudes are somewhat less than elsewhere, with correlations in the ST zone being somewhat weaker, apparently because of the paucity of stations. However, comparing values for the entire hemispheres shows similar correlations for both in all layers except the 850-300 mb layer (where NH values are lower), despite the fact that fewer stations are used to determine the SH average (31 stations for the SH versus 41 for the NH). The errors in the NH evidently arise because many of the stations lie in the interiors of large continents. Globally, the correlation is no less than 0.9 for any layer, even for monthly values.
Angell (1988) shows that the variations in the station data from 1958 to 1987 range over about ±0.5°C in most zones. In the tropics the largest variations are associated with El Niño events and the latter are also evident in the hemispheric and global variations. Along with the standard deviations in Table 3, these give some idea of the size of the signal we are trying to detect.

The seasonal rms errors of 0.2 and 0.4°C (Table 2) are of the same order as the standard deviation (Table 3). Moreover, the rms differences arise from spatial sampling, which has a large systematic component in any given climatic regime, so that merely averaging over longer times is unlikely to reduce the magnitude very much. This is especially relevant whenever trends are being evaluated. If we consider that departures of plus or minus about two times the rms errors are necessary before they could be assumed to be real, then for the NT region, departures would need to exceed 0.4°C at 300–100 mb or 0.8°C at 1000–850 mb.

Therefore, although the correlations between the time series are quite high, quantitative climate changes are not very well measured outside of the tropics. For the NH,
Figure 6. As in Fig. 4, but for the 300–100 mb layer.

SH and global means (Fig. 7) the AK series reflect the true values to a large extent, but with excessive amplitude, especially for 1000-850 mb. Surprisingly, the rms errors are less for the global series than either hemispheric series, especially for the 1000–850 mb layer, indicating considerable cancellation in the hemispheric errors. It may be that this is a coincidence because it is not obvious why it should be the case in general.

4. CONCLUSIONS

The correlations with the true variations are quite high, but rms errors are also quite large and are of the same order as the typical signal we are interested in detecting. We therefore conclude that for the most part, the general low-frequency fluctuations in temperatures in the zones defined by AK are reasonably well picked up by the 63-station network, although the amplitude is typically too large outside of the tropics and trends are apt to be exaggerated.
The above conclusions assume perfect temporal sampling and apply only to the errors arising from spatial sampling alone. Normally, the effects of any missing data at individual stations are mitigated by redundancy in the spatial network. Kidson and Trenberth (1988) examined the effects of missing data on monthly mean general circulation statistics of all kinds. For middle to high latitudes and a 31-day month with twice-daily observations taken as the “truth,” subsampling with (a) regularly once per day, (b) 31 randomly distributed observations or (c) 16 randomly distributed observations, typical rms errors in monthly mean 300 mb geopotential height and lower tropospheric temperature are: (a) 8 gpm and 0.2°C, (b) 20 gpm and 1.0°C and (c) 40 gpm and 2.0°C, respectively, at the individual stations. Averaging the same random temporal sampling errors over, say, nine stations in a latitudinal zone would reduce these errors by a factor of 3. Nevertheless, it can be seen that with a sparse network of stations, requirements for complete temporal sampling become much more demanding.
Table 3. Standard deviation of time series of regionally averaged, seasonal-mean thickness temperature anomalies (non-smoothed time series) from: (a) the AK-stations and (b) the T42 grid. Units are °C.

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ACKNOWLEDGEMENTS

The National Center for Atmospheric Research is sponsored by the National Science Foundation.

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Soil Moisture Content with Global Warming

K. YA. VINNIKOV
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ABSTRACT. The potential greenhouse-gas-induced changes in soil moisture, particularly the desiccation of the Northern Hemisphere continents in summer, are discussed. To check the conclusions based on climate models we have used long-term measurements of contemporary soil moisture in the USSR and reconstructions of soil moisture for the last two epochs that were warmer than the present, namely, the Holocene optimum, 5000–6000 years ago, and the last interglacial, about 125,000 years ago. The analysis shows that there is a considerable disagreement between the model results and the empirical data.

1. INTRODUCTION

In many regions of the Earth precipitation is the major factor limiting crop productivity and influencing other aspects of man's activities. Despite the importance of precipitation, it has been established that modern climatic models have not obtained statistically significant regional precipitation changes from an increased carbon dioxide content in the atmosphere. Furthermore, changes in the moisture content of the active soil layer (usually of 1 m depth) have not been reliably revealed against the background of noise which disguises the climatic changes.

Soil moisture is a comparatively new characteristic of climatic conditions whose properties remain insufficiently known. One of the reasons of growing interest in this quantity is the conclusion that summer desiccation of the continents can take place with global climate warming (Manabe et al., 1981; Manabe and Wetherald, 1987). This conclusion regarding the conditions of developing anthropogenic global warming made it possible to assume a forthcoming climatic catastrophe. Prior to estimating the reliability of this conclusion, let us consider the basic methods for determining soil moisture content and information about its current state.
2. METHODS OF DETERMINING SOIL MOISTURE

The moisture in the active-soil layer has been measured in the USSR since the 1930s at a vast network of agrometeorological stations. At present there are about 3000 stations conducting observations in agricultural fields. Observational data are summarized in reference books wherein they are presented as mean characteristics of soil moisture for areas occupied by different crops averaged over different administrative and economic regions of the country. Analysis of measurement data on soil moisture in agricultural fields has been done in a number of monographs (Verigo and Razumova, 1973; Meshcherskaya et al., 1982; Kelchevskaya, 1983). Unfortunately, soil moisture in fields depends on the type of crops, as well as on agricultural technique; therefore, these measurements are not always representative for the areas adjoining these fields.

Soil moisture measurements for the areas with natural plant cover (usually meadows) began in the USSR in 1967 at hydrometeorological stations. The measurements are conducted each decade by the thermostat-weight method. The results are expressed in the amount of moisture available. The amount of moisture inaccessible by plants is assumed to be constant for each site and is determined experimentally. This inaccessible moisture is subtracted from the measurement data.

As an illustration, Fig. 1 presents a schematic of the relative moisture content of the upper 1-m soil layer for summer and winter based on multiyear measurement data from 50 most-representative stations (Vinnikov and Yeserkepova, 1989). By relative moisture is meant the ratio of soil moisture content to its field capacity when both values are expressed in terms of the amount of available moisture. In northern regions the relative moisture of the active soil layer is more than 100%, because, when the ground water table is close to the surface, the observed value of soil moisture exceeds the value of the field capacity and approaches the value of full capacity.

![Figure 1. Relative wetness of a 1-m soil layer (%). (a) winter, (b) summer. 1 indicates measurements are not available and 2 that the area is overmoistened.]
Soil Moisture Content with Global Warming

In other countries regular networks of soil moisture measurements are not, as a rule, established. Remote sensing methods, including those from satellites, are being developed now, but are not yet widely used. To date information based on them is scanty.

Another method for studying the moisture content of the active soil layer is calculating all the components of its water balance by using mean values of the major meteorological variables. The complex method of Budyko (1971) is most frequently used for this purpose, as well as in modern hydrological calculations (Zubenok, 1976; World Water Balance, 1974). The main advantage of this method is that it satisfies the basic conservation laws (heat and water balance equations), thereby ruling out the possibility of obtaining physically absurd results. The method contains two parameterizations, one for calculating evapotranspiration and the second for calculating runoff. Both parameterizations are well substantiated experimentally. The results obtained by using this method agree satisfactorily with the measured data of soil moisture and runoff.

The third method for obtaining information about soil moisture at present consists of applying atmospheric general circulation models. Until recently, the estimates obtained by this method have not been compared with measurement data; therefore, it was impossible to judge their validity.

3. COMPARISON OF MODEL-SIMULATED SOIL MOISTURE WITH OBSERVATIONS

Figure 3 shows a comparison between the mean soil moisture over four regions of the USSR (depicted schematically in Fig. 2) obtained by the GFDL (Manabe and Wetherald, 1987) and OSU models (Schlesinger and Zhao, 1989) with the measurement data. Comparisons of the values similarly averaged over the indicated regions show that the model calculations agree poorly with the measurement results (Vinnikov and Yeserkepova, 1989). It is particularly noteworthy that the model estimates of summer soil moisture content over the four regions turned out to be very underestimated, that is, close to zero, which is obviously absurd for these regions. There are three reasons for this poor agreement:

1. Insufficiently realistic model simulation of other characteristics of the modern climate;
2. Insufficient validity of the parameterizations used in models to calculate water balance components of the active soil layer; and
3. Limited representativeness and accuracy of measurement data.

It might be thought that in spite of all the shortcomings of the available empirical data, they give us the only basis for checking the validity of the model results.

Analysis of the empirical data shows that in the regions where the ground water table is not deep, the soil moisture and its annual variation depend to a large extent on the average position and variation in the ground water table. Subsequently, the description of these processes will probably be incorporated into climatic models, which will make them more realistic.

Figure 3 also shows, for each of the four regions, the empirical estimates of the linear soil moisture trend for the 1972–1985 period. This is the period that is most fully covered by data. It is characterized by a linear trend of the Northern Hemisphere mean air temperature of about 0.2°C/decade.

In the first three regions in the latitudinal belt of 50 to 60°N, the rate of mean annual soil moisture content increase was 1.5–2 cm/decade, the positive trend taking
place throughout all months of the year. In more southward regions the soil moisture trend is almost absent.

Thus, over the 1972-1985 period, in spite of the rapid increase in the Northern Hemisphere mean temperature, the soil moisture in these regions increased instead of decreased. At the same time, precipitation increased almost everywhere. Simultaneous changes in the elevation of the Caspian Sea also indicate the reliability of the trend estimates obtained.

4. MODEL SIMULATIONS OF CO$_2$-INDUCED SOIL MOISTURE CHANGES

Let us now consider what changes in soil moisture have been predicted by climate models with global warming. Some original studies and reviews present the results of simulating the potential changes in soil moisture content induced by doubling the atmospheric carbon dioxide concentration in five models: GFDL (Manabe and Wetherald, 1987), OSU (Schlesinger and Zhao, 1989), UKMO (Wilson and Mitchell, 1987), GISS (Hansen et al., 1984) and NCAR (Washington and Meehl, 1984). There is good agreement among the results obtained by these models concerning soil moisture changes over winter and during the year on the average. However, in the summer season the effect of continental desiccation is obtained by the first three models, while the other two do not show it. In comparing the model results to each other, it is impossible to establish where the truth is (Schlesinger and Mitchell, 1987). In this situation, as always, the only method for checking the correctness of theoretical inferences is verifying them against empirical data.
The available data on direct soil moisture measurements for the USSR territory have been demonstrated above and, on their basis, the ability of two models to simulate the modern soil moisture content have been examined. These data allow the preliminary conclusion to be made that increasing the mean annual soil moisture content with global warming is not accompanied by decreasing soil moisture in summer.

Subsequently, on the basis of available measurement data, a direct intercomparison of soil moisture calculation methods used in different models should be carried out.

5. COMPARISON WITH PALEOCLIMATES

Another source of empirical information for testing the ability of climatic models to estimate potential changes in soil moisture is the data on climatic reconstructions for the past warm epochs used as analogues of future climate (Anthropogenic Climate Changes, 1987). In Vinnikov and Lemeshko (1987) and Vinnikov et al. (1989) estimates of soil moisture have been obtained by Budkyo's complex method mentioned above for the Holocene climatic optimum (5–6 KA ago), with a mean global temperature 1°C above
The present one, and for the last (Riss-Würm or Mikulino) interglacial (125 KA ago), with the mean global temperature by about 2°C above the present. The results, presented in Fig. 4, show that for the Holocene climatic optimum soil moisture in summer was somewhat less than at present in a number of mid-latitude regions, with the southern boundary of the belt of decreased soil moisture in North America corresponding to about 40°N latitude and in Eurasia to about 50°N latitude. The most noticeable decrease in soil moisture is in northern Europe and eastern North America and makes up about 1–1.5 cm of water amount in a 1-m soil layer (10% of the present). Increasing soil moisture is most significant in the zones of insufficient moistening in Central Asia; however, estimates for these regions are preliminary.

For the last interglacial the estimates are obtained only over Eurasia north of 40°N, since for other regions paleoclimatic data are unavailable. Soil moisture content decreased in summer northward from 55–60°N. However, this decrease is small and does not cover the entire continent. Of most interest is the increasing moisture content in a 1-m layer of soil by 2–2.5 cm (up to 10–30% of the modern normal) in the forest, forest-steppe and steppe zones of Eurasia.

Thus, the past warm epochs considered are characterized by some decrease in soil moisture in wet regions and an increase in the regions with an arid climate.

Summer desiccation over all the continents did not take place in these warm epochs. The scale of changes for these epochs is comparable with the scale of those changes in
the soil moisture that are revealed in the present-day measurement data for the USSR territory over a 15-year period.

6. CONCLUSION

The available model estimates do not allow realistic inferences to be made about the forthcoming soil moisture changes. In this connection only the comparison of model results with empirical data can serve as a criterion of the validity of the model estimates. The empirical soil moisture data available in the USSR have not been used sufficiently yet for developing a parameterization of the hydrological processes in the active soil layer.

In analyzing paleoclimatic reconstructions of soil moisture for the Holocene optimum and the last Interglacial we do not detect changes in soil moisture that could be considerable and unfavorable for the biosphere in the extratropical part of the Northern Hemisphere continents.

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ABSTRACT. The climatology of wet and dry regimes in five different regions of the contiguous United States is examined using station records of daily precipitation for the period 1931–86. The regions were selected to be representative of different climatological synoptic regimes, and hence are associated with different large-scale atmospheric forcing patterns. The results show that the precipitation-producing processes operating during the wettest and driest years are substantially different, with wet years generally displaying large increases in the heavier precipitation categories compared to the dry years. These differences may be related to changes in the large-scale atmospheric circulation patterns associated with the development of persistent oceanic anomalies associated with variations in air-sea exchange processes over the Pacific Ocean and, depending on the particular region of the country, in the Atlantic Ocean as well.

It is shown that sequences of wet and dry years are rather common throughout the country. The 1930s and 50s saw prolonged periods of severe, and in some instances, extreme drought conditions in many parts of the country. There does not appear to be any secular trend in the frequency of either wet or dry regimes through the analysis period. Wet conditions prevailed in much of the country in the early 1980s; a return to drier conditions in the past couple of years, therefore, appears to be just a normal swing to an opposite regime.

1. INTRODUCTION

The severe heat and drought of 1988 in the United States and southern Canada has fueled public interest in global climate change. Public statements by some leading scientists that such “anomalous” weather could be a preview of things to come as a consequence of the “greenhouse effect,” has helped catapult the issue of global climate change into the forefront of the world’s environmental problems. Predictably, events since the summer
of 1988 have conspired to cast some public doubt regarding what many now consider to have been premature claims that the earth’s climate was beginning to show signs of “greenhouse fever.” This debate is not likely to be resolved anytime soon.

The purpose of this study is to compare the frequency distributions of daily precipitation in different parts of the country associated with the driest and wettest quintiles of total annual precipitation. The aim is to gain some insight into the changes in atmospheric circulation patterns associated with very different climatic regimes (very wet and very dry years).

The analysis will be restricted to the area of the contiguous United States, but the concept is applicable elsewhere, being similar to the development of climate scenarios that use particular sets of selected values for compositing. Since even the contiguous U.S. is a very large region containing a variety of climatic types, we will look at several different subregions with more homogeneous climates. The characteristic nature of the driest and wettest precipitation regimes in these areas is described. Emphasis is placed on the differences in synoptic patterns between these regimes, and on possible links with the annual cycle. Some suggestions are made regarding approaches to detection of greenhouse-gas-induced climate change.

The section that follows will detail the type and sources of data, and the methodology that was used in the study. Section 3 presents the analysis results; a discussion of the principal findings and their implications for detecting climatic change is given in Section 4, and a summary is provided in Section 5.

2. DATA AND METHODS

A data set of daily precipitation totals at more than 700 stations in the Lower 48 States was obtained from the National Climatic Data Center, NOAA in Asheville, NC. Figure 1 shows the locations of stations which had continuous records for at least 90% of the 56-year period 1931–86, which has been used as reference (the large majority of stations contain serially complete records). The data actually spanned a variable period of years, many going back before 1931. However, for the sake of maintaining adequate spatial coverage, we have restricted the analysis to the period 1931–86.

It is felt that the range of synoptic regimes covered by the 56-year analysis period is adequate for this study, as it contains a good mix of extremely wet and dry years. It also suffices to illustrate a tendency for dry years to cluster, as occurred during the 1930s and 1950s, with sequences of generally wet years developing in other decades (see Diaz, 1983). In many parts of the country the 1980s have been among the wettest on record.

Simple error-checks were performed, such as ensuring there were no negative values, and daily totals that exceeded four standard deviations from the 56-year mean were compared to nearby stations for corroboration. For example, if no nearby stations were reporting unusual daily totals, these values were flagged and a subjective determination was then made as to whether to retain or reject those data.

Two approaches were taken to minimize the effects of inhomogeneities in the data record and enhance the associations between precipitation and circulation patterns (see, Englehart and Douglas, 1985). First, the average daily precipitation was computed for each of 23 regional subdivisions (see Fig. 2) based on the individual station totals. Then, the frequency of occurrence of different precipitation threshold amounts was determined, and the contribution of these precipitation classes to the yearly total at each station was calculated. All of the analyses presented here are based on the daily regional totals. Furthermore, by also focusing our attention on differences between the driest and wettest years, corresponding to the upper and lower quintiles, it is felt that any errors or biases introduced by station and instrument moves, and other problems has been minimized.
Figure 1. Map showing the distribution of climate stations used in this study. Data for the period 1931–86.

The regional boundaries were determined using several criteria, and were originally drawn on a map of the state climate divisions; hence, the boundaries necessarily follow those contours. The objective discriminant measures included the presence of similar topographical and vegetative characteristics, and similar annual cycles of temperature and precipitation. These regions were used by Karl et al. (1988a) to produce a temperature summary for each of the 23 regions.

Klein and Bloom (1987) have shown that, on average, from 40% to 45% of the seasonal precipitation variance in the United States can be attributed to the occurrence of different circulation patterns of the 700 mb monthly height field. Spatially, the areas with the largest reduction of precipitation variance are found in the far West, and portions of the Southeast and Northeast U.S. Areas with a minimum of reduced variance are located in the continental interior, particularly over the northern Plains.

The characteristics of the seasonal changes in the frequency distribution of precipitation between wet and dry years for different climatic regions of the U.S. provide clues as to the changes in atmospheric circulation patterns that are typically associated with such anomalous periods. They may, therefore, be useful, in a climate analog sense, for identifying potential types of circulation patterns that may become more or less frequent.
if the climate begins to change in response to the increases in greenhouse gases. Another purpose was to ascertain what connections, if any, the occurrence of such extreme precipitation regimes have to the annual cycle.

As background to the analysis, the distribution of precipitation associated with the first four harmonics of the annual cycle of mean monthly precipitation for each of five regions examined in this study. The contribution to the total annual variance of monthly precipitation accounted for by the first 3 or 4 harmonics of the annual precipitation cycle varies spatially across the country, but exceeds 90% of the variance in most areas (Horn and Bryson, 1960; see also Hsu and Wallace, 1976). Presumably, changes in precipitation associated with greenhouse warming may manifest themselves most strongly, if they were either to amplify or dampen the characteristic precipitation patterns linked to the seasonal cycle.

To document the differences in precipitation regimes associated with the driest and wettest years of record, the daily precipitation data for each region (see Fig. 1) were divided into five categories. Although the threshold limits were chosen arbitrarily, they are generally indicative of different types of synoptic situations, regardless of season.

The first category measures the number of days each year (1931-86) with no measurable precipitation; the second category (precipitation from 0.01 to 0.1" inclusive) was considered to represent "light" precipitation events; the third (precipitation from > 0.1, to 0.5", inclusive) is taken to represent "moderate" events; the fourth (precipitation > 0.5 to 1", inclusive) is considered to denote "heavy" precipitation events; and the fifth (precipitation > 1") is considered characteristic of "extreme" events. It should be noted that daily station totals exceeding one inch are not that uncommon, but when one aggregates all the station data within each of the 23 regions illustrated in Fig. 2 there are relatively few daily events in most parts of the country which, on average, exceed this threshold.

Figure 2. Regional boundaries of climatic regions used in this study.
3. ANALYSIS

The precipitation time series for five of these subregions (shaded areas in Fig. 2) are presented in Figs. 3–7. The total length of each bar denotes the annual precipitation, expressed in terms of percent of the long-term mean (i.e., the value for that year divided by the 56-year mean). Also shown are the contributions to the annual total from each of the four precipitation categories, also expressed as a percentage of the long-term mean. In this manner the amounts are directly comparable, since they are all referenced to the same quantity (different for each subregion). The continuous line curve denotes the percent of days in each year with no measurable precipitation. It is clear from the record that all regions exhibit considerable interannual as well as multi-year variability.

3.1. Western United States

For the North Pacific Coast (NPC) and the Southern Rockies (SR) regions (Figs. 3 and 4, respectively), the classification limits were set at 85% of normal for the dry and 115% for the wet categories; for the other three regions they were set at 90% and 110%, respectively. The value of the threshold limit is not all that important, so long as it effectively selects a sufficient, but not too inclusive, number of extreme years.

The statistics for the various precipitation categories are given in Table 1. The NPC is a region of high annual precipitation, most of which comes during the winter half-year (Fig. 8). A high frequency of cyclonic storms originating in the North Pacific is prevalent in this area (Klein, 1957; Zishka and Smith, 1980; Diaz and Fulbright, 1981). The region is also affected by the changing phases of the Southern Oscillation through large-scale teleconnections with sea surface temperature and rainfall variability in the tropical Pacific (Emery and Hamilton, 1985; Yarnal and Diaz, 1986). However, as shown in the latter paper, more often than not, this region lies at a kind of nodal zone between regions of more consistent response to the north and south.

Annually, the number of days without precipitation averages to about one day in three. However, since there is a very strong annual precipitation cycle (the first harmonic of the annual cycle accounts for about 97% of the monthly variance about the annual mean), about 70–75% of the winter half-year period records some measurable precipitation, whereas only about a fourth of the summer days may be rainy.

During the wettest years, the number of days without precipitation is not that much different from the average, but it represents close to 20% fewer dry days compared to the corresponding proportion for the dry category. The contribution from the "light" category is very nearly the same in both sets of years, and we can conclude that changes in the frequency of these synoptic events will not have much of an impact on the annual mean.

The proportion of the annual precipitation contributed by the largest ("extreme") precipitation category is among the highest of any of the 23 subregions illustrated in Fig. 2 (around 12%). In fact, the "heavy" and "extreme" categories account collectively for 45% of the mean annual total (see Fig. 9). The differences in the contribution by the "moderate," "heavy" and "extreme" categories is, however, significant as shown in Table 1 and Fig. 9. The change is greatest in the two highest categories, where differences in the corresponding contributions to the mean annual total are nearly a factor of two for events in the 0.5–1" range, while the contribution during wet years by daily events exceeding one inch is greater by nearly a factor of five compared to the corresponding contribution during dry years.
Figure 3. Time series of annual precipitation for the North Pacific Coast region. Values are expressed as a percentage of the 56-year mean annual total. Also shown are the contribution to each year's total from four daily precipitation categories (see text). The continuous black line denotes the percentage of days with no measurable precipitation.

Figure 4. As in Fig. 3, except for the Southern Rockies region.
SOME CHARACTERISTICS OF WET AND DRY REGIMES

Eastern Prairies

- No Precipitation (25%)
- Mean Precipitation = 43.88"
- Number of Stations = 65

(13%) Light = 0 - 1"
(59%) Moderate = 1 - 5"
(24%) Heavy = 0.5 - 10"
(3%) Extreme = > 10"

Figure 5. As in Fig. 3, except for the Eastern Prairies region.

South Coastal Plains

- No Precipitation (25%)
- Mean Precipitation = 47.35"
- Number of Stations = 50

(12%) Light = 0 - 1"
(56%) Moderate = 1 - 5"
(25%) Heavy = 0.5 - 10"
(5%) Extreme = > 10"

Figure 6. As in Fig. 3, except for the South Coastal Plains region.
The region called the Southern Rockies approximately encompasses the area of the Colorado Plateau. It is affected both by the annual increase in cyclonic storms originating in the North Pacific during winter, and by the development of a summer "monsoon" in connection with strong summer heating, which results in a bimodal rainfall distribution. The summer maximum is the larger of the two maxima (Fig. 10). The area is affected in early summer by mesoscale disturbances from the region along the Pacific coast of Mexico, and in late summer by intrusions of Gulf of Mexico moisture and, occasionally, by eastern Pacific tropical storms. The first two harmonics of the annual cycle of mean monthly precipitation account for about 75% of the mean monthly variance about the 12-month mean (Table 2). The SR region is one of the few areas in the U.S. where the semiannual component is of larger amplitude than the first harmonic (40% versus 35%). The region can be classified as semi-arid, with a little less than 15 inches of annual precipitation. During the period 1931–86 there were no days when the average daily precipitation of the 22 stations in this region exceeded the one-inch threshold assigned to the "extreme" category, and on average only about 2% of the annual precipitation is contributed by the "heavy" category. In this, as in the other large regions in the continental interior, the low annual precipitation and the large regional areas help create minimum values for these last two categories. However, inspection of the precipitation statistics for individual stations in these areas demonstrates that the frequency of occurrence of large daily precipitation totals is generally quite low, and that the typical contributions of such large daily amounts to the annual total tend to be minor.
### Table 1. The wettest and driest years for the five different U.S. climate regions illustrated in Fig. 2, and the proportion (in percent) of average precipitation contributed by different daily amount categories (see text). Also shown is the percent of days during the year with no measurable precipitation.

**REGION: NORTH PACIFIC COAST**

Number of stations = 15  Mean annual precipitation = 53.66 in.

Wet years: R > 115%, Dry years: R < 85%

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<th>% of days with no precip.</th>
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<th></th>
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<td>Heavy (0.5–1&quot;)</td>
<td>Extreme (&lt;1&quot;)</td>
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<td>48</td>
<td>47</td>
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</tbody>
</table>

*Out of 56 values, 1 represents the highest value in that category, 56 represents the lowest.*
Table 1. Continued.

**REGION: SOUTHERN ROCKIES**  
Number of stations = 22 Mean annual precipitation = 14.46 in.  
Wet years: R > 115%, Dry years: R < 85%

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<th>% of days with no precip.</th>
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*averaged rank not computed since only 16 of 56 years recorded precipitation in this category.*
### Table 1. Continued.

**REGION: EASTERN PRAIRIES**

Number of stations = 65  Mean annual precipitation = 43.88 in.
Wet years: R > 110%, dry years: R < 90%

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<th>% of days with no precip.</th>
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<td>Long-term ave.</td>
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#### Wet Years

<table>
<thead>
<tr>
<th>Year</th>
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<th>% Heavy</th>
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#### Dry Years

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<td>13.4</td>
<td>51.2</td>
<td>16.3</td>
</tr>
<tr>
<td>median</td>
<td>29.3</td>
<td>13.4</td>
<td>51.0</td>
<td>15.9</td>
</tr>
<tr>
<td>ave. rank</td>
<td>11</td>
<td>30</td>
<td>46</td>
<td>47</td>
</tr>
</tbody>
</table>
Table 1. Continued.

**REGION: SOUTH COASTAL PLAINS**
Number of stations = 50 Mean annual precipitation = 47.35 in.
Wet years: R > 110%, Dry years: R < 90%

<table>
<thead>
<tr>
<th>% of days with no precip.</th>
<th>% of mean precipitation from</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Light (0.01–0.1&quot;&quot;)</td>
</tr>
<tr>
<td>Long-term ave.</td>
<td>25.7</td>
</tr>
</tbody>
</table>

**Wet Years**

<table>
<thead>
<tr>
<th>Year</th>
<th>No. of Days</th>
<th>Light</th>
<th>Moderate</th>
<th>Heavy</th>
<th>Extreme</th>
</tr>
</thead>
<tbody>
<tr>
<td>1932</td>
<td>19.7</td>
<td>12.3</td>
<td>70.7</td>
<td>21.4</td>
<td>7.1</td>
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<td>1940</td>
<td>27.4</td>
<td>11.0</td>
<td>54.7</td>
<td>43.5</td>
<td>5.9</td>
</tr>
<tr>
<td>1944</td>
<td>23.6</td>
<td>13.5</td>
<td>59.4</td>
<td>35.8</td>
<td>4.8</td>
</tr>
<tr>
<td>1945</td>
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<td>13.2</td>
<td>66.1</td>
<td>30.9</td>
<td>3.1</td>
</tr>
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<td>1946</td>
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<td>62.0</td>
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<td>4.7</td>
</tr>
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<td>1957</td>
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<td>53.2</td>
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</tr>
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<td>48.3</td>
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<td>72.7</td>
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<tr>
<td>1979</td>
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<td>9.6</td>
<td>62.4</td>
<td>37.4</td>
<td>16.3</td>
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<tr>
<td>1982</td>
<td>23.6</td>
<td>12.7</td>
<td>63.8</td>
<td>26.8</td>
<td>8.6</td>
</tr>
<tr>
<td>1983</td>
<td>24.4</td>
<td>13.7</td>
<td>54.3</td>
<td>32.5</td>
<td>15.4</td>
</tr>
</tbody>
</table>

| ave. | 23.6 | 12.1 | 62.7 | 34.9 | 8.3 |
| median | 23.7 | 12.3 | 62.4 | 34.8 | 7.1 |
| ave. rank | 35  | 37  | 16  | 11  | 19 |

**Dry Years**

<table>
<thead>
<tr>
<th>Year</th>
<th>No. of Days</th>
<th>Light</th>
<th>Moderate</th>
<th>Heavy</th>
<th>Extreme</th>
</tr>
</thead>
<tbody>
<tr>
<td>1931</td>
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<td>12.5</td>
<td>41.1</td>
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<tr>
<td>1955</td>
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<td>1956</td>
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<td>1962</td>
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<td>55.6</td>
<td>19.4</td>
<td>2.2</td>
</tr>
</tbody>
</table>

| ave. | 29.4 | 13.6 | 48.2 | 19.1 | 2.7 |
| median | 28.5 | 14.0 | 48.9 | 19.4 | 2.2 |
| ave. rank | 14  | 23  | 45  | 43  | 39 |
### SOME CHARACTERISTICS OF WET AND DRY REGIMES

**Table 1. Continued.**

**REGION: COASTAL SOUTHEAST**  
Number of stations = 17 Mean annual precipitation = 50.21 in.  
Wet years: R > 110%, Dry years: R < 90%

<table>
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<th>% of days with no precip.</th>
<th>% of mean precipitation from</th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Light (0.01–0.1&quot;)</td>
<td>Moderate (0.1–0.5&quot;)</td>
<td>Heavy (0.5–1&quot;)</td>
<td>Extreme (&gt;1&quot;)</td>
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<td>Long-term mean</td>
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<td>53.7</td>
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<td>8.8</td>
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<td>9.9</td>
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<td>51.0</td>
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<td>9.0</td>
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<td>11.0</td>
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<td>14.3</td>
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<td>10.0</td>
<td>58.2</td>
<td>34.3</td>
<td>16.4</td>
</tr>
<tr>
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<td>10.2</td>
<td>57.2</td>
<td>33.9</td>
<td>14.3</td>
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<td>Dry Years</td>
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<td>ave.</td>
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</tr>
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<td>ave. rank</td>
<td>15</td>
<td>30</td>
<td>48</td>
<td>43</td>
<td>37</td>
</tr>
</tbody>
</table>
Figure 8. The annual cycle of mean monthly precipitation for the North Pacific Coast region. Also shown is the sum of the contributions to the annual cycle by its first four harmonics.

Table 2. The variance of mean monthly precipitation about the 12-month mean for the five regions shown in Fig. 2, and the contributions to variance from the first four harmonics of the annual cycle.

<table>
<thead>
<tr>
<th>Region</th>
<th>NPC</th>
<th>SR</th>
<th>EP</th>
<th>SCP</th>
<th>CS</th>
</tr>
</thead>
<tbody>
<tr>
<td>variance (in²)</td>
<td>10.2</td>
<td>0.29</td>
<td>0.25</td>
<td>0.52</td>
<td>2.25</td>
</tr>
<tr>
<td>% of total variance</td>
<td>99.8</td>
<td>92</td>
<td>90</td>
<td>88</td>
<td>99.6</td>
</tr>
<tr>
<td>1st harmonic</td>
<td>97</td>
<td>35</td>
<td>66</td>
<td>73</td>
<td>67</td>
</tr>
<tr>
<td>2nd harmonic</td>
<td>2.6</td>
<td>40</td>
<td>9</td>
<td>5</td>
<td>29</td>
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<tr>
<td>3rd harmonic</td>
<td>0.5</td>
<td>16</td>
<td>11</td>
<td>9</td>
<td>1</td>
</tr>
</tbody>
</table>

From inspection of Table 1 and Fig. 11 it is seen that in the SR region, dry years experience about one-third more dry days than during wet years. This constitutes the biggest such change of any of the five regions examined here. In dry years, heavy precipitation events contribute on average 1% or less of the total (referenced against the long-term mean) and under a fourth of the corresponding contribution from this category during wet years. Significant differences are also evident in the “moderate” category, where in
Figure 9. Histogram distribution of precipitation by amount category for the North Pacific Coast region. Values, expressed in percent of the long-term mean, are shown for the set of dry and wet years (see text), and for “normal.” Values at upper right give the long-term annual mean, standard deviation and coefficient of variation.

Wet years the contribution by these events to the annual total is nearly twice as large as during dry years. Light precipitation events also contribute an additional 20% of the mean total during the wet years compared to the dry years. This is also the only region of the five shown in Fig. 2 where this precipitation category constitutes an important difference between wet and dry years.

The remarkable string of consecutive wet years from 1973–86 (Table 1) led to, among other things, the rapid rise of the Great Salt Lake (Kay and Diaz, 1985), and to record high flows of the Colorado River. The year with the largest contribution to the annual total from daily average precipitation amounts in the 0.1 to 0.5" (moderate) category was 1986 (with 12% of the annual mean, see Fig. 4). Klein and Bloom (1987) have shown that for areas roughly west of the continental divide, wet and dry seasons (especially winter) are essentially characterized by opposite 700 mb height patterns. In wet years, below-normal heights prevail over the eastern Pacific, and vice versa for dry years. Figure 12 depicts the differences in 700 mb height between the winters of 1974–79 and 1980–84. It suggests that the increase in precipitation experienced in the U.S. West during these years was a result of greater troughing and associated cyclonic activity in the eastern Pacific. Some summers during these years were also quite wet, further enhancing the annual totals.
Figure 10. As in Fig. 5, except for the Southern Rockies region.

Figure 11. As in Fig. 9, except for the Southern Rockies region.
In many parts of the contiguous United States the 1980s have been much wetter than average. A plot of the area-weighted annual precipitation for the continental U.S. illustrates this point (Fig. 13). From this perspective, the drought of 1988 does not appear to be an extreme occurrence. For example, during the 1950s there was a 5-year period in which precipitation for the continental U.S. was consistently below the long-term mean, and the last year of that sequence was even drier than 1988.

Diaz (1983) analyzed regional sequences of dry and wet years in the United States for the period 1895-1981 using monthly values of the Palmer Drought Severity Index (see, also, Karl and Koscielny, 1982; Karl, 1986). It was found that in certain parts of the country the initiation and termination of drought sequences tend to occur, preferentially, at certain times of the year. Figure 14, taken from Diaz (1983, hereafter referred to as D83) illustrates this point. For instance, in the "Northwest" region which contains the NCP region (see Fig. 2), drought initiation occurs more frequently during October–January, the time of the normal seasonal increase in precipitation (Fig. 8). Very few droughts originate during the summer dry season. Although drought termination in the U.S. Northwest does not display a strong tendency to occur at any particular time of the year, Fig. 14 shows that both an early start or an extended wet season usually accompanies the termination of a drought period.
The SR region in this study falls partially within the West and Southwest regions used in D83. Drought initiation tends to occur preferentially from about December to May. Interestingly, although the major precipitation maximum in the SR region occurs in August (Fig. 10), very few droughts originate in the summer, suggesting that a dry winter season is more important in drought development. Drought termination, however, tends to occur more often in the August-September period, suggesting that a wet summer "monsoon" season is often critical for eliminating a pre-existing dry spell.

There are also interesting differences between the wettest and driest years in California, comprised of regions (2) (South Pacific Coast or SPC) and (4) (California Interior Valleys or CIV) shown by the hatched areas in Fig. 2. Annual precipitation is under 15 inches in SPC and under 25 inches in CIV, most of which falls during the winter half-year. Nevertheless, it is clear that what generally separates the wet from the dry years is the occurrence of very intense precipitation in the former and a dearth of such events in the latter (Figs. 15 and 16).

The Southern Steppes region (shown as hatched in Fig. 2) is also of interest, as it is influenced by synoptic conditions prevailing over both the Pacific and Gulf of Mexico/Atlantic regions. Figure 17 indicates that in wet years precipitation events in the "heavy" and "extreme" categories contribute an additional 25% of the annual mean to the yearly total compared to dry years. In addition, during wet years, the "moderate" category contributes more than twice as much precipitation as during dry years.
3.2. Eastern United States

The "Eastern Prairies" (EP) region comprises the area west of the Appalachians extending from about Chattanooga, TN to Akron, OH in the east, to Davenport, LA and Little Rock, AR in the west. Figure 5 displays the annual precipitation totals showing the contributions to each annual total from each of the four precipitation categories. As is the case elsewhere, wet and dry years both tend to occur in clusters. For example, 5 of the 11 driest years listed in Table 1 occur from 1934 to 1943, and three more occur sequentially from 1952-54. Five of the 12 wettest years occur from 1945-51, and 6 of them occur from 1973-83.

Inspection of Table 1 reveals that the driest years experienced, on average, about 8 more days with no precipitation compared to the corresponding average for the wettest years; this is about a third more dry days compared to the long-term mean. There is little difference between dry and wet years in the contribution to the annual totals from "light" precipitation events; however, substantial differences do arise in the respective contributions from "moderate", "heavy" and "extreme" categories (an additional 14%, 15% and 5% of the annual mean, respectively, during the wet years). This points to changes in the frequency of occurrence of circulation patterns operating to either enhance or suppress synoptic activity and precipitation processes in the region (see, Diaz, 1981; Diaz and Fulbright, 1981). The analysis by Klein and Bloom (1987) suggests a circulation pattern during wet years with anomalous southerly flow from the Gulf of Mexico up the
Mississippi River Valley, coupled with upper-level troughing over the Rocky Mountains to the west (see, also, Diaz, 1990).

The EP region is located in the "Central Region" used by D83. Figure 14 shows that drought onset is more frequent from May to November than during the other half of the year; while drought termination tends to occur more frequently in January–February (wet winters), April–June (wet springs), and August–September, the latter, possibly representing an extension of the spring-early summer rainfall maximum (Fig. 18). The critical periods for both onset and termination of drought occur in the region’s planting and growing season, which suggests increased agricultural sensitivity to climatic variability.

The South Coastal Plain (SCP, Fig. 6) is a region of high annual precipitation comprising the lowland belt extending from the southern end of the Appalachian chain to central Texas, but recessed away from the immediate Gulf coast area. It is contained mostly within the "South" region used in D83. Precipitation peaks occur in the spring and in the late fall-early winter period (Fig. 19), though the first harmonic of the annual cycle of mean monthly precipitation still accounts for three-quarters of the variance about the 12-month mean (Table 2).

The drought of record in this region occurred from 1951-56, when, with the exception of 1952, all the years recorded less than 90% of the mean annual precipitation (Fig. 6 and Table 1). Wet years occurred sequentially from 1945-47, 1973-75 and, recently, in 1982-83. The differences in the contributions to the annual total by the different precipitation classes amount to an additional 15% of the long-term mean for the "moderate"
category, about the same additional amount from the "heavy" category and about 5\% more from the "extreme" category. There is not much difference in the contribution from the "light" category and about 5--6 more dry days are recorded in dry versus wet years.

Klein and Bloom (1987) obtained a pattern of 700 mb heights for wet and dry seasons in this region similar to that for the EP region. Namias (1983) and Chang and Wallace (1987) showed that the development of positive height anomalies in the mid-troposphere over North America normally accompanies the establishment of dry conditions in the southcentral United States. Figure 14 shows that the months of June to September are strongly preferred for drought onset, while drought termination tends to occur more frequently from about September to December, when the seasonal changes in the atmospheric circulation over the continent help force a breakup of the anomalous summer ridge. It should be emphasized, however, that in any one year the occurrence of such drought-inducing circulation patterns is not uncommon, though years in which drought conditions prevail are those in which such patterns tend to exhibit substantial temporal persistence. Nevertheless, the existence in the climatic record of long-lived (multi-year) dry and wet regimes in most regions of the country, would appear to make the task of assigning cause-and-effect relationships to future drought- or wet-spell development to greenhouse gas factors a difficult task.

The last region studied is named the Coastal Southeast (CS) and like the SCP region it is also an area of high annual precipitation with a strong seasonal component (Fig. 20). The first harmonic of the annual cycle of mean monthly precipitation accounts for
Figure 17. As in Fig. 9, except for the Southern Steppes region.

Figure 18. As in Fig. 8, except for the Eastern Prairies region.
SOME CHARACTERISTICS OF WET AND DRY REGIMES

Figure 19. As in Fig. 8, except for the South Coastal Plains region.

Figure 20. As in Fig. 8, except for the Coastal Southeast region.
two-thirds of the monthly mean variance about the 12-month average; the semi-annual component and the third annual harmonic each contribute an additional 10% of the monthly mean variance (see Table 2). A secondary precipitation maximum in March accounts for this situation. Figure 7 illustrates the tendency, which is also found in the other regions, for wet and dry years to occur in close proximity to each other. Dry weather was prevalent in the early to mid-1950s, with the occurrence of three consecutive dry years from 1954–56, and from 1977–81, which resulted in severe water shortages in the region, particularly in Florida. Wet years were prevalent in the middle 1940s and around 1960. Consecutive wet years were recorded in 1947–48, 1959–60 and 1982–83.

Table 1 shows that, on average, about 5–6 more dry days per year are experienced during dry-versus-wet years. The contribution from “light” precipitation events is roughly equal in both sets of years, but an additional 10% of the mean annual total is provided during wet years from “moderate” events, 15% more is contributed by “heavy” events, and another 10% comes from the “extreme” category. At least for the Florida portion of this region, a substantial fraction of its June-to-October rainfall comes from organized tropical disturbances. Figure 14 shows that aside from the month of April (which may be a statistical fluke), no one calendar month is more likely than the others to have a drought initiated. However, drought termination is more likely to occur during the wet season of summer and early fall.

### 3.3. Seasonal Differences

The separate seasonal contribution to the annual precipitation totals during wet and dry years was calculated for all 23 regions to study seasonality effects more closely. The results are summarized in Table 3 which identifies those seasons in which the wet/dry differences exceed 10% of the annual mean total. As expected, those seasons associated with maxima in the annual cycle experience large differences between wet and dry years. An interesting finding is that in most regions the equinoctial seasons generally experience the largest relative differences in precipitation amounts between regimes. In particular, the autumn season, which in the contiguous United States is nearly as dry as the winter season, displays large and consistent wet-versus-dry differences across the country.

It should be noted that the wet period of the early 1980s in the continental U.S. was associated with very wet autumns. Kay and Diaz (1985) found that precipitation anomalies in the fall season preceding the seasonal rise of the Great Salt Lake was the single most important statistical predictor of lake-level changes. Wet autumn seasons were also a significant factor in the occurrence of relatively high levels of the Great Lakes during the 1980s.

Lastly, we examined the characteristics of interannual variability of annual precipitation for each region. We looked at both the absolute value of the year-to-year changes, as well as the amplitude of the annual cycle for each year over the 56-year record. No statistically significant trends with time were found in either component. By contrast, there is evidence of decreasing diurnal temperature range in the United States over the past few decades (Karl et al., 1988a,b).

### 4. DISCUSSION

We have seen that the climatology of wet and dry precipitation regimes in various regions of the contiguous United States are characterized by large differences in the contribution of different precipitation classes to the annual total. Wet years are associated with much higher totals from higher daily precipitation amounts which occur over relatively large areas, as well as with smaller annual totals of dry days. Except for the Southern Rockies
### Table 3. Regional differences between wet and dry years. An “X” denotes a difference greater than 10% of the mean annual total precipitation. The average seasonal contribution to the annual mean is shown in parentheses.

<table>
<thead>
<tr>
<th>Region</th>
<th>Winter</th>
<th>Spring</th>
<th>Summer</th>
<th>Autumn</th>
</tr>
</thead>
<tbody>
<tr>
<td>#1 NPC</td>
<td>X (47%)</td>
<td>(23%)</td>
<td>(5%)</td>
<td>X (26%)</td>
</tr>
<tr>
<td>#3 NC</td>
<td>(42%)</td>
<td>(23%)</td>
<td>(7%)</td>
<td>X (29%)</td>
</tr>
<tr>
<td>#2 SPC</td>
<td>X (54%)</td>
<td>X (25%)</td>
<td>(2%)</td>
<td>(19%)</td>
</tr>
<tr>
<td>#4 CIV</td>
<td>X (51%)</td>
<td>X (28%)</td>
<td>(2%)</td>
<td>X (18%)</td>
</tr>
<tr>
<td>#5 ENC</td>
<td>(36%)</td>
<td>(23%)</td>
<td>(12%)</td>
<td>X (27%)</td>
</tr>
<tr>
<td>#6 GB</td>
<td>X (31%)</td>
<td>X (27%)</td>
<td>(19%)</td>
<td>X (23%)</td>
</tr>
<tr>
<td>#7 SD</td>
<td>X (27%)</td>
<td>X (15%)</td>
<td>(33%)</td>
<td>X (27%)</td>
</tr>
<tr>
<td>#8 NR</td>
<td>(27%)</td>
<td>(27%)</td>
<td>(21%)</td>
<td>X (26%)</td>
</tr>
<tr>
<td>#9 SR</td>
<td>(22%)</td>
<td>X (22%)</td>
<td>X (31%)</td>
<td>X (24%)</td>
</tr>
<tr>
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<td>(9%)</td>
<td>X (33%)</td>
<td>X (40%)</td>
<td>X (17%)</td>
</tr>
<tr>
<td>#11 SS</td>
<td>(12%)</td>
<td>X (26%)</td>
<td>X (36%)</td>
<td>X (26%)</td>
</tr>
<tr>
<td>#12 NPL</td>
<td>(9%)</td>
<td>X (29%)</td>
<td>(42%)</td>
<td>X (21%)</td>
</tr>
<tr>
<td>#13 SPL</td>
<td>(14%)</td>
<td>X (30%)</td>
<td>X (33%)</td>
<td>X (24%)</td>
</tr>
<tr>
<td>#16 GL</td>
<td>(17%)</td>
<td>(25%)</td>
<td>(32%)</td>
<td>X (26%)</td>
</tr>
<tr>
<td>#18 NA</td>
<td>(20%)</td>
<td>(26%)</td>
<td>(30%)</td>
<td>X (24%)</td>
</tr>
<tr>
<td>#14 SCP</td>
<td>(25%)</td>
<td>(29%)</td>
<td>(24%)</td>
<td>(22%)</td>
</tr>
<tr>
<td>#15 GC</td>
<td>(22%)</td>
<td>X (24%)</td>
<td>(28%)</td>
<td>X (26%)</td>
</tr>
<tr>
<td>#17 ep</td>
<td>(22%)</td>
<td>(29%)</td>
<td>(27%)</td>
<td>(22%)</td>
</tr>
<tr>
<td>#19 SÅ</td>
<td>(23%)</td>
<td>(27%)</td>
<td>(29%)</td>
<td>(21%)</td>
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<tr>
<td>#20 NP</td>
<td>(22%)</td>
<td>(26%)</td>
<td>(27%)</td>
<td>(25%)</td>
</tr>
<tr>
<td>#21 SP</td>
<td>(24%)</td>
<td>(26%)</td>
<td>(29%)</td>
<td>X (21%)</td>
</tr>
<tr>
<td>#22 CN</td>
<td>(23%)</td>
<td>X (25%)</td>
<td>(27%)</td>
<td>(25%)</td>
</tr>
<tr>
<td>#23 CS</td>
<td>(19%)</td>
<td>(22%)</td>
<td>(39%)</td>
<td>X (20%)</td>
</tr>
</tbody>
</table>

region, the contribution from “light” precipitation amounts was not much different during wet years compared to dry years. The changes in the “no precipitation” category do not appear to be as significant as the differences in the contribution from the “moderate” to “extreme” categories.

Namias (1983) has argued persuasively that persistent dry regimes over the North American continental interior result from the development of anomalous circulation patterns that strongly resemble natural teleconnection patterns (Namias, 1981), but are reinforced locally by such factors as loss of soil moisture, intense surface heating and large-scale subsidence. These patterns may be linked, dynamically, to North Pacific circulation features, sea surface temperature and tropical rainfall patterns. Hence, it may be difficult to distinguish greenhouse-gas-induced climatic change from the occurrence of such circulation features, unless in the future the frequency of occurrence of such drought-producing anomalous ridges becomes much greater than in the past.

Namias has also noted that the causes of drought may differ depending upon area. The varying distributions of monthly precipitation between wet and dry periods from one
U.S. region to another indicate that both synoptic-scale and mesoscale interactions may be occurring. They are in turn modulated by the development of persistent, and possibly preferred, modes of variability of the larger-scale atmospheric circulation whose structure is seasonally dependent (see Diaz, 1986; Barnston and Livezey, 1987).

Regarding the problem of detection of climatic shifts and their attribution to the greenhouse effect, one possible approach would be one of showing that: (a) the climatic statistics over a suitably long period, say one to two decades (or more), are sufficiently different from previous observations to be suggestive of a different statistical population; and (b) the pattern of anomalous circulation present in the different climatic regime is consistent with numerical GCM projections with either doubled or greatly increased greenhouse gas concentrations (see, Hansen et al., 1989).

It is further suggested that model results for the control and increased greenhouse gas simulations be examined to see how the frequency distribution of daily rainfall events changes with time. The results of Hansen et al. (1989) suggest that the frequency of drought of all magnitudes increases over the world's land areas. The increases in drought frequency should become statistically highly significant against a null hypothesis of no change in the frequency during the 1990s for moderate to severe droughts, and by the 2000s for the most extreme categories. Time will tell.

Further studies are planned of the relationship between interannual variability in different parts of the U.S., the changes in atmospheric circulation associated with these anomalies, and their relationship to the natural variations in the climatological patterns associated with the annual cycle.

5. CONCLUSIONS

The historical observational record for the United States indicates that wet and dry regimes tend to cluster in time and that considerable interannual variability is also to be expected. A technique used in this study was to compare a set of the driest and wettest years of record in the period 1931–86 using daily precipitation data from different regions of the country that are sensitive to different climatological synoptic regimes. The daily precipitation amounts were divided into five categories (including no precipitation), and the individual class contributions to the annual totals for these years were calculated.

It is clear that temporally persistent, large-scale atmospheric circulation patterns are the dominant factor producing the observed extreme differences in annual precipitation in those regions; some phase locking to the annual cycle also appears to be involved.

There is likely to be increased emphasis on regional climate change projections resulting from enhanced atmospheric greenhouse gas concentrations. Recent modelling studies (e.g., Hansen et al., 1989) suggest that there may well be increased interannual variability in precipitation over middle latitude land areas, with greater frequency of extreme events (both dry and wet spells).

REFERENCES


Data on Present-Day Precipitation Changes in the Extratropical Part of the Northern Hemisphere

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ABSTRACT. 100-year time series of spatially averaged annual precipitation and precipitation for the warm period of the year (May-September) for 12 regions of the USSR, Europe and North America are analyzed. It is shown that for land within 30-70°N the precipitation trend was about 6%/100 year, the increase in precipitation amount being a maximum in the Eastern Hemisphere north of 55°N.

1. INTRODUCTION

One of the most significant climate characteristics determining and even limiting man's activity is the precipitation amount and its seasonal distribution over the continents. Therefore, precipitation became the first meteorological element to be systematically observed as far back as ancient times. By the beginning of the 20th century, precipitation was measured at more than 20,000 stations all over the world, and during recent years there have been more than 40,000 stations and 140,000 raingauges. Nevertheless, only a small amount of these data has been recorded to be computer readable and accessible for climatological investigations. It seems that more than 95% of this information has never been used to analyze climatic change.

The results of analyzing time series and mean monthly and annual precipitation patterns (Gandin et al., 1976; Mooley, 1973; World Water Balance,..., 1974; Shver, 1976; Kagan, 1979; Sevruk, 1986) can be summarized as follows:

(i) Observational time series from an individual station are not correlated in time, that is, after eliminating seasonal variations of the mean and variance they represent white noise.
(ii) The two-parameter gamma distribution with density \( p(x) = C_1 x^{\alpha-1} \exp(-\lambda x) \) satisfactorily approximates the distribution of monthly and annual precipitation practically everywhere.

(iii) The field of precipitation is small-scaled. Over the plains in the extratropical part of the Northern Hemisphere the correlation radius of monthly and annual precipitation is several hundred kilometers, the influence of microscale variability being great. Therefore the approximation formula for the spatial correlation function is most often

\[
R(p) = c_2 \exp(p/r),
\]

where \( 0.5 < c_2 < 1 \), \( 0.2 < r < 1 \), and \( p \) is distance in \( 10^3 \) km.

(iv) The precipitation measured by a network of raingauges needs correction to balance the water budget and to use in model calculations. As a rule, in these measurements the level of actual precipitation is lowered. The data from northern countries with a prolonged snow cover period, and from countries with a high standard altitude of gauge installation, require the greatest corrections. The USSR, where the height of gauge installation is 2 meters, is among such countries. In some countries (for example, in the USSR and Finland), the methodology of precipitation observations has been improved during the last 100 years. As a result, the homogeneity of the time series of observed atmospheric precipitation sums has been disturbed.

Long ago in meteorology and hydrology the problem of averaging atmospheric precipitation over an area was considered (Thiessen, 1911). Considerable work has been carried out in India (Parthasarathy et al., 1987) and in the Soviet Union (Apasova and Gruza, 1982; Ledneva and Mescherskaya, 1977). However, the studies of Bradley et al. (1987) and Diaz et al. (1989) proved to be the most valuable in terms of the amount of data involved and the extent of the globe analyzed. The volume of information processed proved to be so large that these authors were able to suppress the white-noise component prevailing in raingauge data and, thereby, reveal the trends in precipitation. It being possible to relate these trends to model computations of the current anthropogenic climate change, the importance of the results obtained in these studies cannot be overestimated. In particular, in Bradley et al. (1987) a positive precipitation trend during the whole century was found for the 35-70°N zone. A considerable part of this zone is occupied by the USSR. Therefore, accounting for the difficulties of interpreting atmospheric precipitation data over this region, we decided to repeat the Bradley et al. investigation using our available data. Furthermore, the mean indices used in the two mentioned studies above are also difficult to interpret. The aim of the present work was to obtain time series of annual precipitation averaged over the area of the USSR and other regions of the extratropical part of the Northern Hemisphere, as well as to verify of the results obtained by Bradley et al. (1987). At the same time, it was decided to obtain corresponding precipitation series for the warm period of the year (May-September). The analysis of these data is of special interest both from the viewpoint of verifying the conclusions on annual precipitation sums (during the warm period, the precision of precipitation observations is considerably higher than during the period with solid precipitation), and for comparing the results obtained with model estimates indicating a summer drying of the continents during global warming (Manabe et al., 1981).
2. OBSERVATIONS AND ANALYSIS METHODOLOGY

Precipitation has been averaged over the 12 regions presented in Fig. 1. To avoid the effect of missing data, the averaging was made not for the precipitation values themselves, but for the anomalies from the normals estimated from the data for the period 1921–1960 (the period selected by Bradley et al., 1987).

Spatial averaging was carried out by the method of polygons (Thiessen, 1911; Kagan, 1979). According to this method, in year \( t \) with data available at \( n \) stations of a region and its \( \delta \)-environment, the weight of station \( i \), \( p_i(t) \), is proportional to \( S_i(t) \), that is, the area of locus of sites in the region, the distance from which to station \( i \) is the least as compared with other stations having observations in this year, with

\[
\sum_{i=1}^{n} p_i(t) = 1 .
\]

The disadvantage of this method is that the weight of a “favorably situated” station can be very large. At the same time, numerical experiments show that the error of averaging...
when using Thiessen's method only slightly exceeds the error of optimum averaging (Kagan, 1979). This disadvantage is compensated for by the large number of stations in the USSR and Europe involved in the analysis. Theoretical estimates of the precision of spatial precipitation averaging based on the results generalized by Gandin et al. (1976) and Kagan (1979) ensure a 5-10% accuracy of estimating the integral of measured annual precipitation over all regions selected in the Eastern Hemisphere, as well as over regions in the Western Hemisphere south of 55°N. Therefore, there are no real reasons to reject the usage of the precipitation data themselves in the spatial averaging in favor of indices that are difficult to interpret.

When using estimated changes in the spatial integral of precipitation measured at regional stations for practical purposes, it should be born in mind that the scale of these changes needs correction for the difference between measured and actual precipitation (as a rule, the former is smaller). For regions with large altitudinal variation, vertical precipitation gradients should also be taken into account. In the work "World Water Balance...." (1974) the fields of actual mean long-term precipitation values (normals) were estimated, all necessary corrections having been made for this purpose. As a first approximation, these corrections can be introduced as a factor equal to the ratio of the normals from the above work averaged over the region in question to the integral of the measured precipitation normals over the region estimated for the observational network used by us. Data for the analysis have been taken from "World Weather Records" and Neushkin (1975). The former archive provides 300 stations for Europe and North America with necessary usage conditions: the initiation of observations not later than 1951 and the end not earlier than 1984. Then these data have been supplemented up to 1987 from "Monthly Climatic Data for the World."

The precipitation archive for the USSR contains observational raingauge data from more than 600 stations over 1891-1988. Time series of precipitation have been reduced to the readings of Tretjakov's gauge and corrected to account for the changes in observational methods. (Since 1966, corrections for wetting of the raingauge bucket have been introduced into observational series.) We had serious grounds for checking the results of Bradley et al. (1987) for the USSR, since these two procedures necessary for restoring the homogeneity of the precipitation series were absent in their work. For the USSR territory, not only precipitation sums were averaged, but also annual precipitation indices - percentiles of the corresponding gamma distributions computed for each individual station, as introduced in Bradley et al. (1987). The analysis showed that differences in time variations of spatially averaged anomalies of annual precipitation sums and their indices are negligible even for an arid region such as Middle Asia (Vinnikov et al., 1990). Estimates of the $\eta$-parameter of the two-parameter gamma distribution for annual precipitation are usually greater than 10. With such magnitudes of $\eta$ it is not easy to distinguish between the shape of this distribution and the normal distribution; therefore, it is possible to average directly the annual precipitation over most of the USSR and interpret the result as integral precipitation over the area at the mean altitude of meteorological stations in the region. For precipitation amount in the period May-September, the value of $\eta$ turns out to be close to unity at some stations in the 35-70°N zone (for example, in Middle Asia). This fact does not prevent spatial averaging of the precipitation amount itself; however, the accuracy of such averaging proves to be considerably less.

3. RESULTS

Figure 2 shows the time series of precipitation averaged over 5 individual regions of the USSR, as well as the weighted mean for four western regions of the country and the entire territory under study. In the same figure the linear precipitation trends are presented. In 3
Figure 2. Variations in the mean-annual precipitation in the USSR territory: (1) Zone 60-70°N, 30-60°E (North European USSR); (2) Zone 45-60°N, 25-60°E (European USSR south of 60°N); (3) Zone 50-70°N, 60-90°E (Western Siberia); (4) Zone 37-50°N, 52-85°E (Central Asia); (5) Zone 50-70°N, 90-140°E (Eastern Siberia); (6) An average over the first four zones (USSR territory west of Enisei); (7) An average over all five zones (USSR territory).
out of 5 regions, and for the country as a whole, these trends are statistically significant (Table 1). Attention is also drawn to the last decade. Unfortunately, despite the general increase in annual precipitation amount over the country, the geographical and seasonal distributions of the changes were not favorable for agriculture in the USSR. For example, in 1981 a severe drought took place in the grain-growing regions of the USSR, annual precipitation amount being large.

The analysis of the time series of annual precipitation indices for the USSR territory showed that their time variations correspond to the graphs shown in Fig. 2 (Vinnikov et al., 1990).

Estimates of the variation of annual precipitation for non-USSR regions, as well as means for the zone 35-70°N and other parts of the world, are depicted in Figs. 3 and 4. Table 1 presents statistical characteristics of the time series of spatially averaged annual precipitation for the extratropical part of the Northern Hemisphere. σ is the standard deviation and β the estimated parameter of linear trend for 1891–1987 using the model $y(t) = \alpha + \beta t + \varepsilon(t)$.

<table>
<thead>
<tr>
<th>Region</th>
<th>Normal (mm)</th>
<th>$\sigma_y$ (mm)</th>
<th>$\beta$ (mm/100 yr)</th>
<th>$\hat{\beta}$ (%/100 yr)</th>
<th>$\frac{\sigma^2}{\beta}$ (%)</th>
<th>Ratio $\beta/\sigma^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Scandinavia</td>
<td>790</td>
<td>72</td>
<td>99</td>
<td>13*</td>
<td>87</td>
<td>4.1</td>
</tr>
<tr>
<td>Western Europe, south of 55°N</td>
<td>750</td>
<td>57</td>
<td>-41</td>
<td>-5*</td>
<td>98</td>
<td>2.0</td>
</tr>
<tr>
<td>N.W. USSR</td>
<td>510</td>
<td>47</td>
<td>67</td>
<td>13*</td>
<td>85</td>
<td>4.3</td>
</tr>
<tr>
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<td>500</td>
<td>49</td>
<td>17</td>
<td>4</td>
<td>100</td>
<td>&lt;1.0</td>
</tr>
<tr>
<td>Western Siberia</td>
<td>440</td>
<td>32</td>
<td>64</td>
<td>14*</td>
<td>69</td>
<td>6.7</td>
</tr>
<tr>
<td>Central Asia</td>
<td>220</td>
<td>36</td>
<td>10</td>
<td>4</td>
<td>100</td>
<td>&lt;1.0</td>
</tr>
<tr>
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<td>420</td>
<td>28</td>
<td>35</td>
<td>8*</td>
<td>88</td>
<td>3.7</td>
</tr>
<tr>
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<td>1170</td>
<td>87</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>America, north of 55°N</td>
<td>500</td>
<td>41</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
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<tr>
<td>America, 45-50°N</td>
<td>780</td>
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<td>49</td>
<td>6*</td>
<td>95</td>
<td>2.6</td>
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<td>670</td>
<td>57</td>
<td>57</td>
<td>9*</td>
<td>94</td>
<td>2.9</td>
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<td>US, 25-35°N, east of 100°W</td>
<td>1190</td>
<td>137</td>
<td>179</td>
<td>15*</td>
<td>88</td>
<td>3.8</td>
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<tr>
<td>USSR, west of Enisei</td>
<td>400</td>
<td>23</td>
<td>35</td>
<td>9*</td>
<td>83</td>
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<tr>
<td>USSR</td>
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<td>35</td>
<td>9*</td>
<td>74</td>
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<tr>
<td>Europe except USSR</td>
<td>760</td>
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<td>6</td>
<td>1</td>
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<td>45</td>
<td>53</td>
<td>7*</td>
<td>91</td>
<td>3.4</td>
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<tr>
<td>Land, 35-70°N without northern Canada and China</td>
<td>570</td>
<td>20</td>
<td>34</td>
<td>6*</td>
<td>78</td>
<td>5.3</td>
</tr>
<tr>
<td>Mean for 11 regions, north of 35°N</td>
<td>600</td>
<td>14</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

*Estimate of $\beta$ significantly differs from zero at the 95% level.
Figure 3. Variations in the mean-annual precipitation over the non-USSR territory of the extratropical part of the Northern Hemisphere: (1) Scandinavia; (2) Zone 37-55°N, 10°W-25°E (Western Europe, south of 55°N); (3) The Atlantic, north of 55°N; (4) North America, north of 55°N; (5) Zone 45-55°N, 50-130°W; (6) US, 35-45°N; (7) US, south of 35°N and east of 100°W.
Variations in the mean-annual precipitation for the extratropical part of the Northern Hemisphere: (1) USSR; (2) Europe, except for the USSR; (3) North America, zone 35-55°N; (4) Continental Northern Hemisphere, zone 35-70°N, without northern Canada and China; (5) mean for 11 regions, north of 35°N.

From these data a conclusion can be drawn that there is a general precipitation increase in the zone 35-70°N during the last 100 years. Thus, one of the main conclusions of Bradley et al. (1987) has been confirmed. A linear trend of $\beta = 34\,\text{mm}/100\,\text{yr}$ or $6\%/100\,\text{yr}$ (which is more correct than mm/100 yr estimate because of the above-mentioned scale corrections) for land regions within 35-70°N (Northern Canada excluded) describes about 20% of the variance of the time series of annual precipitation.

Figures 5-7 show the time series of precipitation for the same regions, but for the warm period (May-September). As for the annual precipitation amount, linear trends are given in cases when they are significant at the 95% level. Table 2 contains the estimates of the statistical characteristics of these series, analogous to those in Table 1. A conclusion can be drawn that regularities of annual precipitation variations are also confirmed by the data for the warm period of the year. Attention is focussed on the fact that the range of precipitation variations during the warm period over land areas within 35-70°N (without Northern Canada) of 4%/100 year practically coincides with the corresponding estimate.
for annual precipitation, thus confirming it and ensuring that the observed trends are not
due to improvements introduced by national meteorological services in solid precipitation
gages.

The linear precipitation trend estimated for the entire 100-year period for annual
precipitation was not statistically significant at the 95% level in only two out of ten
regions. It should be noted that all of these statistically insignificant regions are southern
regions in the 35-70°N zone. In one more southern region (Western Europe, south of
55°N) the precipitation trend proved to be negative. Therefore, a suspicion arose that in
the 35-55°N zone the secular precipitation variations are of a special character. Averaging
was made over that part of Eurasia in the 35-70°N zone including the areas are hatched
in Fig. 1.

The analysis carried out revealed that, as a whole for the zone 35-70°N of the East-
ern Hemisphere, no noticeable changes in precipitation amount occurred either from May
to September or during the entire year (the portion of precipitation variance described
by linear trends in negligibly small here). Therefore, increased air temperature in the
hemisphere and the resulting increase in evaporation should have promoted aridization of
this zone, thus confirming the model calculation by Manabe et al. (1981). In the 35-55°N
zone of the Western Hemisphere, the annual precipitation increased slightly during the
During the period under consideration, the mean-annual temperature $T$ in the Northern Hemisphere increased at a mean rate of $0.5^\circ C/100$ year (Budyko and Izrael, Table 2. Statistical characteristics of the time series of spatially averaged precipitation for the May-September period of the extratropical part of the Northern Hemisphere. $\sigma$ is the standard deviation and $\beta$ the linear trend parameter in the model $y(t) = \alpha + \beta t + \epsilon(t).$

<table>
<thead>
<tr>
<th>Region</th>
<th>Normal (mm)</th>
<th>$\sigma_y$ (mm)</th>
<th>$\beta$ (mm/100 yr)</th>
<th>$\beta$ (%/100 yr)</th>
<th>$\alpha_{\beta}$ (%)</th>
<th>Ratio $\beta/\alpha_{\beta}$</th>
</tr>
</thead>
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<tr>
<td>Scandinavia</td>
<td>340</td>
<td>40</td>
<td>40</td>
<td>12*</td>
<td>94</td>
<td>2.9</td>
</tr>
<tr>
<td>Western Europe, south of 55°N</td>
<td>290</td>
<td>28</td>
<td>-20</td>
<td>-7*</td>
<td>98</td>
<td>2.0</td>
</tr>
<tr>
<td>N.W. USSR</td>
<td>280</td>
<td>30</td>
<td>19</td>
<td>7</td>
<td>98</td>
<td>1.7</td>
</tr>
<tr>
<td>European USSR, south of 60°N</td>
<td>260</td>
<td>33</td>
<td>8</td>
<td>3</td>
<td>100</td>
<td>1.0</td>
</tr>
<tr>
<td>Western Siberia</td>
<td>270</td>
<td>23</td>
<td>14</td>
<td>5</td>
<td>98</td>
<td>1.7</td>
</tr>
<tr>
<td>Central Asia</td>
<td>90</td>
<td>19</td>
<td>-5</td>
<td>-6</td>
<td>100</td>
<td>&lt;1.0</td>
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<tr>
<td>Eastern Siberia</td>
<td>290</td>
<td>21</td>
<td>13</td>
<td>5</td>
<td>98</td>
<td>1.8</td>
</tr>
<tr>
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<td>410</td>
<td>35</td>
<td>-</td>
<td>-</td>
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<td>-</td>
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<tr>
<td>North America, north of 55°N</td>
<td>250</td>
<td>22</td>
<td>-</td>
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<td>-</td>
</tr>
<tr>
<td>North America, 45-55°N</td>
<td>350</td>
<td>39</td>
<td>28</td>
<td>8*</td>
<td>98</td>
<td>2.0</td>
</tr>
<tr>
<td>US, 35-45°N</td>
<td>300</td>
<td>32</td>
<td>19</td>
<td>6</td>
<td>99</td>
<td>1.6</td>
</tr>
<tr>
<td>US, 25-35°N east of 100°W</td>
<td>610</td>
<td>81</td>
<td>88</td>
<td>14*</td>
<td>93</td>
<td>3.1</td>
</tr>
<tr>
<td>USSR, west of Enisei</td>
<td>220</td>
<td>14</td>
<td>7</td>
<td>3</td>
<td>99</td>
<td>1.5</td>
</tr>
<tr>
<td>USSR</td>
<td>240</td>
<td>11</td>
<td>9</td>
<td>4*</td>
<td>96</td>
<td>2.3</td>
</tr>
<tr>
<td>Europe except USSR</td>
<td>300</td>
<td>20</td>
<td>0</td>
<td>0</td>
<td>100</td>
<td>&lt;1.0</td>
</tr>
<tr>
<td>America, 35-55°N</td>
<td>320</td>
<td>28</td>
<td>23</td>
<td>7*</td>
<td>96</td>
<td>2.4</td>
</tr>
<tr>
<td>Land 35-70°N without northern Canada and China</td>
<td>280</td>
<td>12</td>
<td>12</td>
<td>4*</td>
<td>93</td>
<td>2.8</td>
</tr>
<tr>
<td>Mean for 11 regions north of 35°N</td>
<td>280</td>
<td>10</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

*Estimate of $\beta$ significantly differs from zero at the 95% level.
Figure 6. As in Fig. 3, except for the period May-September.

1987; Budyko, 1988). If we suppose the existence of a causal relationship between the global warming process and the growth of precipitation for the extratropical part of the hemisphere, then for a considered part of the 35-70°N zone we arrive at the result: \( \frac{\delta P}{\delta T} \approx 12\% / 1^\circ C \), which can be compared to estimates made by climate models. From these models it follows that such a causal relationship does exist, but the estimate of \( \frac{\delta P}{\delta T} \) is about 2.5% / 1° C in the model of Manabe and Wetherald (1987) and about
4.2%/1°C in the model of Schlesinger and Zhao (1989). Such a considerable discrepancy needs further study. Among its causes can also be anthropogenic factors not related to global temperature change, such as an increase of atmospheric pollution with trace gases, especially in high latitudes of the Northern Hemisphere, construction of a large number of water reservoirs and increased regions of artificial irrigation. All these factors contribute to intensify the moisture cycle over continents and, as a result, could promote an increase in total precipitation there.

The contribution of a "lucky" choice of time period for estimating trends also cannot be excluded. The last ten years do look rather unusual – against the background of a sharp temperature increase, no warming was observed in the Arctic and the meridional temperature gradient did not decrease, but instead increased. This resulted in abnormally high precipitation amounts in the regions in question. Therefore, the nonstationary response of moisture conditions to the current global warming process, observed during the last century, can in principle differ considerably from the stationary response to doubling atmospheric CO$_2$ content that has been computed by the GFDL and Oregon State University models.

Table 3 contains estimated changes of annual precipitation for a 1°C global air temperature change obtained by: (1) the models for 5 land regions of the extratropical part of the Northern Hemisphere induced by a doubling of the atmospheric CO$_2$ concentration (Manabe and Wetherald, 1987; Schlesinger and Zhao, 1989); (2) paleoclimatic reconstructions of the Holocene climatic optimum (Budyko and Izrael, 1987); and (3) empirical data using linear trends, as well as the estimates based on data spanning the last decade from (Budyko and Groisman, 1989), and empirical estimates of climate sensitivity to current changes in atmospheric CO$_2$ content given by Vinnikov and Groisman (1982). Without discussing in detail the reliability of each method for obtaining the estimates, let us note their following common features:
(i) the coincidence of the signs of the estimates, with one exception, shows the common tendency of total precipitation growth in the extratropical part of the Northern Hemisphere with global warming;
(ii) empirical estimates of precipitation changes are considerably higher than those obtained by climate models;
(iii) discrepancy even between signs of estimates of future changes in one of the most important agricultural regions of the world within the US and southern Canada indicates the necessity of more detailed investigation of the processes causing changes in moistening of this region.

Table 3. Estimated variations of annual precipitation (%) with a 1°C global warming.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>USSR</td>
<td>3</td>
<td>4</td>
<td>12</td>
<td>4</td>
<td>15</td>
<td>17</td>
</tr>
<tr>
<td>Europe without the USSR</td>
<td>2</td>
<td>5</td>
<td>5</td>
<td>9</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>North America, north of 55°N</td>
<td>5</td>
<td>6</td>
<td>8</td>
<td>47</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>North America, 35-55°C</td>
<td>1</td>
<td>4</td>
<td>-5</td>
<td>19</td>
<td>12</td>
<td>15</td>
</tr>
<tr>
<td>Land, 35-70°N without northern Canada and China</td>
<td>2</td>
<td>4</td>
<td>5</td>
<td>11</td>
<td>10</td>
<td>12</td>
</tr>
</tbody>
</table>


REFERENCES


Schlesinger, M. E., and Z.-C. Zhao, 1989: Seasonal climatic changes by doubled CO₂ as simulated by the OSU atmospheric GCM/mixed-layer ocean model. *J. Climate*, 2, 459-495.


Monitoring Tropospheric Water Vapor Changes Using Radiosonde Data

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ABSTRACT. Significant increases in the water vapor content of the troposphere are expected to accompany temperature increases due to rising concentrations of the greenhouse gases. Thus it is important to follow changes in water vapor over time. There are a number of difficulties in developing a homogeneous data set, however, because of changes in radiosonde instrumentation and reporting practices. We report here on preliminary attempts to establish indices of water vapor which can be monitored. The precipitable water between the surface and 500 mb is the first candidate. We describe our method for calculating this quantity from radiosonde data for a network very similar to the network Angell uses for detecting temperature trends (see Angell, 1990). Preliminary results suggest that the noise level is low enough to detect trends in water vapor at the individual stations. While a slight increase in global water vapor is hinted at in the data, and the data suggest there may have been a net transfer of water from the Southern Hemisphere to the Northern Hemisphere, these conclusions are tentative. We also discuss the future course of this investigation.

1. INTRODUCTION

Water vapor is the most abundant “greenhouse gas” in the atmosphere and contributes the most warming of any of the greenhouse gases. Over half the warming expected from increases in CO₂ and the other greenhouse gases is attributed to concomitant increases in water vapor, since the warming increases evaporation and the capacity of the air to hold moisture (see, for example, Hansen et al. 1984). Trenberth et al. (1987) estimate that doubling the CO₂ concentration would increase the water vapor by about 20%, a figure similar to that obtained by Schlesinger and Zhao (1989). Hansen et al. (1984) estimate a 33% increase in water vapor from a doubled CO₂ experiment. This makes
water vapor a candidate for monitoring, along with the other radiatively active gases. Because water vapor is not well-mixed in the troposphere and has a mean residence time of only about 10 days, it is much more difficult to detect global changes of this gas than of the well-mixed gases.

Most estimates of global water vapor have been based on evaluating radiosonde humidity data. Bannon and Steele (1960) presented a nearly global analysis of precipitable water for the 1951–55 period, and Sellers (1965) used this analysis to estimate latitudinal- and global-average values. Tuller (1968) prepared a global estimate for 1964–68. Shands (1949), Reitan (1960) and Lott (1976) have prepared estimates for the U.S. for different periods. More recent estimates using radiosonde data are included in Peixoto and Oort (1983). Trenberth et al. (1987) have suggested changes in surface pressure as a way of estimating global water vapor changes. They point out, however, that their method may not be useful for estimating interannual changes because the noise level is too high. They also discuss the difficulties of using analyses of model fields of moisture and underlying topography for estimating moisture changes, as methods of computing the fields have changed with time. In the future, data from satellites may be the most practicable method of estimating global water vapor and its changes. An example of what can be expected when microwave techniques become operational can be found in Liu (1987). Although these previous estimates are for different time periods, changes from results of one study to another cannot be taken as evidence of a true temporal change in water vapor because of differences in analysis procedures.

The goal of the present effort is the establishment of a data base from which future changes in water vapor can be detected, should they occur. This paper is a report of progress toward that goal. Although there are several candidate indices of tropospheric water vapor, the emphasis here will be on the precipitable water in the surface-to-500 mb layer. Section 2 describes some of the problems of compiling a homogeneous record from the available radiosonde observations. Section 3 describes the data and their treatment. Section 4 gives some comparisons with the work of others. Section 5 shows samples of the data. Section 6 shows some changes in precipitable water over the 1973–88 period. The final section discusses the results and indicates the direction the work will take.

2. PROBLEMS

There are many problems in constructing a homogeneous record of water vapor changes from past radiosonde data. Improved instrumentation has been put into use and this can affect the record. Probably even more important have been changes in reporting procedures, particularly in handling low-humidity observations. Table 1 gives some of the changes that have taken place in U.S. instruments and practices which make it difficult to extend the record backward without considerable adjustments. The records from other countries are likely to be at least as troublesome as the U.S. record.

An illustration of the effects of such changes is shown in Fig. 1, which is reproduced from Angell et al. (1984). This was an examination of the long-term humidity variations at two U.S. stations, Brownsville, Texas and Great Falls, Montana. Prior to about 1965, nearly half the relative humidity observations at 500 mb were reported as “missing,” whereas after that date few observations were missing. This change occurred with the introduction of the carbon element into the radiosonde, and low-humidity values began to be reported. Prior to then “motorboating,” that is, relative humidity values generally less than 20%, were reported as missing. The dashed line in Fig. 1 shows the reported relative humidities before 1965 and the solid line during this period shows values adjusted by assuming all missing observations had relative humidities of 20%. This adjustment probably overestimates the humidity and some of the missing data may have been from
other causes, but the adjusted curves seem far more likely to represent the true variations. A similar effect, leading to an apparent decrease in humidity, may have occurred at Wakkanai, Japan (to be shown later).

Table 1. Some changes in U.S. radiosonde practices relevant to humidity measurements.

<table>
<thead>
<tr>
<th>Date</th>
<th>Practice</th>
<th>Replaced</th>
</tr>
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<tbody>
<tr>
<td>1943</td>
<td>Lithium chloride element</td>
<td>Hair hygrometer</td>
</tr>
<tr>
<td>1946</td>
<td>Constant pressure data</td>
<td>Constant height</td>
</tr>
<tr>
<td>1948</td>
<td>Saturation with respect to water assumed at all temperatures</td>
<td>Saturation with respect to ice assumed at temperatures below freezing</td>
</tr>
<tr>
<td>1957</td>
<td>Observations at 00/12 UT</td>
<td>Observations at 03/15 UT</td>
</tr>
<tr>
<td>ca. 1965</td>
<td>Carbon element</td>
<td>Lithium chloride element</td>
</tr>
<tr>
<td>1965</td>
<td>RH &lt; 20% = 19%</td>
<td>RH &lt; 20% = “missing”</td>
</tr>
<tr>
<td>ca. 1973</td>
<td>“New” housing</td>
<td>“Old” housing</td>
</tr>
<tr>
<td>ca. 1973</td>
<td>“Motorboating”</td>
<td>RH &lt; 20% = 19%</td>
</tr>
</tbody>
</table>

Another change in about 1972–73 occurred after it was realized there was a radiation effect on the humidity sensor. The housing of the element permitted a warming of the element and thus the reported relative humidities were too low. The housing was redesigned to reduce this error, so there could be an apparent increase in humidity after these housings came into use. This effect may be the source of the apparent rise in humidity after the early seventies seen in Fig. 1.

It also appears that reporting low-humidity values as motorboating is not always consistent. There is evidence that such observations are sometimes simply “missing” and thus are indistinguishable from instrument malfunction. It is the practice of the U.S. to report the humidity as missing when the air temperature is below –40°C, regardless of the signal from the humidity sensor. Certain other nations seem to follow this practice, but not all. These procedures will lead to a bias in the data toward high humidities, as these will be the only ones reported. This effect will increase with height, particularly in the higher latitudes where low temperatures are common. It is for these reasons that we terminated the calculations at 500 mb, although even at 500 mb there may be some bias toward the high values. (The bias can even be present occasionally in the winter at high latitudes at higher pressure levels.) Above 500 mb the bias can be quite pronounced.

3. DATA

The analyses are based on radiosonde data as compiled and sent over the teletype network by the National Meteorological Center (NMC). Worldwide radiosonde observations since 1973 have been archived at the National Climatic Data Center in Asheville, NC, and these data were obtained from 1973 to mid-1987, after which the data were extracted from the NMC teletype messages directly. The available records do not always have all the observations that were made. In particular, most of the Southern Hemisphere stations have data only for one time of day, and there are a few Northern Hemisphere stations for which this is also true.
Figure 1. Variation in year-average relative humidity (%) at mandatory pressure surfaces at Brownsville and Great Falls from 1958 to 1980, expressed as a deviation from the mean of the values for 1965 to 1980. The dashed lines show the indicated relative humidity before the adjustment based on the assumption that missing observations had relative humidities of 20%.

We used the reported temperature, dew-point depression and pressure at mandatory and significant levels from a limited global network. The stations used were based on the 63-station network used by Angell and Korshover (1983), which were selected to provide as uniform coverage as possible with a small network, given the distribution of observing stations. (Three stations were eliminated because their records had too many gaps: Khartoum, Cayenne and South Pole.) The stations are shown in Fig. 1 of Angell (1991); Appendix 1 gives the names, locations and elevations of the stations, as well as the observing times used. The dew-point depression was converted to specific humidity (q) at each reported data point up to 500 mb, and the q-profile was vertically integrated to give precipitable water (PW). PW between the surface and the 850 mb level (if the station pressure was greater than 850 mb), and from the surface to 700 mb, were also calculated. In addition, the temperature, dew point, specific humidity and relative humidity were tabulated for each observation at the surface and at the mandatory levels. We also evaluated the mean layer temperatures (thicknesses) between 850 mb, 700 mb and 500 mb. The details of the calculations are given in Appendix 2.
When the dew point was reported as "motorboating," we assumed the relative humidity was 15% and calculated the specific humidity accordingly. If the dew point was missing at any mandatory level up to and including 700 mb, not motorboating, the entire sounding was discarded; if it was missing at 500 mb, the sounding was retained up to only 700 mb.

Despite some quality control of the data at NMC, erroneous data sometimes appeared. We established limits on the temperatures so that the more egregious data were rejected. If these questionable data were at mandatory levels below 500 mb, the entire sounding was rejected; if at 500 mb only, that level was rejected. If a significant level's data failed the test, that data point was rejected.

There are gaps in the records of individual stations. The main reason for gaps is the inability of the NMC to acquire the data, probably because the data were not available at the time of collection. Other reasons are the above-mentioned reporting of relative humidities as "missing" when the temperature is below −40°C, and "missing" may be sent as low humidities at other times. It is possible that some instruments in use are poor at recording low values. In particular, Indian stations had a poor record during the drier winter months, and many observations are missing at these stations.

To form a monthly average we required at least 3 observations for each observation time (00 UT or 12 UT) or a nominal 10% of the possible observations. In practice the recovery rate was much better. About 80% of the stations had better than 50% of the months with more than 75% of the observations available. The North American and European stations were the best, with the USSR stations not far behind. Exceptions were the very high-latitude stations in winter when there is little moisture. In general, Southern Hemisphere station records have more gaps than Northern Hemisphere stations. The coverage usually improved with time. The quality of the records is probably more a function of NMC's collection practices than the zeal of the individual stations.

One would hope that the causes of the missing data are not such as to bias the monthly averages. This may be the case if the data are missing because NMC could not acquire them, but that does not appear to be so when the data are missing at only the highest level. An illustration of this is shown in Fig. 2. The data shown are the mean January values at Wakkanai, Japan from 1973 to 1989, the solid line being the surface-to-500 mb PW and the dashed line the surface-to-700 mb PW. The bar graphs give the number of observations going into the respective averages. Clearly some change in instrumentation or reporting procedure, or both, took place in 1981. The reporting of 700 mb humidities improved slightly after that time, but there was a dramatic increase in the number of 500 mb reports. This change accompanies the apparent drop in surface-to-500 mb PW. It appears that only the high-humidity data were reported in the early part of the record. This is the kind of problem that one must watch for in evaluating moisture data.

4. ESTIMATE OF GLOBAL PRECIPITABLE WATER

Although the main purpose is to develop one or more indices of water vapor whose values can be used to detect long-term changes, we wished to make an estimate of total precipitable water (PW) to compare with other estimates of this quantity. Because the network of stations is sparse, we would not expect our estimate to be as accurate as those made with more stations, but we would hope to be in reasonable agreement. To check this we examined the data for the period 1985–87 in some detail. The monthly means of the data for each station were used to form seasonal means, the seasons being DJF, MAM, etc. The seasonal means were then averaged for the three years for each station, and the individual station data were aggregated into zonal values. The zones were North and South Polar (60°–90°), North and South Temperate (30°–60°), North
Figure 2. Time changes of the January PW at Wakkanai, Japan, 1973–1989. Figure 2a shows the surface-to-500 mb PW (solid line) and the surface-to-700 mb PW (dashed line). There were no 500 mb data available in 1977 and 1981. Figure 2b shows the number of available observations each month; open bars for the surface-to-700 mb, solid bars for surface-to-500 mb.

and South Subtropics (10°–30°) and Equatorial (10°N–10°S). These zones were further aggregated, with weightings appropriate to the area of the zones, into hemisphere values and a global value.

There are some biases in the distribution of stations and so some possible bias in the “climate” of the network. The North Temperate Zone is comprised of stations that are in the northern part of the zone; in fact none is south of 45°N. This means that the mean temperature, and therefore quite likely the water vapor, is underestimated in this zone. On the other hand, the South Temperate Zone has no station south of 47°S, so that the temperature and water vapor there are likely slightly overestimated. The South Polar Zone consists of stations along the coast of Antarctica, and so the temperature and water vapor will be overestimated. Because there is very little water vapor in this region anyway, there will be only a small effect on the estimate of total vapor.

Another bias occurs in the elevation of the stations, particularly in the Equatorial Zone. Both Bogota (elevation 2560 m) and Nairobi (elevation 1820 m) are in this zone. This will produce a cooler and drier estimate of the zone than is probably correct. The lower elevations of the South Polar stations will produce a warmer and more moist estimate there. All in all, the Northern Hemisphere would likely be cooler and slightly drier, and the Southern Hemisphere warmer and more moist, than the true climate. Globally we probably underestimate the total moisture between the surface and 500 mb.

The greatest source of bias in our estimate of total water vapor is the termination of the calculations at 500 mb, for reasons explained above. There are several estimates of the fraction of water vapor above 500 mb. Reitan (1960) estimated about 6% above
the 48 contiguous states of the U.S. Liu (1986) estimates 4.3\%, but says this is based on limited data. From Bannon and Steele (1960) we also estimate about 6\%, and so in subsequent comparisons with other estimates we have increased the surface-to-500 mb PW by 6\%, although it would be safer to say the true value probably lies between 4 and 8\%, and would be a function of latitude.

Our global estimate of the PW between the surface and 500 mb is 2.29 cm, which becomes 2.43 cm with the 6\% added. This value is close to the Trenberth et al. (1987) adjusted estimate of 2.57 cm, which also agrees with the Peixoto and Oort (1983) estimate of 2.57 cm. The 2.43 cm estimate is very close to Sellers’ (1965) estimate from the Bannon and Steele (1960) data which give 2.47 cm.

Trenberth et al. (1989) calculated that, on an annual basis, the Northern Hemisphere has 0.16 cm more water vapor than the Southern Hemisphere, whereas we calculate the difference to be 0.12 cm. Since we underestimate slightly the Northern Hemisphere and overestimate the Southern Hemisphere (because the latitudes of our stations are too far north in both Temperate Zones), the agreement is good. Trenberth et al. also estimate that the global atmosphere has 0.39 cm more water vapor in July than in January; we calculate a difference of 0.34 cm between JJA and DJF. Again the agreement is good. These comparisons give confidence that our calculations reflect the atmosphere fairly well despite the small number of stations.

It is worth noting that our global-mean surface temperature of 13.1°C is somewhat lower than the value of 14.9°C given by Oort (1983), but higher than the value of 12.3°C in Sellers (1965). Our estimate of the mean global specific humidity is 9.6 g/kg, whereas Oort’s is 10.1 g/kg. These lower values compared to Oort’s are consistent with the biases noted above. Both Oort’s values and ours show that the surface of the globe is warmest in JJA and that the surface moisture is largest in this season.

5. PRELIMINARY RESULTS

Eight station records have been selected to illustrate the data and their variations. The stations, from north to south, are: Barrow, Kiev, Brownsville, Trivandrum, Majuro, Tahiti, Adelaide and Gough. Trivandrum was chosen to illustrate a station with a poor record. Adelaide’s record was poor in the early years but improved with time. The others generally had fairly complete records, although Majuro and Tahiti had data only for 00 UT. The data presented are the surface-to-500 mb PW in all cases.

Figure 3 shows the monthly means of the 00 UT data for 1986. The error bars represent twice the standard error, based on the number of available observations for the particular months. Based on this record alone, one would be hard put to argue that there was any annual cycle at Trivandrum. Annual cycles are clearly evident at the other stations, although they are most pronounced at Barrow, Kiev and Brownsville, as would be expected from their continental locations.

The curves of Fig. 3 can be compared with those in Fig. 4, where the 00 UT monthly averages have been averaged themselves from 1976 through 1986 to give an estimate of the mean annual cycle. (There are more gaps in the records before 1976 and some missing months in 1987.) The error bars are based on the number of months that went into the averages, usually 11, except at Trivandrum. The 12 UT curves are virtually identical to those shown, except for Trivandrum where the 12 UT generally are more moist, and Brownsville in the summer where the 12 UT is also slightly more moist. If one overlays the curves in Figs. 3 and 4, it would show that 1986 was a reasonable sample of the years, and that 1986 reproduced the mean annual cycle reasonably well. Trivandrum, as usual, is something of an exception.
Figure 3. Monthly 00 UT surface-to-500 mb PW at 8 selected stations for 1986. The error bars show twice the standard errors of the monthly estimates.

Figure 5 shows samples of the annual average PW as a function of time for the 1973–88 period. The solid lines are based on the 00 UT soundings and the dashed lines on the 12 UT soundings. The stations illustrate several features of the records. There is very little year-to-year variation in the polar regions where the PW values are small. At Barrow there is no discernable difference between the two times, where the observations are made at 1400 and 0200 local time, and at Kiev where the observations are also at 1400 and 0200 local. In fact, the records show little difference between the two observation times except at Trivandrum (observations at 0530 and 1730 local). Adelaide presents an interesting picture; in the early part of the record the data are sparse but the record improves with time. The interannual variability seems to lessen as the number of observations per month increases.
Figure 4. Average annual variation of the 00 UT surface-to-500 mb PW at 8 selected stations, based on the 1976–86 period. The error bars show twice the standard errors.

As a preliminary look at possible trends, the data from the individual stations, or at least those with an adequate record, were tested with the Spearman rank statistic, a statistic that measures randomness against the alternative of a trend (WMO, 1966). The annual-mean surface-to-500 mb PW values at 55 stations were tested for trend. Upward trends significant at the 95% level were found for 13 stations, while an additional 8 stations had upward trends significant at the 67% (roughly one sigma) level but not 95%.
For the downward trending stations, 5 were significant at the 95% level and another 9 between 67% and 95%. No trends were detected at the other 20 stations. Of those stations displayed in Fig. 5, Brownsville, Trivandrum and Majuro had significant (95%) upward trends while Kiev had an upward trend significant above 67% but not at 95%. Adelaide and Gough had downward trends significant above the 67% level but not at 95%.
Figure 6 shows the pattern of the trends. The stations with single plus or minus signs had trends significant between 67% and 95%, and the double sign indicates significance at better than 95%. The distribution does not appear to be completely random. The equatorial region north of the Equator seems to have experienced an increase in PW during this period. Both polar regions show a tendency for a decrease, albeit weakly, as do some of the Southern Hemisphere stations outside the equatorial region. Significant increases are also evident over western North America.

The increases in the PW in the lower latitudes is supported in a recent study of the PW between the 700 and 500 mb levels (Hense et al., 1988). They show much the same distribution of trends, as Fig. 6, for the common regions. Their record is from 1965 to 1984 and used tabulated mean monthly dew points at 700 and 500 mb rather than the individual soundings. They also have a few more stations in the western tropical Pacific than we. The pattern of Fig. 6 suggests that water vapor may have increased globally over the period studied. The indicated increases occurred in regions of high PW, while the decreases were in regions of lower PW. Figure 6 also suggests the possibility of a net transfer of water vapor from the Southern Hemisphere to the Northern tropics during this period. Of course, more years of data will help clarify the situation.

Figure 6. Trends in precipitable water at the stations for the period 1973–1986. The plus and minus signs indicate increasing and decreasing PW, respectively. Double signs indicate the trend was significant at the 95% level or better, while single signs indicate significance between 67% and 95%. A zero indicates significance was less than 67%. No symbol by a station indicates insufficient data to test.
6. DISCUSSION

The work so far shows that it is possible to develop records of the surface-to-500 mb PW at individual stations that have fairly low noise levels so that trends can be detected at these stations. The tentative finding of an increase in PW during 1973–1988, at least in the Northern Hemisphere tropics, is consistent with the rise in temperature during this period (Angell, 1991). Such trends may well reflect the timing of the ENSO events that occurred during this time, rather than the signature of a long-term trend.

Other indices should be examined, including the surface-to-700 mb layer PW and the 850 mb specific humidity and mean dew point. The latter is not a perfect reflection of mean water vapor, as discussed in Appendix 2, but the ease of acquiring such data suggests examining their use. Furthermore, these data allow one to examine relationships between changes of relative humidity and temperature, and changes of specific humidity and temperature.

One would like to extend the study, both in space and time. A sobering conclusion of this examination, however, is that such an enterprise will require scrutinizing individual records for discontinuities in instrumentation and procedure. To look back beyond 1973 may well require development of techniques to adjust the data for these changes, and indeed this may be necessary for some post-1973 data. U.S. practices are not always the practices of other nations.
APPENDIX 1

The names of the stations, their latest WMO Identification (I.D.) number, latitude, longitude and elevation are given in Table A1. The observation times used are indicated in the appropriate column by an “X.” Data were sometimes available for the other time, but we list here the time when the record was best.

Table A1. List of stations.

<table>
<thead>
<tr>
<th>NAME (Country)</th>
<th>WMO I.D.</th>
<th>LAT (deg.)</th>
<th>LONG (deg.)</th>
<th>ELEV (m)</th>
<th>OBS. TIME</th>
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APPENDIX 2. CALCULATION OF PRECIPITABLE WATER FROM RADIOSONDE DATA

The radiosonde sends a signal that is proportional to the ambient relative humidity. The instrument is calibrated to give the ratio of the ambient vapor pressure (e) to the saturation vapor pressure over water (es), rather than ice, at all temperatures. The relative humidity is converted to dew-point depression at the ambient temperature and this is the variable that is reported.

We enter the dewpoint (Td, degrees Celsius) into the equation

\[
e_s = 6.1121 f \exp\left(\frac{17.502 T_d}{240.97 + T_d}\right) ,
\]

where \( f \) is the "enhancement factor," a correction to adjust the pressure of pure vapor at saturation over water for the mixture of air and vapor. This factor is approximated by

\[
f = 1.0007 + 3.46P(10^{-6}) ,
\]

where \( P \) is the total pressure.

Equations (1) and (2) are from Buck (1981) who used formulations by Wexler (1976) which were based on then recent measurements of the appropriate quantities.

Equation (2) is an approximation to the correct form which includes a very weak dependence on temperature. As used here it implies an increase of about 0.3% over the uncorrected value. If \( f \) is taken as 1.0, Eq. (1) will give values that are quite close to values from the Smithsonian Tables (List, 1968) which are commonly used.

For each point in a sounding the values of \( P \) and \( e_s \) (from Eq. (1)) were entered into Eq. (3) for the specific humidity, \( q \),

\[
q = \frac{0.622 e_s}{P - 0.378 e_s} ,
\]

Finally, the precipitable water (PW) was obtained by integrating the values of \( q \) for that sounding according to

\[
PW = \frac{1}{g} \int q \, dP .
\]

The integration was carried out assuming \( q \) is linear in \( P \) between each significant level so that in practice PW between level \( i \) and \( i + 1 \) was

\[
PW = \frac{1}{g}[(q_i - q_{i+1})/2](P_i - P_{i+1}) .
\]

With \( g = 980 \text{ cm/s}^2 \), PW has the units of \( \text{g/cm}^2 \); when divided by the density of water, taken as 1 \( \text{g/cm}^3 \), PW is expressed as centimeters of water (q has the units of \( \text{g/g} \) but in the tables and text we express q in \( \text{g/kg} \)). For each sounding we calculated the PW from the surface to 850 mb (if the surface pressure was higher than 850 mb), from the surface to 700 mb and to 500 mb.

The assumption of linearity between data points may lead to an overestimate of the PW. The q-P curves often suggest an exponential decrease with decreasing pressure, so a linear assumption may tend to overestimate the integral. Since significant levels are supposed to be reported in the soundings if temperature \( T \) or relative humidity departs from linearity, this should be a minimal effect.

In those cases where the dew point was reported as "motorboating," that is, at relative humidities less than 20%, we expressed the dew-point depression as
\[ T - T_d = 23.6 + 0.183T - 0.00034T^2 , \]

where \( T_d \) is the dew-point temperature. This assumes a relative humidity of 15\% for those situations.

Different studies have made somewhat different assumptions in the process of calculating PW. We mention here a few of them and attempt to estimate the effects, almost always small, on the results. As noted above, the inclusion of the enhancement factor, \( f \), adds about 0.3\% to the values of \( e_s \) and hence to PW. Some studies have used the mixing ratio, \( w \), instead of \( q \). The specific humidity, \( q \), can be as much as 2 to 3\% less than \( w \) at high temperatures, but a more likely overall value is in the range of 1\%. Sometimes the approximation \( q = e_s/P \) is used in which case \( q \) can be underestimated by as much as 1–2\% but is more likely less than 1\%. All these errors are greatest at high humidities, high temperatures and low pressures.

Because of the ease of calculation, sometimes the mean monthly dew point has been used to estimate the mean monthly vapor pressure and hence the mean monthly specific humidity. The saturation vapor pressure-temperature relationship is not linear; the relation leads to a systematic underestimate of the mean \( q \). A. Taylor (personal communication) has pointed out that the difference between the true mean \( e_s \) and the \( e_s \) calculated from the mean dewpoint can be estimated from

\[ \frac{1}{2} \frac{d^2 e_s}{dT_d^2} \text{var}(T_d) , \]

where \( \text{var}(T_d) \) is the variance of the dew point. Since the second derivative of the \( e_s, T \) relationship is positive, the mean \( e_s \) is underestimated by using the mean dew point. This effect can be greater than 10\%, and as high as 20\% in some cases at cold temperatures where the effect is generally largest. It can result in overall underestimates of PW of as much as 10\%.

It is also worth noting that the definition used here of relative humidity \( = e/e_s \) is different from the definition, \( w/w_s \), used by some. The difference is largest at the nominal relative humidity of 50\%. At this point \( w/w_s \) can be less than \( e/e_s \) by as much as 0.025, at high temperatures, but a more general value of the difference is about 0.01.
REFERENCES


Observational Evidence of Changes in Global Snow and Ice Cover

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University of Colorado
Boulder, CO 80309-0449
U.S.A.

ABSTRACT. Sources of observational data on recent variations in the seasonal extent of snow cover and sea ice, of the terminal position and volume of alpine glaciers, and of ground temperature profiles in areas of permafrost are briefly reviewed. Recent evidence of changes in these variables is then examined. The extent of seasonal snow cover in the Northern Hemisphere and of sea ice in both hemispheres has fluctuated irregularly over the last 15–20 years with a range of about 10–15 percent in each case. There is no clear evidence of any recent trends, despite general global warming. In contrast, most glaciers retreated and thinned from before the turn of the century until the 1960s and Alaskan permafrost temperatures have risen 2°–4°C per century. Recently, glacier advances have been noted, perhaps in response to increased accumulation. Problems of linking climate forcing and snow/ice responses are discussed.

1. INTRODUCTION

The major features of snow and ice cover that are of significance for global climate are seasonal snow cover, sea ice, land ice and permafrost. The dimensions of these cryospheric components are detailed in Barry (1985), who also discusses their respective roles in the climate system in terms of their surface albedo, cold reserve and insulating properties on the underlying surface.

The sources of information on snow and ice were largely limited to point observations until the advent of global satellite coverage in the mid-1960s. This remains true in the case of permafrost data and to a large degree also for land ice, except for the basic inventory of glacier occurrence. Some pre-satellite observational records are discussed to provide a longer time perspective, but the limited spatial coverage and uncertain representativeness of such observations must be recognized. These four variables are discussed in turn, beginning with land ice and permafrost for which the observations are primarily surface based.
2. LAND ICE

The formal collection of observations on glacier extent began under the auspices of the International Glacier Commission (now International Commission on Snow and Ice) in 1894. Most records of changes in land ice extent refer to this century; the European Alps, where the changes of a few glaciers have been documented over several centuries, is an exception. Measurements of ice volume and changes in mass balance are considerably more scarce and, prior to the beginning of the International Hydrological Decade (1965-1974), there are records for only 37 glaciers worldwide. Standardized glacier data are now assembled and reported by the World Glacier Monitoring Service in Zurich, Switzerland (Haeberli et al., 1989).

Ice volume can be determined on the basis of airborne radio-echo sounding transects of bedrock and ice surface elevation; these now cover much of Greenland but only about 25 percent of Antarctica. Conventional mass balance studies are performed by measuring winter accumulation and summer ablation, but the field techniques are slow, approximate and inapplicable to large ice sheets where iceberg calving is a major contributor to mass loss. For this reason we do not know definitively whether the Greenland and Antarctic ice sheets are growing or shrinking. The Greenland ice sheet seems close to a balanced state, while computations for the Antarctic ice sheet indicate that it may be growing (Polar Research Board, 1985).

Data on glacier balance and volume change spanning more than 50 years are available for about 25 glaciers (Meier, 1984; Makarevich and Rototaeva, 1986) located mostly in mid-latitudes of the Northern Hemisphere. Direct measurements of mass balance obtained from pits and snow stakes on the glacier surface began only in 1945/46 (Wood, 1988), but are now reported for about 75 glaciers in the Northern Hemisphere. There are more numerous and longer records of glacier advance and retreat (Haeberli and Müller, 1989), but these are an unreliable indicator of volume change. Oerlemans (1988) proposes normalization of records of glacier length by scaling, using the ratio of glacier area to the mean width of the glacier tongue as a characteristic length. Müller (1988) correlated changes in length of Swiss glaciers with data on mass balance and meteorological data. He found that a model relating length changes to summer temperature at a mountain station and annual precipitation was satisfactory for only about half of the 69 glaciers studied. This is mainly a result of the variable geometry of glaciers and their differing response times. For this reason composite statistics of such advances and retreats for mountain ranges are commonly used to provide a general measure of the regional response of glaciers to climatic fluctuations. Figure 1 illustrating the trends observed over the last hundred years in the Alps, shows a high degree of similarity between Austria and Switzerland. On a broader scale, Röthlisberger (1986) shows a good agreement between glacier trends in the Alps and in New Zealand over the last 5000 years and, more generally, between the two hemispheres as shown in Fig. 2. The comparison is based on glaciers of comparable size, with the amplitudes scaled against the AD 1980 minimum and the maximum neoglacial advances between 3700 BP and the present. The distance between the location of neoglacial and present termini averages 1-3 km. It should be noted that the recession phases are generally less well determined than the advances. Also, the resolution of the last one thousand years is higher than for the earlier record, especially in Europe. Prior to about 100 AD, there is a roughly 500-year interval between the main phases of ice extension.

For the 1955-80 period, Makarevich and Rototaeva (1986) show 27 percent of 104 North American glaciers advancing and 53 percent retreating, whereas for the northern Tien Shan (210 glaciers), the Altai (135 glaciers) and the Caucasus (63 glaciers), only 2-8 percent were advancing and 82-94 percent retreating. Zhang (1986) reports that, over the interval 1969-1975, 33 percent of about 400-450 glaciers worldwide were advancing and
Figure 1. Changes in glacier length, mass balance and climate variables in the Alps, 1890–1986 (from Patzelt, 1987). (a) Percentage of glaciers advancing (black) or stationary (unshaded column) in the Austrian Alps. The inverse scale indicates retreating glaciers. (b) As in (a) for the Swiss Alps. (c) Mean specific mass balance (cm). (d) Temperature of the May–September ablation season at mountain stations; individual years and 10-year running-mean values (heavy line). (e) Decadal-mean departures of mean-annual precipitation from the 1931–60 mean in glacierized mountain ranges of the eastern Alps; the dashed line shows values for 1950–80 for stations with annual totals >1500 m, the dotted line for stations with <800 mm.
Figure 2. Glacier fluctuations in the Northern and Southern Hemispheres over the last 11,000 years. Each scale shows fluctuations relative to the 1980 minimum on the left and maximum neoglacial advances since 3700 BP on the right. [From Röthlisberger, 1986.]

57 percent retreating, while for 195 glaciers in the Kunlun, Tien Shan, Qilian, Himalaya, Karakoram and other ranges, 23 percent were advancing and 48 percent retreating. The majority of some 550 glaciers, worldwide, showed retreat rates of 1–10 m/yr during the 1960–80 period.
The above statistics may conceal some recent trends. Wood (1988) shows that the percentage of glaciers worldwide that are observed to be advancing annually increased from about 7 percent (of 270 glaciers) in the early 1960s to 50 percent of some 450 glaciers in the late 1970s (Fig. 3). Most of these glaciers are located in the European Alps, Iceland and, since about 1970, in North America and the Soviet Union. Mass balance measurements for 53 glaciers in the same countries over the same interval show a more complex trend (Fig. 4). For both the European Alps and all countries, the percentage of glaciers with a positive balance peaked in the late 1960s, exceeding 70 percent of glaciers in the Alps, declined in the mid-1970s, especially in the Alps, but rose to another peak for the Alps in the 1970s, while declining further for all countries. This pattern strongly suggests a shift away from the predominantly negative mass balances and retreating glacier fronts that characterized most alpine glaciers during the period from around 1900 until the 1960s–1970s, only briefly interrupted about 1920 (Meier, 1984; Patzelt, 1985; Vallon et al., 1986; Haeberli et al., 1989).

The spatial consistency of such trends is an important issue. Vallon et al. (1986) demonstrate from four long (reconstructed) records for 1896–1977 for the IGAN glacier (Urals), Dzhankuat (Caucasus), Folgefonna (Norway) and Sarennes (French Alps) that there was consistent agreement on an interval of positive mass balance in the early 20th century that led to a brief interval of glacier advance around 1920 and negative balance around 1940–50 causing retreat. However, the South Cascade Glacier, Washington, USA, located in a maritime climatic regime showed an almost entirely negative record. Similar continental/maritime contrasts are found within Norway for glaciers on either side of the climatic divide (Lauman et al., 1988).

The climatic implications of these trends of glacier extent and mass balance are complex. Positive mass balances may reflect cooler summers with less ablation, increased winter accumulation from snowfall, or a combination of the two. Moreover, the movement of the terminus is dependent on the glacier size and type, its hypsometric profile
and its geographical location (maritime/continental, polar, mid-latitude, tropical) (Furbish and Andrews, 1984; Müller, 1988). Patzelt (1985) considers that increased winter accumulation and cooler summers reinforced one another in the Alps between 1910–20 and between 1965–1980 (see Fig. 1b–d). Mayo and Trabant (1984) suggest that in southern Alaska increased winter snowfalls associated with recent milder winters were responsible for recent positive mass balance years.

Observed changes in glacier length and thickness have been used to derive changes in climate from a two-dimensional, time-dependent model (Smith and Budd, 1981), but the solutions are not unique. Conversely, changes in ice masses that may result from greenhouse-gas-induced warming have been predicted using simple energy balance models by Kuhn (1981) and Oerlemans (1986). However, such work is still in an early stage of development.

3. PERMAFROST

Permafrost may occur where mean-annual air temperatures (MAAT) are less than \(-1^\circ C\) or \(-2^\circ C\), and it is generally continuous where MAAT < \(-7^\circ C\). Information on permafrost occurrence has been assembled by geographers and geologists from field observations, with extensive recent data collected through mining exploration, engineering construction activities, and geophysical surveys in the Arctic and sub-Arctic. Reliable maps of permafrost extent date only from about 1950 and, therefore, large-scale temporal changes are scarcely detectable. Permafrost thickness is known from boreholes drilled in connection with mining or other engineering activities, although much of this information remains proprietary. Thickness can also be estimated from knowledge of the geothermal temperature gradient and the mean-annual ground (or air) temperature. Geothermal data
for Canada are routinely published by the Earth Physics Branch of the Department of Energy, Mines and Resources, but there are no other comparable readily accessible archives (Barry, 1988).

As a result of the long history of Russian and later Soviet exploration, settlement and scientific study in Siberia, there is considerable information on Eurasian permafrost extent and thickness. Conditions in Tibet and western China, however, are less well known. Permafrost study began only in the 1940s in North America. Information for mountain areas is sketchy (Harris, 1986; Pêwé, 1983) and offshore subsea permafrost in the Arctic is still being mapped (Vigdorchik, 1980; Blasco, 1983). The mapping of permafrost distribution and thickness relies heavily on climatic predictive methods. Conditions beneath the Greenland and Antarctic ice sheets are known primarily by inference and ice modelling calculations.

Borehole temperature measurements in permafrost can serve as an indicator of net changes in temperature regime. Gold and Lachenbruch (1973) infer a 2–4°C warming over 75–100 years at Cape Thompson, Alaska, where the upper 25 percent of the 400 m thick permafrost is unstable with respect to an equilibrium profile of temperature with depth (for the present mean-annual surface temperature of −5°C). In the coastal plain of Alaska similar data imply a variable, but widespread, warming of 2–4°C at the permafrost surface during the twentieth century (Lachenbruch and Marshall, 1986; Lachenbruch et al., 1988). The warming extends to 75–100 m depth. The longest air temperature record in the Alaskan Arctic, at Barrow, shows a slight cooling trend since 1921. However, spatially average values for stations in the North American sector of the Arctic showed a strong rise (5.4°C/100 yr) from 1880–1925, followed by cooling of 0.6°C/100 yr. Lachenbruch et al. (1988) point out the similarity of the net warming to the measured rise in permafrost temperature. Fuller understanding of this relationship requires better information on other possible changes—in snow cover, on the seasonal distribution of the warming and, at some sites, on perturbations of the surface associated with the drilling of the boreholes. On the continental scale, present-day climatic conditions and the theoretical annual ground temperatures needed for permafrost to exist, indicate that most of the permafrost occurring in North America is now unstable and subject to degradation (Harris, 1986).

4. SNOW COVER OBSERVATIONS

Four characteristics of a snow cover are important for the present discussion, namely, extent, depth, water content and surface albedo. The first and last are of primary importance for large-scale climate, whereas the other two are principally of hydrological significance.

Standard meteorological observations include the 6- or 12-hourly synoptic weather reports of snowfall and once-daily determinations of snow depth on the ground. However, synoptic reports of snowfall and snow depth are difficult and expensive to retrieve because they are included within large-volume data bases of station synoptic reports. Surveys of snow depth and water content at snow courses are also conducted at about monthly intervals or continuously (via snow-pressure pillows) by other organizations; in the United States these surveys are carried out by the Soil Conservation Service and cooperating groups on a state-by-state basis. Nearly all of these records begin in this century. There are no standardized annual, global summaries of snow cover duration and average maximum depths. Although various national summaries or maps exist, the time intervals for which these data are available differ greatly and the criteria used to define particular parameters are seldom identical (Barry and Armstrong, 1987). A little-recognized problem is the incompatibility of accumulated snowfall water equivalents on
the ground (at open or forested sites) with those measured by precipitation gauges at climatological stations (Goodison, 1981). Measurement of the density of new snow is not standard procedure at the majority of Canadian stations, for example. Precipitation gauge design, shielding devices used to minimize wind effects, and measuring techniques also may differ considerably between countries.

The major sources of snow cover charts and surveys are described in Crane (1979). Weekly maps of winter snow depth for the United States (U.S. Department of Commerce, 1935–present) have been digitized for 120°–70° W by Walsh et al. (1982) for the period 1947–82 and used to study the effects of snow cover anomalies on monthly climate. Daily statistics are less readily available, although station data are published for Canada (Atmospheric Environment Service, Canada, 1961). Robinson (1987) examined twentieth century snow cover over the Great Plains and Midwest of the United States and found an increase in January snow cover over the past forty years. Foster (1989) analyzed dates of snow melt at Arctic stations and found a trend towards earlier snow disappearance since about 1950, perhaps related to drier conditions. In contrast, there were irregular trends at Canadian Arctic stations and no change since the mid-1930s at a station in Finland.

The value of satellite remote sensing of snow and ice was recognized early in the development of weather satellites and such systems now provide the primary tool for routine global observations of the cryosphere. Satellites afford frequent repetitive mapping of global snow cover, and data collection is possible in the absence of illumination and in the presence of clouds (using microwave measurements). Satellite relay, or meteor-burst radio relay, also permit data collection from platforms installed in remote locations.

Snow-cover mapping from satellite imagery began operationally in 1966. The National Environmental Satellite [now Data and Information] Service (NESDIS) prepares hemispheric maps of snow and ice at weekly intervals. The maps have been digitized as snow/no snow for 7921 gridboxes covering the Northern Hemisphere for November 1966 through December 1980 by Dewey and Heim (1981). Some deficiencies of this data set are noted by Wiesnet et al. (1987): the charts from 1966 through 1974 did not consistently map Himalayan snow cover; there are occasional extensions of snow cover beyond the southern limits of the map; the seasonal northern limit of illumination for the satellite sensors in the visible wavelengths limits polar coverage; and scattered mountain snows are omitted due to the coarse grid resolution.

A snow-cover atlas has been produced from these data by Matson et al. (1986). A time series of snow extent for 1966–1987 (Fig. 5) shows no evident trends. Dewey (1987) analyzes monthly snow-cover frequency from the NOAA-NESDIS data and finds that the autumn onset of snow cover is several weeks later than depicted by earlier studies. However, he notes that this may be attributable to the different sources of data rather than to a true change.

The use of satellite passive microwave data to monitor global snow-cover is currently being explored (Rango, 1980; Kunzi et al., 1982; Foster et al., 1984; Chang et al., 1987). At present this is still a research technique, but a capability to map global-scale snow extent has been demonstrated. Corrections need to be developed to the algorithms for vegetation cover and terrain effects (Schweiger et al., 1987). Snow-water equivalent may also be determinable in dry snow environments, but further work on this is required.

While numerous records of snow cover exist, there are several deficiencies which affect their usefulness. The satellite-based record is relatively brief and intercomparisons are needed with conventional surface observations to assess the spatio-temporal variability more comprehensively. A further problem is caused by the general absence of compact digital files of snow cover data.
OBSERVATIONAL EVIDENCE OF CHANGES IN GLOBAL SNOW AND ICE

The relationships between ground-based or satellite-based data and snow parameters needed for climatic modelling or assessment purposes are ill defined. The modeller is primarily interested in surface albedo which depends upon snow depth, age, water content and type of vegetation cover. Neither surface nor space observations of snow cover can yet provide the information desired in a simple straightforward manner. More work on this problem is required, although there may be no single best solution.

5. SEA ICE

Sea ice limits have long been observed by ships, and harbors have reported the dates of the appearance and disappearance of coastal ice. Historical sea ice records have been used as climatic indicators by Lamb (1977) and others, although the early observations present many problems of interpretation (Barry, 1986). Ice conditions are now reported regularly in the marine synoptic observations, by special aerial reconnaissance flights (as coded observer reports, photography and remote sensing data), and by coastal radar. Since the early 1970s, satellite remote sensing data, particularly NOAA Very High Resolution Radiometer and Nimbus passive microwave data, have been routinely incorporated into the U.S. Navy-NOAA Joint Ice Center weekly ice charts for the Northern and Southern Hemispheres. Maps of sea ice generally depict the boundaries for various ice concentration classes (0–10/10) and they may also distinguish different age categories and the degree of ridging intensity (World Meteorological Organization, 1970).

The longest historical records are those of drift-ice duration on the coasts of Iceland (Koch, 1945). Lamb (1977, p. 583) tabulates these data, with some additions, annually from AD 1600 to 1975; the series is believed to be complete from 1780. A new analysis of the Icelandic data has recently been completed and a decadal ice index presented for 1601 to 1780 (Ogilvie, 1984). A related record of the Storis drift (ice from the East...
Greenland Current) northward along the west coast of Greenland exists from 1820–1930 (Speerschneider, 1931; see Lamb, 1977). The annual winter maximum extent of Baltic Sea ice has been tabulated from 1720 to 1956 (Betin and Preobazhensky, 1959; summarized by Lamb, 1977, p. 586–589), and a brief analysis of these records for 1830–1951 has been made by Jurva (1952).

Numerous chart series now provide both monthly and approximately weekly maps of ice conditions in virtually all seasonally ice-covered sea areas (Barry, 1986). The longest series was that published for the Arctic Ocean by the Danish Meteorological Institute (1901–56) for 1901–1939 and 1946–1950. The monthly ice limits for the North Atlantic from 1901 have been digitized by Kelly (1979). Problems with the early twentieth century ice information are illustrated by the data on Arctic Ocean ice extent presented by Zakharov (1981). The 57-year record of ice area for late August (smoothed by a 5-year binomially weighted moving average) has been correlated with similarly smoothed June–August temperature data for 65°–85°N (Kelly and Jones, 1981). There is a strong inverse correlation (−0.81) for 1951–80, but the correlation drops to −0.13 for the whole period 1924–80 (Barry, 1983). One must conclude either, that one or both sets of data are unreliable due to the limited spatial coverage of the observations in the Arctic prior to about 1950, or that the strong statistical ice-climate relationship for 1951–80 is fortuitous and the variables are essentially unrelated. The Soviet ice data apparently are derived from the Danish charts for the Arctic, probably supplemented by other records for the Siberian Northern Sea Route.

Reasonably consistent data on ice extent are available in the Navy-NOAA weekly charts for the Arctic (from 1972) and the Antarctic (from 1973) (Fig. 6). Neither graph shows any evidence of long-term trends. The information on ice concentration on these maps is of variable quality. The Navy-NOAA maps for the Antarctic have been digitized to provide data sets of ice extent versus longitude and ice area by several different workers. Despite the common source of these data sets, there are distinct differences between them in the estimates of monthly ice areas, especially in the transition seasons (Sturman and Anderson, 1984). This illustrates the necessity to document carefully the procedures used to generate data sets likely to be used by others.

The availability of passive microwave data since 1973 provides the first estimates of ice concentration and ice type. Atlases have been prepared for both polar regions from the single-channel ESMR data (Zwally et al., 1983; Parkinson et al., 1987). The Scanning Multichannel (dual polarization) Microwave Radiometer (SMMR) data from Nimbus 7 with ten channels of radiometric data have provided new information on total ice concentration, the multiyear ice fraction, and surface temperature and its trends (Gloersen and Campbell, 1988). Particularly interesting is their suggestion of decreases in the global extent of ice-covered ocean, and also of open water areas for 1978–87, in contrast to the absence of change in total ice area (Fig. 7). However, sensor drift cannot yet be ruled out. This type of data will continue to be available in the future via the Special Sensor Microwave Imager (SSM/I) system on satellites of the Defense Meteorological Satellite Program (Weaver et al., 1987).

6. CLIMATIC TRENDS AND THE CRYOSPHERIC EVIDENCE

In considering cryospheric evidence for climatic trends, care must be taken first to distinguish between quantities that respond rapidly to change, such as freshwater ice and seasonal snow cover, and those that are known to respond to climatic forcing with a long time lag, that is, large ice masses and permafrost thickness. Second, it must be recognized that the response to a given climate forcing may differ regionally in magnitude, or even
in sign, as well as in its timing. Most climatic anomalies show spatial patterns of positive and negative sign on subcontinental and smaller scales.

Records of surface air temperature from the late nineteenth century through the present show a warming trend beginning about 1880–1890. A continuing steady increase is apparent in the Southern Hemisphere, where the trend is nearly monotonic, as well as in low latitudes (Elsaesser et al., 1986; Jones et al., 1986b; Hansen and Lebedeff, 1987). In the Northern Hemisphere, however, a general warming beginning about 1890 culminated around 1940, with subsequent cooling until the mid 1960s, followed by a more recent rise to higher levels than before (Jones et al., 1986a; Hansen and Lebedeff, 1988). Despite assertions that these trends are consistent with the anticipated effects of increasing concentrations of CO₂ and other greenhouse gases, there remain some serious reservations. Model simulations for a CO₂ doubling predict an amplified temperature
response in polar regions. However, over the 1947–86 interval, warming was concentrated over northwestern North America, the Soviet Union and north Africa, with cooling over most of the Northern Hemisphere oceans, eastern North America, and western Europe (Jones et al., 1987). The pattern is one suggestive of circulation anomalies and air mass advection effects. In contrast, Jones et al. (1987) found widespread warming during 1947–86 in the Southern Hemisphere, with maxima off Antarctica.

Regional averages of precipitation are less subject to high spatial variability. Nevertheless, Bradley et al. (1987) demonstrate increases in mid-latitudes, but decreases in low latitudes of the Northern Hemisphere, over the last 30–40 years compared with the long-term average (1880–1940). This distribution is broadly consistent with model predictions for doubled CO₂ concentrations. In the Southern Hemisphere land areas, records both for mid-latitude and tropical zones spanning the last 100 years show increased annual precipitation since 1960 compared with 1921–60 (Diaz et al., 1989). They suggest that these trends may relate to stronger winds and recent increases in sea surface temperature.

In light of the noted recent trend of Northern Hemisphere temperatures, it is striking that the 20-year record of hemispheric snow-cover extent shows no apparent response, being characterized by irregular fluctuations. Local and regional trends (cf. Dewey, 1987; Robinson, 1987) show a not-unexpected pattern of both increases and decreases. The records of hemispheric sea-ice extent for both polar regions also fail to show any consistent signature analogous to the hemispheric temperature. However, on a regional scale, ice responses to temperature trends are clear. For example, the duration of the ice season along the Finnish coast has decreased 20–30 days/century over the period of record, 1830–1984 (Leppäranta and Seina, 1985). Also running means of lake-ice duration in southern Finland show a close parallelism with air temperatures at Helsinki since the
mid-nineteenth century (Palecki and Barry, 1986). That work, and other studies (Anderson, 1987) suggest temperature changes of ±1°C correspond to a ±5 day change in ice season length. Such findings confirm the sensitivity of cryospheric indicators that have a short response time.

Alpine glaciers and permafrost adjust more slowly, and in a complex, integrated fashion. Both provide clear evidence of regional and even hemispheric responses to decadal to century time-scale climatic fluctuations. Evidence for worldwide glacier retreat from the mid-or-late nineteenth century maximal positions is abundant (Grove, 1988), but as shown earlier, its climatic interpretations are complicated. Glaciers need to be categorized according to their reaction time to climatic variables, and according to their climatic setting, before changes in their length and mass balance can be interpreted climatically. Ground temperature profiles in the continuous permafrost zone are similarly indicative of slow trends, but their interpretation requires knowledge of the ground conditions, as well as of any natural or human-induced changes in vegetation cover.

Much analytical work remains to be done to improve our understanding of the varying responses of snow and ice cover variables to changes in climate in order to use such indicators for purposes of long-term climate monitoring, or to assess likely cryospheric responses to future changes in the climate system.

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Observations of Long-Term Tide-Gauge Records for Indications of Accelerated Sea-Level Rise

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ABSTRACT. Long-term tide-gauge records have been examined for indications of accelerated sea-level rise. An initial evaluation of 21 records was made, using least-squares linear regression. Four of the longest records were selected for more formal statistical analysis. The regional-mean sea-level curve for west-central Europe displays an upswing after 1900. However, individual stations vary considerably over short distances. Results from other regions are less conclusive. Application of univariate and multivariate techniques to the four longest records yields strong evidence for the presence of a non-global, nonlinear component. However, the three longest European records show some common nonlinear features, implying the presence of a regional component. There is some weak statistical evidence for a common changepoint around 1895 in the long-term European records, particularly for Amsterdam and Brest.

1. INTRODUCTION

The global-mean sea-level (SL) rise, derived from tide-gauge records since 1880, ranges between 1–2 mm/yr (Warrick and Oerlemans, 1990). Spatially and temporally coherent increases are observed (Gornitz and Lebedeff, 1987; Barnett, 1984, 1988) in spite of a high level of noise caused by vertical land movements, ocean-atmosphere processes and incomplete data coverage (Gornitz, 1990). The rise in SL is compatible with estimates of alpine glacier melting (Meier, 1984), and thermal
expansion of the upper layers of the oceans (Gornitz et al., 1982; Wigley and Raper, 1987). The global SL trend of the last 100 years could therefore represent a eustatic change associated with the climate warming over this period (Hansen and Lebedeff, 1987; Jones et al., 1986).

Evidence for an accelerated rise could provide further support for a relation between the recent SL trends and global warming. There is some suggestion that the SL trend has accelerated within the last 50 years (Barnett, 1984, 1988; Gornitz and Lebedeff, 1987). On the other hand, the regional responses are quite variable. While sea levels in North America rose faster in the last 50 years relative to the preceding 50 years, other regions, such as Europe, Australia and the Pacific, exhibited the reverse behavior (Gornitz and Lebedeff, 1987). Barnett (1988) found a more rapid rise in SL within the last 50 years in 4 out of 7 geographic regions examined. Thus, although the global and some regional SL trends appear to have accelerated within the last 50 years, the overall picture is far from uniform, and the significance is open to question.

It can be argued that 50 years may be too short an interval to detect a meaningful increase in the SL trend. The acceleration, if any, may have begun earlier during the 19th century. Some support for this view comes from examination of the two longest tide-gauge records available, those of Stockholm and Amsterdam, which go back at least 200 years. Although Stockholm is rising, due to ongoing glacial rebound, and Amsterdam is sinking, both stations (corrected for isostatic factors) showed comparable increases in sea-level change, beginning around 1850. Similar increases were noted for several long-term German records as well (Mörner, 1973). The change in the relative SL curve of Stockholm, between 1774–1884 and 1885-1984 was found to be $+1.01 \pm 0.30$ mm/yr (Ekman, 1988), close to that of the estimated eustatic rise.

This paper undertakes an exploratory statistical analysis of several long-term tide-gauge records of relative sea level. Most studies take the approach that the analysis is limited by the period over which a near-global data set is available. Here, by contrast, the available record length is maximized for a correspondingly smaller set of stations. Although substantial interregional variability exists (e.g., Barnett, 1984, 1988), sea-level curves are frequently coherent and in phase over large segments of the coast (Hicks and Crosby, 1974; Chelton and Enfield, 1986; Sturges, 1987), so that a limited number of stations may be representative of much wider areas.

The study can be divided into two parts: an initial evaluation of a number of records, using least-squares linear regression, and a more formal statistical analysis of four of the longest-record stations. We seek to determine whether these long-term records show an increase in the rate of SL rise and, if so, whether a common onset of change can be established.

Nearly all of the tide-gauge stations examined have time series which begin prior to 1880 (Table 1). However, only three stations have records that start before 1810, and another five before 1860. The longest records for Asia and Australia are only 100 years. Other regions, such as South America and Africa, have no long-term records. The geographic distribution is heavily weighted toward Europe. Therefore, the results obtained from this study are more valid for Europe, and to a lesser extent North America, than for the world as a whole.

2. METHODS

2.1. Least-Squares Linear Regression

The relative sea-level curve for any station represents a composite of eustatic, isostatic and tectonic components. During the geologically short period of the last few centuries, vertical land movements due to glacio-isostatic and tectonic processes can be expected to
Table 1. Linear trends of annual mean sea level, mm/yr.

<table>
<thead>
<tr>
<th>Region</th>
<th>Station</th>
<th>Start and End Dates</th>
<th>Entire Series</th>
<th>Series Since 1880</th>
<th>Late Holocene Trend</th>
</tr>
</thead>
<tbody>
<tr>
<td>Scandinavia</td>
<td>Stockholm, Sweden*</td>
<td>1774–1984</td>
<td>-4.23 ± 0.08**</td>
<td>-3.85 ± 0.18</td>
<td>-6.39</td>
</tr>
<tr>
<td></td>
<td>Nedre Sodertalje, Sweden</td>
<td>1869–1965</td>
<td>-3.38 ± 0.20</td>
<td>-3.38 ± 0.23</td>
<td>-6.39</td>
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<td>Helsinki, Finland</td>
<td>1879–1976</td>
<td>-2.88 ± 0.22</td>
<td>-2.88 ± 0.22</td>
<td>-3.75</td>
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<tr>
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<td>Lyokki, Finland</td>
<td>1858–1935</td>
<td>-5.35 ± 0.27</td>
<td>5.04 ± 0.37</td>
<td>-3.75</td>
</tr>
<tr>
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<td>Lypertti, Finland</td>
<td>1858–1935</td>
<td>-4.98 ± 0.25</td>
<td>-4.68 ± 0.36</td>
<td>-3.75</td>
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<tr>
<td></td>
<td>Regional mean</td>
<td></td>
<td>-3.83 ± 0.09</td>
<td>-3.62 ± 0.19</td>
<td></td>
</tr>
<tr>
<td>West Central Europe</td>
<td>Brest, France*</td>
<td>1807–1970</td>
<td>0.91 ± 0.07</td>
<td>1.59 ± 0.15</td>
<td>0.15</td>
</tr>
<tr>
<td></td>
<td>Aberdeen, Scotland</td>
<td>1862–1965</td>
<td>0.57 ± 0.10</td>
<td>1.02 ± 0.11</td>
<td>N.A.</td>
</tr>
<tr>
<td></td>
<td>Amsterdam, Holland*</td>
<td>1700–1940</td>
<td>0.55 ± 0.03</td>
<td>1.68 ± 0.25</td>
<td>0.82</td>
</tr>
<tr>
<td></td>
<td>Vlissingen, Holland</td>
<td>1862–1985</td>
<td>1.12 ± 0.13</td>
<td>1.94 ± 0.12</td>
<td>0.82</td>
</tr>
<tr>
<td></td>
<td>Hoek van Holland</td>
<td>1864–1985</td>
<td>2.35 ± 0.09</td>
<td>2.24 ± 0.11</td>
<td>0.82</td>
</tr>
<tr>
<td></td>
<td>Ijmuider, Holland</td>
<td>1872–1985</td>
<td>1.41 ± 0.15</td>
<td>1.80 ± 0.14</td>
<td>0.82</td>
</tr>
<tr>
<td></td>
<td>Den Helder, Holland</td>
<td>1832–1985</td>
<td>1.26 ± 0.60</td>
<td>1.37 ± 0.10</td>
<td>0.82</td>
</tr>
<tr>
<td></td>
<td>West Terschelling, Holland</td>
<td>1887–1985</td>
<td>1.26 ± 0.12</td>
<td>1.26 ± 0.12</td>
<td>0.82</td>
</tr>
<tr>
<td></td>
<td>Harlingen, Holland</td>
<td>1865–1985</td>
<td>1.43 ± 0.10</td>
<td>1.32 ± 0.11</td>
<td>0.82</td>
</tr>
<tr>
<td></td>
<td>Delfzijl, Holland</td>
<td>1865–1985</td>
<td>1.54 ± 0.11</td>
<td>1.85 ± 0.12</td>
<td>0.89</td>
</tr>
<tr>
<td></td>
<td>Regional mean</td>
<td></td>
<td>0.31 ± 0.03</td>
<td>1.42 ± 0.07</td>
<td></td>
</tr>
<tr>
<td>North America</td>
<td>New York City, U.S.A.*</td>
<td>1856–1986</td>
<td>2.74 ± 0.07</td>
<td>2.90 ± 0.11</td>
<td>2.17</td>
</tr>
<tr>
<td></td>
<td>San Francisco, U.S.A.</td>
<td>1855–1986</td>
<td>1.35 ± 1.10</td>
<td>1.47 ± 0.13</td>
<td>1.32</td>
</tr>
<tr>
<td></td>
<td>Mean</td>
<td></td>
<td>2.05</td>
<td>2.19</td>
<td></td>
</tr>
<tr>
<td>Australasia</td>
<td>Aden</td>
<td>1879–1969</td>
<td>1.20 ± 0.10</td>
<td>1.20 ± 0.10</td>
<td>N.A.</td>
</tr>
<tr>
<td></td>
<td>Bombay, India</td>
<td>1879–1962</td>
<td>1.29 ± 0.12</td>
<td>1.32 ± 0.13</td>
<td>N.A.</td>
</tr>
<tr>
<td></td>
<td>Kidderpore, India</td>
<td>1881–1929</td>
<td>-4.65 ± 1.94</td>
<td>-4.65 ± 1.94</td>
<td>N.A.</td>
</tr>
<tr>
<td></td>
<td>Sydney, Australia</td>
<td>1897–1982</td>
<td>0.63 ± 0.13</td>
<td>0.63 ± 0.13</td>
<td>~0</td>
</tr>
<tr>
<td></td>
<td>Mean</td>
<td></td>
<td>-0.38 ± 2.90</td>
<td>-0.38 ± 2.90</td>
<td></td>
</tr>
</tbody>
</table>

* Stations for which additional statistical tests have been made.
** Standard error of trend (Hicks and Crosby, 1974).
N.A. Data not available.
have remained constant. Therefore, any sudden, marked change in the SL curve can be attributed to a change in the eustatic component. In this section we are primarily interested in changes in trends between different time intervals for the same station; therefore, the relative SL curves have not been corrected for land motions.

Since the relative SL curves, taken from various sources, have not been referenced to a common datum, they are normalized by setting the average for the period 1951–1970 equal to zero. Linear trends are calculated by simple least-squares regression on both the raw data and on 5-year running means for: (1) the entire time series, (2) the record since 1880, and (3) 30-year intervals (1979–1950, 1949–1920, 1919–1890, etc). At both tails, the few extra years falling outside the nearest 30-year interval have been added in. Results are summarized in tables and plots.

The relative sea-level (RSL) curve for any station can be decomposed into a global component common to all records (the eustatic component), and a local component, specific to that particular record. The global and local components may each exhibit both linear and nonlinear behavior. In practice, without further information, it is impossible to separate these various components. Therefore, the approach taken is to identify the nonlinear features common to all records, and also to test for a common changepoint. The records examined in greater detail for the period 1801–1984 are those for Amsterdam, Brest, Stockholm and New York. These four records are examined both individually and jointly.

2.2. Univariate Analysis

The nonlinear component, \( X_{it} \), of the relative sea-level (RSL) curve, is obtained by subtracting the linear component, as obtained by ordinary least-squares regression, from the sea-level curve, \( Y_{it} \). \( X_{it} \) is referred to as the residual relative sea-level curve,

\[
X_{it} = Y_{it} - (c_i + d_i t)
\]

where record \( i = 1, \ldots, p \), time \( t = 1, \ldots, n \), and \( c_i \) and \( d_i \) are the estimated intercept and slope of the linear trend for record \( i \).

\[
X_{it} = G_{it} + L_{it} + e_{it}
\]

where \( G_{i} \) is the nonlinear global component, \( L_{it} \) is the nonlinear local component and \( e_{it} \) is a random error term. The nonlinear trends in the residual RSL curves are estimated by locally weighted regression (Cleveland, 1979).

The statistical significance of the residual trends is tested, using the null hypothesis that \( G_{i} + L_{it} = 0 \), or, in other words, that any residual is random noise (\( e_{it} \)). Under normal theory assumptions, the statistic

\[
F = \frac{(RSS_0 - RSS)/(r - 1)}{RSS/(n - r)}
\]

has an F distribution, with \((r-1)\) and \((n-r)\) degrees of freedom under the null hypothesis, where \( RSS_0 \) is the residual sum of squares in the null model, \( RSS \) is the residual sum of squares for fitting the trend by locally weighted regression, and \( r \) is the equivalent number of parameters used in the locally weighted regression.

An assumption underlying statistical inference based on locally weighted regression is that the errors \( e_{it} \) are serially independent. For example, if the errors are positively serially correlated, then the F test described above will not be strictly valid and will tend
to give a significant result more often that it should. To check for serial correlation, the autocorrelation functions for the residuals around the estimated trends were obtained.

2.3. Multivariate Analysis

The joint behavior of the four stations is now assessed. The hypothesis is tested that all four residual RSL records have the same nonlinear trend, and that differences in the estimated trends arise from random error. The null model

\[ X_{it} = G_t + e_{it}, \quad i = 1, ..., p, \quad t = 1, ..., n \]  

is tested against the alternative

\[ X_{it} = G_t + L_{it} + e_{it}, \quad i = 1, ..., p, \quad t = 1, ..., n \]  

The weighted-residual sum of squares for the fitted models under the null and alternative hypotheses are compared. The choice of weights depends on differences in variances and any cross-correlation between the records.

The estimated trend under Eq. (4) is found by applying locally weighted regression to a linear combination of observations of residual relative sea level in each period. The combination is

\[ X_t = \sum_{i=1}^{p} W_{it} X_{it}, \quad t = 1, ..., n \]  

The weight \( W_{it} \) is zero if the observations for record \( i \) is missing in period \( t \). The remaining weights are found by generalized least-squares regression (Seber, 1977). For period \( t \)

\[ w = (I' S^{-1} I)^{-1} I' S^{-1} \]  

where \( w \) is the row vector of weights, \( I \) is a column vector of 1’s, and \( S \) is the estimated covariance matrix between the available records. The dimensions of \( w \), \( I \) and \( S \) are 1-by-\( q \), \( q \)-by-1, and \( q \)-by-\( q \), where \( q \) is the number of available observations in period \( t \).

Finally, a particular form of common nonlinear behavior is tested, namely, a systematic acceleration in relative sea level occurring at the same time in all records. A simple model of such an acceleration is the two-phase linear regression model (Hinkley, 1969; Solow, 1987). Under this model

\[ Y_{it} = a_{oi} + b_{oi} + e_{it}, \quad t = 1, ..., r_i \]

\[ a_{ii} + b_{ii} + e_{it}, \quad t = r_i + 1, ..., n_i \]

where

\[ c_i = (a_{oi} - a_{ii})/(b_{ii} - b_{oi}) \]

is the changepoint for record \( i \). To ensure continuity, the changepoint should fall between the last observation in the first phase and the first observation in the second phase. Without this restriction the model will have a discontinuity at the changepoint.

The null hypothesis that the records have a common changepoint (i.e., that \( c_i = c \), \( i = 1, ..., p \)) is tested against the alternative that at least one changepoint is different from the rest. Under the null hypothesis, the pre- and post-change slopes do not have to be
the same for all records. The null model can be fit by choosing the value of \( c \) that minimizes the weighted residual sum of squares for fitting model (8) with \( c_i = c \) to each record. The alternative model can be fit by choosing the values of \( c_i \) (i = 1, ..., p) that minimize the residual sum of squares for fitting model (8) to each record separately.

3. RESULTS

3.1. Linear Regression

Table 1 summarizes the least-squares regression slopes over the entire period of observation, and since 1880, for the tide-gauge stations examined in this study. Slopes for both the raw data and the 5-year running means are very similar, thus only the former are tabulated. Sea-level is rising in most regions, except for Scandinavia, which is still undergoing glacial isostatic rebound. Late Holocene SL trends, where available, are listed for comparison. On average, the latter are 0.5 mm/yr less than the trends of the last few centuries, for the corresponding localities. For this set of stations, the late Holocene trends are 0.9 mm/yr less than those from 1880 to the present.

For many stations, the SL trend since 1880 is somewhat greater than that of the entire series. The increase is especially marked for several of the longer European records such as Stockholm, Brest, Amsterdam and Aberdeen, and lends some support to the introductory observations.

Figures 1-14 present plots of selected regional-mean and individual smoothed (5-year running mean) sea-level curves, with slopes drawn for the entire time series, and for 30-year intervals. The regional-mean sea-level curve for west-central Europe shows a relatively flat slope until around 1860, a minimum around 1890, and a sharp rise after 1900 (Fig. 1). The regional trend for the entire time series is \( 0.31 \pm 0.03 \) mm/yr, which increases to \( 1.42 \pm 0.07 \) mm/yr after 1880 (Table 1 and Fig. 1). Trends on 30-year intervals are < 1.2 mm/yr before ~1890, and > 1.4 mm/yr thereafter. A steep rise in the slope occurs after 1860 in Amsterdam and after 1890 in Brest (Figs. 2 and 5). Individual Dutch stations fall into two groups, characterized by the RSL records for Vlissingen (Fig. 3) and Den Helder (Fig. 4), respectively. The former show a marked minimum around 1890, the latter a relatively constant linear trend. Aberdeen (Fig. 6) also shows a weak minimum around 1890.

Sea levels are falling in Scandinavia (Fig. 7). The regional mean is \( -3.83 \pm 0.09 \) mm/yr (Table 1). There is no marked change in the 30-year slopes over time, except for Stockholm, which shows less negative trends during the last 100 years than before (Fig. 8; see also Ekman, 1987).

Sea-level slopes for the U.S. stations are steeper after ~ 1900 than before (Figs. 9 and 10). However, San Francisco displays a temporary low shortly after 1890, whereas New York has a data gap around 1890. San Francisco is located on the seismically active San Andreas fault system; therefore, anomalies in the SL curve could be related to local tectonic movements. However, the minimum at ~ 1890 is also observed in several west European stations, and may be part of a longer-period oscillation (Sturges, 1987).

The Australasian stations have shorter records. They display local short-term variability, such as a minimum between 1920-1950 in the Sydney record, and a rise after 1930 in Aden (Figs. 11 and 12). Bombay also shows a rise after 1930.
Figure 1. Regional-mean sea-level curve, west-central Europe. The data are smoothed, using the 5-year running means. The least-squares linear trend for the entire record is shown, as well as trends for successive 30-year intervals. The baseline is obtained by setting the average for the period 1951-1920 equal to zero.

Figure 2. Relative sea-level curve, Amsterdam. Other information as in Fig. 1.
Figure 3. Relative sea-level curve, Vlissingen.

Figure 4. Relative sea-level curve, Den Helder.
Figure 5. Relative sea-level curve, Brest.

Figure 6. Relative sea-level curve, Aberdeen.
Figure 7. Regional-mean sea-level curve, Scandinavia.

Figure 8. Relative sea-level curve, Stockholm.
Figure 9. Relative sea-level curve, New York.

Figure 10. Relative sea-level curve, San Francisco.
Figure 1'. Relative sea-level curve, Sydney.

Figure 12. Relative sea-level curve, Aden.
3.2. Univariate Analysis

Residual relative sea-level curves for Amsterdam, Brest, New York and Stockholm are plotted in Figs. 13-16, as given by Eq. (1), and the trends in these records determined by locally weighted regression (Cleveland, 1979). An inspection of Figs. 13-16 suggests that the three longest European records are similar: a decrease in residual relative sea level prior to 1900, followed by an increase in residual relative sea level. The New York record is somewhat different. The lack of observations for New York prior to 1856 makes this comparison difficult.

The statistical significance of the individual estimated trends shown in Figs. 13-16 was determined, using the F test given by Eq. (3). The results are given in Table 2. The p-value indicates the level of significance at which the null hypotheses of no trend would just be rejected. Thus, three of the estimated trends are highly significant and that for Stockholm is marginally significant.

Table 2. Significance of estimated nonlinear trends.

<table>
<thead>
<tr>
<th>Station</th>
<th>r</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amsterdam</td>
<td>8.0</td>
<td>0.000</td>
</tr>
<tr>
<td>Brest</td>
<td>8.9</td>
<td>0.000</td>
</tr>
<tr>
<td>New York</td>
<td>7.6</td>
<td>0.001</td>
</tr>
<tr>
<td>Stockholm</td>
<td>10.3</td>
<td>0.065</td>
</tr>
</tbody>
</table>

An estimate of the variance of $e_{it}$ is given by

$$s_t^2 = \frac{RSS_i}{(n_i - r_i)} ,$$

where $RSS_i$ is the residual sum of squares for fitting the trend in record $i$ by locally weighted regression. It is assumed that the variance of $e_{it}$ does not depend on $t$. Figures 13-16 show no evidence of non-constant variability. The estimated standard deviations are shown in Table 3. The estimated standard deviations are quite similar, except for Stockholm, which is twice as variable as the other records. As the power of the F test is inversely related to the variance of $e_{it}$, this may explain in part the relative lack of significance of the estimated trend for Stockholm.

Table 3. Standard deviations of non-linear trends.

<table>
<thead>
<tr>
<th>Station</th>
<th>Estimated standard deviation (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amsterdam</td>
<td>25.3</td>
</tr>
<tr>
<td>Brest</td>
<td>32.2</td>
</tr>
<tr>
<td>New York</td>
<td>25.7</td>
</tr>
<tr>
<td>Stockholm</td>
<td>57.6</td>
</tr>
</tbody>
</table>

As mentioned above, the F test is sensitive to the presence of serial correlation. Therefore, the records were tested for serial correlation by calculating autocorrelation functions for residuals around the estimated trends of Figs. 13-16. The largest estimated
Figure 13. Residual relative sea level, Amsterdam. The curve represents the nonlinear trend estimated by locally weighted regression (see text). Normalized sea-level scale is in cm.

Figure 14. Residual relative sea-level record, Brest.
Figure 15. Residual relative sea-level record, Stockholm. Note the difference in scale from the other stations, due to greater variability.

Figure 16. Residual relative sea-level record, New York.
lag-one correlation was 0.26 for Brest. The presence of serial correlation of this magnitude is not a problem, particularly since the number of observations in each record is large. This weak serial correlation disappears if locally weighted regression is applied with a smaller span (i.e., allowing more variability in the estimate of trend). Thus, a tradeoff exists between complexity in the form of the trend and complexity in the structure of the error process.

3.3. Multivariate Analysis

The cross-correlation functions between all pairs of records are estimated, using the residuals from the estimated trends shown in Figs. 13-16. The largest estimated lag-zero cross-correlation was 0.34 for Amsterdam and Stockholm. The estimated lag-zero cross-correlation for Amsterdam and Brest was 0.24.

The weights given by Eq. (7) used in deriving the linear combinations of residual SL curves of Eq. (6), if all four records are used together, are 0.37, 0.24, 0.37 and 0.02 for Amsterdam, Brest, New York and Stockholm, respectively. Note that the weight for Stockholm is small. This is a consequence of the large variance for Stockholm and of the cross-correlation between Amsterdam and Stockholm. The effect due to correlation is rather small. For example, if correlation is ignored, the weights are 0.36, 0.22, 0.35 and 0.07.

The linear combination found in this way is shown in Fig. 17, along with the trend estimated by locally weighted regression. The estimated trend, not surprisingly, resembles the general trends shown in Figs. 13-16; that is, it shows a decrease prior to 1900, followed by an increase.

Figure 17. Linear combination of the residual relative sea-level curves, shown in Figs. 13-16, with the trend (line) by locally weighted regression.
The statistical significance of this trend was tested in the same way as for the univariate analysis. The estimated trend in Fig. 17 used approximately 11 parameters. The F test gave a highly significant result (p-value = 0.003).

To test Eq. (4) against Eq. (5), the following statistic is used

\[(r_o^T K^{-1} r_o) - (r^T k^{-1} r)\]

where \(r_o\) is the vector of residuals from the estimated trend shown in Fig. 17, \(r\) is the vector of residuals from the individual estimated trends shown in Figs. 13-16, and \(K\) is the covariance matrix among these residuals. Note that the dimensions of \(r_o\) and \(r\) are the same (i.e., both equal to the total number of observations). Note, too, that because no significant serial correlation was found, \(K\) has block-diagonal structure. Under the null hypotheses, the statistic (10) has an approximate chi-squared distribution with degrees of freedom given by the difference in the number of parameters used in the individual fits and the number of parameters used in the common fit. Based on this approximation, the null hypothesis of a common global nonlinear trend was strongly rejected (p-value < 0.001). In summary, the four records of residual relative sea level examined do not have the same trend.

We now test for a particular form of nonlinear behavior, namely, a systematic acceleration in relative sea level occurring at the same time in all records, using the two-phase linear-regression model (Hinkley, 1969; Solow, 1987; see Eq. 8). The hypothesis that all records have the same changepoint, \(c_i\), is examined.

The estimated individual values of \(c_i\) are listed in Table 4. The estimated common value of \(c\) is 1892. The test statistic (10) has a value of 18.0 on 4 degrees of freedom, which indicates that the four records do not have a common changepoint. When the test was repeated for the three European records only, an estimated common changepoint occurred at 1895. The test statistic has a value of 7.4 on 3 degrees of freedom. This value is only marginally significant (p = 0.061). Thus, the statistical test provides weak evidence that the European records have a common changepoint. However, if only Amsterdam and Brest are considered, there would be no significant evidence against the null hypothesis of a common changepoint.

\begin{table}
\centering
\begin{tabular}{ll}
\hline
Station & Estimated changepoint, \(c_i\) \\
\hline
Amsterdam & 1894 \\
Brest & 1897 \\
New York & 1928 \\
Stockholm & 1847 \\
\hline
\end{tabular}
\caption{Individual changepoints of long-period stations.}
\end{table}

4. DISCUSSION

Sea level appears to be rising more rapidly in the last few hundred years, and especially after 1880, than during the last few thousand years. For those stations with matched data, the average difference in SL trend between the entire tide-gauge record and the late Holocene is around 0.5 mm/yr. The average difference increases to around 0.9 mm/yr if the recent series begins at 1880. The latter value, although based on a very
limited sample size, is close to the systematic offset of 1 mm/yr between European tide-
gauge SL trends and late Holocene SL trends (Shennan and Woodworth, manuscript in
preparation), and also to the lower range of estimates of the recent eustatic sea level rise
(as summarized in Warrick and Oerlemans, 1990). The questions considered here are: (1)
do we detect an acceleration in the rate of SL rise during the period represented by the
long time-series tide-gauge record and, if so, (2) can we define a common date when the
acceleration began?

When the tide-gauge records are examined more closely, the results are less than
clear cut. First of all, not all of the European records have similar curves. For example, in a
country as small as the Netherlands, stations can be divided into two groups, one of which
shows essentially linear behavior (Fig. 4) and the other with a marked minimum around
1890 (Fig. 3), as indicated by changes in successive 30-year trends over time. These
two groups do not cluster geographically, nor are they related to any major geological
discontinuity. The Netherlands are located on a deltaic and lagoonal plain at the mouth
of the Rhine River and its distributaries. Late Holocene subsidence, as determined from
C$^{14}$ dating of paleo-sea-level indicators in sediments is 0.89 mm/yr for northern Holland
and 0.82 mm/yr for west and central Holland (based on data in Shennan, 1987). There
is no reason to suspect any sudden, recent changes in the geologic regime. If the SL
anomaly were caused, on the other hand, by some shift in coupled oceanic-atmospheric
phenomena, nearby stations should all have similar SL curves (see, for example Hicks
and Crosby, 1974; Chelton and Enfield, 1986).

Ten North Sea German tide-gauge stations, north of the Netherlands, show a coherent
pattern of rising sea levels, with an average trend of 1.2 mm/yr since 1855. Only Cuxhaven
shows a small minimum in the mean low-water record around 1890. When linear trends
for these stations are calculated, using a sliding window of 50 years, and plotted against
time, there is no consistent pattern in the variation of the 50-year trend over time. Stable,
rising and falling trends are all observed (Jensen, 1984).

Among the Scandinavian stations, only Stockholm (which, however, has a much
longer record than the rest) shows some indication of a decrease in the rate of apparent
sea-level fall (i.e., an increase in the rate of SL rise, assuming constant land uplift over the
last 200 years, Fig. 8). The other Scandinavian stations are characterized by predominantly
linear behavior over the last 120 years.

Statistical tests established significant nonlinear components in the SL records of the
four longest time series. (Nonlinear behavior can be seen, more qualitatively, in Figs. 2, 3,
5, 6, 9-12.) Evidence is weak for a common nonlinear component in all four of the records.
But the three longest European records show some similarities: a decrease in residual sea
level prior to 1900, followed by an increase, whereas the increase in New York begins
much later (1920s). The existence of a common changepoint among the three European
stations at around 1895 is only marginally significant, in a statistical sense. However, the
fact that all three records are changing in the same direction, namely, toward an increase,
has not been taken into account in the two-sided test. Thus, the significance may be more
meaningful than suggested by the formal result. The indication of a common changepoint
for Amsterdam and Brest is more convincing.

What are some possible explanations for the absence of strong evidence of an ac-
celeration in SL rise over the time of the instrumental record?

(1) The assumptions underlying the two-phase regression model are not all met. In par-
ticular, the two-phase regression model assumes a fairly sharp and abrupt change. It is
unlikely that the rate of SL rise should change suddenly. If the transition period is short
relative to the record length, the two-phase regression model will still work well. On
the other hand, if the transition period is long, the model will tend to identify a changepoint toward the middle rather than the beginning of the transition period. Finally, if the transition period begins toward the end of the record, the changepoint may be missed altogether (Solow, 1987).

In addition, the two-phase regression model, as applied here, assumes, at most, a single change, although, in principle, it can be extended to account for two or more changepoints.

Another problem is the possible presence of weak periodicity in residual sea-level curves. The sensitivity of the two-phase model to periodicity depends on both the period and amplitude of the periodic component. The two-phase model will work, if the period is short relative to the record length and if the amplitude is small relative to the signal (Solow, 1987). Tests for periodicity in the records are in progress.

(2) The acceleration may have begun much earlier, over 200 years ago, beyond the period of tide-gauge records. If so, this would place the recent sea-level history at odds with other indications of recent climate change. Land and ocean temperature data record a global warming trend since at least 1861 (Jones et al., 1986; Hansen and Lebedeff, 1987.) Mountain glaciers have been retreating, worldwide, since the mid- to late-19th century (Grove, 1988). North American tree-ring records indicate a general warming trend since 1850 (Jacob et al., 1985; Jacoby and D’Arrigo, 1987). The boreal forest-tundra transition in North America has shifted north during the 19th century (Ball, 1986). Cadmium/calcium ratios in Pacific corals have been correlated with sea surface temperatures. When SSTs are high, Cd/Ca ratios in the coral are low, and vice versa. A plot of Cd/Ca ratios in Pacific corals from 1600 to the present shows high and variable Cd/Ca ratios until around 1850, when the ratios become low and relatively constant. The Cd/Ca record correlates well with $\delta^{18}O$ from an ice core in the Peruvian Andes, suggesting a warming since the 1850s (Shen, 1988).

There is some suggestion that current rates of sea-level rise may have remained fairly constant over the last few centuries. Dating of geologically recent peats in Chesapeake Bay (Froomer, 1980; Leatherman, 1989, priv. comm.) and historical records of maximum flood levels on the German North Sea coast (Rohde, 1980) do not indicate a substantial change in the rate of SL rise during this period. However, additional data are needed to substantiate this finding and to resolve the apparent contradiction with the warming data.

5. CONCLUSIONS

The results of this exploratory analysis can be summarized as follows. Least-squares regression slopes on SL data show that sea-level trends of the last few centuries, and particularly since 1880, are higher, on average, than the corresponding late Holocene trends. Thus, an acceleration in the rate of sea-level rise began at some point within the last 1000 years. Limited data from opposite sides of the Atlantic Ocean suggest that current rates of SL rise may not have changed substantially in the last 200-300 years (Froomer, 1980; Rohde, 1980). However, examination of long-term tide-gauge data indicates a more complex situation. The regional-mean sea-level curve for west-central Europe shows a sharp rise after 1900 (Fig. 1). Yet, individual stations exhibit great variability in the pattern of sea-level change over relatively small distances (e.g., Vlissingen versus Den Helder; Figs. 3 and 4). Except for Stockholm (Fig. 8), the Scandinavian stations do not reveal a systematic decrease in the rate of apparent SL fall (i.e., given constant uplift, an increase in the rate of SL rise). The two North American stations show more rapid SL rise in the 20th century than in the 19th, but the onset of change occurs at different times
Evidence from Australasian stations is less definitive, in part due to the relatively short record length.

The findings of the preliminary univariate and multivariate analysis are now summarized. First, there is strong evidence of nonlinear behavior in the records of the four longest-term stations. That is, with the possible exception of the record for Stockholm, a linear trend is inadequate to describe the behavior of relative sea level. Second, this nonlinear behavior is not identical in all four records. That is, all the nonlinear behavior cannot be attributed to a global component; the local components also exhibit some nonlinear behavior. However, the nonlinear components of the three longest European records appear to share some common features. This suggests the presence of a regional component in the relative sea-level curves. Third, if the two-phase linear model can be taken as representative of the gross nonlinear trends in relative sea level, then the long-term European records (and Amsterdam and Brest in particular) show evidence of a common changepoint around 1895. Although the test that was applied is two-sided, the fact that the estimated change is in the direction of more rapid change in relative sea-level lends further support to the hypothesis of a regional component. These findings are supported by Woodworth (1990) who finds a regionally coherent acceleration of around 0.4 mm/yr per century, over the last 2–3 centuries, in northwest Europe.

The statistical significance of these results is only approximate, given the small sample size. The apparent presence of a regional component in the relative sea level curves underlines the importance of sampling over widely spaced locations.

ACKNOWLEDGMENTS

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The Southern Oscillation and Northern Hemisphere Temperature Variability

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ABSTRACT. The Southern Oscillation (SO) is the best defined and understood mode of interannual climate variability. The extreme phases of the SO have been identified with global-scale variations in the atmosphere/ocean circulation system and with the modulation of monsoon precipitation on the global scale. While SO-related precipitation has been the subject of several studies, the magnitude of the SO-related temperature variability on the global scale has not been well documented. In this paper we provide an estimate of the SO-related temperature variability in the context of monitoring global warming related to the increase in greenhouse gases. This analysis suggested that traditional time series of hemispheric and global temperature anomalies for the calendar year may confuse interannual temperature variability associated with the SO and perceived "climate trend." Analyses based on calendar-year data are likely to split the effects of the SO-related temperature variability over two years. The Northern Hemisphere cold season (October through March) time series may be more appropriate to separate the SO-related effects on the hemispheric temperature from other modes of variability. Mean interannual temperature anomaly differences associated with the extremes of the SO are estimated to be 0.2°C for the October-to-March season in the Northern Hemisphere. In areas directly linked to the SO, the mean interannual differences amount to over 0.5°C. The SO cannot account for all the variability in the hemispheric times series of surface temperature estimates, but the SO signal must be properly accounted for if these time series are to be understood.

1. INTRODUCTION

The potential of greenhouse warming has focused considerable attention on estimates of global surface temperature. Time series of surface air temperatures (e.g., Jones et al., 1988) and sea surface temperatures (e.g., Folland and Parker, 1989) have been used as estimates of the global temperature in efforts to identify significant trends. The extraordinary sequence of above-normal temperatures during the 1980s has been particularly
noteworthy. Whether these recent global temperature anomalies are within the range of "normal" climate variability or whether they are the first harbingers of a significant warming trend has been the subject of lively discussions both within and outside of the scientific community (e.g., Wigley et al., 1989).

It may seem surprising, therefore, that very little has been published to define the magnitude and limits of "natural climate variability" in the context of the global warming problem. In this paper we describe a first-order attempt to estimate the temperature variability associated with the strongest-known mode of interannual variation, namely, the Southern Oscillation (SO). While some investigators have acknowledged that the SO may influence estimates of global surface temperature (Jones et al., 1988), there appear to be no readily available published quantitative estimates of the magnitude of the SO surface-temperature signal relative to the global estimate.

The extremes of the SO are characterized by global-scale shifts in atmospheric circulation features and with a marked, high-amplitude modulation of the equatorial Pacific sea surface temperature anomalies. Large areas with sea surface temperature anomalies greater than +3°C have been observed with the warm, El Niño/Southern Oscillation (ENSO) phase of the oscillation (Quiroz, 1983), while negative anomalies of comparable magnitude have been associated with the cold, or high-index phase of the oscillation (Ropelewski and Halpert, 1989). Since the SO tends to persist in either extreme for up to 12 months, SO-related sea surface temperature anomalies undoubtedly have a direct effect on the global surface temperature. A simple calculation indicates that the SO-related sea surface temperature anomalies can account for between 0.1°C and 0.2°C of the annual temperature anomaly for every 1°C of sea surface temperature anomaly. This estimate is based on the assumption that the effective SO-related surface temperature anomaly is confined to within 7.5° latitude of the equator.

Empirical studies have shown, however, that the SO-related temperature teleconnections can extend to higher latitudes, thus the simple estimate given above must be modified. In particular, there is a strong SO-related temperature pattern in northwestern North America, and a weaker pattern of opposite sign in the southeast United States (Ropelewski and Halpert, 1986a). In general, SO-related effects also appear at land surface stations within approximately 20° latitude of the equator, as well as in southern Africa and regions of Australia (Fig. 1).

2. DATA

Since the distribution of land surface data is sparse in the Southern Hemisphere, the discussion to follow is limited to the Northern Hemisphere land areas only. However, since the SO has such a pervasive influence on the few land stations available in the Southern Hemisphere (Fig. 1.), we expect the SO-temperature influence to be at least as great as, and likely to be even greater than, documented here for the Northern Hemisphere.

Land observations of surface temperature for the Northern Hemisphere were analyzed onto a 2° latitude by 2° longitude grid. This grid, area weighted, forms the basis of the hemispheric temperature time series discussed in this paper. No interpolation was performed and thus the time series shown here represent temperature anomalies only for those areas that contain data points. This is equivalent to assuming zero temperature anomaly where there is no data, and thus the time series will differ somewhat from those produced by several other investigators. The data are limited to currently active weather observation locations that regularly report monthly averages in the Monthly Climate Data for the World. Even though these data represent less of a hemispheric coverage than those presented in the more complete data set of Jones et al. (1988), the time series presented here and the Jones et al. (1988) time series have a correlation of 0.93 for the annual
Figure 1. The amplitudes and phase of ENSO composite ranked temperature plotted as vectors. The vectors are based on a 24-month harmonic fitted to ranked temperature composites for ENSO episodes. The phases and amplitudes of these vectors are indicated by the harmonic dial in the figure.

hemispheric temperature anomaly estimates over the 1951 to 1988 period. Thus, while we do not claim to present yet another time series of hemispheric temperature anomalies for the study of global climate trends, these data appear to be sufficient to provide an estimate of the magnitude of the SO-related temperature anomaly.

3. ANALYSIS

The annual temperature anomaly time series for the 1951 to 1988 period (Fig. 2) shows no apparent relationship between the extremes of the SO and the annual Northern Hemisphere temperature anomaly. The mean-annual hemispheric temperature anomalies for either extreme of the SO turns out to be nearly zero (Table 1). On the other hand, large temperature anomalies have been linked with ENSO over substantial portions of the global (Ropelewski and Halpert, 1986a,b). In fact, the temperature anomalies for the Northern Hemisphere in 1987 exhibit a classic ENSO-temperature pattern with large areas of 1°C to 2°C sea surface temperature anomaly in the Pacific basin, a large area of positive temperature anomalies over northwestern North America, and a smaller area of weak negative anomalies over the Gulf of Mexico (Fig. 3).

This seeming paradox can be resolved if we consider that ENSO-related temperature anomalies tend to reach a maximum during the latter half of the year, generally in association with the "mature" phase of the SO. This tendency is clearly indicated by the phases of the harmonic vectors (Fig. 1) which show a marked preference to center about the January of the year following a warm episode. A re-examination of the annual time series (Fig. 2) shows that all but two of the years following the nine ENSO in the series have relatively large positive temperature anomalies, but annual anomalies may be misleading.

The SO influence on Northern Hemisphere temperature anomaly will be masked for any annual average which splits the cold half of the year. Time series of hemispheric
temperature anomaly estimates for the October-through-March period (Fig. 4) clearly show the influence of the SO. The mean temperature anomalies associated with both the high-index and ENSO phases of the SO are on the order of 0.1°C (Table 1). Since high- and low-index episodes tend to occur in pairs, interannual mean hemispheric temperature fluctuations on the order of 0.2°C can be associated with the extremes of the SO during the cold half of the year. If the analysis is confined to the Northern Hemisphere winter only, the interannual variability resulting from swings in the SO can amount to 0.3°C. The magnitudes of these interannual differences are equivalent to about one standard deviation of temperature anomaly.

**Table 1.** Northern Hemisphere mean temperature anomalies for the high (cold) and low (ENSO) index phases of the Southern Oscillation.

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<tr>
<td>SO PHASE</td>
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<td>Low (warm)</td>
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4. DISCUSSION

Large-amplitude variations in the mean temperature anomalies occur for those areas that are related to the SO, either directly or through teleconnections (northwestern North America, the Gulf Coastal regions of the United States and Mexico, and within 20° latitude of the Equator). For these areas the mean SO-related interannual variability is greater than 0.5°C (Table 1). The SO influence will extend to larger areas for any of the gridded analyses, (e.g., Jones et al., 1989; Hansen and Lebedeff, 1987). Thus the potential influence of the SO on the gridded analysis of global temperature anomalies is likely to be even larger than indicated here.

Since the SO-temperature influence tends to span the calendar year, any analysis of the calendar-year annual temperature aliases the SO influence, that is, the SO influence is arbitrarily split between two calendar years. It is desirable to abandon calendar-year
time series in the analysis of the global temperature anomalies in favor of an averaging period more consistent with the temporal scales associated with the naturally modes of climate variability. If 12-month anomalies are desirable, then the period April through March, a Southern Oscillation “year,” or October through September, to catch the mature phase of ENSO, may be more appropriate periods than the calendar year, but it’s not clear that there is a scientific basis for presenting time series based on 12-month periods. Nonetheless, a time series of the October-through-September global temperature anomaly estimates (Fig. 5) clearly identifies positive temperature anomalies with the SO, including the extreme warmth of the two recent ENSO years, 1982–83 and 1986–87. Since a relatively large pool of ENSO-related warm water persisted in the central Pacific through February-March, 1988, a portion of the 1987 temperature anomaly should also be attributed to the most recent warm episode. The high-index phase of the SO can be associated with the cold anomalies of 1955, 1964, 1970, 1973 and 1975. Thus, both extremes of the SO need to be considered in the interpretation of global and hemispheric temperature-anomaly time series.

We have confined our discussion to the modern era, in part because the SO extremes since 1950 can be clearly identified. It is the Pacific Ocean basin-scale episodes that are most important in the context of hemispheric and global temperature analysis. In the early part of the record ENSO's have been identified almost solely on the basis of records associated with the traditional El Niño area of the eastern Pacific. Desser and Wallace (1987) point out that there is not always a one-to-one relationship between the local El Niños and the basic-scale warmings associated with major swings in the SO. For the period from 1882 to the present, indices based on the pressure difference between Tahiti and Darwin may be used in defining the SO phase, for example, Ropeleowski and Jones (1987). Earlier in the record it may be difficult to identify, and thus remove, the SO portion of interannual variability from the time series of hemispheric and global temperature estimates.
In summary, we have noted that the SO is the best-documented mode of interannual climate variability. Its effect on the time series of global and hemispheric temperature estimates must be understood and quantified if these time series are to be useful for the identification of climate trends. This brief paper outlines a first-order estimate of the mean magnitude of SO influence on estimates of global temperature anomalies. The SO influence for any given warm or cold episode will be a function of the evolution of individual episode and also a function of the episode’s magnitude. For the purposes of monitoring the global temperature, a measure of the SO strength must be related to the extent and magnitude of the sea surface temperature anomaly associated with the extreme of the SO. It is clear that the effects of the SO cannot be ignored in attempts to analyze and interpret time series of hemispheric and global temperature anomaly estimates.

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Recent Climate Changes in the Northern Hemisphere

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ABSTRACT. The consistency of analyzed changes in surface wind stress, sea level pressures and surface temperatures between 1980–86 and previous periods indicates the reality of statistically significant and substantial climate changes in the Northern Hemisphere, especially over the North Pacific, on decadal time scales. Cooling in North Pacific sea surface temperatures and warming along the west coast of North America and Alaska are ascribed mainly to the changes in thermal advection associated with a deeper and more extensive Aleutian Low.

1. INTRODUCTION

During the course of computing and evaluating a new surface wind stress climatology over the global oceans (Trenberth et al., 1989), a comprehensive comparison between the new results and those of Hellerman and Rosenstein (1983) (henceforth H&R) was carried out and revealed some rather startling differences in several regions. The H&R climatology was based upon analysis of individual ship observations from historical records until about 1977. This climatology is expected to be reasonably reliable within the constraints of the methodology (including the choice of drag coefficient formulation), except where sampling was clearly a problem, in particular, south of about 35°S where H&R estimate standard errors in their mean stresses of over 0.25 dynes cm⁻². The new climatology is based upon surface winds from the global analyses of the European Centre for Medium Range Weather Forecasts (ECMWF) for the period 1980–1986. This has the strong advantages of global coverage and regular 12-hour sampling, which removes temporal sampling as an issue, even for individual months. The disadvantage is that results depend entirely on the veracity of the analysis system in reproducing the true wind fields.

As expected, substantial differences were found over the southern oceans. In addition, however, very substantial differences were found over the North Atlantic and North Pacific Oceans. Further investigation with sea level pressures (SLPs) and sea surface temperatures (SSTs) led us to conclude that these differences are mostly real and a consequence
of large changes in the atmospheric circulation on decadal time scales. This note reports on the changes, the evidence supporting their reality, and implications for climate change in general.

2. DATA AND METHODS

Seven years (1980–86) of monthly mean wind stresses have been computed on a 2.5° × 2.5° grid based upon twice-daily 1000 mb wind analyses from ECMWF. As noted above, the analyses provide much better coverage and temporal sampling over much of the world’s oceans, especially in the Southern Hemisphere (SH), than conventional ship data alone. Since the analysis method uses four-dimensional data assimilation, the surface wind analysis fully incorporates all the SLP measurements, for instance from buoys, through a sophisticated variant of the geostrophic relation. The analyses also include relevant past information that has been effectively carried forward in time using the numerical weather prediction model which then provides the first guess for the new analysis.

An evaluation of the ECMWF analyses and intercomparisons with analyses from the U.S. National Meteorological Center (NMC), by Trenberth and Olson (1988), have shown that the ECMWF fields are more suitable globally owing principally to better analyses over the SH and a more complete set of analyses. There are obvious problems in the tropics owing to difficulties in analyzing the divergent, three-dimensional wind field, but in the extratropics the fields are considered fairly reliable, although perhaps somewhat conservative in the sense that very rapidly developing or intense disturbances are apt to be underestimated.

We used the 1000 mb-level wind data, which is generally not expected to be the same as that at the surface. However, the manner of analysis of ship wind data at ECMWF prior to 9 September 1986 meant that the ship winds were effectively assigned to the 1000 mb-level and the 1000 mb winds are considered to be the most appropriate ECMWF “surface” wind product prior to that date. Because of the change in September 1986, only data up till that date have been included in the climatology.

The drag coefficient used has stability and wind speed dependence as specified by Large and Pond (1982). Harrison (1989) has also used this formulation in an analysis of ship data otherwise similar to that of H&R, whereas H&R used the Bunker (1976) formulation. Trenberth et al. (1989) note that the drag coefficient using the Large-Pond formulation is considerably lower than that used in most applications. At wind speeds of 20 m s⁻¹, Bunker (1976) and H&R assign $C_D \approx 2.3 \times 10^{-3}$ versus $1.8 \times 10^{-3}$ for Large and Pond. At wind speeds of 10 m s⁻¹, H&R assign $C_D = 1.65 \times 10^{-3}$ versus $1.14 \times 10^{-3}$ for Large and Pond. Harrison (1989) consequently found his wind stresses to be 20 to 30% less than H&R owing to his use of the Large and Pond formulation. However, we find differences with H&R to be less, apparently because of compensating effects of the different temporal and spatial sampling arising from the use of the ECMWF analyses.

The full details of all the computations and further discussion of the data sets are given in Trenberth et al. (1989).

For SLP we use monthly mean fields based on daily historical analyses dating from 1899. Discussion of the quality and problems with this set is given by Trenberth and Paolino (1980). The worst problems occur before 1924 and there are big discontinuities over high orography owing to changes with time in procedures to correct surface pressures to sea level.
3. CHANGES IN THE ATMOSPHERIC CIRCULATION

The annual-mean wind stress is shown in vectorial form in Fig. 1 for the global oceans from both sets of analyses. Vector differences of the mean stress with H&R (Fig. 2) for the annual mean are highly significant over the southern oceans and in the North Atlantic and North Pacific. The zonal-average differences of the eastward component (e.g., see Fig. 5) are in fact remarkably small over most of the oceans, with the exception of the southern oceans. This shows that in spite of the expected reduction in wind stress due to the use of a drag coefficient that is about 25% smaller, the added variance captured by the ECMWF analyses has evidently compensated for a large extent. However, in the

![Figure 1](image_url)

**Figure 1.** Annual-mean wind stress over the global oceans depicted as vectors from: (a) ECMWF, and (b) H&R. The arrow at bottom right corresponds to 10 dynes cm\(^{-2}\).
Northern Hemisphere (NH) the differences are primarily in the meridional wind stress component and are confined to the winter half-year. Figure 3 therefore shows the ECMWF January-mean surface stress field and the differences from H&R. Note the much stronger southerly component over the eastern North Pacific and North Atlantic. In July (Fig. 4) the differences are much less. Some idea of the annual cycle of the differences can be gained from time sections of the annual cycle of zonal averages across all the oceans (Fig. 5).

These differences have implications for the associated oceanic circulation, as can be seen from the differences in the Sverdrup mass transport stream function, shown in Fig. 6 for the annual mean. Here the mass transport has been computed from the curl of the wind stress and is defined such that positive values indicate stronger subtropical gyres in the NH. Thus the differences in Fig. 6 indicate a stronger Kuroshio current off Japan and a stronger and northward-shifted Gulf Stream in the Atlantic.

Further investigation of the differences in the North Atlantic and North Pacific using SLP fields for the 1980–86 period versus means of the 1945–1977 period (Figs. 7 and 8) or the 1926–1977 period (Fig. 9), which were taken as possibly representative of the period of ship data going into the H&R analyses, reveals a statistically significant change in the atmospheric circulation over the oceans. Statistical significance was tested using a t-test based upon the monthly standard deviations within each data set. Both Figs. 8 and 9 show the same pattern, with pressures 7 to 9 mb lower in the Aleutian Low, and about 6 mb higher in the North Atlantic, in the recent seven-year period for January. Both the changes in SLPs and in surface wind stress are consistent with what would be expected from the geostrophic relation.

The other large and statistically significant changes in SLP over land mostly signal the changes in procedure associated with corrections to sea level, and should be ignored.

The changes in SLP are remarkably consistent from November through March. In the North Pacific, maximum differences lie within 40 to 50°N and are −4.5, −6.2, −8.9, −7.5, and −7.9 mb in November to March, respectively. The pattern is more varied in the North
Figure 3. January-mean wind stress over the global oceans depicted as vectors from: (a) ECMWF, and (b) H&R. The arrow at bottom right corresponds to 10 dynes cm$^{-2}$. (c) Differences in the January-mean wind stress over the global oceans between the current study and H&R, depicted as vectors. The arrow at bottom right represents 5 dynes cm$^{-2}$. 
Atlantic. The overall difference, averaged from November through March, is shown in Fig. 10. This serves to emphasize the deeper, by more than 5 mb, and more extensive Aleutian Low, but with a now fairly indistinct pattern in the North Atlantic. The annual-mean SLP differences (not presented) continue to show very statistically significantly lower pressures in the Pacific and the departures are about one annual standard deviation (2.2 mb) below normal, but the pattern in the Atlantic is indistinct.

Of course the apparent consistency of the changes in SLP raises the question of whether they might simply be due to increased data and more sophisticated analysis.

Figure 4. (a) July-mean wind stress over the global oceans depicted as vectors from ECMWF. The arrow at bottom right corresponds to 10 dynes cm\(^{-2}\). (b) Differences in the July-mean wind stress over the global oceans between the current study and H&R, depicted as vectors. The arrow at bottom right represents 5 dynes cm\(^{-2}\).
techniques in recent years. We have performed similar analyses with the NMC products with the same results. Nevertheless, we suspect that some, but not all, of the differences may be accounted for by the different methods and data. Evidence for this is the essentially independent, but consistent change in the wind stress field. Note that because of the nonlinear nature of the surface stress (which increases with wind at a rate of more than the square of the wind speed), the changes in wind in the winter half-year, as indicated geostrophically by the gradient in SLP, dominate the annual wind stress changes.

Further evidence to be factored into the apparent change in circulation comes from independent surface temperature analyses which show both the North Pacific and North Atlantic annual-mean values persistently to be anomalously low in recent years (Jones et al., 1988); see Fig. 11. Over the oceans the surface temperatures have been replaced with SSTs. Of interest are the changes in surface temperatures and how they relate to what would be expected from changes in temperature advection, and perhaps vertical mixing over the oceans, by the anomalous winds.

Results from all three data sets are essentially based upon independent data and methods of analysis, and provide especially strong evidence that the Aleutian Low has
been significantly deeper and extends farther east in the recent period. The pattern of change of SSTs is consistent. The general pattern around the North Pacific basin, notably the warming trend along western North America and the strong cooling between Japan and 160° in mid-latitudes, is consistent with the expected anomalous advection patterns associated with the change in circulation (see van Loon and Williams, 1976, 1977). In addition, increased mixing in the ocean and stronger heat fluxes into the atmosphere would help account for the recent persistently cold SSTs in the North Pacific.

In the North Atlantic, although there has been an increased southerly component to the wind stress in recent years, the pattern of change in SLP, and thus in the winds themselves, varies sufficiently from month to month that changes in advection are not a factor for the annual mean. Overall, the changes in circulation do not explain the apparent cooling trend in the North Atlantic area.

4. CONCLUSIONS

The new climatology of surface wind stress over the oceans for a recent seven-year period has significantly different features from those in H&R, and the independent analysis of sea level pressures supports the view that the differences in the NH are due to real climate variations on a decadal time scale. Such changes have strong implications for the oceans because the two climatologies would result in quite different Gulf Stream and Kuroshio circulations (Fig. 6). Although the oceanic adjustment time scales are fairly long, significant changes would occur over a decade and call into question any assumptions of an oceanic steady state, such as is implied in some of the World Ocean Circulation Experiment strategies for sampling the ocean.

The highly significant and strong changes in the NH winter in the winds and pressures also have implications for interpreting trends in surface temperatures, and are of
Figure 7. Mean sea level pressures in mb (minus 1000 mb) for January: (a) 1980–86, and (b) 1945–77. The contour interval is 4 mb.
Figure 8. Differences in mean sea level pressures for January from 1980--86 versus 1945--77: (a) total in mb, negative values are stippled, and (b) t values, magnitudes greater than 2 are stippled and are statistically significant at about the 5% level.
Figure 9. Differences in mean sea level pressures for January from 1980–86 versus 1924–77: (a) total in mb, negative values are stippled, and (b) t values, values greater than 2 are stippled and are statistically significant at about the 5% level.
considerable interest because of the prospect of global warming associated with the buildup in greenhouse gases in the atmosphere. The results emphasize once more the conclusions of van Loon and Williams (1976, 1977) concerning the important role of the planetary-scale waves in the NH and the spatial unevenness of the changes, as warming is apt to occur in the regions of increased southerlies and cooling in regions of increased northerlies. At the same time, there are added complications from mixing in the oceans and the additional heat storage (as manifest as positive or negative SST anomalies) compared with the land. The depth of the mixing plus the heat capacity of the oceans means that the anomalies may persist long after the conditions that created them have ceased. Inevitably, these factors will complicate any interpretation of the surface temperature changes in terms of global warming.
ACKNOWLEDGEMENTS

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Pre-Instrumental Climate: How Has Climate Varied During the Past 500 Years?

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ABSTRACT. There is a need for a long-term record of global or hemispheric temperature which extends back in time beyond the limited (100–150 year) span of the instrumental record. The extent to which this can be achieved is discussed and attention is focused on climatic (mainly temperature) information that has been derived from historical (documentary) sources. Records from western Europe, East Asia, North America and Africa are examined and the characteristics that they have in common are identified. Although there are many differences among the regional records, all show that the 20th-century warming is unique in the context of the last 500 years. However, many types of documentary evidence tend to emphasize colder intervals and may underestimate warmer periods. The underlying trend in temperature depends on the time period selected and may differ by season. Maximum upward trends are given by a series beginning in the late 19th century and extending through the 1980s. There is little prospect of reconstructing a hemispherically representative record of temperature from historical data alone, and the best strategy may be to combine many different types of proxy data.

1. INTRODUCTION

Although considerable effort has been expended in trying to construct a globally representative record of climatic fluctuations for the last 100–130 years (Bradley et al., 1985; Jones et al., 1986a–c), a similar record for earlier periods remains an elusive goal. Indeed, some would argue that a meaningful "global" record of interannual climatic fluctuations does not exist even for the last 100 years and, therefore, the underlying trend in climate is really not known. At the same time, concerns over anthropogenic impacts on climate are increasing. Whatever the magnitude of these impacts, and their spatial distribution, there is no doubt that they will be superimposed on "natural" climatic processes that may amplify or subdue the anthropogenic effect. Since we need to know what the climate of the future may be, it is axiomatic that we learn more about the climate of the past. Only with such knowledge can we hope to place our contemporary climate in a longer
time perspective and identify any underlying trend or periodicities in climate upon which future climatic changes might be superimposed. Only with such knowledge can we hope to isolate the causes of past climatic fluctuations which may continue to operate in the future and influence the course of forthcoming climatic events.

How can the climatic history of the world over the last few centuries be determined? Clearly, instrumental data become increasingly sparse the further back in time one goes. Most of the world is unrepresented in the instrumental record prior to 1850. In a fairly comprehensive search of long-term climatic data sources, Bradley et al. (1985) could locate only enough data to extend their gridded coverage of continental monthly temperature anomalies to 7% of the Northern Hemisphere in the mid-nineteenth century (Jones et al., 1986a). For oceanic regions, and for the Southern Hemisphere as a whole, the situation is far worse and a spatially extensive network of instrumental data really exists only for the past few decades (Jones et al., 1986b). Nevertheless, analysis of the limited mid-nineteenth century network suggests that if there had been no growth in network coverage through time, the low-frequency nature of climatic changes which were actually observed (by the ever-expanding network that eventually evolved) could have been recognized (cf. Jones et al., 1986a). This is because there is significant spatial coherence in low-frequency climatic fluctuations (at least) which suggests some underlying large-scale forcing. Thus it is not unreasonable to expect that if a network of long-term records of climatic fluctuations can be assembled for many regions of the world (though perhaps not all regions), a meaningful picture of global climatic fluctuations might be constructed. How dense such a network must be has not been determined, and perhaps cannot be a priori because the nature of the climatic signal that one hopes to capture is unknown, both in its spatial and temporal characteristics. However, such a question may be of only academic merit since there is a limited range of potential data sources, and the networks that are realistically probable are likely to be sparse. Since our knowledge of climatic fluctuations over the last few centuries is still quite incomplete, any new pieces that can be added to the puzzle must be welcomed. Once the picture becomes clearer it may be possible to be more selective, but such a situation is still a long way off.

What data are available to place climatic fluctuations of the last century into a longer-term perspective? Continuous, instrumentally recorded temperature measurements extending back into the 18th century exist only for about 50 stations worldwide (90% of which are in western Europe) so we must turn to additional non-instrumental records from which climatic information can be derived. For the period of interest there is a limited range of proxy records that have the potential of providing interannual resolution of past climatic variations (Bradley and Jones, 1990). These are: historical records, tree-growth indices, ice cores, varved lacustrine (and in a few locations, marine) sediments and coral-growth increments. In each field specialized methods have been developed to extract the climatic signal from non-climatic noise (Bradley, 1985). Although these methods differ in detail, they all involve careful dating of the material, calibration with instrumental data (for a period of overlapping record), reconstruction of past climatic conditions (based on the calibration) and some attempt at verification of the reconstruction by independent lines of evidence. Here we will not attempt to review the multitude of paleoclimatic reconstructions in each field, but instead will focus on paleotemperature reconstructions from historical records. Since none of these records was made with the objective of long-term climatic monitoring, there are many uncertainties in using them for such a purpose (DeVries, 1980; Ingram et al., 1981a). Consequently, no single record should be relied upon too much; only by assembling a variety of individual data sets can past climatic fluctuations be determined with any degree of confidence.
2. HISTORICAL DATA

Historical records contain a wealth of information about past climatic conditions. Providing that adequate precautions are taken in checking sources and evaluating non-climatic influences on the records, detailed interannual (and even intra-annual) climatic conditions can be assessed (Bell and Ogilvie, 1978; DeVries, 1980; Ingram et al., 1981b). Unfortunately, there are only a few regions of the world where extensive studies of historical records, in terms of past climatic conditions, have been carried out. These are: western Europe, China and Japan, and some parts of North America and Africa; each of these regions is discussed below. Studies of historical documents from eastern Europe and the Soviet Union will add considerably to the climatic record of the last few centuries (Borisenkov, 1990). Apart from studies of southern Africa by Nicholson (1978, 1979, 1981), there is almost a complete absence of historical climate research for regions in the Southern Hemisphere.

2.1. European Evidence

Detailed studies of European climatic fluctuations based on historical records have been made by Le Roy Ladurie (1971), Pfister (1980, 1981, 1984, 1985, 1988), Alexandre (1986) and Ogilvie (1984). A variety of climate-related data has been used, including agricultural statistics (particularly harvest dates), records of snow occurrence or freezing/thawing events, phenological data and sea-ice occurrence.

European historical records are particularly rich in phenological data which can provide valuable insights into past climatic conditions. Perhaps the most useful of these are the long records of grape harvest dates from vineyards in northeastern France, Germany and Switzerland (Angot, 1885). Recently, over 100 of these series have been analyzed and statistically reduced to a single composite index of harvest dates for western Europe (representing an area centered near Dijon) for the period 1484–1879 (Le Roy Ladurie and Baulant, 1980). Additional (earlier) data for the Dijon (Cote D'Or) region have been presented by Pfister (1988) for the period 1370 to 1525. Since the two data sets overlap for 40 years in the late 16th century, and the two series are highly correlated in this interval (r = 0.89), the earlier Dijon series can be adjusted to the "western Europe" data standard to produce a comprehensive record of grape harvest dates for the last 500 years (1370–1879). Le Roy Ladurie and Baulant demonstrated that the grape harvest dates were inversely related to April to September temperatures in Paris (r = 0.86) based on the 1797–1879 (83-year) period of overlap between the grape harvest data and early instrumental Parisian records. Our own analysis of these data, and additional instrumental records from Strasbourg (1801–1879) gave correlation coefficients of 0.74 and 0.80, respectively, for April–September mean temperatures. Using the regression equation established for the period of overlapping records, a reconstruction of Parisian spring/summer temperatures back to 1370 can be obtained (Fig. 1). This reconstruction suggests that growing season temperatures have declined by 0.75°C over the last 600 years, with temperatures in the late 14th and early 15th centuries 1 to 1.5°C higher than in recent decades. The instrumental record from Paris continues this downward trend, despite the probable influence of urban warming on the 20th-century section of the record (Detwiller, 1978). A similar trend was found in the Strasbourg instrumental record, although this is incomplete for the early part of this century.

The most comprehensive set of European climate-related historical data has been assembled by Pfister (1985). Pfister combines a vast range of documentary evidence to rate the relative warmth or coldness of individual months. His sources include the occurrence and duration of lake freezing, extensive snow cover (or absence thereof), phenological...
records (such as the dates of beech tree flowering or of cherry tree blossoming), tithe-
auction records (related to harvest dates, which reflect early summer temperatures), vine
phenology (first flowering, full flowering), grape-harvest dates, wine yields and quality,
and tree-ring density (lower density indicative of cooler growing season conditions)
a certain degree of reliability in his climatic interpretations. Similar evidence is used
to assess the relative wetness or dryness of individual months and seasons. From these
ratings, "thermal and wetness indices" for each season over the last 450 years have
been assigned (Pfister, 1980, 1981). Although there is an inevitable element of subjectiv-
ity in rating these diverse phenomena in terms of temperature and precipitation, Pfister
demonstrates a reasonably good correlation between his indices and composite records
of temperature and precipitation, based on a network of instrumental data recorded on
the Swiss Plateau in the period 1901-1960 (Pfister, 1980). Using such comparisons, Pfis-
ter converted the indices to temperature and precipitation anomalies from a 1901-1960
reference period (Pfister, 1984). His reconstruction (Fig. 2) clearly demonstrates the un-
usual characteristics of climate in this area during the 19th century, and the pronounced
warming that has occurred since the 1890s. Mean-annual temperatures appear to have
been about 0.25°C below the 1901-1960 mean for most of the last 450 years, with the
coldest episodes (in the 1690s, 1810s, 1840s and 1850s, and the late 1880s) 0.8 to 1°C
lower than 1901-60. A continuous rise of about 1°C in mean-annual temperature over the
last century (from the 450-year minimum in the late 1880s) is a pronounced feature of

Figure 1. Reconstruction of April-to-September mean temperature at Paris, 1370-1879, and instrumental record from 1797-1976. Reconstruction is based on a regression of instrumental data on a composite record of grape harvest dates (see text) for the period 1797-1879 \( r = 0.74; y = -0.731x + 176.7 \), where \( y \) is mean April-September temperature at Paris and \( x \) is mean grape harvest date at a network of vineyards, in days from September 1. Values plotted as anomalies from the average in the period of data overlap (1797-1879).
the reconstruction. Seasonal differences are apparent, however. Data representing summer conditions (shown in Fig. 3, with Paris temperatures reconstructed from grape harvest data) suggest that summer temperatures for much of the 18th century were similar to, or above, the 1901–60 mean, and may have been 0.5°C higher in the mid-16th century. The only cold periods comparable to the late 19th and early 20th century were in the late 1500s and in the 1810s, a period of exceptionally low summer temperatures. Winter temperatures, on the other hand, appear to have been markedly cooler than the 20th century mean (by an average of 0.6°C) for almost the entire 450-year period. Winters were coldest in the 1690s and there has been a general upward trend in temperature since then.

Precipitation anomalies also suggest that the climate of the last one hundred years in this region has been unusual in the context of the past few centuries. Precipitation was generally lower than in this century, with the lowest amounts in the 1820s, and somewhat of an upward trend since then, punctuated only by dry conditions in the late 1940s and early 1950s.

Some of the longest records of climate-related phenomena are available from Iceland, where commentaries about the occurrence of sea ice have enabled a proxy record of temperature to be reconstructed spanning several centuries (Koch, 1945; Berghorsson, 1969; Sigtryggsson, 1972; Ogilvie, 1984; Grove, 1988). Sea-ice occurrence was far more prevalent from 1740–1900 than for most of this century; ice was a minor problem in the early 18th century and during 1640–1680, but was especially severe in the 1680s and 1690s, the 1740s and 1750s, and for much of the 19th century (Fig. 4). Kelly et al. (1987) caution against using the Icelandic sea-ice record as an indicator of more
widespread conditions, but several of these periods (especially the 1690s) also stand out as extraordinarily cold in the instrumental winter temperature record from central England (discussed below) and in Pfister’s winter and spring thermal indices for western Europe (Fig. 2), indicating widespread regional anomalies at these times.

Bergthorsson (1969) attempted to express sea-ice indices in terms of mean-annual temperature in Iceland by regressing the number of months of sea ice per year against mean-annual temperature (Fig. 5a). The resulting reconstruction appears to indicate the exceptional warmth of the 20th century, 1°C higher than prevailing conditions from 1600–1900 (Fig. 5b). However, this reconstruction illustrates a common problem in historical data where an index may be a useful indicator of cold conditions, but provides limited information about warm conditions. As Fig. 5a shows, once sea-ice occurrence drops to one month or less, it is no longer a useful predictor of mean-annual temperature. Consequently, the warm periods in Fig. 5b, before 1846, must be considered as only minimum estimates, and the 20th-century record of temperature in this area may not be quite so anomalous as Bergthorsson’s reconstruction suggests.

The longest instrumental data set is that of Manley (1953, 1974) for central England (subsequently updated by the Climatic Research Unit, University of East Anglia, in their Monthly Bulletin). This record is shown in Fig. 6. The long-term trend for the entire 330-year record amounts to an increase in mean-annual temperature of 0.16°C per century. However, this trend is amplified by the fact that temperatures in the 1690s, near the beginning of the record, were exceptionally low. Although there is considerable documentary evidence for unusually cold conditions in the 1690s (Lamb, 1982), it is clear from Manley’s writings that these early values are based mainly on estimates,
not instrumental data. For example, Manley (1974) states "... before 1723 we have a very troublesome gap ... over the period January 1707 to October 1722 ... this
Figure 5a. Relationship between annual temperature and ice incidence off the coast of Iceland in months per year. Calibration period 1846-1919. [After Bergthorsson, 1969.]

gap was filled by reference to the Utrecht series . . . adjusted as far as possible from consideration of English non-instrumental 'wind and weather' diaries over that period.” In fact, most of the data prior to 1707 is based entirely on the estimated temperature of prevailing air masses. In view of these uncertainties, it is probably wise to view the record before 1723 with caution. Mean-annual temperatures declined from 1723 to the '310s, but have risen steadily since then. Looking at higher-frequency variations, there have been only a few periods with mean-annual temperatures equal to or higher than recent decades: the 1730s, 1865-75 and the 1910s-40s. However, this is not true of seasonal data; winter temperatures were considerably warmer in the 1910s and 1920s, whereas summer temperatures were well below average at that time (Fig. 6). Summer temperatures were highest in the 1770s when winters were among the coldest on record (cf. Fig. 2). Such differences are probably related to regional circulation anomalies. For example, a period of frequent blocking with anticyclonic conditions over Great Britain could result in both unusually warm summers and cold winters. By contrast, in the early 1900s winter warmth and summer cold are likely to reflect more zonal conditions. These strong seasonal differences highlight the dangers of interpreting proxy records or historical data (which are often indicative of a particular season) as representative of overall annual conditions. A combination of data reflecting all seasons is really needed to reconstruct past climate in individual decades (e.g., Pfister, 1984).

A long record of winter temperature for DeBilt, Holland has been presented by Van den Dool et al. (1978) based on a combination of instrumental data (from 1734 on) and records of Dutch canal freezing dates and winter barge trips between Haarlem and Amsterdam. Overlapping records enable the historical data to be calibrated in terms of winter temperature so that a composite record from 1634–1977 can be constructed. This is shown in Fig. 7 with the central England winter temperature record of Manley (1974) and Pfister’s (1984) winter temperature estimates. Although there are clear differences in
the high-frequency realm, the three records are remarkably consistent in depicting broad-
scale changes in the winter climate of western Europe since the 17th century. The 1690s
stand out as a brief period of exceptionally low temperatures, followed by relatively mild
conditions in the early 18th century. Temperatures subsequently fell to the end of the 18th
or early 19th century, and have steadily risen since then, interrupted by exceptionally cold
periods in the 1810s, 1840s and/or 1850s and the 1890s.

2.2. Chinese Evidence

Chinese historical records are a particularly rich source of climate-related information
(Wang and Zhang, 1988). Records were kept by official Imperial observers, as well as
at the local level, permitting comprehensive spatial and temporal reconstructions to be
obtained for the past several centuries, or (for selected regions) even longer periods (e.g.,
State Meteorological Administration, 1981). Much recent work has been reviewed by
Figure 6. The instrumental record of winter, summer and annual temperature in central England, expressed as departures from the 1751–1800 period mean and filtered with a 21-term Gaussian filter (data from Manley, 1974). Note that data before 1723 are based largely on estimates and comparisons with records from Labrijn, Holland.

Zhang (1988) and Zhang and Crowley (1989). The vast majority of studies have focused on the long records of floods and droughts which have affected Chinese society over the centuries, but a few temperature reconstructions have also been published. It is worth noting that Chinese annual-temperature fluctuations over the last century mirror annual-temperature fluctuations for the northern hemisphere continents as a whole, and so low-frequency changes in long-term Chinese records may be a useful index of larger scale conditions (Bradley et al., 1987).

In a very comprehensive study, Zhang (1980, 1988) examined over 1200 local histories and more than 4400 historical writings to construct a record of severe winter occurrence in eight different sub-regions of China. The reconstructions have been correlated with instrumental data to assess variations in winter temperature back to 1470. Winter conditions were rated by compiling records of river and lake ice occurrence, the freezing of wells, presence and extent of sea ice, snowfall frequency, snow depth and duration of snow cover, and damage to citrus crops or coconut groves. Mild conditions were typified by the absence of such severe conditions and by the early blossoming of peach and plum trees and other agricultural and phenological indicators. By comparing overlapping periods of winter temperature indices and instrumental data for Shanghai and Hankou (Wuhan) a rough index of temperature anomaly can be derived in which a change of two units in the thermal index approximates a change in winter temperature of 1°C (Zhang, 1988). A composite record for 5 regions of eastern China between 22–38°N and 108–122°W is shown in Fig. 8, based on decadal averages for each region. Zhang's reconstructions can be compared with an independent investigation of the frequency of cold winters (though perhaps not based on entirely independent sources) conducted by Zhang and Gong (1979). Both studies point to the periods 1500–1530, 1601–1700 and
1810–1900 as generally colder than in the 20th century, but with milder conditions in the mid-18th century. There is some correspondence with European data in terms of a relatively mild period in the mid-18th century and a colder 19th century, but apart from that, there are few similarities between the winter temperature records of China and Europe.

Historic summer temperature variations at Beijing have been estimated by Zhang and Liu (1987). Using the observed relationship between rainy-day frequency and July temperatures during the period of instrumental records, these authors have reconstructed July mean and maximum temperatures back to 1724 using historical records of rainy-day occurrence (Fig. 8). This reconstruction indicates lowest summer temperatures from 1780 to 1820 and from 1885–1900, with mild conditions from 1825–1875. Like the winter record, there is only minor correspondence with the European evidence; the cold period at the start of the 19th century began earlier in Beijing and the mid-19th century was considerably warmer in Beijing than in Europe. Further studies of temperature change are in progress and it will be of interest to see if these initial reconstructions are subsequently confirmed.
2.3. Japanese Evidence

Like China, Japan has a wealth of historical documents containing information about climatic conditions of the past. One of the most carefully studied sets of material is that of the Tsugaru feudal clan who recorded daily weather conditions at Hirosaki (northern Honshu, 40.5°N, 142.5°E) almost continuously from 1661-1868 (Maejima et al., 1983). From these records, Maejima and Tagami (1984) estimate that temperatures in the 20th century were generally warmer than for most of the preceding 250 years. Relatively mild conditions prevailed in the latter half of the 17th century and from 1741-1780 (also noted in China). However, from 1821-1880 conditions were "very cold" in winters (again, as in China), with heavy snowfalls common. Summers were coldest (with frequent rains) from 1781-1850. These conclusions reinforce Yamamoto's (1971) study of heavy snowfall years and river and lake freezing occurrences in southern Japan (near Lake Biwa) from the 17th to the 19th century. Exceptionally cold conditions were common in the 1750-1850 interval and especially in the early 1820s. Mean January temperatures from 1801 to 1850 are estimated to have been 1° to 2°C lower than in the 1890-1960 period (Yamamoto, 1971). Rice harvest data suggest a similar July temperature anomaly at this time. This seems to support the Chinese winter reconstructions of Zhang (1988), though summer temperatures in Beijing appear to have been quite warm after 1825.

Using records of the freezing dates of Lake Suwa, near Tokyo, Gray (1974) reconstructed December-to-February average temperatures from 1440-1950 (Fig. 9). This record gives a somewhat different picture than the reconstructions of Yamamoto and Maejima et al. Gray's reconstruction of Tokyo winter temperatures would indicate that the 1820s-1860s were relatively mild, with the highest winter temperatures of the last
Figure 9. Winter (DJF)-mean temperatures reconstructed for Tokyo, based on the freezing dates of Lake Suwa. Values expressed as departures from the long-term mean. (After Gray, 1974).

500 years occurring in the 1850s and 1860s. Conditions were below the long-term mean from 1440–1820, apart from the first two decades of the 1500s and the period 1700–1750 (which was already noted as having been mild in both eastern China and western Europe). It is not easy to reconcile much of Gray’s reconstruction with other Japanese historical records, and it would appear that the calibration of Lake Suwa freezing dates may need to be re-evaluated.

2.4. North American Evidence

Historical records of past climatic conditions in North America have received relatively little attention. The main exceptions are studies of Hudson Bay archives which have yielded much useful climatic information (e.g., Moodie and Catchpole, 1975; Catchpole et al., 1976; Catchpole and Ball, 1980; Catchpole, 1980; Rannie, 1983) and the more limited studies of documentary records from New England (Baron and Gordon, 1985). Figure 10 shows the record of “first breaking” of ice in the estuary at Moose Factory on the shores of James Bay (51°19′N, 80°44′W) from 1738 to the late 19th century (Catchpole et al., 1976) compared to the average date in recent decades. For the first half of the 19th century, and in the mid-18th century, freezing dates were generally a week earlier and break-up dates ten days to two weeks later, particularly in the 1810–1820 decade. This is supported by a similar freeze-up/break-up record from the Red River near Winnipeg, based on a variety of historical data (Rannie, 1983). Correlation of these data with contemporary instrumental data suggests that spring and fall temperatures were 2.5°C lower in the 19th century compared to the 20th century, resulting in the “ice-covered period” lasting 3 weeks longer in the 19th century. It is of interest that the ice break-up record shows similar fluctuations to the winter and spring temperature record reconstructed for Switzerland by Pfister (1984), indicating some large-scale teleconnection between the two areas.

Although many other parameters of climate in the 18th and 19th centuries have been documented for the Hudson Bay region (e.g., days with snowfall, frequency of rain and days with thunder; Catchpole and Ball, 1980), no overall pattern comparable with other regions can yet be discerned. The same is true of similar studies in New England (Baron and Gordon, 1985). Further analysis is required to place these reconstructions in a broad-scale perspective.
Figure 10. The record of "first breaking" of ice in Moose Bay (near Moosonee, James Bay) from 1738 (after Catchpole et al., 1976) compared to the average of recent decades (Julian day 116 = April 26).

2.5. African Evidence

Historical data for Africa generally involve observations related to the abundance or absence of rainfall. At present there are very few observations that can be interpreted in terms of temperature (Nicholson, 1978). Certainly, there is strong evidence that the 16th to 18th centuries were generally more humid than recent decades throughout much of the Soudano-Sahelian region, although drought appears to have prevailed in the 1680s and in the early to mid-18th century (1738–1756) (Nicholson, 1981). Drought has been more frequent and more severe since then, particularly in the early 19th century (1800s–1830s) and throughout the 20th century. Indeed, rainfall was relatively low across much of Africa in the early 1800s and again in the early 20th century (Nicholson, 1981). Such trends towards more arid conditions suggest that temperatures may have increased throughout tropical Africa over the last three centuries, associated with increased aridity, but no quantitative estimates of how much warming may have occurred are yet available. Evidence of more frequent occurrence of snow and freezing temperatures in South Africa and eastern Namibia before the mid-19th century supports this notion (Nicholson, 1981), but further studies are required.
3. TRENDS IN CLIMATE RECORDS

What can these reconstructions tell us about trends in temperature over the past few centuries? Fitting a linear trend to the data reveals an overall upward trend in central England mean-annual temperature from 1723 to 1987 of approximately 0.1°C per century (the slope of the regression line). However, stepping through the record at one-year intervals and calculating the (linear) trend up to 1987 from each new point gives an upward trend of between 0.1°C and 0.4°C per century (depending on the date when one starts the analysis), except for the last 100 years when a downward trend is apparent (Fig. 11). The maximum upward trend is manifested with a record beginning in the early or late 19th century. An examination of seasonal data reveals further complexity; overall winter temperature trends are generally more strongly positive, but become strongly negative after about 1890; the trend from 1897–1987 is downward at −0.08°C per decade. Summer temperature trends are quite the opposite of those in winter—weakly negative when the overall period is considered, then increasingly positive until the early 1900s. A

![Figure 11. Linear trends (°C per century) in winter, summer and annual temperature data from central England, calculated year by year from the year at which the trend value is plotted to 1987. Minimum period for trend calculation is 50 years (1938–1987). Trends that exceed 95% significance fall outside the range delimited by the dashed lines.](image)
similar analysis of temperature estimates from Pfister's thermal indices reveals a positive (upward) trend in temperature when considering the overall records from the 16th century to 1975, and from the late 19th century to 1975 (Fig. 12). However, over shorter intervals the trends are generally less positive, and actually negative over the last 50 years for summer months, as was observed for the central England temperature record. Winters, on the other hand, show a strong upward trend in the last 50 years. These records indicate that: a) the underlying trend in climate depends on the frame of reference chosen, and b) underlying trends may not be the same in all seasons. Thus determining the magnitude of the underlying trend in temperature, on which anthropogenic changes will be superimposed, is not a simple task.

4. CLIMATE OF THE LAST 500 YEARS: THE HISTORICAL EVIDENCE

There is unanimity in the available historical reconstructions that the 20th-century warming is unique in the context of the last 500 years, and it is necessary to go back to the early 15th and 14th centuries (1330–1450) to find evidence of temperatures similar to, or higher than, in recent decades. However, it is worth noting that many historical indices (particularly those currently available from East Asia) are based largely on the occurrence of extreme cold events, and the absence of such extremes does not necessarily imply unusually warm conditions. It is thus difficult to assess from these data whether temperatures ever reached 20th-century levels during the 18th and 16th centuries when conditions appear to have been "less severe" than at other times. Individual years in some areas clearly stand out as exceptionally warm (e.g., in Europe: 1540, 1473, 1420, 1336, 1331, 1270; Pfister, 1988), but whether these anomalies were only regional, or of more global, significance is not known at present. In fact, a comparison of individual years and decades from one region to another often shows little consistency in the magnitude, or even the sign of the anomalies. This is partly due to the use of different reference periods, and because reconstructions are generally presented as a running mean, in an attempt to smooth over the inadequacies of converting historical records to quantitative temperature estimates. Nevertheless, there appears to be a consistent signal from many regions that the 1690s, the 1810s and 1820s and the 1890s were exceptionally cold in the context of the last 500 years. Indeed, in many areas, almost the entire 19th century was cold, in stark contrast to conditions which have developed in the subsequent 100 years. Attention should be focused on identifying when temperatures were similar to 20th-century averages; in general terms it appears that such conditions have not prevailed for at least 400 years, though individual years and perhaps even decades (e.g., 1730s, 1780–1805) during the last 500 years may have been relatively warm, in some, if not all, seasons. Overall, interdecadal temperature fluctuations do not appear to have been more than ±1°C seasonally in most regions, although the rise in temperature from the low levels of the 1810s to the high point of the 1980s may be around +2°C in mean-annual temperature for many regions. There is little prospect of reconstructing, from historical data alone, a hemispherically representative record of temperature for the last 500 years without a significant increase in information about paleotemperature conditions in the tropics. There is even less hope of reconstructing southern hemisphere paleotemperatures from historical data since the written records are fewer and shorter, and the oceanic areas are so much more extensive. The best prospect for paleotemperature reconstruction over the past few centuries lies in the combined use of several different types of proxy data, to include historical data, tree-growth indices, ice-core data and sedimentary deposits. Only with such an approach can the required geographical coverage be achieved and the results from the many diverse approaches to paleoclimatic reconstruction be tested against each other.
Figure 12. Linear trends in Pfister's (1984) thermal-index-based temperature estimates for Switzerland, year by year to the end of record (1975). Minimum period for calculation of trend is 46 years (1930–1975). Trends that exceed 95% significance fall outside the range delimited by the dashed lines.
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Part 4:

Forcing Other Than by Greenhouse Gases: What Has Caused the Variations in the Observed Climatic Record
The Effects of Solar Variability on Climate*

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ABSTRACT. It has been hypothesized for at least a century that some of the observed variance in global temperature records arises from variations in solar output. Theories of solar-variability effects on climate could not be tested directly prior to satellite measurements because uncertainties in ground-based measurements of solar irradiance were larger than the solar variations themselves. Measurements by the Active Cavity Radiometer (ACRIM) onboard the Solar Max satellite and by the Earth Radiation Budget (ERB) instrument onboard Nimbus 6 are now available which indicate solar-constant variations are positively correlated with solar activity over an 11-yr solar cycle, and are of order \pm 1.0 W m\(^{-2}\) relative to a mean solar constant of \(S_0 = 1367 \text{ W m}^{-2}\), \(\Delta S/S_0 \approx \pm 0.07\%\). For a typical climate sensitivity parameter of \(\beta = S_0 \Delta T/\Delta S \approx 100 \text{ C}\), the corresponding variations in radiative equilibrium temperature at the Earth's surface are \(\Delta T_e \approx \pm 0.07 \text{ C}\). The realized temperature variations from solar forcing, \(\Delta T\), can be significantly smaller because of thermal damping by the ocean.

I consider effects of solar variability on the observed and projected history of the global temperature record in light of this data using an upwelling-diffusion ocean model to assess the effect of ocean thermal inertia on the thermal response. The response to harmonic variations of the 11-yr sunspot cycle is of order \(\Delta T \approx \pm 0.02 \text{ C}\), though the coupling between response and forcing is stronger for long-term variations in the envelope of the solar cycle which more nearly match the thermal response time of the deep ocean (e.g., the 80-yr Gleissberg cycle). Nonetheless, solar variability effects are estimated to have been small compared with the 0.5\text{ C} warming observed over the past century and the increased rates of global warming observed in recent years. It can be concluded from this analysis that variations in solar output are unlikely to significantly alter the warmings projected for the next century from anthropogenic greenhouse gases.

1. INTRODUCTION

Global climatic change is thought to arise primarily from four factors (Hoffert and Flannery, 1985): (i) variations in solar luminosity; (ii) variations in planetary albedo associated

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with changing amounts of aerosols or dust, surface reflectivity and cloudiness distributions; (iii) variations in amounts of infrared absorbing gases in the atmosphere \((\text{H}_2\text{O}, \text{CO}_2, \text{O}_3\) and various trace gases); and (iv) internal feedbacks among elements of the climate system. This report will focus on what can be learned from direct satellite measurements of solar irradiance fluctuations from 1980-1984 about the contribution of solar variability to the global surface temperature history of the Earth.

2. GLOBAL TEMPERATURE HISTORIES AND SOLAR VARIABILITY

Considerable effort has been expended in recent years to describe the temperature history of the Earth from instrumental records over the past century. Figure 1, for example, illustrates two recent reconstructions of surface temperature anomalies (relative to the year 1980) extending back in time over a century which have been developed independently by Jones et al. (1986) at the University of East Anglia, UK, and Hansen and Lebedeff (1988) at the Goddard Institute for Space Studies in New York. The Jones et al. curve is based on area-weighted averages of both land and sea records, with sea surface temperatures from the Comprehensive Ocean-Atmosphere Data Set (COADS) corrected for the transition from the sailing ship "bucket" temperatures to the water inlet temperatures of steamships; 1980-1984 temperatures are from NOAA observations adjusted for compatibility with earlier data. The Hansen and Lebedeff curve is based on slightly different data sources and methodologies, but exhibits essentially the same trends: a systematic global warming of some 0.5°C per century superimposed on considerable variability on interannual to decadal time scales. Both analyses indicate the rate of global warming has increased over the past several years, perhaps signaling the emergence of the fossil fuel greenhouse effect.

![Figure 1. Smoothed global-mean surface air temperatures: 1861-1984.](image-url)
To establish whether, and how much, of this variation is due to anthropogenic greenhouse gas emissions, it is necessary to know the relative contributions of other factors. As discussed below, space-based irradiance observations are necessary for an accurate assessment of solar variation effects on surface temperature. Figure 1 shows that direct measurements of solar irradiance incident on the Earth from platforms above the atmosphere are very recent in comparison with the instrumental record. The question addressed here is the degree to which one can attribute the historical variance of global-mean temperature to solar variability in light of the relatively small interval of the historical temperature record over which space-based irradiance measurements are available.

A number of modeling studies have appeared in recent years aimed at explaining global temperature records in terms of multiple driving mechanisms (Schneider and Mass, 1975; Robock, 1978, 1979; Vinnikov and Groisman, 1979; Hansen et al., 1981; Gilliland, 1982; Gilliland and Schneider, 1984; Reid, 1987). All of these to some extent allow for an effect from solar luminosity variations using assumed correlations with ground-based observables such as sunspots. Nevertheless, the contribution to the temperature signal from variations in the "solar constant" $S_o$ remains controversial. Despite longstanding proposals that solar variability associated with sunspot cycles has a significant impact on climate over the past 100 years, some recent assessments dispute this strongly (Pittock 1978, 1983; Budyko et al., 1986). Interestingly, the measurements of solar irradiance from satellites which bear directly on this issue have not yet been incorporated in transient climate models.

The main problem with surface-based radiometers is that owing to uncertainties in atmospheric scattering and absorption effects they can measure variations in the solar constant only to an accuracy of 1-2% (Newkirk, 1983). The change in global-mean surface temperature corresponding to a solar irradiance anomaly $\Delta S/S_o$ persisting long enough for the planet to come into a new radiative equilibrium -- perhaps decades, as a result of the ocean's thermal inertia -- can be estimated from the steady-state climate sensitivity parameter, $\beta \equiv S_o \partial T/\partial S \approx 100^\circ C$ (Hoffert and Flannery, 1985). Thus an uncertainty of $\Delta S/S_o \approx \pm 1-2\%$ corresponds to an upper bound uncertainty in the effect of solar variations on global temperature anomalies of $\Delta T \approx \beta \Delta S/S_o \approx \pm 1-2^\circ C$. This is much greater than the variance of the instrumental global temperature record from all physical mechanisms over the past 100 years (Fig. 1). Thus ground-based measurements clearly have inadequate resolution to assess solar-fluctuation effects on climate.

The fault is not with the instruments. Present-day radiometers can measure solar irradiance with a precision of $\pm 0.002\%$ and a long-term accuracy better than 0.1% (Willson, 1984), the range needed for solar variability studies. But to exploit that capability one has to get the detector above the sensible atmosphere, point it in the right direction, and be confident that its calibration will not drift for many years. Certainly, applications to understanding climatic change have been a major motivation for recent solar flux monitoring programs from spacecraft. The past decade has seen the beginnings of extraterrestrial long-term solar monitoring programs including the Earth Radiation Budget (ERB) instruments onboard the NIMBUS 6 and 7 satellites (Hickey et al., 1981), and more recently the Active Cavity Radiometer Irradiance Monitor (ACRIM) onboard the Solar Maximum Mission (SMM) satellite (Willson et al., 1981; Willson and Hudson, 1981; Willson, 1984).

3. SPACE-BASED IRRADIANCE MEASUREMENTS AND SOLAR ACTIVITY

Apart from short-duration NASA sounding rocket and Skylab experiments (Eddy, 1979), continuous data sets of solar flux measured from above the atmosphere are mainly from the NIMBUS 6 and 7 ERB satellite experiments, launched in mid-1975 and late-1978,
respectively, and from the Solar Maximum Mission ACRIM instrument in orbit since early 1980 (Willson, 1984). Some problems with other instruments on “Solar Max” were corrected in orbit in April 1984 by astronauts from the ill-fated Space Shuttle Challenger, but a fairly continuous ACRIM record exists from 1980 onwards.

The ERB/NIMBUS 6 was a simple detector comprised of a blackened flat plate attached to a thermopile incapable of electrical self-calibration. It relied on prelaunch calibrations to relate its observations to SI units (W m⁻²), and had too wide a field of view to resolve the solar disk to better than 4°. Willson (1984) estimates ERB/NIMBUS 6 measurement uncertainties of ΔS/S₀ ~ 0.2%, too large for reliable estimates of irradiance fluctuation effects on climate. The ERB radiometer on the follow-on NIMBUS 7 launched in late 1978 is a superior detector capable of self-calibration. But the most accurate irradiance monitoring in space to date is from the self-calibrating SMM/ACRIM, with a long-term accuracy estimated by Willson (1984) as ΔS/S₀ < ±0.1%. In a flight test, three ACRIM sensors agreed to within 0.04% of their average result. The precision of the data, about ±0.002%, is higher than its accuracy. The time resolution of ACRIM raw data samples (< 1 s) is much shorter than what is required for climate analysis, so some level of averaging is needed for data analysis.

While speculations on sunspot effects on irradiance and climate have been made for hundreds of years (Lamb, 1972), the SMM/ACRIM data permit realistic assessments for the first time of correlations between irradiance and solar surface features observable from the Earth’s surface (Hoyt and Eddy, 1982, 1983). The most studied features are sunspots - dark, cool regions on the sun’s visible surface, or photosphere, whose numbers have varied with an ≈ 11-year cycle since they have been observed continuously by telescope beginning in the early 17th Century (Eddy, 1979). Indices of sunspot activity include the so-called Wolf daily sunspot number, Nₗ, and the number of daily sunspot groups, N₉. The number of sunspots/group is of the order of 10 (Hoyt and Eddy, 1983); N₉ ~ Nₗ/10. The Zurich sunspot number, which is dominated by the number of sunspot groups, is also used.

The various sunspot indices tend to trend together; they approach a maximum when 100 or more individual spots are found on a solar hemisphere at one time, and a sunspot minimum, or quiet sun, when few or none are seen for months at a time. The last sunspot maximum was around 1980, hence the term Solar Maximum Mission (SMM) for the Sun-observing satellite launched that year. The fact that ERB/Nimbus 7 began monitoring irradiance two years before the sunspot maximum permits an assessment of possible irradiance-sunspot correlations. An overlapping record of ACRIM and ERB measurements is available beyond 1980.

In analyzing the early ACRIM data, Willson et al. (1981) focused on short-term solar irradiance fluctuations over the first 153-day period in 1980. The major new finding from early ACRIM data was that reductions in solar constant as much as ΔS/S₀ ~ -0.2% were found over timescales of 5-8 days as sunspot groups passed over the solar disk. This short-term anticorrelation is opposite in sign to the usual assumptions made by climate modelers that long-term luminosity variations are positively correlated with sunspots (Robock, 1979). Physical models for the short-term irradiance deficit are based on the idea that the “dark” sunspots create temporary blockage of emerging solar flux, which must be re-radiated over timescales of a month or more (Hoyt and Eddy, 1982). We show next that short-term anticorrelations of irradiance with sunspot groups may reverse in sign on monthly timescales when the five-year ACRIM record is used.

Figure 2a shows monthly averages of the first five years of ACRIM irradiance data. In contrast to the short-term anticorrelation of irradiance with the area of sunspot groups crossing its surface, the monthly mean ACRIM data trends downward along with sunspots since the sunspot maximum in 1980. The negative, deviant irradiance point at about April
Figure 2. (a) SMM/ACRIM Irradiance Data: 1980-1984 (Willson, 1985). (b) NCAR sunspot group data: 1980-1984 (Gilliland, 1985).
1984 is probably insignificant; it occurred at the time of SMM repair and represents little real data. (Note also that only 11 months of data were available in 1980.) Sunspot numbers alone are often used in irradiance correlations for climate models. Robock (1979), for example, employed the positive correlation $\Delta S/S_o \sim 0.0052 <N_w>$ (%) to drive a climate model for the "Little Ice Age" in the northern hemisphere extending back to the sixteenth century, where $<N_w>$ is the Wolf sunspot number of the 11-year cycle smoothed out. We now know, however, that such relations are based on surface irradiance observations of insufficient accuracy.

Other potentially relevant observables on the solar disk are available for the past 100 years from routine observations published by solar observatories. For example, the sunspot umbra is its dark, central core whose mean brightness is $\sim 0.25$ of the surrounding photosphere, and the penumbra, a somewhat less dark region surrounding the umbra with a brightness $\sim 0.75$ of the surrounding photosphere. The relative contrast (brightness - 1) of these zones is $\sim -0.75$ for the umbra and $\sim -0.25$ for the penumbra. Hoyt (1979) proposed, on largely empirical grounds, a possible correlation between the umbra/penumbra ratio and the historical northern hemisphere surface temperature record from 1881 to 1980 of Jones et al. (1982). Also distinguishable from the darker photospheric background are irregular, bright patches, called faculae or plage. These "anti-sunspots" emit energy fluxes higher than the background levels of solar radiation. Their mean facular contrast is quite low, $\sim +0.03$ (Hoyt and Eddy, 1982), which makes the bright zones less obvious than sunspots, although their areal extent is greater. Both types of features are related to the underlying solar magnetic field activity through fluctuations in magnetic field lines, or "tubes," which penetrate the visible surface.

Based on corrected area-weighted contributions of light and dark areas on the photosphere, an expression can be written for solar irradiance incorporating projected surface areas of the umbra and penumbra of sunspot groups, $u_i$ and $p_i$, the facular area $f$, and a correction factor for photospheric limb darkening, $C(\theta) = 0.36 + 0.84\cos\theta - 0.20\cos^2\theta$, where $\theta$ is the angle between the radius vector to the central point of the Sun and the line of sight to the spot group (Hoyt and Eddy, 1982),

$$\frac{\Delta S}{S_o} = 0.03f - \sum_{i=1}^{N_g} C(\theta)(0.75u_i + 0.25p_i).$$

From this formula one expects an anticorrelation between sunspots and irradiance if the area-weighted cooling from dark sunspots dominates the heating effect from faculae area. If, on the other hand, the bright faculae dominate, and if the facular area scales with sunspot area, then a positive sunspot/irradiance correlation should result. However, if the bright faculae and dark sunspots have comparable opposing magnitudes over given averaging periods, the net sunspot/irradiance correlation could very well be weak or insignificant, particularly so if there are unrelated factors influencing brightness variations. When Hoyt and Eddy (1982) applied their model to corrected umbral, penumbral and facular areas observed over the past 100 years spanning 10 solar cycles, the computed irradiance curve showed minima during sunspot maxima, suggesting a long-term anticorrelation. This is opposite to the long-term positive correlation assumed between sunspots and irradiance assumed by some climate modelers, and is opposite to the secular trend of the ACRIM data itself (see below).
4. STATISTICAL ANALYSIS OF ACRIM/NCAR DATA

Figure 3 is a scatter diagram showing the monthly mean ACRIM irradiance data of Fig. 2a in W m\(^{-2}\) versus the monthly mean sunspot group number \(<N_G>\) of Fig. 2b. As a first step in the statistical analysis, the regression line through the data shown in Fig. 3 was computed using a least square best-fit routine. (The April 1984 point, while shown in Fig. 3, is not used in computing the regression line.)

A positive irradiance/group number correlation was found for the monthly data, albeit with appreciable scatter around the trend line. The standard deviation of irradiance was \(\pm 0.56\) W m\(^{-2}\), corresponding to only \(~16\%\) of the irradiance variation predictable by \(<N_G>\). The group, rather than the Wolf, sunspot number was used for consistency with Willson et al.'s (1981) findings on short-term variability. More elaborate correlations such as the Hoyt and Eddy (1982) one based on umbral, penumbral and facular areas would presumably do better. Our objective at this point, however, was simply from gross statistical analysis to determine whether a time scale could be found at which the bulk correlation of irradiance with sunspot groups switches from negative to positive, perhaps arising from re-radiation after some time lag \(\tau_{\text{lag}}\).

To do this we computed the cross-correlation coefficients, \(R\), which compare deviations about the means of the irradiance and group number time series. A value of \(R = +1\) implies that the relative magnitudes and signs of deviations of one time series can be used to predict the behavior of the second time series; a value of \(R = -1\) implies that deviations in one data set are comparable in magnitude but opposite in sign to the other.

\[
\text{Mean trend line:~~} \Delta S/S_0 = 6.05 \times 10^{-5} <N_G> \\
S_0 = 1366.87 \text{ W m}^{-2}
\]

Figure 3. ACRIM monthly irradiance data versus NCAR sunspot groups: 1980-1984.
that is, they are anticorrelated. One can predict the behavior of one time series from the other with a confidence level of \(R^2 \times 100\)%%. To isolate possible effects of re-radiation from facular areas correlated with sunspots at earlier times, we introduced a time lag variable, \(\tau_{\text{lag}}\), such that \(R^2(\tau_{\text{lag}})\) is the confidence level with which we can predict the behavior of one time series at time \(t + \tau_{\text{lag}}\) from the behavior of the other time series at time \(t\). The cross correlation coefficients versus lag time over the 1980-1984 time frame varied smoothly in the range of \(-14\) months < \(\tau_{\text{lag}}\) < +14 months, with a peak at \(\tau_{\text{lag}} \sim 6\) months. Throughout this range the formal predictability was only \(100R^2 < 25\)%%, too small to indicate a significant lagged correlation of irradiance with earlier sunspots.

The whole-sun radiance measured calorimetrically by ACRIM is the integral of the incident component of energy flux emanating from all areas of the solar surface. Our statistical analysis did not include effects of bright features which may vary independently of present and past sunspots; this may explain the low correlation.

Foukal and Lean (1988) studied the slow changes in solar irradiance using ERB and ACRIM data. They found a low-amplitude (0.04% - 0.07%) variation over timescales of 4 to 9 months which is well-correlated with changes in total facular radiations from an 81-day smoothing of daily data. They attribute the observed dimming of the Sun mainly to a decreasing contribution to the irradiance by a network of unbalanced bright photospheric magnetic elements. However, the contributions of sunspots and large faculae are nearly balanced, that is, the blocking effect of sunspots is compensated by the enhanced irradiance of “large” faculae. Our lack of a significant lagged correlation discussed above is consistent with their contention that “the excess radiation from these bright magnetic elements cannot represent the re-radiation of blocked sunspot heat flux.”

On the other hand, there is no firm basis for back calculations of solar effects on climate prior to the satellite era unless the faint-network radiation can be related to a pre-satellite observational data base. This will be discussed again later.

The downward irradiance trend of the ACRIM data is consistent with similar findings of solar cooling at about the same rate by the NIMBUS 7 ERB instrument during this period. (Both data sets are plotted versus time in Fig. 3 of Fröhlich, 1987). In the 1980-1984 period both the measured SMM/ACRIM and the measured NIMBUS 7/ERB irradiance data trend downward at a rate of about 0.01-0.02%/year (Willson, 1984). During the same epoch all indices of solar activity, including counts of groups (Fig. 1, middle panel) also decline, since the period embraces the declining phase of an 11-year solar cycle whose sunspot maximum was a wide peak spanning 1980-82. Thus one expects an apparent, if possibly accidental, positive correlation. The question is whether this reflects the behavior of the Sun over longer time frames. The answer is problematic because ERB/NIMBUS 7 irradiance values have decreased monotonically since early 1979, well before the peak of the 11-year activity cycle (Hickey et al., 1981). That is, the NIMBUS data decline more or less monotonically while the sunspot numbers rise to a maximum (1980) and then fall, apparently refuting a straightforward correlation between measured irradiance and any simple sunspot index.

These uncertainties underscore that models for solar climatic effects based on correlations between irradiance and surface observables prior to 1980 should be viewed with caution, particularly since a number of correlations that have appeared previously in the literature are now known to be inconsistent with the space-based observations. We will review a number of these in the concluding section, where we also consider the implications for transient climate change of a new irradiance correlation developed by Schatten (1988) which does appear consistent with both solar physics and satellite data. But first we consider what can be learned about the global thermal-response to the early ACRIM measurement themselves using a simple transient-response climate model.
THE EFFECTS OF SOLAR VARIABILITY ON CLIMATE

In addition to mean values, tables of solar observations from 1874 to 1981 (Hoyt and Eddy, 1982) are available which give the number of sunspot groups/day ($N_G$) and Wolf sunspot number ($N_W$), as well as the projected and corrected umbral area ($u$), whole spot area ($w = u + p$) and facular area ($f$). Areas are normally given in units of $10^{-6}$ of the solar disk. R. Gilliland (1985) of the National Center for Atmospheric Research has supplied us with monthly mean values of these parameters from 1980-1984, enabling an analysis over the ACRIM time frame. Figure 2b shows the variation of monthly mean sunspot groups/day, $<N_G>$, from Gilliland's NCAR data over the same time frame as the ACRIM data of Fig. 2a. The declining trend of both monthly mean irradiance and monthly mean sunspots from 1981-1984 is quite evident.

5. CLIMATE MODEL RESPONSE TO SOLAR FORCING: 1980-1984

The world's oceans exert a kind of "thermal-flywheel" effect on all external climatic forcing including solar irradiance fluctuations. To study the influence of ACRIM irradiance data on the response of global-mean surface temperature $T_s(t)$ we used the upwelling-diffusion, one-dimensional ocean/transient climate model of Hoffert et al. (1980). The salient features of the model are reviewed below.

A useful reference condition for transient climate studies is the equilibrium temperature, $T_e$, corresponding to the steady-state surface temperature at solar flux $S$, planetary absorptance $a$, and atmospheric carbon dioxide concentration $c$. At the reference values of $S_o$, $a_o$ and $c_o$, $T_o \equiv T_e$. An increase in any of the forcing parameters by $\Delta S = S - S_o$, $\Delta a = a - a_o$, or $\Delta c = c - c_o$ tends to create a new equilibrium surface temperature,

$$T_s = T_o + \beta_T \left[ \frac{\Delta S}{S_o} + \frac{\Delta a}{a_o} \right] + \beta_c \ln \left[ 1 + \frac{\Delta c}{c_o} \right],$$

where $\beta_T \approx 108^\circ C$ and $\beta_c \approx 3.6^\circ C$ are climate sensitivity parameters (Hoffert and Flannery, 1985). For a planet with zero thermal inertia, $T_s(t) = T_e(t)$. In the real world the $T_s(t)$ response is delayed and modified by oceanic mixing and storage in ways which depend on the $T_e(t)$ forcing.

The transient climate model used here computes heat capacity and internal mixing effects on $T_s(t)$ of an ocean mixed layer of depth $h = 75$ m and thermal relaxation time $\tau \approx 4$ yr, overlying a deep ocean upwelling at $w \approx 4$ m yr$^{-1}$ with eddy diffusivity $\kappa \approx 2000$ m$^2$ yr$^{-1}$. Considerations leading to these numerical values and to the model itself are discussed in Hoffert et al. (1980). The evolving surface temperature is computed by numerical solution of the differential equation

$$\frac{dT_s}{dt} = \frac{[T_e(t) - T_s]}{\tau} + \frac{1}{h} \left[ \kappa \frac{\partial T}{\partial z} + w(T - T_p) \right]_{z=h},$$

where $T_p$ is the temperature of polar bottom water. The last term in large brackets on the right-hand side is the rate of temperature change from heat transfer with the deep ocean at the mixed layer/thermocline interface. If the term is small, then heat is trapped in the mixed layer and only superficial heating of the oceans needs to occur for climate to re-equilibrate with an altered surface heat balance. If the term is large, then warming of the ocean's surface cannot occur until the ocean warms from top to bottom. To evaluate the complete ocean model in transient evolution, the $T_s$-equation is integrated numerically simultaneously with a coupled 1-D, upwelling-diffusion energy transport equation for $T(z,t)$ in the deep ocean,
\[ \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left[ \kappa \frac{\partial T}{\partial z} + w(T - T_p) \right], \]

where \( z \) is depth below the mixed layer, and the bottom boundary condition on this equation is \( \kappa \frac{\partial T}{\partial z} + wT = wT_p \) at \( z = h_d \). The partial differential equation for oceanic internal temperature, \( T(z,t) \), is solved by finite differences over 81 fifty-meter-thick layers between the mixed layer/thermocline interface and the sea floor at \( h_d = 4050 \) m.

To calculate the irradiance effect from 1980-1984, we forced the system with \( T_s(t) = T_o + \beta_T \Delta S(t)/S_o \), where \( t \) is the time from a hypothetical initial state in 1980, \( T_o \) is the 1980 surface temperature, \( S_o = 1368.4 \) W m\(^{-2}\) is the monthly mean ACRIM value for early 1°C, and \( \Delta S(t)/S_o \) was specified by the monthly mean ACRIM irradiance data of Fig. 2a. The polar sea temperature was held constant at \( T_p = 1^\circ C \) during the run. The initial vertical ocean temperature profile used to start the calculation was specified by

\[ T(z,0) = T_p + (T_o - T_p)e^{zw/\kappa}, \]

that is, a pre-existing oceanic steady state was assumed in 1980. This assumption was imposed by lack of prior data, but should be viewed with caution in light of a possible "history effect" (see below).

The surface temperature response is illustrated in Fig. 4. Also shown is the equilibrium temperature forcing and the global surface temperatures from 1975-1985 blown up from Fig. 1. It is evident in Fig. 4 that for the climate sensitivity parameter used here \( (\beta_T \sim 108^\circ C) \), the ACRIM irradiances correspond to short-period (monthly) fluctuations of equilibrium temperature of order \( \sim 0.1^\circ C \). The predicted longer-term trend over the five years is cooling, but considerably smoothed relative to the forcing and with fluctuations damped to the \( \sim 0.01^\circ C \) level by the ocean's thermal inertia. Interestingly, although the smoothed Jones et al. (1986) temperature data indicate a warming in the early 1980s, the more recent 5-year running-mean global temperatures of Hansen and Lebedeff (1988) show a pronounced cooling dip from 1982-1984, followed by a very recent global warming.

The magnitude of the solar-induced effect is generally smaller than the observed temperature signal, but could still contribute a significant part of the signal over time. It is tempting, but probably premature, to attribute Hansen and Lebedeff's cooling dip to the irradiance decline.

The computed response reflects the interplay of oceanic mixing and storage to modulate the imposed solar signal. The total response might look different with a different initialization, as the memory of prior heat anomalies in the ocean is reflected in the superimposed transient solutions. But since the model equations are linear, our solar-induced transient solution initialized at a steady state would simply add to prior history, greenhouse-gas and albedo-change components.

Hansen et al. (1986) suggested that the solar cooling implied by satellite measurements might have been compensated for by anthropogenic greenhouse-gas warming over comparable time periods, and that such effects might even continue for some time. Wigley (1987, personal communication) estimates that for roughly the five-year period from 1979-1984, the total reduction in infrared cooling to space from the observed CO\(_2\) increase of \( \sim 6 \) ppm, and from other anthropogenic greenhouse gases, was \( \sim 0.26 \) W m\(^{-2}\). Since the observed irradiance decline over this period of \( \sim 0.1\% \) corresponds to a planet-wide-average decrease of absorbed solar flux of \( \sim 0.24 \) W m\(^{-2}\), a short-term cancellation of effects on the radiation balance is indeed possible, bearing in mind that the uncompensated effect over these 5 years is in any event small.
6. IMPLICATIONS FOR TRANSIENT CLIMATE MODELS

I have already referred to the pre-ACRIM positive irradiance-sunspot-number correlations used for example by Schneider and Mass (1975) and Robock (1979) in climate models as being based on insufficiently accurate observations. Hansen et al. (1981) used the Hoyt (1979) umbra/penumbra ratio correlation for the solar component to improve predictions by their model of the local peak around 1940 of global temperature (see Fig. 1). This correlation has not to our knowledge been tested against extraterrestrial irradiance measurements, and Hoyt (1979) himself states: "The high cross-correlation between northern hemisphere temperature anomalies and the umbral/penumbral ratio may be a mathematical oddity without physical meaning." Gilliland and Schneider (1984) modeled the effects of solar forcing in a transient climate model with a sinusoidal term based on assumed solar radius cycle of 76-year period with phase and amplitude arbitrarily adjusted to fit temperature data. However, their "best fit" of solar forcing to surface temperature histories contradicts satellite observations, since Gilliland and Schneider (1984) show a rise in solar forcing over the 1980-1984 time frame when both SMM/ACRIM and ERB/NIMBUS 7 instruments measured declining irradiance trends.

It is now clear that correlations grounded in accurate (space-based) observations are needed to reliably assess solar-variation effects on climate, but some care is needed with regard to matching these to the time-scales of interest. At the time, the Hoyt and Eddy (1982) short-term anticorrelation seemed well-motivated by the ACRIM data. It produced a 90% correlation with sunspot blocking (10% short-term storage) based on ACRIM data.
over the first year of Solar Max operation (Hoyt and Eddy, 1983). To extend it over long
periods, one needs corrected umbral, penumbral and facular areas (these are available for
the past ~ 100 years). But, when this extension was done by Hoyt and Eddy (1982) for
the period April 1974-October 1981, their predictions of monthly mean solar irradiance
(computed from their sunspot and facular radiation model and observed projected sunspot
areas) had minima of order 0.1% during periods of maximum solar activity. This anti-
correlation of long-term solar activity with irradiance is in the same direction as the
short-term blockage effect of Willson et al. (1981), but opposite to the five-year trend of the
ACRIM and ERB data! The discrepancy can only mean that other features on the solar surface correlated with the solar cycle brighten sufficiently to compensate for
short-term sunspot blockage effects of Hoyt and Eddy (1982). The findings of Foukal and
Lean (1988) suggest these features are faint-network faculae distributed over the large
solar surface.

The blockage factor of Hoyt and Eddy has nonetheless proven useful in decipher-
ing the ACRIM record. Livingston et al. (1988) showed that the ACRIM irradiance
record corrected for sunspot blockage is well-correlated with ground-based irradiance
spectrophotometry of Fraunhofer absorption lines of certain elements in the sun's atmo-
sphere. In particular, the strength of the mid-photospheric manganese 0.5394 µm line
tracks the corrected ACRIM irradiance almost perfectly. (Because line strength is de-
termined relative to the local blackbody continuum, no correction for sunspot blocking
is needed to show any global variability.) These comparisons independently confirm the
variability of the corrected ACRIM signal, indicate that its source is (probably network)
faculae, and indicate that the corrected irradiance follows the 11-year activity cycle.

Evidently, the key to success in modelling climatically significant irradiance vari-
ations over solar cycles is the way one handles the residual irradiance from network-facular
radiation: A positive residual gives a brighter sun as the sunspot number increases. A
recent solar irradiance model in this vein whose predictions overlap the ACRIM time
frame was developed by Lean and Foukal (1988), who model irradiance variations due to
solar magnetic activity during the past three solar cycles. They find irradiance residuals
(irradiance minus the blocking effects of sunspots) to be well-correlated with the 10.7 cm
microwave flux which represents radiation from bright facular elements. Assuming that
this correlation holds over the past three solar cycles, and by using estimates of sunspot
contrast and number in the past, they reconstruct the irradiance time series from 1954-
1985. The match between the data trend and their model over the overlapping period
appears to be quite good.

Because the 1-D ocean analysis of the previous section was intentionally limited to
the ACRIM time frame, it did not include the history effect that would develop from a
"running start" in which the memory of earlier forcing cycles was present. Prior simu-
lations of the response of a coupled atmosphere/ocean transient model have shown that
the amplitude of the global temperature response to harmonic radiative forcing of fixed
amplitude increases as the period of the forcing period becomes large compared with the
mixed layer thermal damping time (Hoffert et al., 1980). Moreover, if one is aiming at
a transient climate model on the time scale of the hundred-year instrumental record, it
may be necessary to account for changes in the peak number of sunspots during a given
cycle, a quantity which itself varies over still longer periods. Reid (1987), for example,
has recently linked the variation in sea surface temperature (SST) to the envelope of the
11-year cycle of solar activity (this is called the Gleissberg cycle and has a period of
~ 80 years). The solar constant was varied in phase with the envelope of the sunspot
number, with the amplitude of the solar constant assumed proportional to the sunspot
number on the envelope. Using this variation of solar constant with time as the forcing
term in the transient one-dimensional, upwelling-diffusion model of Hoffert et al. (1980),
Reid reconstructs SST's from 1860-1980.
To assess the implications of these effects for transient climate models, we worked with a recent, and fairly comprehensive, model for solar irradiance variability developed by Schatten (1988). His model includes: (i) effects of solar active regions (sunspots and large faculae), (ii) effects of global faculae (polar and network) and (iii) long-term variations in the upper envelope of sunspot number. The latter effect sets the so-called "quiet sun" solar constant $S_o$ associated with few or no active regions and little or no global field or faculae. The ratio $\Delta S/S_o$ (where $\Delta S$ is the deviation from $S_o$) is computed as the sum of two factors, one accounts for the blocking effect of sunspots and the other for the enhanced irradiance from faculae. Faculae are due to both active regions and global sources. The effect of the meridional drift of the active regions across the solar photosphere is considered when computing the irradiance variation. This model also predicts irradiance to be positively correlated with the 11-year cycle of sunspot activity. As shown in Fig. 5, the ACRIM irradiance trend from 1980-1984 is matched quite well. Moreover, Schatten (1988) used the sunspot prediction of Schatten and Sofia (1987) for the projected solar constant variation through 1997. Figure 5 also shows this prediction from 1987-1997 and its associated standard error.

We used Schatten's (1988) model of solar irradiance variation to drive our transient climate model from 1976 more than one solar cycle into the future. Figure 6 shows the equilibrium temperature forcing and the model's response. In general, the imposed solar

Figure 5. Solar irradiance model from 1976-1997 from a model incorporating effects of global faculae and active cavity regions as a function of solar latitude and time. The projection into the next sunspot cycle is based on Schatten and Sofia's (1987) prediction of sunspot number; the solar constant for zero solar activity was normalized to 1366.5 W m$^{-2}$ for comparison with ACRIM data. The model curves were supplied by K. H. Schatten of NASA/Goddard Space Flight Center (personal communication), and differ from those in Schatten (1988) insofar as they include a more realistic and gradual change of polar faculae during the cycle.
signal is both damped and phase-lagged by the thermal inertia of the oceans, and the response exhibits an upward drift as it seeks to oscillate about the mean equilibrium temperature of the forcing. (This would occur asymptotically for a large enough number of identical cycles.) Hoffert et al. (1980) have shown that a 65-70% reduction of the amplitude of the response to identical 11-year sunspot cycles is caused by mixed layer damping. From Fig. 6 it is seen that the peak of the response is ~ 0.05°C which is consistent with this estimate. Since the amplitude of the response will increase if the forcing period becomes large (while keeping the forcing amplitude constant), a model driven by the ~ 80-year Gleissberg cycle will exhibit a temperature response that will be damped to a smaller extent.

Note that the “historical” temperatures of Jones et al. (1986) and Hansen and Lebedeff (1988) in Fig. 6 are smoothed over interannual variations and contain the effects of atmospheric aerosol and greenhouse-gas variations as well as solar forcing. The variance in the observed signal is substantially larger than that predicted by solar effects alone, but does not appear inconsistent with such effects. The (linear) thermodynamic upwelling-diffusion ocean model results discussed here indicate that currently available satellite data are sufficient to rule out a major solar-variation effect on surface temperature in the short term (but longer-term effects are still possible and may even be quite important). We should also emphasize that our 1-D ocean model may not be entirely appropriate for analyzing the local response to short time scale forcing, particularly in low thermal inertia continental interiors, and the possibility of localized nonlinear amplification cannot be ruled out. Moreover, we cannot exclude the possibility that nonlinearities in the climate system amplify and modulate imposed forcing in ways not captured by the current linear models (Gaffen et al., 1986).

In light of these results, it seems extremely important to continue monitoring the variations of solar flux from space in the coming decades, perhaps from a dedicated instrument on the proposed NASA Earth Observing System (EOS). Since some transient

![Figure 6](image_url)  
**Figure 6.** Equilibrium and transient global temperature response from the transient climate model for Schatten's (1988) solar forcing over the present and subsequent solar cycle.
climate models predict the emergence of a fossil fuel carbon dioxide warming signal in this time frame, it would be particularly useful to know the components of the signal that can be attributed to other effects, including solar irradiance changes. Hopefully, future generations will have the data needed to resolve the transient effects of our dynamic Sun on climate, as long time-series irradiance monitoring from space becomes an operational fact of life.

ACKNOWLEDGEMENTS

We thank Tom Wigley for drawing our attention to the "history effect" and its possible importance. This work was supported at New York University by the National Aeronautics and Space Administration (Solar Maximum Guest Investigator Program) under Grant NAG 5-503 and by the Department of Energy (Carbon Dioxide Research Division) under Grant DE-FGO2-85EK60350.

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The Volcanic Contribution to Climate Change of the Past 100 Years

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ABSTRACT. Volcanic eruptions which inject large amounts of sulfur-rich gas into the stratosphere produce dust veils which last several years and cool the earth’s surface. At the same time these dust veils absorb enough solar radiation to warm the stratosphere. Since these temperature changes at the earth’s surface and in the stratosphere are both in the opposite direction to the hypothesized effects from greenhouse gases, they act to delay and mask the detection of greenhouse effects on the climate system.

A large portion of the global climate change of the past 100 years may be due to the effects of volcanoes, but a definitive answer is not yet clear. While effects over several years have been demonstrated with both data studies and numerical models, long-term effects, while found in climate model calculations, await confirmation with more realistic models. In this paper chronologies of past volcanic eruptions and the evidence from data analyses and climate model calculations are reviewed.

1. INTRODUCTION

Since 1784, when Benjamin Franklin suggested that the Laki eruption in Iceland in 1783 might have been responsible for the abnormally cold winter of 1783–4 (Franklin, 1784), emissions from volcanoes have been implicated as a possible cause of weather and climate variations. Conventional wisdom (Ramanathan, 1988; Self and Rampino, 1988) holds that the effects of volcanoes on climate have not yet been demonstrated, but other work has claimed that volcanoes can be important causes of hemispheric temperature changes for several years following large eruptions, and even on a 100-year time scale when their cumulative effects are taken into account.

Other forcings of climate change on the 100-year time scale include greenhouse gases, solar constant variations, internal (to the climate system) oscillations, and random, stochastic variations. The problem of identifying a volcanic signal in the past is the same as the problem of identifying a greenhouse signal in the past—separating out the effects of other potential forcings. Because the climatic signal of volcanic eruptions is of
approximately the same amplitude as that of El Niño/Southern Oscillation (ENSO), and because there have been so few large eruptions in the past century, it has been difficult to separate the volcanic signal from that of other simultaneous climatic variations. However, it has recently been demonstrated (Angell, 1988; Nicholls, 1988; Mass and Portman, 1989) that the ENSO signal in the past climatic record partially obscures the detection of the volcanic signal for surface air temperature on a hemispheric, annual-average basis. Angell (1988) and Mass and Portman (1989) have conducted preliminary studies in which the volcanic signal has been extracted more clearly.

It is beyond the scope of this paper to give a comprehensive review of the effects of volcanoes on climate. Excellent reviews of this type include Lamb (1970), Toon and Pollack (1980), Toon (1982), Ellsaesser (1983), Asaturov et al. (1986) and Kondratyev (1988). Theoretical studies of the radiative effects include Pollack et al. (1976) and Harshvardhan (1979). In this paper several new aspects of the problem of relating volcanic eruptions to climate will be addressed, including the problem of volcanic indices, methods of conducting superposed-epoch analyses, and the relationship of volcanic eruptions to ENSO events.

In order to determine the effects of past volcanic eruptions, it is first necessary to have a record of the past eruptions that have had an impact on the climate. This problem of volcanic indices is discussed first. Next, various analyses of past data are compared. Then climate model calculations are reviewed. Finally, the collective evidence is discussed and conclusions are presented.

2. VOLCANO INDICES

It has become clear in the last decade (e.g., Rampino and Self, 1984) that the effect of a volcano on climate is most directly related to the sulfur content of emissions that reach into the stratosphere, and not to the explosivity of the eruption. These sulfur gases convert to small sulfate particles, which persist for several years in the stratosphere and efficiently scatter the incoming sunlight, reducing the direct and total solar radiation reaching the ground.

To investigate the effects of volcanic eruptions on climate, it would be desirable to have a volcanic index that is proportional to the physical effect of the volcanic dust veil on climate, namely, the net radiation deficit. If the index is incomplete in its geographical or temporal coverage, if it assumes that surface air temperature drops after an eruption and uses this information to create the index, or if it is a measure of some property of volcanic eruptions other than their long-term stratospheric dust loading, it will be unsuitable for this type of study. All volcanic indices produced so far suffer from one or more of these problems. Yet if the various deficiencies of each index are kept in mind, they can be used cautiously which, as discussed in the next section, has not been the case in many instances.

The first extensive modern compilation of past volcanic eruptions is the classic study of Lamb (1970), updated by Lamb (1977, 1983). Lamb created a volcanic Dust Veil Index (d.v.i.) specifically designed for analyzing the effects of volcanoes on “surface weather, on lower and upper atmospheric temperatures, and on the large-scale wind circulation” (Lamb, 1970, p. 470). The methods used to create the DVI are described by Lamb (1970), and in more detail by Kelly and Sear (1982), and include historical reports of eruptions, optical phenomena, radiation measurements (for the period 1883 onward), temperature information, and estimates of the volume of ejecta.

The formula for the d.v.i. includes a term $E_{max}$ which gives an estimate of the fraction of the globe covered by the dust veil. In order to compare the amount of material emitted from volcanoes it is convenient to present $DVI = d.v.i./E_{max}$, which “indicates the magnitudes of the eruptions as dust producers without regard to the area over which
the dust veil may have been spread by the general circulation of the atmosphere” (Lamb, 1970, p. 500), as is done in Table 1. Although the DVI for the Mt. St. Helens eruption of 1980 was 500, and DVI for El Chichón of 1982 was 800, $E_{\max}$ for Mt. St. Helens was 0.3, while $E_{\max}$ for El Chichón was 1. Therefore, Lamb’s (1983) estimate of the relative climatic effect of the two volcanoes was different by a factor of more than 5.

Lamb’s DVI has often been criticized (e.g., Bradley, 1988) as having used climatic information in its derivation, thereby resulting in circular reasoning if the DVI is used as an index to compare to temperature changes. In fact, for only a few eruptions between 1763 and 1882 was the Northern Hemisphere (NH) averaged DVI calculated based solely on temperature information. Robock (1981a) created a modified version of Lamb’s DVI which excluded temperature information. When used to force a climate model, the results did not differ significantly from those using Lamb’s original DVI, demonstrating that this is not a serious problem.

Mitchell (1970) also produced a time series of volcanic eruptions for the period 1850–1968 using data from Lamb. As discussed by Robock (1978, 1981a), the Mitchell volcanic compilation for the NH is more detailed than Lamb’s because Lamb excluded all volcanoes with DVI < 100 in producing his NH annual average DVI (Table 7(a), p. 526). Thus Mitchell’s volcanic series has proven to be very useful as a climatic volcanic index.

More recently a comprehensive survey of past volcanic eruptions (Simkin et al., 1981) produced the Volcanic Explosivity Index (VEI) (Newhall and Self, 1982) which gives a geologically based measure of the power of the volcanic explosion. Unfortunately, this index has been used without any modification in many studies (see next section) as an index of the climatological impact of volcanoes. A careful reading of Newhall and Self (1982), however, will find the following quotes: “We have restricted ourselves to consideration of volcanological data (no atmospheric data),” and “Since the abundance of sulfate aerosol is important in climate problems, VEI’s must be combined with a compositional factor before use in such studies.” In their Table 1, Newhall and Self (1982) list criteria for estimating the VEI in “decreasing order of reliability,” and the very last criterion out of 11 is “stratospheric injection.” For VEI of 3, this is listed as “possible,” for 4 “definite,” and for 5 and larger “significant.” If one attempts to work backwards and use a geologically determined VEI to give a measure of stratospheric injection, serious errors can result. Not only is this the least reliable criterion for assigning a VEI, but it was never intended as a description of the eruption which had a VEI assigned from more reliable evidence.

Eruptions with a high VEI may also have a large stratospheric impact, such as Tambora (1815, VEI = 7) or Krakatau (1883, VEI = 6), but three recent examples demonstrate the danger in using the VEI for climate studies. Mt. St. Helens in 1980 had a high VEI of 5, and while it had a large local temperature impact (Robock and Mass, 1982; Mass and Robock, 1982), it had a negligible stratospheric impact (Robock, 1981b). On the other hand, Agung in 1963 and El Chichón in 1982, had a very large stratospheric impact (Robock, 1983a) but a smaller VEI of 4. Several studies (discussed below) have been done using the VEI as an index for the climatic effect of volcanoes, and then excluded Mt. St. Helens as a special case. This example raises the question of the possibility of other special cases in the past for which we do not have the additional information as in this case.

Schönwiese (1988) has even created a Smithsonian Volcanic Index (SVI) which takes 10 to the VEI power, and includes volcanoes with VEI of 3 and greater. This is clearly not justified. As mentioned by Newhall and Self (1982), by combining information about the typical type of eruption that each volcano produces with the VEI, it may be possible to produce a “climatic VEI,” but it will probably be necessary to include additional information to produce a good index of the climatic impact of past eruptions.
Table 1. Data studies of the effects of volcanic eruptions on climate.

<table>
<thead>
<tr>
<th>Year</th>
<th>Volcano</th>
<th>VEI</th>
<th>DVI</th>
<th>Study*</th>
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</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>A</td>
<td>AK</td>
</tr>
<tr>
<td>pre-1800</td>
<td></td>
<td>1</td>
<td>18</td>
<td></td>
</tr>
<tr>
<td>mid 1800s</td>
<td></td>
<td>3</td>
<td>3</td>
<td>5</td>
</tr>
<tr>
<td>1855</td>
<td>Cotopaxi*</td>
<td>2</td>
<td>700</td>
<td>T</td>
</tr>
<tr>
<td>1872</td>
<td>Merapi</td>
<td>2</td>
<td>80</td>
<td>T</td>
</tr>
<tr>
<td>1875</td>
<td>T (A) Jokull (Askja)</td>
<td>5</td>
<td>1000</td>
<td>X</td>
</tr>
<tr>
<td>1883</td>
<td>Krakatoa</td>
<td>6</td>
<td>1000</td>
<td>X</td>
</tr>
<tr>
<td>1886</td>
<td>Tarawera</td>
<td>5</td>
<td>800</td>
<td>X</td>
</tr>
<tr>
<td>1888</td>
<td>Bandai</td>
<td>4</td>
<td>500</td>
<td>X</td>
</tr>
<tr>
<td>1890</td>
<td>Bogoslof*</td>
<td>2</td>
<td>170</td>
<td>N</td>
</tr>
<tr>
<td>1892</td>
<td>Awu</td>
<td>2</td>
<td>100</td>
<td>T</td>
</tr>
<tr>
<td>1902</td>
<td>Pelee</td>
<td>2×4</td>
<td>100</td>
<td>N</td>
</tr>
<tr>
<td></td>
<td>Soufriere</td>
<td>4</td>
<td>300</td>
<td>X</td>
</tr>
<tr>
<td></td>
<td>Santa Maria</td>
<td>6</td>
<td>600</td>
<td>X</td>
</tr>
<tr>
<td>1907</td>
<td>Ksudach</td>
<td>5</td>
<td>500</td>
<td>X</td>
</tr>
<tr>
<td>1912</td>
<td>Katmai (Novarupta)</td>
<td>6</td>
<td>500</td>
<td>X</td>
</tr>
<tr>
<td>1921</td>
<td>Puyehi</td>
<td>4</td>
<td>200</td>
<td>X</td>
</tr>
<tr>
<td>1932</td>
<td>Chup (Azul)</td>
<td>5</td>
<td>70</td>
<td>X</td>
</tr>
<tr>
<td>1947</td>
<td>Hekla</td>
<td>4</td>
<td>70</td>
<td></td>
</tr>
<tr>
<td>1953</td>
<td>Mt. Spurr</td>
<td>4</td>
<td>7</td>
<td>N</td>
</tr>
<tr>
<td>1953</td>
<td>Hibok-Hibok</td>
<td>2</td>
<td>missing</td>
<td>N</td>
</tr>
<tr>
<td>1955</td>
<td>Rancho, Puyehi</td>
<td>4</td>
<td>30–40</td>
<td>X</td>
</tr>
<tr>
<td>1956</td>
<td>Bezymianni</td>
<td>5</td>
<td>30</td>
<td>X</td>
</tr>
<tr>
<td>1963</td>
<td>Agung</td>
<td>4</td>
<td>800</td>
<td>X</td>
</tr>
<tr>
<td>1965</td>
<td>Taal</td>
<td>4</td>
<td>10–15</td>
<td>X</td>
</tr>
<tr>
<td>1965</td>
<td>Awu</td>
<td>4</td>
<td>200</td>
<td>W</td>
</tr>
<tr>
<td>1968</td>
<td>Fernandina</td>
<td>4</td>
<td>50–100</td>
<td>X</td>
</tr>
<tr>
<td>1970</td>
<td>Hekla</td>
<td>3</td>
<td>2</td>
<td>X</td>
</tr>
<tr>
<td>1970</td>
<td>Beerenberg</td>
<td>3</td>
<td>90</td>
<td>X</td>
</tr>
<tr>
<td>1974</td>
<td>Fuego</td>
<td>4</td>
<td>250</td>
<td>W</td>
</tr>
<tr>
<td>1980</td>
<td>Mt. St. Helens</td>
<td>5</td>
<td>500</td>
<td>X</td>
</tr>
<tr>
<td>1982</td>
<td>El Chichón</td>
<td>4</td>
<td>800</td>
<td>X</td>
</tr>
</tbody>
</table>

Continuous volcanic record? X X X
Used VEI? X X X X X X X X
Used DVI? X X X X X X X X X
Used ice acidity? X X X X X
Used radiation data? X X X X
Used lunar eclipse? X
Normalized? X X X
ENSO-corrected? X X

* See Table 2.
DVI = d.v.i./E_{max} of Lamb (1970)
* Mitchell calls doubtful, but used it.
+ not used, because of obscuration by other 1902 eruptions.
Table 2. Abbreviations used in Table 1.

<table>
<thead>
<tr>
<th>Study abbreviations</th>
<th>Time period</th>
<th>Type of climate data</th>
<th>Averaging time</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>1880-1986</td>
<td>SST, SAT, upper air</td>
<td>season</td>
</tr>
<tr>
<td>AK</td>
<td>1740-1983</td>
<td>SAT</td>
<td>season</td>
</tr>
<tr>
<td>B</td>
<td>1851-1984</td>
<td>SAT</td>
<td>month</td>
</tr>
<tr>
<td>H</td>
<td>79-1913</td>
<td>SAT</td>
<td>season</td>
</tr>
<tr>
<td>K</td>
<td>1300-1986</td>
<td>SST, cool summers, famines</td>
<td>season</td>
</tr>
<tr>
<td>KS</td>
<td>1881-1980</td>
<td>SAT</td>
<td>month</td>
</tr>
<tr>
<td>L</td>
<td>1602-1900</td>
<td>SAT, tree rings, ice</td>
<td>season</td>
</tr>
<tr>
<td>M</td>
<td>1845-1959</td>
<td>SAT</td>
<td>5 years</td>
</tr>
<tr>
<td>MP</td>
<td>1873-1985</td>
<td>SST, SAT, PR, PS</td>
<td>month</td>
</tr>
<tr>
<td>MS</td>
<td>1755-1973</td>
<td>SAT</td>
<td>year</td>
</tr>
<tr>
<td>S</td>
<td>1881-1980</td>
<td>SST, SAT</td>
<td>month</td>
</tr>
<tr>
<td>SC</td>
<td>1750-1985</td>
<td>SAT</td>
<td>year</td>
</tr>
<tr>
<td>SE</td>
<td>1755-1977</td>
<td>SAT</td>
<td>year</td>
</tr>
<tr>
<td>T</td>
<td>1811-1975</td>
<td>SAT</td>
<td>4 months</td>
</tr>
<tr>
<td>X</td>
<td>1500-1982</td>
<td>cold summers</td>
<td>season</td>
</tr>
<tr>
<td>Y</td>
<td>1951-1972</td>
<td>SAT</td>
<td>season</td>
</tr>
</tbody>
</table>

*SST = sea surface temperature, SAT = surface air temperature, PR = precipitation, PS = surface pressure.

Ice core analysis (e.g., Lyons et al., 1990) can give the chemistry and particle content of well-dated layers which can give a measure of the important volcanic parameters. Xu (1988) has actually used the Acidity Index of Hammer et al. (1980) as a volcanic index for comparison to climatic data. Unfortunately, small local eruptions can give as large a signal as distant large eruptions. By comparing the acidity and particle records from Greenland, Antarctic, and tropical ice cores, it may be possible to produce a global or hemispheric record from signals that appear simultaneously in all three regions, or in the tropics and one of the high-latitude cores.

Radiation measurements of the transmission of the direct solar beam give indications of the atmospheric turbidity. By combining measurements from many locations to eliminate local influences, Pollack et al. (1976), Pivovarova (1977), and Bryson and Goodman (1980) presented time series of radiation, interpreted as the volcanic loading of the atmosphere. Xu (1985) created a volcanic index based on radiation data, and Mass and Portman (1988) also give other sources of actinometric data. Because each of these data sets is incomplete spatially and temporally, and because local influences may not have been completely eliminated, they are by themselves not sufficient as a measure of volcanic influence on the atmosphere. By combining all the available radiation information, however, they would be a valuable input to a volcanic index.

Another source of information comes from lunar brightness during eclipses (Keen, 1983). In addition, lidar measurements (e.g., McCormick and Osborn, 1986), balloon sampling (e.g., Hoffmann and Rosen, 1984a,b), and aircraft sampling (e.g., Sedlacek et al. 1983) can all now give detailed measurements of the stratospheric aerosol concentration. Satellite data from the SAM II (e.g., McCormick and Brandl, 1986) and SAGE measurements (e.g., McCormick, 1987) also give measures of stratospheric aerosols. An instrument designed for measuring ozone, TOMS on Nimbus 7, can also pick up the signal of sulfur dioxide from volcanic eruptions, and in fact has been used by Krueger.
(personal communication) to identify the source of the “mystery cloud” of early 1982 as the Nyamuragira eruption of 26 December 1981 in eastern Zaire.

Until a good climatic-volcanic index is developed, all previous studies using inadequate indices must be evaluated cautiously. Since DVI, VEI and acidity index are correlated (Schönwiese, 1988), the results presented below are in partial agreement even if based on different indices. An objective, quantitative measure of the effects of volcanoes on climate, however, will require a better volcanic index.

3. ANALYSES OF PAST TEMPERATURE CHANGE

There have been many studies in the past attempting to link climatic changes with large volcanic eruptions. [Lamb (1970) even took temperature drops as indications of the size of volcanic eruptions, in a few cases without any other evidence, when creating his volcanic index.] These studies (summarized in Tables 1 and 2) range from case studies of a single large eruption or a few eruptions, to comparisons of time series of temperature to the timing of eruptions, to superposed-epoch analyses combining the signals of many eruptions. The different studies distinguish themselves from each other by their choices of volcanoes (or volcanic indices), temperature data sets (usually air at the surface, but also upper air and sea surface), time resolution, analysis technique (especially whether climatic data are normalized by their standard deviation for monthly data), and treatment of the ENSO signal. In this section the effects of the 1815 eruption of Tambora are discussed, and then analyses of more recent eruptions are compared.

3.1. Tambora – Cause of the “Year Without a Summer”?  

The book by Stommel and Stommel (1983), which is subtitled “The Story of 1816, The Year Without a Summer,” presents the fascinating story of the severe weather disruptions in New England and Western Europe which also resulted in 1816 being called, “Eighteen Hundred and Froze-to-Death” and “Poverty Year” (Humphreys, 1940). Stommel and Stommel’s book includes the stories of the record price of grain in London, Mary Shelley’s writing Frankenstein being influenced by the terrible summer weather on the shores of Lake Geneva, and the killing summer frosts in the United States and Canada (Robock, 1984c). Stommel and Stommel conclude that the case cannot be proven that the great eruption of Tambora in 1815 was responsible for the extreme weather of the next year because evidence was available only from a small region of the globe (eastern North America and Western Europe).

Recent studies, however, present new evidence of climate effects in both China and India in 1816, although Kondo (1988) found no evidence of effects in Japan. Hameed et al. (1989) found evidence in Chinese documents of abnormally cold weather from the winter of 1815–16 through the summer of 1817, manifesting itself in crop failures and snow in June, 1816. Sigurdsson and Carey (1988) point out that bad harvests in India in 1816 led to a famine which was followed by a serious cholera outbreak. In the next two decades the cholera spread to Europe and Asia as the greatest pandemic of the century.

One aspect of the 1816 events that is not widely recognized is the significant volcanic eruptions in each of the four years preceding Tambora (Sabrina in 1811, Soufriere and Awu in 1812, Vesuvius in 1813, and Mayon in 1814). Thus any effects felt in 1816 were the cumulative effect of five years of enhanced stratospheric aerosol loading. Stothers (1984) and Rampino and Self (1982), who presented detailed geological descriptions of the Tambora eruption, found a NH temperature depression of approximately 0.7°C in 1816 from a limited network of stations. Humphreys (1940) similarly found a depression of about 1.0°C in 1816. Both studies and Groveman and Landsberg (1979) found that the
NH temperature was cool for several years before 1816, and then rose rapidly, by more than 1°C, during the next 10 years. The antecedent cooling can be easily explained by the effects of the preceding volcanoes, but the subsequent strong warming is more difficult to understand. How good were the temperature records? Was this a response to a dust-free atmosphere? Were internal oscillations becoming dominant? Was there strong tropical ocean forcing of the climate system during this period? If so, was it made stronger by the volcanoes? Quinn et al. (1987) report no ENSO events from 1814 to 1828, but their record may be incomplete.

Thus the case of the volcanic eruptions of 1811–1815 and the severe weather of 1816 is strongly suggestive of the large potential short-term effect of volcanoes on climate. One individual case cannot prove the relationship, however, since other causes of interannual variability were undoubtedly playing a part simultaneously, but with unknown amplitude. Next, studies which combine the effects of several eruptions (although none with as large a stratospheric impact as Tambora) and study the effects on temperature with the improved data network of the past 100 years are presented.

3.2. Comparative Studies

Humphreys (1940), Yamamoto et al. (1975), Angell and Korshover (1985), Kondo (1988), Angell (1988), and Xu (1988) present time series of volcanic eruptions and climate change and comment on their correspondence. The volcanoes used by each are shown in Table 1. In each case the evidence is suggestive of a cause and effect, with varying degrees of agreement. All use surface air temperatures, except Kondo also used reports of crop failures and famines in Japan, Xu used reports of cold summers in China, and Angell also used sea surface and upper air temperatures.

Although visually comparing time series can suggest agreements, the superposed epoch technique, discussed next, can objectively filter out other effects and give a quantitative measure of the volcanic effect. Of course, if some other cause of climate variation is correlated with volcanic eruptions, the superposed-epoch technique will not remove it. This seems to be the case with ENSOs, and Angell (1988) showed that, by removing a signal correlated with sea surface temperature (SSTs) in the tropical Pacific with a 6-month lag, the volcanic signal is made clearer. This argument still may suffer from circular reasoning, since the SST is also part of the signal being measured.

3.3. Superposed-Epoch Analyses

In superposed-epoch analysis a key date is identified for each volcanic eruption, the resulting temperature data are superposed, and the average is then used to measure the signal of volcanic eruptions. This has been done for data-averaging periods on time scales ranging from five years (Mitchell, 1961) to one year (Mass and Schneider, 1977; Schönwiese, 1988) to one season (Taylor et al. 1980; Angell and Korshover, 1985; Lough and Fritts, 1987) to one month (Self et al. 1981; Kelly and Sear, 1984; Sear et al. 1987; Bradley, 1988; Mass and Portman, 1989).

When a one-month averaging period is used, the months are counted starting with the month of the eruption, so that if two volcanoes occurred in different years, say one in April and one in August, then the effects after three months would be an average of July and November data. Thus, if there is a seasonal component to the response of the climate system, it cannot be identified with this technique. Recognizing that climate variability is larger in the winter, in order to avoid averaging large and small variations together, Kelly and Sear (1974), Sear et al. (1987), and Bradley (1988) looked at normalized monthly surface air temperatures, with the temperature anomalies divided by the standard deviation
of temperature for that month. However, this analysis technique gives less weight to winter temperature fluctuations and also works to obscure the seasonal cycle of the temperature response.

Mass and Portman (1989) have removed an ENSO signal, in a manner similar to Angell (1988), and find a definite volcanic signal.

3.4. Seasonal Cycle Analysis

It has been shown by Robock (1983b) that the sea ice/thermal inertia feedback is responsible for the amplification of climate response in high latitudes during winter for equilibrium climate simulations with an energy-balance climate model. This also explains the results Manabe and Stouffer (1980) obtained with a general circulation model (GCM). For transient experiments with volcanic eruptions (Robock, 1981b, 1984a; discussed below) and for nuclear winter forcing (Robock, 1984b; Vogelmann et al. 1988), it has also been shown with an energy-balance model that the sea ice/thermal inertia feedback causes an amplification of the seasonal cycle when the climate system cools, resulting in more cooling in the polar regions in winter. Yamamoto et al. (1975) also found winter polar enhancement of the volcanic cooling for several eruptions.

Since volcanic eruptions are thought to result in cooling of the climate system for a few years, Robock (1985) presented a preliminary analysis in which the amplitude of the seasonal cycle in high latitudes is examined by doing an analysis of surface temperature that compares all months from different years. This analysis also solves another problem of previous studies, namely, that large volcanic eruptions can sometimes occur close to each other in time, and a superposed-epoch analysis must make the assumption that the year or years before the key date (date of the eruption) represent a normal climate. Since the climate system is constantly fluctuating, Robock (1985) examined the overall level of volcanic forcing and compared it to the corresponding response of the climate system.

Robock (1985) presented a volcanically weighted temperature variation, but it seems more straightforward to present the correlation coefficient between temperature variations and a volcanic index. An example of such a calculation is shown in Fig. 1, which shows the correlation between the volcanic dust veil index from Mitchell (1970) and the surface air temperature over land for 1881 through 1981 from the Russian surface temperature data set (Robock, 1982) for a lag of two years (volcanic data with temperature two years later). Although the ENSO signal has not been removed and the volcanic and temperature data sets are not optimal, it can still be seen that the correlation is negative for almost all months and latitudes (showing cooling after eruptions), and that the correlation is amplified at the North Pole in winter. It is expected that if the same analysis were done with the ENSO signal removed from an improved temperature data set, such as the global set from the University of East Anglia, (Jones et al. 1988), and with an improved volcanic index, that it would be possible to establish a typical volcanic signal in a more definitive manner than before.

3.5. Stratospheric Effects

Even though the data record for the stratosphere is shorter than that for the troposphere, large warming beyond the level of the quasi-biennial oscillation has been measured in the stratosphere following the Agung and El Chichón eruptions (e.g., Labitzke et al., 1983; Parker and Brownscombe, 1983; Angell and Korshover, 1983; Quiroz, 1983, 1984;
Figure 1. Zonal average of correlation between surface air temperature (Robock, 1982) and volcanic dust veil index (Mitchell, 1970) for land grid points only, for 1881–1981, for a lag of two years. The large correlation at the pole in the winter, the small correlation at the pole in the summer, and the negative correlation at almost all latitudes and months lend support to the proposed effect of volcanoes on climate.

Fujita, 1985; Wendler and Kodama, 1986). This is in the opposite direction to the anticipated effects from greenhouse cases which will be “virtually certain” to cause large stratospheric cooling (National Research Council, 1987). Thus these large stratospheric effects of volcanoes, while lasting only a few years, will have to be dealt with when modeling or interpreting climatic data in attempts to identify a greenhouse signal.

4. CLIMATE MODEL CALCULATIONS

4.1. Energy-Balance Models

Energy-balance climate model calculations by Schneider and Mass (1975), Oliver (1976), Robock (1978, 1979, 1981b, 1984a), Gilliland (1982), and Gilliland and Schneider (1984) have all shown cooling effects due to volcanic eruptions for several years. As mentioned above, Robock (1984a) demonstrated the winter polar enhancement of cooling due to the sea ice/thermal inertia feedback, the same pattern seen in the observations (Fig. 1).

Robock (1978, 1979, 1981b) demonstrated a long-term effect of volcanic eruptions with the warming of the 1920s and 1930s resulting from the lack of significant eruptions. Oerlemans (1988) coupled an energy-balance climate model with a glacier model and found that about half of the observed long-term behavior of glaciers for the past 100 years can be explained by volcanoes and half by greenhouse gases. Porter (1981) had previously found a relationship between glacier advances and volcanoes for the past 100 years. The models used for these calculations had only a 75m mixed-layer ocean, and
these longer time scales of the climate system may depend on deep ocean circulation (Hansen et al. 1985). Experiments with coupled atmospheric and oceanic GCMs will be necessary to confirm this result.

4.2. Radiative-Convective Models

Hansen et al. (1978) and Vupputuri and Blanchet (1984) have used radiative-convective models to calculate the vertical distribution of the climatic effect of the Agung and El Chichón eruptions, respectively. They both found cooling at the surface and warming in the stratosphere, which corresponds to the observations mentioned above.

4.3. Zonally-Averaged Model

MacCracken and Luther (1984) used a zonally-averaged dynamic climate model to calculate the vertical and latitudinal response to the El Chichón eruption. They found cooling at the surface in agreement with the energy-balance calculations of Robock (1984a), but in addition found intriguing precipitation and circulation anomalies caused by a shift in the ITCZ which they suggested may be related to El Niño generation.

4.4. General Circulation Model (GCM) Simulations

Hunt (1977) presented the first GCM calculation of the effects of a volcano on climate, but it was done with a crude model and did not examine seasonal effects. Hansen et al. (1988) recently performed time-dependent simulations of climate from 1960 through 2050 with a GCM at the NASA Goddard Institute for Space Studies (GISS). In three different simulations with different amounts of greenhouse gases, the effects of a total of 12 large and 9 small volcanoes were shown to cause cooling for several years. Although Hansen et al. (1988) present hemispheric, annual-average results of these simulations which show cooling from the volcanoes, the seasonal and latitudinal signals of the volcanoes have not been analyzed in their output. In addition, the simple ocean model used precluded the precise determination of long-time-scale effects.

It is obvious that only with GCM studies of the effects of volcanic eruptions will the subtle interactions of the climate system, including possible ENSO or monsoon relationships, and the geographical distribution of effects be determined. Currently ongoing experiments include those of Graf (1989) at the Max Planck Institute in Hamburg and of Hansen, McCormick and Pollack (Hansen et al. 1988) at GISS in New York.

5. DISCUSSION AND CONCLUSIONS

5.1. Short-term Effects

The overall impression of the weight of both observational and modeling results is that large volcanic eruptions, which produce substantial stratospheric dust veils, produce global-average surface cooling for several years and stratospheric warming. The cooling is enhanced at the poles in the winter. Finer time- and spatial-scale effects are difficult to identify due to other competing climate variations.
5.2. Long-term Effects

The case for the effects of volcanoes on long-term (decadal and century time scales) climate change is suggestive, but not as strong as that for short-term effects. As a cause of the warming of the 1920s and 30s, the lack of volcanic aerosols in the atmosphere is certainly a prime candidate. But the rapid warming during the decade after Tambora (1815–1824) must be adequately explained before the long-term effects are completely understood. Certainly a period of enhanced volcanism (of the type which produces climatic effects) in the near future would work in the opposite direction from the greenhouse effect in both the troposphere and stratosphere, and thereby complicate the interpretation of the causes of the future course of climate.

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Chaos, Spontaneous Climatic Variations and Detection of the Greenhouse Effect

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ABSTRACT. We illustrate some of the general properties of chaotic dissipative dynamical systems with a simple model. One frequently observed property is the existence of extended intervals, longer than any built-in time scale, during which the system exhibits one type of behavior, followed by extended intervals when another type predominates. In models designed to simulate a climate system with no external variability, we find that an interval may persist for decades. We note the consequent difficulty in attributing particular real climatic changes to causes that are not purely internal. We conclude that we cannot say at present, on the basis of observations alone, that a greenhouse-gas-induced global warming has already set in, nor can we say that it has not already set in.

1. INTRODUCTION

In the minds of many of us who are gathered here, the most important question concerning greenhouse gases is not whether they will produce a recognizable global warming, but when will they do so? Probably we take it for granted that, barring some catastrophe that halts or overwhels the accumulation of carbon dioxide and other constituents, the warming predicted by theoretical studies will eventually occur. The apparent upward trend of global-average temperature during the most recent century, and the unusually warm and dry weather that has invaded parts of the world during parts of the most recent decade, have led some of us to speculate that the greenhouse warming is already being felt. Figure 1, which is transcribed from the cover of a recent report of the National Climate Program (1988), is typical of the graphical documentations of the situation that faces us. In this talk I wish to examine the basis for speculating that the greenhouse effect is not the main cause of what we have been experiencing and, particularly, that the suggested warming is due to processes purely internal to the atmosphere and its immediate surroundings.
2. CHAOS

Except for those living in special regions of the globe, including some tropical areas, everyone is familiar with the day-to-day and week-to-week irregular alternations between heat and cold, or sunshine and rain, that typify our weather. These fluctuations are one manifestation of what has recently been called "chaos." The term "chaos" currently has a variety of accepted meanings, but here we shall use it to mean deterministically, or nearly deterministically, governed behavior that nevertheless looks rather random. Upon closer inspection, chaotic behavior will generally appear more systematic, but not so much so that it will repeat itself at regular intervals, as do, for example, the oceanic tides.

Before proceeding to an acceptable working definition of chaos, we shall discuss a particular example. Consider the system of three equations

\[
\frac{dX}{dt} = -Y^2 - Z^2 - aX + aF
\]
\[
\frac{dY}{dt} = XY - bXZ - Y + G
\]
\[
\frac{dZ}{dt} = bXY + XZ - Z
\]

in the three dependent variables X, Y and Z, and the independent variable t representing time. This system was originally introduced as a simplified model of large-scale atmospheric flow (Lorenz, 1984a), but our purpose for presenting it here is simply to illustrate some of the properties of chaotic systems in general. We do not claim or intend that it should reproduce atmospheric behavior in any quantitative sense.

The equations have been scaled so that the time unit is 5 days. They may easily be "solved" numerically. We have let \( a = 0.25, \ b = 4.0, \ F = 8.0 \) and \( G = 1.0 \), and have used a fourth-order Taylor-series integration scheme with a time step of 0.025 units, or 3 hours. For initial conditions we have chosen \( X = Y = Z = 1.0 \). Figure 2 shows four consecutive one-year segments of the time series for X.

![Figure 2. Variations of the variable X during four consecutive years in a numerical solution of the three-variable system. The time scale is in months.](image-url)
We see that \( X \) may behave quite differently during different intervals of several months, but that the types of behavior, or "regimes," are limited in number. Near the beginning and the end of the first year and the middle of the second year, and through much of the fourth year, \( X \) undergoes large-amplitude fluctuations with successive maxima or minima typically separated by 3 or 4 weeks. In much of the third year there are weaker oscillations with a period close to 2 weeks. During brief intervals there are still weaker oscillations with maxima about a week apart. Although there is a high probability that one of these three regimes will prevail at any particular moment, changes from one regime to another occur at virtually random intervals, and in no obvious fixed order. Such a combination of short-term regularity and long-term irregularity is shared by many chaotic systems.

The regimes will show up more clearly in a time series of some quantity chosen to reveal them. One such quantity is the standard deviation of \( X \) within an interval long enough to include several maxima and minima. In Fig. 3 we show, on a more compressed horizontal scale than in Fig. 2, the standard deviation \( \sigma \) of \( X \) within running 45-day intervals, for twelve consecutive years. The regimes of large-amplitude and smaller-amplitude fluctuations show up plainly as periods where \( \sigma \) is near 0.8 or 0.4, while

![Graph showing variations of \( \sigma \), the standard deviation of \( X \) within the 45-day interval centered at the indicated time, during three consecutive 4-year intervals in a numerical solution of the three-variable system. The first four years are the same as those in Fig. 2. The time scale is in years.](image-url)
the still weaker oscillations correspond to brief minima near or below 0.3. The large-amplitude fluctuations evidently dominate during the second 4-year segment, but become less prevalent during the third.

The cause of the irregularity in Figs. 2 and 3, and of chaos in general, is instability with respect to small-amplitude perturbations, which manifests itself as sensitive dependence on present conditions. This means that solutions originating from nearly identical states will evolve in due time into considerably, and sometimes unrecognizably, different states. In Fig. 2 the behavior during July of year 2, for example, is a close analogue of the behavior during April of year 1. The segments of the solution following these months remain similar for several more weeks, but then go their own ways. Without sensitive dependence, the entire remainder of year 2, and everything thereafter, would have had to repeat what happened fifteen months earlier, and October of year 3 would also have been an analogue of April of year 1. In short, the solution would have been periodic and easily predictable. Our definition of a chaotic system will be one whose general behavior exhibits sensitive dependence.

Note particularly that in the 3-variable system, F and G, which represent external forcing, are constants. The shorter-period fluctuations and the oscillations between regimes are therefore spontaneous. The longest built-in time scale is $l/a$, or 20 days, but the regimes can persist for months. Variations with unexpectedly long periods are frequent features of chaos.

3. LONG-TERM SPONTANEOUS VARIATIONS

The regimes appearing in Figs. 2 and 3 are distinguished mainly by the manner in which X oscillates. They show no great difference in the mean value of X, and no tendency to last for many years, and it is not obvious from this example that chaos has much to do with our present problem.

The situation is quite different with another model (Lorenz, 1984b), with 27 variables, also representing atmospheric flow, but incorporating more physical processes. The physical variables include the mixing ratios of water vapor and liquid water. The atmosphere and the ocean exchange heat and water through evaporation and precipitation, and the model produces its own clouds, which reflect incoming solar radiation, while both phases of water absorb and re-emit infrared radiation. Figure 4 shows model-produced sea-level air temperatures, averaged over the globe and over one-year intervals, for 400 successive years (see Lorenz, 1986). We see upward and downward trends lasting through one or several decades, resulting in a total range of about 2°K.

The resemblance of Fig. 4 to records of real global-mean temperature is hard to overlook. Almost certainly, if the real variations had taken this course, we would at various times have been seeking the causes, and would at least have seriously considered the possibility that these causes were external. Yet in the 27-variable model there is no variability of any external feature, or of any atmospheric constituent other than water. The sea-surface temperature field is not prespecified; it is a dependent variable controlled by the model, and its globally averaged variations closely parallel those of globally averaged air temperature.

The long-period variations produced by the model can be traced to a positive cloud-albedo feedback process, which can be eliminated by replacing the variable albedo by a constant. In the model, higher temperatures tend to suppress some of the cloudiness, thereby allowing the solar heating to raise the temperature still more. The heating does not, however, cause the temperature to run away, or to level off at an extreme value; the migratory synoptic systems, which fluctuate chaotically with periods of weeks or months, act as a sort of random forcing for the globally averaged conditions.
The point is not that the model is a good representation of the atmosphere. It is not. In particular, the real atmosphere may very well not possess an important cloud-albedo feedback process. The point is that, in chaotic dynamical systems in general, very-long-period fluctuations, much longer than any obvious time constants appearing in the governing laws, are capable of developing without the help of any variable external influences. The atmosphere and its surroundings constitute a chaotic dynamical system, and we cannot without careful investigation reject the possibility that this system is one where spontaneous long-period fluctuations occur.

Computers have now reached the point where long-term integrations with rather large models are economically feasible. As one example, we show in Fig. 5 a 100-year series of annual-mean global-mean temperatures produced by the GISS Model II (Hansen et al., 1988) in a control run with no variable external influences. Again the solution is patently chaotic, and bears a reasonable resemblance to recent real-world events (compare Figs. 1 and 5). Of special interest for the present discussion, the model consists of nearly 100,000 equations. Spontaneous climatic-scale variations are evidently not restricted to models with only a few degrees of freedom.

Unfortunately, recognizing a system as chaotic will not tell us all that we might like to know. It will not provide us with a means of predicting the future course of the system. It will tell us that there is a limit to how far ahead we can predict, but it may not tell us what this limit is. Perhaps the best advice that chaos theory can give us is not to jump at conclusions; unexpected occurrences may constitute perfectly normal behavior.
4. DETECTION OF GREENHOUSE WARMING

In view of these considerations, how are we to know when a stronger greenhouse effect is finally making its presence felt? First, we must realize that we are not looking for the onset of the effect. Presumably there is no critical concentration of CO$_2$ or some other gas beyond which the greenhouse effect begins to operate; more likely absorption of terrestrial re-radiation by CO$_2$ varies continuously, even if not truly linearly, with the concentration. This concentration has been steadily increasing for many years; hence, if the effect exists at all, its onset must have occurred long ago. What we are looking for is the time when the effect crosses the threshold of detectability.

It has sometimes been objected that in dealing with this problem we have relied too heavily on theory, but I would maintain that the problem cannot be wholly dissociated from theoretical considerations. Imagine for the moment a scenario in which we have traveled to a new location, with whose weather we are unfamiliar. For the first ten days or so the maximum temperature varies between 5° and 15°C. Suddenly, on two successive days, it exceeds 25°C. Do we on the second warm day, or perhaps on the first, conclude that somebody or something is tampering with the weather? Almost surely we do not; we are quite familiar with this sort of behavior, and we take it for granted that this is what the weather often does.

Consider now a second scenario where a succession of ten or more decades without extreme global-average temperatures is followed by two decades with decidedly higher averages; possibly we shall face such a situation before the 20th century ends. Does this scenario really differ from the previous one? We may feel that it does; for example, we
may believe that if the atmosphere is subjected to similar external influences over separate long intervals, say decades, the average conditions should be similar, with the short-period fluctuations tending to cancel. If so, our conclusions have been reached through theory, that is, through what we believe is demanded by the physical laws, even though the theory may be qualitative and not worked out in detail. Certainly no observations have told us that decadal-mean temperatures are nearly constant under constant external influences. If we discard all theoretical considerations, we cannot distinguish between the two scenarios.

At this point we may, in the second scenario, turn to statistical procedures. We may introduce a null hypothesis, which could say that the mean of the population of decadal mean temperatures to which the last two observations belong is not different from the mean of the population to which the earlier observations belong. We would then seek the probability that a discrepancy as large as the one that we have observed would occur, if the null hypothesis is true. If the probability is rather small, we would be likely to reject the null hypothesis, and conclude that the populations do indeed have different means. If the probability is large, the populations may still have different means, but we will lack a basis for concluding that they do.

Let us note, then, that we could introduce a similar null hypothesis in the first scenario. We might easily be led to reject the null hypothesis, incorrectly, if we should fail to take into account the high persistence of daily maximum temperatures, which renders the number of independent observations far less than the total number. That daily maximum temperatures are indeed persistent is readily verified from many years of records.

Returning to the second scenario, should we assume that decadal mean temperatures are also highly persistent? Our observations are insufficient to yield an answer, but we may turn to theory. The high persistence revealing itself in Figs. 3–5 suggests the possibility that real atmospheric decadal-mean temperatures are persistent; at least, it indicates that there is no obvious theoretical reason for hypothesizing that they are not persistent, no matter what intuition might tell us. There is a good chance, then, that in a real situation resembling the second scenario, we might not be able to reject the null hypothesis, that is, we might have to say that any change in the climate, including a change brought about by the greenhouse effect, has yet to cross the threshold of detectability.

If our only evidence were observational, we might have to pause at this point, and wait for more years of data to accumulate. However, since we do have theoretical results, and since, in fact, the entire greenhouse effect would have remained unsuspected without some theory, we can put the theory to use. Different models agree reasonably well as to the increase in globally averaged sea-level temperature that would be produced by a prescribed increase in CO₂ concentration. We are therefore equally justified in introducing a second null hypothesis, which would say that the difference between the means of two populations, one to which the earlier decadal mean temperatures belong, and one to which the later ones belong, is not different from the numerical value that the consensus of the theoretical studies stipulates. Again, we can ask whether anything as unusual as the difference between the observed sample means would be likely to have occurred, if the new null hypothesis is correct. Again, there is a good chance that we might lack sufficient evidence for rejecting the new hypothesis.

This somewhat unorthodox procedure would be quite unacceptable if the new null hypothesis had been formulated after the fact, that is, if the observed climatic trend had directly or indirectly affected the statement of the hypothesis. This would be the case, for example, if the models had been tuned to fit the observed course of the climate. Provided, however, that the observed trend has in no way entered the construction or operation of the models, the procedure would appear to be sound.
What we would conclude, then, if the second scenario is realistic, is that, although we may lack sufficient direct evidence that an increased greenhouse effect is influencing our climate, we just as surely lack direct evidence that it is not. If the effect is important, we may have to wait a few years to verify that it is, but, by the same token, if it is not important, we may have to wait a few years to verify that it is not. The implications of such a conclusion for future decision making speak for themselves.

What does chaos have to do with these claims? Without chaos, the numerical solutions summarized in Figs. 2-5 would have been periodic, repeating at regular intervals. Unless these periods proved to be extremely long, successive decadal or even annual means would have exhibited very little variability, and the qualitative theoretical conclusion that similar external conditions must produce similar long-term averages could not be so easily refuted. In that event, we might already be able to reject the original null hypothesis, and conclude that we are presently experiencing a climatic change, which in turn might be due to an increased greenhouse effect.

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Part 5:

Comparison of Model Simulations and Observations: Has a Greenhouse-Gas-Induced Climatic Change Been Detected?
Natural Climate Variability in a Coupled Model

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ABSTRACT. Multi-century simulations with a simplified coupled ocean-atmosphere model are described. These simulations reveal a wide range of variability on decadal and longer time scales, in addition to the dominant interannual El Niño/Southern Oscillation signal that the model originally was designed to simulate. Based on a very large sample of century-long simulations, it is nonetheless possible to identify distinct model parameter sensitivities that are described here in terms of selected indices. Preliminary experiments motivated by general circulation model results for increasing greenhouse gases suggest a definite sensitivity to model global warming. While these results are not definitive, they strongly suggest that coupled air-sea dynamics figure prominently in global change and must be included in models for reliable predictions.

1. INTRODUCTION

Whether we choose to define climate as the average of conditions occurring over a single season, or over a million years, we observe that it is continually changing. For many regions the normal march of the seasons each year represents the largest climatic signal. The ultimate driving for the annual cycle is the cyclical variation of incoming solar radiation, a process external to the climate system. The same has been argued for climate variations on the time scale of ice ages. In either case, the actual expression of the climatic fluctuation at any particular location depends not only the external forcing itself, but also (and perhaps more importantly) on the complex set of internal feedbacks operating on that external forcing. In addition to such externally driven climatic variations, we also observe phenomena having no association with external forcing, which must be regarded as strictly internal modes of the climate system. A prominent case in point is the El Niño/Southern Oscillation (ENSO) phenomenon, operating primarily on interannual (2–10 year) time scales. This phenomenon is of particular interest as it stems fundamentally
from the dynamical interaction of the atmosphere and ocean, thereby embodying physics crucial to all longer-term climate variability as well. Moreover, ENSO is of great importance because of its impacts. For some regions, especially in the tropical Pacific sector, the climate signal associated with ENSO far exceeds that of the annual cycle, and is indeed the primary signal of interest from a socio-economic point of view. To a lesser extent, ENSO has been associated with significant climate events on a worldwide basis, and it figures prominently in measures of global climate such as mean surface temperature.

Turning to the issue of anthropogenic climate change, we are faced with the sobering results of several modeling studies, which suggest a decadal-scale climate response to reasonable estimates of increases in greenhouse gases (e.g., Hansen et al., 1988; Schlesinger and Mitchell, 1985; Washington and Meehl, 1984; Manabe and Wetherald, 1975). Some of these studies have addressed the issue of how the extremes of the annual cycle might change. At least as important, however, is the issue of how the important internal modes of climate variability such as ENSO might be altered in a greenhouse scenario. It might be argued that, given the impact of interactive ocean-atmosphere dynamics on global climate as manifested in ENSO, any attempt to simulate longer-term climate change without including these processes explicitly is open to question. The related issue of detection of climatic signals in observations also must be viewed in the context of natural variability on interannual and longer time scales. Unless the range and the characteristic patterns of natural climatic variability are understood, it is impossible to make meaningful assessments of externally induced climate change.

In this paper we will address the issue of natural climate variability associated with tropical ocean-atmosphere interaction, focusing on the dynamics of ENSO. The questions of interest include the following: What are the characteristic patterns of this variability? Can the same processes responsible for ENSO on interannual time scales also give rise to lower-frequency variability? What is the sensitivity to various physical parameters, and specifically, how might the variability be altered in an environment of increasing concentrations of greenhouse gases?

2. BACKGROUND

Our study is based on a coupled model of the tropical Pacific which, although simplified in many respects, produces aperiodic interannual oscillations bearing many features in common with those observed (Zebiak and Cane, 1987). It is one of several models that together have led to a converging view of ENSO as an internal oscillation of the tropical ocean-atmosphere system (see also Cane and Zebiak, 1985; Schopf and Suarez, 1988; Hirst, 1988; Battisti, 1988).

Both the atmospheric and oceanic components of the model describe perturbations about the mean climatological state, with the monthly climatology specified from observations (Rasmusson and Carpenter, 1982). The atmospheric dynamics take the form of the steady, linear shallow-water equations on an equatorial beta-plane, with the circulation forced by a heating anomaly which depends partly on local heating associated with sea surface temperature (SST) anomalies and partly on the low-level moisture convergence. The latter effect is nonlinear because latent heating occurs only when the total wind field is convergent, and this is a function of the background mean convergence as well as the calculated anomalous convergence.

The ocean model is set up in a rectangular basin extending from 124°E to 80°W, and 29°S to 29°N. The dynamics begin with the shallow-water equations, but are modified by the addition of a shallow frictional layer of constant depth 50 m in order to simulate the intensification of wind-driven currents near the surface. The thermodynamics describe
the solution of temperature anomalies in the model surface layer. All terms, including nonlinear advection components, are retained. Finally, temperature anomalies below the base of the surface layer are parameterized in terms of vertical displacements of the model thermocline (identified with movements of the layer interface).

3. VARIABILITY WITHIN A MULTI-CE NTURY MODEL SIMULATION

Figure 1 shows a time series of NINO3-averaged SST anomaly (90°W-150°W, 5°N-5°S) from a 1024-year run of the model. Inspection of this time series reveals a wide range of behavior: there are well-marked periods of large-amplitude, rather regular oscillations, and other periods of more chaotic, smaller-amplitude variability. Also of note are the period of nearly twenty years with negligible anomalies (near year 450) and the periods of similar length with anomalies of only one sign. The other indices of Fig. 1 describe these properties quantitatively, and were devised as follows. The 1024-year (monthly) time series was divided into 51 segments, each of 24-years length (overlapping each other by 4 years). Within each segment, the mean, standard deviation, and power spectrum were computed. (A Welch window was applied prior to FFT calculation.) The spectral band with largest power was determined, as well as the fraction of total (interannual) variance contained in that band. Shown are the standard deviation, dominant frequency, and fraction-of-variance indices. The latter is a crude measure of the degree of regularity,

![Figure 1](image-url)
as nearly periodic oscillations result in large variance accompanying the dominant frequency, and chaotic variations result in variance spread among many or all bands. Two regimes are apparent, one with large variance, high regularity, and a dominant 4-year period, and the other with lower variance, mixed periods, and low regularity. Thus, the indices provide a quantitative measure of the behavior patterns conspicuous in the original time series.

The distributions of the indices (among the 51 realizations) are shown in Fig. 2. The standard deviation and fraction-of-variance distributions are notably broad, reflecting the distinct regimes. The favored frequency band centers on 4 years, but adjacent bands are also represented (3–6 years). For purposes of comparison, these same indices were computed from the single 18-year realization of observed SST anomalies between 1970 and 1987 (Climate Analysis Center analysis), and are indicated on the distribution plots. The observed values fall well within the range of the model-derived distributions, showing that, at least for this period, the temporal characteristics of observed interannual variability are consistent with those of the model.

To examine the spatial structures, we computed EOF's of model fields based on monthly values from the 1024-year simulation. Figure 3 shows the first four EOF's of the SST anomalies. The first EOF, with 83% of the total variance, is clearly the mature El Niño signal. The others represent various phases of the life cycle of model warm and cold episodes. Corresponding EOF's derived from observed SST anomalies (1970–1987) are shown in Fig. 4. Though there is correspondence between the patterns, two biases of the model are apparent: the underestimation of variability in the South American coastal region and in the vicinity of the dateline at the equator. Also, whereas the first 4

![Figure 2. Distributions of mean sea surface temperature (°C), standard deviation, dominant frequency (years⁻¹), and fraction-of-variance indices from the 1024-year simulation. The values of each index computed from observed sea surface temperature between 1970 and 1987 are indicated by vertical line segments.](image-url)
EOF's of the model fields contain 93% of the total variance, only 68% of the variance is accounted for in the observed fields. Partly this reflects the absence of non-ENSO signal (that is, "noise") in the model fields, but in addition, the model anomaly patterns are more consistent from event to event than in nature.

Figures 5 and 6 show the corresponding EOF's of the model-simulated wind stress and thermocline depth anomalies. In both cases, the first EOF again represents a mature warm event configuration. The model wind stress is realistic with respect to the westerly anomalies along the equator, but unrealistic in having off-equatorial easterlies in the eastern Pacific. Compared to other fields, the variance distribution of the thermocline field is broad, reflecting the fact that this field is characterized by features propagating at various speeds depending on location.

4. PARAMETER-DEPENDENCE OF MODEL VARIABILITY

Five parameters of the model were chosen for a detailed parameter study. They reflect the oceanic equivalent depth (Par1), the sharpness (Par2) and the amplitude (Par3) of the mean thermocline, the strength of atmospheric heating associated with SST anomalies (Par4), and the atmospheric friction (Par5). A detailed treatment of these parameters can be found in Zebiak and Cane (1987). For this study, each of the five parameters was allowed to assume three values, including the standard one, a 5% decrease, and a 5% increase, in all possible combinations. This amounts to $3^5$ or 243 simulations, each of which was run for 100 years (starting with the same initial conditions). The purpose of this study was to determine whether rather modest parameter changes could significantly
change the model behavior, and if so, in what sense. Because of the large number of simulations, it is possible to assign statistical significance to the results, avoiding the uncertainties inherent in comparisons of individual realizations. After breaking each 100-year simulation into four segments, the characteristics were evaluated using the four indices defined above. To examine the individual effects of each parameter, the total set of results was grouped into three subsets corresponding to the three values of that parameter. Each of the subsets then contained 324 realizations. The cumulative distributions of each of the four indices were computed for each of the three subsets, and then compared (Fig. 7). In this form, the significance of differences in the distributions can be assessed using the Kolmogorov-Smirnov statistic. The 99% confidence interval derived in this manner is displayed in each distribution plot; there is only a 1% probability that two distributions (of size 324) differing by more than this amount represent the same process.

All five of the parameters produce significant changes in at least one of the indices, and the net effect of each parameter is unique. In the case of Par1, increasing values produce smaller mean and standard deviation, and a tendency toward higher frequencies. Increasing Par2 results only in decreasing the standard deviation on the high end (that is, decreasing the amplitude of the largest warm/cold events). Increasing Par3 acts to increase the mean (marginally) and the standard deviation, while giving a more dominant 4-year period. For Par4, the result is larger mean and standard deviation, and smaller fraction-of-variance (more chaotic). Finally, increasing Par5 leads to much reduced mean and standard deviation, a greater likelihood of dominant periods less than 4 years, and more regular oscillations. All of these effects are consistent and identifiable amidst the background of natural variability, showing a genuine sensitivity to externally determined factors making up the background climate state.
Figure 5. As in Fig. 3, except for model wind stress anomalies.

Figure 6. As in Fig. 3, except for model thermocline depth anomalies.
Figure 7. Cumulative distributions of mean SST, standard deviation, dominant frequency and fraction-of-variance indices from the set of 243 simulations with variations in five model parameters. Across each row, the total set of results in each plot has been stratified according the three distinct values of the indicated parameter. The heavy solid line represents the distribution for the subset with 5% reduction of the parameter; the dashed line represents the subset with the standard value, and the dotted line the subset with 5% increase. Also indicated in each plot (vertical bar) is the separation interval beyond which any pair of distributions may be considered distinct at the 99% level.
5. POSSIBLE GREENHOUSE SCENARIOS?

Given the results of the generalized sensitivity experiments, we have reason to suspect that there could be substantial effects from externally induced changes in environmental conditions. What about the particular changes envisioned in a greenhouse warming scenario? Though at this stage we can only speculate as to what they would be, previous studies provide general guidance. We conducted two prototype experiments, one in which the mean oceanic and atmospheric temperature was increased by 1 degree Celsius everywhere, and another which additionally allowed for an increase in specific humidity and a decrease in oceanic stratification in the upper thermocline. These choices cannot be justified in any rigorous fashion, and may well not represent a realistic greenhouse warming scenario, but hopefully can give a sense of the possible range of response.

The first scenario (scenario I) simplistically assumes a uniform 1 degree warming of the ocean and atmosphere. The largest effect of this is an increase in anomalous latent heat flux at the ocean surface (for a given wind and SST anomaly). A 1000-year simulation with this change (Fig. 8) shows a marked departure from the standard model simulation (Fig. 1). There are now longer periods of very large and regular warm events. This is reflected in the distributions of statistical indices (as defined above), shown in Fig. 10. Compared to the standard model simulation, the mean and standard deviation of interannual fluctuations in east Pacific SST are larger, and the 4-year quasi-periodicity even stronger.

Scenario II is an attempt at a more complete (if still incorrect) depiction of environmental changes. Along with an increase of 1 degree in SST, the mean specific humidity is assumed to increase (by 5%), so that the atmospheric boundary layer moisture convergence efficiency increases (by the same amount). Also, increased evaporation anomalies are accounted for consistently in the oceanic surface-layer heat budget. We additionally assume that the stratification in the upper thermocline of the equatorial ocean decreases (by 5%). This is based on the results of previous studies (e.g., Hansen et al., 1988) which indicate a larger warming at higher latitudes. From this one can postulate that the subtropical waters which feed into tropical thermocline may warm more than the tropical surface waters, thus decreasing the near-surface stratification. Lacking any strong evidence, we assume no systematic change in the structure or magnitude of tropical surface winds.

The results of the Scenario II simulation are shown in Fig. 9. The behavior is more chaotic, and less energetic than even the standard model. A new feature also appears: multi-decadal sequences of continuously warm SST anomalies. The substantial decrease in interannual variance is evident in the index distributions (Fig. 10), along with a wider range in dominant frequency (especially toward lower frequencies) and reduced degree of regularity. In some ways the Scenario II tendencies seem to run counter to those of Scenario I. The changes of Scenario II, to a certain degree, are acting in opposition. Whereas the increases of surface temperature and humidity acts to enhance interannual variability, the decrease in oceanic stratification acts to diminish it. Evidently, the latter effect dominates in this particular scenario.

6. CONCLUSIONS

Multi-century simulations with the Zebiak and Cane (1987) coupled model show a large degree of natural variability on many time scales. Much of this variability consists of transitions between identifiable regimes which persist from a few years to more than a century. Associated with this temporal variability are characteristic spatial patterns, as revealed by EOF analyses. The dominant patterns have the structure of mature warm events, with largest SST anomalies in the eastern equatorial Pacific, and largest wind
anomalies in the central equatorial Pacific. In terms of a global climate signal, this is a pattern that would be associated with tropospheric temperature anomalies largest in the tropical latitudes. To the extent that this differs from observed patterns, or those expected from other processes, it should be possible to distinguish between them. In particular, if climatic trends are detected in which high-latitude regions exhibit the largest signal, then tropical ocean-atmosphere processes are unlikely to be the cause. Present expectations, based on atmospheric general circulation model (GCM) simulations, are that high-latitude regions will experience the largest changes in a greenhouse scenario. However, this result may be incorrect inasmuch as tropical ocean-atmosphere interaction has not been included.

In response to changes in various physical parameters, the model behavior changes significantly, beyond the range of natural variability. This could be determined with
confidence by having a large number of century-long simulations. These results can be viewed from two perspectives. From one point of view, the individual sensitivities imply that the model could be tuned, assuming we knew what to tune against. Given the degree of natural variability in the model, and presumably in nature, it is hard to see how to do so in a meaningful way. At the same time, the sensitivities suggest that the characteristics of interannual variability may be expected to change in response to externally imposed variations in climate.

Two simulations were run specifically with greenhouse-warming scenarios in mind. In accord with the generalized sensitivity experiments, the model behavior exhibited significant changes in response to rather modest adjustments of the background climate. Note that the two simulations are quite different. We place no great significance on

Figure 9. Time series of NINO3-averaged SST anomaly from a 1000-year simulation of the coupled model using Greenhouse Scenario II (see text).
Figure 10. Distributions of statistical indices as in Fig. 2, from standard run (heavy solid line), Greenhouse Scenario I run (dotted line), and Greenhouse Scenario II run (dashed line).

these particular simulations, as the imposed changes in the background state are highly uncertain. Moreover, there is always a question when a model is extended to a parameter range outside the one used to verify it. This is an issue for GCM's, and a greater issue for this more parameterized climate model. All that can be said with confidence is that these numerical experiments reinforce the plausible idea that, associated with reasonable estimates of externally induced climate change, one should expect a potentially important feedback from tropical ocean-atmosphere processes. This has the further implication that model simulations of climate change must include these processes in order to give reliable indications of how the real climate system will change.

ACKNOWLEDGEMENTS

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Internally Generated Natural Variability of Global-Mean Temperatures

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ABSTRACT. Quantitative frequency-domain and time-domain estimates are made of an important aspect of natural variability of global-mean temperatures, namely, passive internal variability resulting from the modulation of atmospheric variability by the ocean. The results are derived using an upwelling-diffusion, energy-balance climate model. In the frequency domain, analytical spectral results show a transition from a high-frequency region in which the response is determined by the mixed-layer heat capacity and is independent of the climate sensitivity (time scales less than around 10 years), to a low-frequency region in which the response depends only on the climate sensitivity. In the former region the spectral power is proportional to \( f^{-2} \), where \( f \) is the frequency, while in the latter the power is independent of frequency. The range of validity of these results depends on the components of the climate system that are included in the model. In this case these restrict the low-frequency results to time scales less than about 1,000 years. A qualitative extrapolation is presented in an attempt to explain the observed low-frequency power spectra from deep-sea-core \( \delta^{18}O \) time series. The spectral results are also used to estimate the effective heat capacity of the ocean as a function of frequency. At low frequencies, this can range up to 50 times greater than the heat capacity of the mixed layer.

Results in the time domain are obtained by solving the model equations numerically. The results show that substantial 100-year time-scale fluctuations in global-mean temperature may occur purely as a result of the passive modulation of atmospheric variability by the oceans, independent of external forcing and/or ocean circulation changes. Century time-scale trends of up to 0.3°C per 100 years are possible. Some of the simulated results show a striking similarity with those from the 100-year coupled ocean-atmosphere GCM simulation of Stouffer et al. (1989). This similarity suggests that appropriate longer coupled O-A GCM simulations could lead to far more substantial 100-year time-scale trends than so far seen in such experiments.

The derived magnitude of passive internal variability, when considered in conjunction with data uncertainties and other aspects of natural variability (i.e., those associated with ocean circulation changes and/or natural external forcing factors), makes it difficult to estimate the climate sensitivity empirically from the observed warming that has occurred over the past 100 years. If greenhouse-gas forcing were
the sole cause of the observed warming, then the implied sensitivity for a warming of $0.5^\circ$C would be an equilibrium change of around $1.5^\circ$C for $2\times CO_2$. If passive internal variability and uncertainties in the data and model are accounted for, the sensitivity could be less than $1^\circ$C or more than $4^\circ$C. The range of possible values is even higher if other aspects of natural variability are considered.

1. INTRODUCTION

The climate system is known to be highly variable on all time scales. It is possible to divide the causes of this variability into external and internal factors (e.g., Robock, 1978). External factors are those that are either external to the planet (such as changes in solar output, or in the spatial and seasonal signature of incoming solar radiation due to Earth-orbital changes) or derived from sources external to the climate system (such as volcanic eruptions, which modify the planet's radiation balance by injecting material into the stratosphere). Internal factors are those within the climate system, which we define here as the atmosphere, the oceans and those parts of the cryosphere that respond on time scales up to a century. Internally generated variability usually involves interaction among different components of the climate system.

In Wigley and Raper (1990) we distinguish between passive and active internal variability. The former has its primary driving force in the atmosphere and arises through the modulation of atmospheric variability by the ocean in a strictly passive mode. In contrast, active internal variability, which may also be stimulated by the atmosphere, involves a change in the ocean circulation, either the vertical, thermohaline circulation and/or the horizontal current system. The main purpose of this paper is to provide quantitative estimates of the magnitude of passive internal variability.

2. THE SPECTRUM OF PASSIVE INTERNAL VARIABILITY

In a previous paper (Wigley and Raper, 1990) we derived an analytical expression for the power spectrum of global-mean, ocean mixed-layer temperature variations as a function of the power spectrum of some arbitrary external forcing. To do this we used a simple, global-mean, energy-balance climate model, in which the vertical transport of heat in the ocean was parameterized as an upwelling-diffusion process. (For further details on such models see Hoffert and Flannery, 1985). The parameters that determine the model's response, $AT(t)$, to external forcing, $AQ(t)$, are the climate sensitivity $\lambda^{-1}$, where $\lambda$ is the feedback parameter, and the ocean's vertical diffusivity, $\kappa$, upwelling rate, $W$, and mixed layer depth, $h$. It is assumed that the temperature of the high-latitude sinking water, which drives the ocean's thermohaline circulation and which leads to the model's upwelling, does not change. Such changes are characterized by a parameter $\pi$ which is defined in a later section of this paper. The model equations are

$$\phi p c h \frac{dAT}{dt} + \lambda AT = AQ - \Delta F$$

$$\Delta F = -\phi p c \left[ \kappa \frac{\partial^2 \Delta \theta}{\partial z^2} + W \Delta \theta \right]_{z=0}$$

$$\frac{\partial \Delta \theta}{\partial t} - W \frac{\partial \Delta \theta}{\partial z} = \kappa \frac{\partial^2 \Delta \theta}{\partial z^2}$$

where $\phi$ ($\approx 0.7$) is a factor accounting for land/sea differences in heat capacity, $p$ and $c$ are the density and specific heat capacity of ocean water, $\Delta F$ is the flux of heat out of the...
mixed layer into the deeper ocean, and $\Delta \theta(z,t)$ is the temperature change at depth $z$ below the mixed layer. For the frequency-domain calculations, accounting for the land by simply scaling the heat capacity term by $\phi$ is probably the most severe of the approximations made in deriving Eqs. (1)-(3). It may lead to unrealistic results at high frequencies (i.e., time scales of order a year or less).

The solution to the model equations in the frequency domain is

$$S_T = S_Q \omega_{-2} \left\{ \left[ 1 - \frac{\tau}{2h} + \frac{\tau}{2h} \sqrt{\frac{v^2 + W^2}{2}} \right]^2 + \left[ 2\pi f \tau + \frac{\tau}{2h} \sqrt{\frac{v^2 - W^2}{2}} \right]^2 \right\}^{-1},$$

where $S_Q(\omega)$ is the power spectrum of the forcing at frequency $\omega$ (yr$^{-1}$), $S_T(\omega)$ is the power spectrum of the mixed-layer temperature, $\tau$ is a mixed-layer time scale,

$$\tau = \phi \rho c h / \lambda,$$

($\tau \approx 2$-10yr for $\lambda$ between 0.3 and 1.0°C/Wm$^{-2}$) and $v$ is a characteristic velocity defined by

$$v^4 = W^4 + 64\pi^2 f^2 \kappa^2.$$

[Note that there is a typographical error in Wigley and Raper (1990). The correct exponent for $\lambda$ is as given in Eq. (4) above.]

Equation (4) shows that the thermal inertia of the ocean leads to a reddening of the spectrum, as pointed out by Hasselmann (1976). Whatever the forcing spectrum is, the response spectrum $S_T$ will have a greater preponderance of power at lower frequencies. Essentially, the ocean, acting as a passive heat reservoir, amplifies the lower frequencies relative to the higher.

Equation (4) is a general result. What we require is the spectrum of internal variability, specifically that component which arises from internal variability of the atmosphere. Because the above formulation decouples the atmospheric and oceanic parts of the climate system, atmospheric variability, reflecting changes in the radiation balance of the atmosphere, may be considered as a forcing term, $\Delta Q(t)$ in Eq. 1. If we assume the forcing spectrum to be white noise (i.e., $S_Q$ is a constant), the resulting temperature spectrum $S_T$ shown in Fig. 1 is then the spectrum of passive internal variability.

At the high-frequency end, the frequency dependence of this spectrum is $\omega^{-2}$ and the spectrum is determined solely by the heat capacity of the mixed-layer. If this is denoted by $C$, where

$$C = \phi \rho c h,$$

then Eq. (4) reduces to

$$S_T = S_Q/(2\pi C \omega)^2.$$

The frequency range over which the $\omega^{-2}$ behavior prevails is dependent on $\lambda$, and is defined by $\omega \gg 1/2\pi \tau$ (effectively, $\omega > 1/4\tau$). The cutoff point is about 1/8 yr$^{-1}$ for $\lambda^{-1} = 0.3°C/Wm^{-2}$ to 1/30 yr$^{-1}$ for $\lambda^{-1} = 1.0°C/Wm^{-2}$. 
Therefore, for forcings on time scales less than about 10 years, the response of the climate system is virtually independent of the climate sensitivity. This means that the climate sensitivity cannot be inferred from observational data on the climate response to short-period forcings, such as those due to individual volcanic eruptions and the seasonal insolation cycle, nor from information on interannual or inter-decadal variability. We will confirm this latter result below using a more complete model, by showing that the variance of temperature fluctuations on annual and decadal time scales depends only weakly on the climate sensitivity (see Table 1).

At the low-frequency end, the spectrum is flat and Eq. (1) reduces to

\[ S_T = \frac{S_Q}{\lambda^2} \]

The response to forcings on time scales greater than about 100 years (for \( \lambda^{-1} = 0.3^\circ\text{C}/\text{Wm}^2 \)) to 500 years (for \( \lambda^{-1} = 1.0^\circ\text{C}/\text{Wm}^2 \)) is therefore determined almost solely by the climate sensitivity.

Since the high- and low-frequency behavior of passive internal variability depends largely on the mixed-layer heat capacity and the climate sensitivity, respectively, it is clear that processes controlling the heat flux out of the mixed layer (i.e., ocean mixing processes below the mixed layer) are important over only a relatively narrow band of time scales, 10-500 years. This happens to be a range of considerable practical significance.

3. BEHAVIOR AT VERY LOW FREQUENCIES

The results given in the previous section are limited by the way we have defined the climate system. On time scales greater than those of the longest ocean time scale (of order 1000 years), one would expect other, more slowly acting components of the climate system to have an influence. Observational evidence indicates that this is the case.

The observed background spectrum of ocean temperature variability based on \( ^8\text{O} \) data from deep-sea cores (Shackleton and Imbrie, 1990) clearly differs from the theoretical spectrum of Eq. (4) on time scales of 1000 years and greater, and parallels a low-frequency continuation of the \( f^{-2} \) part of the theoretical spectrum. Passive, internally generated variability arising from the variability of the atmosphere does not contribute significantly to observed variability on these very long time scales. How then does the observed variability arise?

A partial answer can be given by generalizing the above results to cover other, more slowly varying components of the system. On time scales greater than a millennium, for example, large ice sheets must be considered as interacting parts of the climate system. In the above, the \( f^{-2} \) section of the spectrum extends to frequencies of order \( 1/4 \tau \), where the mixed-layer time scale, \( \tau \), is proportional to its heat capacity per unit area. Since large ice sheets have a heat capacity per unit area at least two orders of magnitude greater than for the mixed layer, the \( f^{-2} \) part of the spectrum might be expected to extend to time scales of order 1000s of years.

Because of the separation of upper-ocean and ice-sheet time scales, an analogy can be drawn with the theoretical results presented above. Just as white-noise forcing across the whole frequency domain is converted by interaction with the ocean to the reddened spectrum of Fig. 1, so interaction between the flat, low-frequency part of this spectrum (which is essentially a limited-domain, white-noise region) is converted to a redder (\( f^{-2} \)) spectrum by interaction with the slower ice-sheet component of the climate system. This is the classic response to random external forcing for a system with a simple linear feedback term. Deviations from linearity would cause deviations from the \( f^{-2} \) spectrum, just as the effects of diffusivity (which introduce mildly nonlinear behavior in the ocean case) affect the slope of the spectrum at intermediate (10-100 year) time scales in Fig. 1.
Table 1. Statistical characteristics of model-generated natural variability for different model parameters. h is the mixed-layer depth (m), \( \pi \) is the ratio of the sinking-water temperature change to the global-mean temperature change, \( \Delta T_{2x} \) is the climate sensitivity (°C) and \( \kappa \) is the diffusivity (cm\(^2\)/sec). This table is an expansion of the one given in Wigley and Raper (1990), giving a wider range for \( \Delta T_{2x} \) (for \( h = 70 \) m only) and showing the sensitivity of the results to the assumed mixed-layer depth (\( h = 70 \) m and \( h = 110 \) m) and to the assumed \( \pi \) value (\( \pi = 0 \) and 1). \( r_1 \) is the lag-1 autocorrelation coefficient, \( \sigma_1 \) is the overall standard deviation of the annual time series, and \( \sigma_{10} \) is the standard deviation of 10-year means. Values for \( \kappa = 0 \) also have W set equal to zero and correspond to a mixed-layer-only ocean model.

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This effect, just like the oceanic effect, can extend only to a certain low-frequency limit determined by the ice-sheet time scale, beyond which the spectrum should again be flat. In fact, the observed $\delta^{18}O$ spectrum behaves like $f^{-2}$ at much lower frequencies than this. Arguing by analogy again, one might expect still slower components of a general climate system (e.g., involving the lithosphere) to transform the spectrum towards the $f^{-2}$ form at yet lower frequencies. These extensions of the results presented here are, however, highly speculative, and explaining the $f^{-2}$ behavior of the climate system over such a wide frequency range still presents quite a challenge.
4. THE EFFECTIVE HEAT CAPACITY OF THE OCEAN

The concept of an effective heat capacity for the ocean is one that has been used qualitatively in the literature, but never rigorously defined. We can do this by comparing the general spectrum with that for a mixed-layer ocean. The latter is obtained simply by putting $\kappa = 0$ in Eq. (4) to give

$$S_T = S_0 \lambda^{-2} \left\{ 1 + (2\pi f)^2 \right\}^{-1},$$

(10)

which, by Eqs. (5) and (7), may be written in terms of the mixed-layer heat capacity, $C$, as

$$S_T = S_0 \lambda^{-2} \left\{ 1 + (2\pi C/\lambda)^2 \right\}^{-1}.$$  

(11)

By comparing Eq. (11) with Eq. (4) we can define an effective heat capacity as

$$C^* = \frac{C}{2\pi f} \left\{ \frac{S_0 \lambda^{-2}}{S_T} - 1 \right\}^{\frac{1}{2}}.$$  

(12)

$C^*$ clearly depends on $f$, through both the $2\pi f$ term and the $S_T$ term in Eq. (12). Examples of the $f$-dependence are given in Fig. 2. At low frequencies the effective heat capacity may be as large as 50 times the heat capacity of the mixed layer.

Although the general form for $C^*$ is algebraically complicated, the asymptotic form for small $f$ is more manageable, viz.

$$C^* \to C_0^* = C \left\{ \left[ 1 + \frac{\kappa}{hW} \right]^2 + \frac{2\kappa^2}{thW^3} \right\}^{\frac{1}{2}}.$$  

(13)

Equation (13) shows that $C^*$ has an upper bound determined by the model parameters, including the climate sensitivity (through $\tau$). Since the second term inside the square root is generally much larger than the first, Eq. (13) may be approximated by

$$C_0^* \approx \frac{\kappa}{W} \left\{ \frac{2\lambda \phi \rho c}{W} \right\}^{\frac{1}{2}}.$$  

(14)

This result shows that, as expected, the asymptotic effective heat capacity is larger for larger diffusivity and for smaller upwelling rate. $C_0^*$ is also larger for smaller climate sensitivity. These results apply to $C^*$ as well as $C_0^*$. The dependence on mixed-layer depth, however, depends more critically on frequency. As the time scale increases, allowing greater and greater interaction with the deeper ocean, the dependence of $C^*$ on $h$ decreases. At very low frequencies, Eq. (14) shows that $C_0^*$ is independent of the mixed-layer depth.

5. PASSIVE INTERNAL VARIABILITY IN THE TIME DOMAIN

In the time domain, important results can be obtained analytically (e.g., Wigley and Schlesinger, 1985; Watts, 1985; Lebedeff, 1988; Morant and Watts, 1990), but in the context of natural variability generated by stochastic forcing it is more informative to
proceed through numerical integration of the equation. In doing this we use a similar, but more realistic, model than the one employed for our analytical calculations, an upwelling-diffusion model in which the Earth is divided into land and ocean ‘boxes’ in each hemisphere (see Wigley and Raper, 1987, 1990). To facilitate comparison with greenhouse studies, the climate sensitivity is specified in terms of the equilibrium warming for a doubling of atmosphere CO$_2$, $\Delta T_{2x} = 1.5-4.5^\circ$C, related to $\lambda$ by

$$\lambda^{-1} = \Delta T_{2x}/\Delta Q_{2x} ,$$  

(15)

where $\Delta Q_{2x}$ is taken as 4.39 Wm$^{-2}$. 

Figure 2. Effective oceanic heat capacity, $C^*$, as a function of frequency ($f$ in yr$^{-1}$), expressed relative to the mixed-layer heat capacity, $C$, showing its dependence on climate sensitivity ($S \equiv \lambda^{-1}$ in $^\circ$C/Wm$^{-2}$) and vertical diffusivity ($\kappa$ in cm$^2$/sec). Other model parameters used: $h = 70$ m, $W = 4$ m/yr, $\pi = 0$. 
Our aim is to determine the magnitude of natural, internally generated decadal-to-century time-scale variations in global-mean temperature. To facilitate this, and to make use of what we know about high-frequency (1- to 10-year time scale) variability we constrain the low-frequency variations to be compatible with observed high-frequency temperature variations. The model is forced with random interannual variations in $Q$; specifically, global-mean Gaussian white-noise forcing. The total variance of the forcing ($\sigma_Q^2$) is then tuned so that the output matches observed high-frequency temperature variations.

A preliminary estimate of $\sigma_Q^2$ can be made from the spectral solution, Eq. (4), using the known high-frequency variance of the global-mean temperature data. The implied value of $\sigma_Q$ is about 1 Wm$^{-2}$. This value is consistent with the observed month-to-month variations in the Earth's radiation budget due to cloudiness variations (Ramanathan et al., 1989) which have a standard deviation of 3.3 Wm$^{-2}$. If month-to-month variations are uncorrelated, then the interannual standard deviation should be $\sim$1Wm$^{-2}$. This may be interpreted as a forcing term provided that the radiation budget changes are not themselves dependent on temperature changes (M. I. Hoffert, personal communication, 1989).

We therefore began by forcing the model with white noise with $\sigma_Q = 1$ Wm$^{-2}$, and then adjusted the forcing slightly so as to obtain a precise match with the observed high-frequency variations (viz, $\sigma_T = 0.063^\circ$C). Because the model cannot possibly simulate the internal variability related to ENSO, the $\sigma_T$ value used here corresponds to the global-mean temperature record with ENSO factored out. (For further justification, see Wigley and Raper, 1990.) Removing ENSO produces slightly lower low-frequency variability than if the "raw" global-mean data were used for calibration. Simulations of 100,000 years length were performed for a wide range of values of the parameters that control model output.

For temperature variations on the 10- to 100-year time scale, which are particularly pertinent to the issue of greenhouse-gas forcing, the spectral results show that the magnitude of internally generated natural variability should have appreciable $\lambda$ dependence. The larger the climate sensitivity, the larger the internal variability. Typical simulated variations in global-mean temperature are illustrated in Fig. 3, which shows two 100-year extracts from a 100,000-year simulation with $\Delta T_{2x} = 4^\circ$C. One of these has been selected to demonstrate that strong century time-scale trends may result from internal effects alone, while the other shows a situation with no noticeable century time-scale trend. Neither of these examples is unusual, as will be shown below by examining the frequency distribution of trends over different periods. In other words, it appears likely that substantial 100-year time-scale fluctuations in global-mean temperature may occur purely as a result of the thermal inertia of the oceans acting in a passive mode, independent of any external forcing changes and independent of active oceanic effects which might be associated with circulation changes.

Figure 3 also shows (upper panel) the 100-year simulation of Stouffer et al. (1989) using a fully-coupled ocean-atmosphere GCM. This model has a climate sensitivity of $\Delta T_{2x} = 4^\circ$C, as assumed in our own simulations. The GCM results are strikingly similar to the simpler model simulations, particularly to the example with little century time-scale trend (Fig. 3, middle panel). Since our model has been calibrated to have the observed high-frequency variability, this similarity demonstrates that the GCM simulates such variability quite well. Whether or not the single 100-year GCM result is a typical output of the model, however, cannot be determined unless a much longer run is performed. The fact that the simple model can produce results with much larger 100-year trends, and the correspondence between the two models on shorter time scales, must be a strong indicator that a longer GCM run would also show larger 100-year trends. The low-frequency
Figure 3. Simulated natural variability of global-mean temperature. The upper panel shows results from the 100-year control run with the coupled ocean-atmosphere GCM of Stouffer et al. (1989). The lower two panels are 100-year sections from a 100,000-year simulation using an upwelling-diffusion model with the same climate sensitivity as the Stouffer et al. model ($\Delta T_{2x} = 4^\circ$C). The upwelling-diffusion model is forced with random interannual radiative forcing changes chosen to match observed interannual variations in global-mean temperature. The consequent low-frequency variability arises due to the modulating effect of oceanic thermal inertia.
behaviour of the GCM may, however, be constrained by the use of flux correction. Until this constraint is lifted, it may well be that the only way to estimate the \(\approx 30\)-year time-scale variability is through the use of relatively simple models.

As noted above and shown in Table 1, the magnitude of our simulated passive internal variability depends on the model parameters. (The results shown in Table 1 are for globally coherent forcing, but they are insensitive to the degree of aggregation of the forcing.) Ten- to one-hundred-year time-scale variability, as judged by any of the measures used here, increases with increasing \(\Delta T_{2x}\), increasing \(\Lambda\) and decreasing \(\kappa\) as predicted by the analytical results. Table 1 also shows the effect of changes in the model's sinking-water temperature through simulations with \(\pi = 0\) (no change) and \(\pi = 1\) (changes equal to those in the global mean). Larger \(\pi\) leads to slightly lower 10- to 100-year time-scale variability. In support of the realism of these simulations, we note that the lag-1 autocorrelations \(r_1\); that is, the correlation between successive annual-mean values) are similar to those observed in the data of Jones et al. (1986).

In all cases, the 90% confidence limits for the 100-year trends are less than the observed warming trend of 0.5°C over the past 100 years (the maximum natural trend range is \(\pm 0.38\)°C for \(h = 110\ m\), \(\kappa = 0.5\ \text{cm}^2/\text{sec}\), \(\pi = 0\) and \(\Delta T_{2x} = 4.5\)°C; see Table 1). Since passive internal variability is the most basic mode of natural variability of the climate system, and since its statistical properties closely match those of the observations, the 100-year trend results indicate that the observed warming trend is highly statistically significant. Based on the results presented in Table 1, the 90% confidence limits for a 100-year trend of 0.5°C range from 0.34-0.66°C to 0.12-0.88°C. The range of possible trend limits arises from the range of possible model parameter values, \(h = 70-110\ m\), \(\kappa = 0.5-2.0\ \text{cm}^2/\text{sec}\), \(\pi = 0-1\) and \(\Delta T_{2x} = 1.5-4.5\)°C.

It is important to note that, although we judge the observed warming trend to be statistically significant, we cannot claim to have detected the enhanced greenhouse effect on the basis of these results, largely because we have not demonstrated a cause-effect relationship. Furthermore, we have only considered one aspect of natural variability. If other factors are considered (external forcing effects and/or changes in the thermohaline circulation, for example), then an even larger portion of the observed global-mean warming could be attributed to non-greenhouse processes, leaving open the possibility that almost all of the observed warming is natural.

But this is only one side of the coin. If we admit the possibility of a natural warming trend of order 0.5°C, we must also allow a possible natural cooling trend of this magnitude. In this case, the greenhouse component of the observed warming could be much larger than 0.5°C, requiring a large climate sensitivity. As noted in Wigley and Raper (1990), although a direct empirical estimate based on the observed warming points to a quite small climate sensitivity, around 1.5°C for \(\Delta T_{2x}\), a consideration of natural variability leads to a wide range of possible values. If only passive internal variability is considered, if the observed warming is assumed to lie in the range 0.4-0.6°C, and if model parameter uncertainties are accounted for, then it can be shown that the range of possible values of \(\Delta T_{2x}\) varies from less than 1°C to more than 4°C (Wigley and Raper, 1991).

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Multivariate Statistical Assessments of Greenhouse-Gas-Induced Climatic Change and Comparison With Results from General Circulation Models

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ABSTRACT. Based on univariate correlation and coherence analyses, including techniques moving in time, and taking account of the physical basis of the relationships, a simple multivariate concept is presented which correlates observational climatic time series simultaneously with solar, volcanic, ENSO (El Niño/Southern Oscillation) and anthropogenic greenhouse-gas forcing. The climatic elements considered are air temperature (near the ground and stratosphere), sea surface temperature, sea level and precipitation, and cover at least the period 1881-1980 (stratospheric temperature only since 1960). The climate signal assessments which may be hypothetically attributed to the observed CO$_2$ or "equivalent" CO$_2$ (implying additional greenhouse gases) increase are compared with those resulting from GCM experiments. In case of the Northern Hemisphere air temperature these comparisons are performed not only in respect to hemispheric and global means, but also in respect to the regional and seasonal patterns. Autocorrelations and phase shifts of the climate response to natural and anthropogenic forcing complicate the statistical assessments.

1. INTRODUCTION

A substantial increase in the atmospheric concentration of a number of trace gases has been observed as a consequence of human behavior. As shown in Fig. 1, the atmospheric carbon dioxide (CO$_2$) concentration has increased over the past 200 years from preindustrial estimates of approximately 280 ppm (Neftel et al., 1985) to a value of 347 ppm in 1986 at the Mauna Loa Observatory, Hawaii (Keeling, 1987). The corresponding radiative effects of additional trace gases such as the chlorofluorocarbons (CFCs), methane (CH$_4$),
nitrous oxide (N\textsubscript{2}O) and tropospheric ozone (O\textsubscript{3}) can be expressed in terms of CO\textsubscript{2} "equivalents" (e.g., Tricot and Berger, 1987) which also contribute to the greenhouse warming (Fig. 1).

The climate response to an atmospheric CO\textsubscript{2} doubling has been simulated by climate models of different complexity. It is only the (three-dimensional) general circulation models (GCMs) that can properly simulate not only the response of the most important climate elements (e.g., temperature and precipitation), but also their regional patterns. Neglecting the "81" experiment in Fig. 2, the result of the recent GCM CO\textsubscript{2} doubling experiments is a globally averaged temperature increase near the ground of 1.3–5.2°C. Simultaneously, a stratospheric temperature decrease (more intense than the temperature increase near the ground) and a global sea level rise (roughly 1 ± 0.5 m within the next 50-100 yr) are predicted (Bolin et al., 1986). Furthermore, subtropical precipitation rates may decrease and the mid-latitude and subpolar precipitation may increase.

It is very important to verify as early as possible this climate response to increasing greenhouse gases projected by GCM experiments by means of observational statistics (DOE, 1985). The first approach to this problem may be to look at some long-term time series of observational climatic data, preferably averaged over an extended area up to the hemispheric or global scale to reduce the natural variability or "noise" (arising from regional circulation peculiarities). Figure 3 presents some annual data examples where the autocorrelation coefficients (one-year time lag), the annual data standard deviations

Figure 1. Measured Mauna Loa, 1958-1986 (Keeling, 1987) and reconstructed (ice-core drillings, Antarctica; Neftel et al., 1985) annual-mean CO\textsubscript{2} concentrations, the latter represented by a double-logarithmic regression corrected by +1 ppm (Schönwiese, 1987). The additional "equivalent" CO\textsubscript{2} concentrations, dotted line, are from Tricot and Berger (1987).
(as a simple measure of "noise"), the linear trends and the linear trend-to-noise ratios are specified. Some of the time series considered actually indicate the trends (at least qualitatively) that are expected based on the GCM experiments, although in general the trend-to-noise ratio remains below a value of three (corresponding approximately to < 99% confidence). Note that in the case of the stratospheric temperature, this ratio can be somewhat increased by computing the stratosphere-troposphere temperature differences. In case of precipitation, the European trend-to-noise ratios (Fig. 3b) are very small and seldom exceed a value of unity. [See also worldwide analyses of precipitation fluctuations by Bradley et al. (1987) and tropical humidity trends as evaluated by Flohn and Hense (1987).]

Any observed climatic trend, however, may be caused by very different forcing factors, anthropogenic and natural, and cannot be *a priori* attributed to the effect of increasing greenhouse gases. Moreover, one should also be aware of some trends that are opposite to the effect expected based on GCM CO₂-increase experiments, for example, the circa 1940–1970 drop in Northern Hemisphere land temperatures (Fig. 3a). Obviously, the anthropogenic climate signal is involved with natural climate variability and can only be detected if it is possible to explain a major part of the overall climate variability and separate the anthropogenic signal from the remaining variability. (For more general discussions of the signal detection problems see elsewhere, e.g., DOE, 1985.)

One way to do this is by the multivariate technique of statistical observational data analysis, in the simplest case, the evaluation of a multiple linear-regression model

\[ y = a_0 + \sum a_i x_i \quad , \] (1)
where $y$ is the (climatic) predictant, $x_i$ are the predictors and $a_i$ ($i = 0, 1, \ldots, k$) the regression coefficients. One may add to Eq. (1) a biregressive term which relates different regression
coefficients to identical forcing factors $x_i$ at different time lags. In any case, Eq. (1) allows the extraction of particular signals as will be demonstrated in the following sections.

Before any multivariate model is used for climate signal assessments, however, it is necessary to set up the underlying univariate relationships. Some cross correlation tables concerning temperature (air and sea surface) and sea level time series and their relationships with volcanic, solar, ENSO and trace-gas forcing parameter time series are published elsewhere (Schönwiese, 1987; Schönwiese and Runge, 1988). The volcanic-forcing parameter time series for the Northern Hemisphere are shown in Fig. 4. The confidence tests of the correlation coefficients must account for both non-Gaussian distributed quantities (e.g., by means of the Fisher transformation) and autocorrelation (reduced degrees of freedom; for methods see, e.g., Schönwiese, 1985). The coherence spectra plotted in Fig. 5 illustrate that some relationships, in particular concerning volcanic and solar forcing, are established predominantly in the long-period part of the spectrum. This may lead to an analysis of low-pass filtered instead of annual or monthly data. Filtering, however, increases autocorrelation. Signal assessments based on low-pass filtered data,
Figure 3b. Similar to Fig. 3a, but European precipitation data (from Birrong and Schönwiese, 1988) based on 301 stations and specified for latitude bands and seasons; t = trend, trend-to-noise ratios in parentheses, τ = time correlation.
Figure 4. TNH data (compare Fig. 3a, before 1851 from Grove-
man and Landsberg, 1979, very uncertain) compared with the
volcanic-activity parameter time series. DVI = dust veil index
(Lamb, 1983, 1985), SVI = Smithsonian volcanic index (evalu-
ated by Schönwiese and Cress, 1988, based on the chronology
edited by Simkin et al. 1981, 1985), SVI* = similar, but implying
a stratospheric particle residence time derived from SVI/AI inter-
comparisons (Schönwiese and Cress, 1988) and AI = acidity index
derived from Greenland ice-core measurements (Hammer et al.,
1980, 1983). All data are annual and 10-year low-pass filtered.
Therefore, should be checked for stability using unfiltered data, and any statistical model "explaining" more or less of the observed climate variance should significantly exceed the variance which is due to autocorrelation. Moreover, moving correlation and coherence analyses reveal that the strength of some relationships is not stable in time; see examples shown in Fig. 6.

In general, most of the statistical problems and misinterpretations are avoided if all correlations are strictly related to a physical basis, preferably on corresponding deterministic model assessments (see, e.g., Hansen et al., 1981). So, the statistical assessments can and should be verified by climate model experiments and vice versa. In particular, the anthropogenic trace-gas trends observed in the atmosphere are hardly appropriate for statistical correlation analyses. Therefore, all statistical assessments concerning this trace-gas forcing should be interpreted from some kind of residual trend analysis which discerns natural variability from the trends that may be hypothetically due to the anthropogenic trace-gas forcing.

In the following some statistical assessments of greenhouse-gas-induced climatic change based on multivariate regression models are presented. In doing this, one should keep in mind that only a few statistical problems have been mentioned in this section, neglecting, for instance, the phase shift problems. The time lags evaluated from cross correlation analyses (mostly a few years in case of volcanic forcing, but roughly 20-25 years in case of trace-gas forcing) are preliminary and all details are open to discussion.

2. DATA BASE

In continuation of some earlier studies (Schönwiese, 1983, 1984, 1987; Schönwiese and Malcher, 1987; Schönwiese and Runge, 1988), but restricted to some aspects which seem to be of interest, the following anthropogenic and natural-forcing parameter time series are used (annual data).
(a) Atmospheric CO₂ concentration, alternatively "equivalent" CO₂ concentration, both as shown in Fig. 1. The CO₂ data before 1958 are calculated by means of a double-logarithmic regression based on the ice-core data from Neftel et al. (1985), and after 1958 on the Mauna Loa observations (Keeling, 1987) are used. In order to get a best fit, the ice core regression data were corrected by +1 ppm. The added "equivalent" data are from Tricot and Berger (1987).

(b) Volcanic activity parameters, four alternative time series (see Fig. 4): stratospheric "dust veil index" DVI from Lamb (1983, 1985); acidity index AI from Hammer et al. (1980, 1983) based on Greenland ice-core reconstructions (only Northern Hemisphere; same index implying regional corrections not used here); "Smithsonian volcanic index" SVI based on the Smithsonian volcano chronology edited by Simkin et al. (1981, 1985) modified by Schönwiese (1988); same index but again modified by Schönwiese and Cress (1988) implying an empirical stratospheric residence time of the volcanogenic particles based on AI/SVI intercomparisons (SVI*).

(c) Solar forcing, six alternative parameters, later reduced to four alternative parameters, mainly based on solar activity (relative sunspot numbers SRN and derived hypotheses discussed elsewhere (Schönwiese, 1983, 1984) and the solar diameter oscillation hypothesis as supposed by Gilliland (1982, 1983) in terms of hypothetical temperature oscillations TSD.

(d) El Niño/Southern Oscillation (ENSO) forcing represented by the tropical Pacific sea surface temperature (SST) anomalies as reconstructed by Wright (1984) or Schneider and Schönwiese (1989), respectively (annual and seasonal data).
These forcing parameter time series are correlated with time series of the following observational climatic time series.

(a) Air temperature near the ground, focused on land areas, Northern Hemisphere TNH-J since 1851, Southern Hemisphere TSH-J (< 62.5°S) since 1858 and global TGL-J (since 1858) mean estimates from Jones et al. (1982), Jones (1985, 1988), alternatively hemispheric and global-mean estimates since 1881 from Hansen and Lebedeff (1987) TNH-H, etc.

(b) Stratospheric temperature since 1960, only Northern Hemisphere, as provided by Angell and Korshover (1984) TST-A (100-30 hPa mean) or Labitzke et al. (1986) TST-L (30 hPa, since 1965).

(c) Sea surface temperatures, hemispheric and global means, data alternatively since 1856 from Folland et al. (1984) SNH-F etc., or since 1854 from Jones (1987) SNH-J etc.

(d) Global-mean temperature near the ground derived from land and marine data, from Jones et al. (1988) TGL-LM.

(e) Global mean sea level data since 1881, alternatively from Barnett (1985, 1987) LGL-B or Gornitz et al. (1982) LGL-G.

In addition to these hemispheric or global and annual averages, in case of the Northern Hemisphere air temperature near the ground (land areas), also gridded monthly data are used (5° latitude and 10° longitude grid; DOE, 1986). A corresponding global analysis using the "box data" from Hansen and Lebedeff (1987) is in preparation.

3. MULTIVARIATE REGRESSION MODEL

Based on an univariate correlation analysis using annual as well as 3-year and 10-year Gaussian low-pass filtered data, in the first approach a simple linear multivariate regression model was evaluated

\[ A = a + bV + cR + dE + eC, \]

where A is any climatic element, V any volcanic parameter, R any solar parameter, E any ENSO parameter and C the CO₂ or "equivalent" CO₂ forcing. All of these parameters V, R, E and C are alternatively used (only one volcanic, solar, etc. parameter in each regression). One gets a matrix of regression coefficients explaining the observed climatic time series variance to a certain extent. Using four volcanic, four solar, two ENSO and two trace-gas alternative parameter time series one gets 64 combinations for one climate element (multiple correlation and signal assessments).

In the second approach, time lags (of the climatic elements with respect to the forcing parameters, the latter leading) are introduced. Figure 7 presents two examples of alternative reproductions of the observed Northern Hemisphere air temperature (land areas) fluctuations, using 10-year low-pass filtered data as reproduced by the different multiple regression models. It is found that the observed temperature decrease during approximately 1940–1970 can be hypothetically explained by the volcanic SVI* or, alternatively, by the solar TSD parameter. (For detailed discussion see Schönwiese and Runge, 1988). Figure 8 shows similar reproductions for the Southern Hemisphere, Fig. 9 for the global land and marine data, and Fig. 10 for the stratosphere.

As soon as the regression coefficients are known, any climate signals, observed or projected, can be computed. These signals describe the climate response to any forcing. For instance, a CO₂ doubling signal S, related to the air temperature near the ground, is obtained from
Figure 7. Observed TNH (see Fig. 3a, Northern Hemisphere) fluctuations (anomalies, solid line), and multivariate statistical reproductions using SVI* (volcanic), TSD (solar) and CO₂ or CO₂E (= equivalent, see Fig. 1) forcing. The dotted line implies a CO₂E (leading) phase shift of 20 years (SVI* and TSD phase shift 5 years), shown 10-year low-pass filtered. The best fit statistical model explains 81% of variance (multiple correlation coefficient rm = 0.900), in the case of unfiltered annual data implying El Niño forcing 58% (rm = 0.764).

\[ S = A_{\text{max}} - A_{\text{min}} = f(V,R,E,C_{\text{max}} = 600 \text{ ppm}) - f(V,R,E,C_{\text{min}} = 300 \text{ ppm}) \]

where V, R and E are the long-term averages of the natural forcing parameters and C the CO₂ concentrations. Note that these signals describe the response of any climate element only with respect to one selected forcing parameter; however, this signal assessment is based on a multivariate statistical analysis (using the multiple regression coefficients). In the case that the observed standard deviation s of the (in general annual) climatic data is defined to represent the climatic "noise" N (= s), S/N values are assessments of the related signal-to-noise ratios specifying the signal confidence levels.

4. SOME RESULTS

Only a few results can be presented here. Table 1 summarizes some of the signal assessments concerning the mean hemispheric and global data where the signals are evaluated from the multivariate analysis of 10-year or 3-year low-pass filtered data. The Tables specify the following.
Figure 8. Similar to Fig. 7, but TSH data from Hansen and Lebedeff (1987; Southern Hemisphere) and without phase shift. This model explains 80% of variance ($r_m = 0.897$), in case of unfiltered annual data implying El Niño forcing 46% ($r_m = 0.678$).

(a) "Industrial" signals using the observed preindustrial ($C_{\text{min}}$) and recent ($C_{\text{max}}$) atmospheric trace-gas concentrations; in case of CO$_2$, $C_{\text{min}} = 280$ ppm and $C_{\text{max}} = 347$ ppm; in case of "equivalent" CO$_2$, $C_{\text{min}} = 280$ ppm and $C_{\text{max}} = 386$ ppm (see Fig. 1).

(b) "Projected" signals, that is, extrapolations based on the observational statistics; in case of CO$_2$, $C_{\text{min}} = 300$ ppm and $C_{\text{max}} = 600$ ppm; in the "equivalent" case, $C_{\text{min}} = 300$ ppm and $C_{\text{max}} = 900$ ppm.

Note that in the "industrial equivalent" case the same observed temperature (or sea level) change is attributed to the increasing trace-gas concentrations as in case that "only CO$_2$" is considered. The introduction of the "equivalent" approach, therefore, means less sensitivity of the climate response to the greenhouse-gas forcing. In consequence, using only the CO$_2$ time series, the climate response is overestimated. The "equivalent" approach, however, can be used to assess a "reduced" and, therefore, more correct CO$_2$ contribution to the climatic greenhouse effect. This is done in the way that the "equivalent" CO$_2$ time series is used for the computation of the regression coefficients, but the assessment of the "reduced" CO$_2$ signals is based only on the CO$_2$ contribution.
Figure 9. Similar to Fig. 7, but global land and marine data (TGL-LM) and CO$_2$E (leading) phase shift 25 years. This model explains 72% of variance ($r_m = 0.847$), in case of unfiltered annual data implying El Niño forcing 59% ($r_m = 0.768$).

Figure 10. Observed stratospheric temperature anomalies (see Fig. 3a) and statistical reproduction using SVI, SRN and CO$_2$E forcing (without phase shift; abbreviations see Figs. 3a, 4 and 5), unfiltered annual data. This model explains 54% of variance ($r_m = 0.736$).
In Table 1 every signal assessment is based on 12 combinations (using only three volcanic parameters: DVI, AI and SVI*) and 24 additional combinations introducing the ENSO parameters. The uncertainties (±) are the signal standard deviations due to the use of different natural forcing parameters. Taking into account these uncertainties, it is concluded that – on a global average and concerning the air temperature of the land areas near the ground – a temperature rise of 0.5-1.0°C (since preindustrial time) may be hypothetically attributed to the “combined” greenhouse effect (“equivalent” approach, until 1986) and a contribution in the order of 0.3-0.6°C may be due to the CO₂ increase. In the case of the SST assessments, in particular using the data from Jones et al. (1988), these statistical assessments are quite similar, but in case of the sea level, these assessments strongly depend on the underlying climatic data: 13-17 cm using the data from Gornitz et al. (1982), but 25-30 cm using the data from Barnett (1987; same magnitude as the observed overall sea level increase or even more).

Table 1. Extremely abbreviated results concerning the multivariate statistical assessments of the climate signals forced hypothetically by the anthropogenic increase of the greenhouse gases; 10-year or, if "*" indicated, 3-year low-pass filtered data. (Explanations and abbreviations see text; ± specifies the standard deviations arising from the different regression models.)

<table>
<thead>
<tr>
<th>Record</th>
<th>Period</th>
<th>Mean mult. correlation</th>
<th>&quot;Industrial&quot; signals (in K)</th>
<th>Reduced to CO₂</th>
<th>Projected signals (in K)</th>
<th>&quot;Equivalent&quot;</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>only CO₂</td>
<td>&quot;equivalent&quot; (S/N)</td>
<td></td>
<td>CO₂ doubling</td>
<td></td>
</tr>
<tr>
<td>TNH-J</td>
<td>1881–1980</td>
<td>0.76</td>
<td>0.6±0.2 (2.4)</td>
<td>0.4±0.1</td>
<td>3.2±0.8</td>
<td>3.5±1.2</td>
</tr>
<tr>
<td>TNH-J</td>
<td>1881–1980</td>
<td>0.78</td>
<td>0.6±0.2 (3.0)</td>
<td>0.5±0.1</td>
<td>3.3±0.8</td>
<td>4.1±1.3</td>
</tr>
<tr>
<td>*TNH-J</td>
<td>1881–1980</td>
<td>0.70</td>
<td>0.6±0.2 (2.6)</td>
<td>0.4±0.1</td>
<td>3.5±0.9</td>
<td>3.8±1.0</td>
</tr>
<tr>
<td>*TNH-H</td>
<td>1881–1980</td>
<td>0.75</td>
<td>0.6±0.2 (3.2)</td>
<td>0.5±0.2</td>
<td>4.4±1.0</td>
<td>4.6±1.5</td>
</tr>
<tr>
<td>TSH-J</td>
<td>1881–1980</td>
<td>0.86</td>
<td>0.6±0.1 (4.1)</td>
<td>0.5±0.1</td>
<td>3.5±0.6</td>
<td>4.1±0.9</td>
</tr>
<tr>
<td>TSH-J</td>
<td>1858–1980</td>
<td>0.85</td>
<td>0.6±0.1 (4.3)</td>
<td>0.5±0.1</td>
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<td>4.3±0.4</td>
</tr>
<tr>
<td>*TSH-J</td>
<td>1881–1980</td>
<td>0.81</td>
<td>0.6±0.2 (4.3)</td>
<td>0.5±0.1</td>
<td>3.8±0.8</td>
<td>4.4±1.0</td>
</tr>
<tr>
<td>*TSH-H</td>
<td>1881–1980</td>
<td>0.72</td>
<td>0.6±0.1 (3.6)</td>
<td>0.4±0.1</td>
<td>2.7±0.2</td>
<td>3.3±0.4</td>
</tr>
<tr>
<td>TGL-J</td>
<td>1881–1980</td>
<td>0.81</td>
<td>0.6±0.2 (3.2)</td>
<td>0.4±0.1</td>
<td>3.4±0.7</td>
<td>3.8±0.8</td>
</tr>
<tr>
<td>TGL-J</td>
<td>1858–1980</td>
<td>0.83</td>
<td>0.6±0.1 (3.6)</td>
<td>0.4±0.1</td>
<td>3.3±0.5</td>
<td>4.0±0.6</td>
</tr>
<tr>
<td>TGL-J</td>
<td>1881–1980</td>
<td>0.81</td>
<td>0.6±0.2 (3.2)</td>
<td>0.4±0.1</td>
<td>3.4±0.7</td>
<td>3.8±0.8</td>
</tr>
<tr>
<td>*TGL-J</td>
<td>1881–1980</td>
<td>0.78</td>
<td>0.6±0.2 (3.4)</td>
<td>0.5±0.1</td>
<td>3.6±0.8</td>
<td>4.1±0.9</td>
</tr>
<tr>
<td>*TGL-H</td>
<td>1881–1980</td>
<td>0.77</td>
<td>0.6±0.2 (3.3)</td>
<td>0.5±0.1</td>
<td>3.6±0.8</td>
<td>4.1±1.1</td>
</tr>
<tr>
<td>SNH-F</td>
<td>1881–1980</td>
<td>0.63</td>
<td>0.6±0.2 (2.9)</td>
<td>0.3±0.2</td>
<td>2.3±0.9</td>
<td>2.8±1.3</td>
</tr>
<tr>
<td>SSH-F</td>
<td>1881–1980</td>
<td>0.64</td>
<td>0.6±0.1 (3.9)</td>
<td>0.3±0.1</td>
<td>2.2±0.3</td>
<td>2.7±0.4</td>
</tr>
<tr>
<td>SGL-F</td>
<td>1881–1980</td>
<td>0.66</td>
<td>0.6±0.1 (3.3)</td>
<td>0.3±0.1</td>
<td>2.3±0.5</td>
<td>2.8±0.7</td>
</tr>
<tr>
<td>*SGL-F</td>
<td>1881–1980</td>
<td>0.60</td>
<td>0.6±0.1 (3.5)</td>
<td>0.3±0.1</td>
<td>2.4±0.4</td>
<td>3.0±0.5</td>
</tr>
<tr>
<td>SNH-J</td>
<td>1881–1980</td>
<td>0.75</td>
<td>0.6±0.2 (3.6)</td>
<td>0.4±0.2</td>
<td>3.1±1.0</td>
<td>3.6±1.4</td>
</tr>
<tr>
<td>SSH-J</td>
<td>1881–1980</td>
<td>0.74</td>
<td>0.6±0.1 (3.5)</td>
<td>0.4±0.1</td>
<td>2.7±0.3</td>
<td>3.2±0.4</td>
</tr>
<tr>
<td>SGL-J</td>
<td>1881–1980</td>
<td>0.78</td>
<td>0.6±0.1 (3.5)</td>
<td>0.4±0.1</td>
<td>2.9±0.6</td>
<td>3.4±0.9</td>
</tr>
<tr>
<td>*SGL-J</td>
<td>1881–1980</td>
<td>0.74</td>
<td>0.6±0.1 (3.4)</td>
<td>0.4±0.1</td>
<td>2.8±0.8</td>
<td>3.3±1.1</td>
</tr>
<tr>
<td>LGL-B</td>
<td>1881–1980</td>
<td>0.95</td>
<td>27±4 (5.9)</td>
<td>17±2</td>
<td>127±13</td>
<td>154±22</td>
</tr>
<tr>
<td>LGL-G</td>
<td>1881–1980</td>
<td>0.87</td>
<td>15±2 (5.2)</td>
<td>10±1</td>
<td>70±6</td>
<td>85±10</td>
</tr>
</tbody>
</table>
The statistical extrapolation to a CO₂ doubling or corresponding “equivalent” situation leads to a temperature increase of 2.7-4.4°C (CO₂) or 3.0-5.2°C (“equivalent”, “combined”). The CO₂ assessments are in fair agreement with the climate model (GCM) projections as shown by Fig. 2. The corresponding sea level assessments (70-170 cm), however, may lead to overestimations when compared with the recent expert statements. The signal assessments are nearly the same if annual, instead of low-pass filtered, data (including extreme autocorrelation) are used.

One of the major problems arising in this context is that in Fig. 2 equilibrium climate model projections compared with observational statistics which describe the non-equilibrium reality. It is, however, difficult to assess a realistic phase shift of the climate in respect to the greenhouse-gas forcing. Based on a statistical cross correlation analysis and statistical best fit strategies (see Figs. 7-9) a magnitude of roughly 20 years may be supposed (Schönwiese, 1987). Following this hypothesis the statistically assessed temperature rise in case of a CO₂ doubling (air temperature near the ground, global average) would amplify from 2.7-4.4°C (as stated above) to 3.9-6.0°C (see Table 2). Another problem is the nonlinearity of the climate response. Assuming a logarithmic relationship as discussed elsewhere, this temperature response would be damped by a factor of approximately 0.7, which would compensate, more or less, the phase shift effect.

Table 2. Similar to Table 1, but implying a phase shift of 20 years (gas concentrations leading) and only land based air temperatures.

<table>
<thead>
<tr>
<th>Record</th>
<th>Period</th>
<th>Mean mult. correlation</th>
<th>“Industrial” signals (in K)</th>
<th>Projected signals (in K)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>only CO₂ “equivalent”</td>
<td>Reduced to CO₂ “equivalent”</td>
<td></td>
</tr>
<tr>
<td>TNH-J</td>
<td>1881–1984</td>
<td>0.84</td>
<td>0.8±0.1</td>
<td>0.9±0.1 (3.5)</td>
</tr>
<tr>
<td>*TNH-J</td>
<td>1881–1983</td>
<td>0.74</td>
<td>1.2±0.2</td>
<td>1.3±0.2 (5.1)</td>
</tr>
<tr>
<td>TSH-J</td>
<td>1881–1984</td>
<td>0.91</td>
<td>1.0±0.1</td>
<td>1.2±0.1 (6.6)</td>
</tr>
<tr>
<td>*TSH-J</td>
<td>1881–1983</td>
<td>0.84</td>
<td>1.2±0.2</td>
<td>1.3±0.2 (7.4)</td>
</tr>
<tr>
<td>TGL-J</td>
<td>1881–1984</td>
<td>0.88</td>
<td>0.9±0.1</td>
<td>1.0±0.1 (5.0)</td>
</tr>
<tr>
<td>*TGL-J</td>
<td>1858–1983</td>
<td>0.82</td>
<td>1.2±0.2</td>
<td>1.3±0.2 (6.4)</td>
</tr>
</tbody>
</table>

Table 3 specifies similar “industrial” and “projected” signals derived from the global-mean temperature series combining land and marine data (Fig. 9). Without phase shifts, CO₂ doubling signals of 2.0–5.2°C are derived, in the “equivalent” case 1.8-7.0°C. Assuming a phase shift of 25 years (trace-gas concentration), these signals amplify to 3.6-5.6°C (CO₂ only) or 5.9–7.9°C (“equivalent”). In case of the stratospheric temperature, a simple extrapolation of the linear trend leads to a CO₂ doubling signal of roughly 10°C.

Although the hemispheric and global signals assessed by both climate model experiments and multivariate statistics are of a similar order of magnitude, this may be by chance. The error probability decreases considerably, however, if also the seasonal-regional patterns of the signal assessments are similar. Figure 11 shows two examples for the Northern Hemisphere air temperature (near the ground, CO₂ doubling signals). In contrast to a previous study (Schönwiese and Malcher, 1987; this study also includes
Table 3. Similar to Tables 1 and 2, but global-mean land and marine data.

<table>
<thead>
<tr>
<th>Record</th>
<th>Period</th>
<th>Mean mult. correlation</th>
<th>&quot;Industrial&quot; signals (in K)</th>
<th>Reduced to CO$_2$ (S/N)</th>
<th>Projected signals (in K)</th>
<th>&quot;Equivalent&quot; CO$_2$ doubling</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Without phase shifts</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGL-LM</td>
<td>1881–1980</td>
<td>0.78</td>
<td>0.8±0.4</td>
<td>0.8±0.5 (4.7)</td>
<td>0.5±0.3</td>
<td>3.6±1.6</td>
</tr>
<tr>
<td>*TGL-LM</td>
<td>1881–1980</td>
<td>0.72</td>
<td>0.7±0.2</td>
<td>0.7±0.3 (3.9)</td>
<td>0.4±0.2</td>
<td>3.1±1.0</td>
</tr>
<tr>
<td><strong>Including phase shifts</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGL-LM</td>
<td>1881–1984</td>
<td>0.87</td>
<td>1.0±0.2</td>
<td>1.2±0.2 (7.3)</td>
<td>0.8±0.1</td>
<td>4.6±1.0</td>
</tr>
<tr>
<td>*TGL-LM</td>
<td>1881–1983</td>
<td>0.80</td>
<td>1.1±0.1</td>
<td>1.2±0.2 (7.4)</td>
<td>0.8±0.1</td>
<td>4.8±0.5</td>
</tr>
</tbody>
</table>

*phase shifts: trace-gas concentrations 25 years, volcanic parameters 2-year (forcing parameters leading)

---

**Figure 11.** Seasonal-latitudinal pattern of CO$_2$ doubling signals derived and extrapolated from the multivariate regressions (mean of all forcing alternatives) as described in the text, left side without and right side implying ENSO forcing.
a regional signal-to-noise discussion), the maximum signals appear in spring instead of winter. Except for this seasonal peculiarity, the statistical mean-latitudinal results are again very similar to some GCM experiment results (for details see Schönwiese and Malcher, 1987; Schönwiese and Runge, 1988). The regional aspect, however, reveals very pronounced differences; see Figs. 12 and 13 (for details see again Schönwiese and Runge, 1988; the maximum signal-to-noise ratios appear in spring and summer, 50°-70°N).

In the case of precipitation the situation is much more complicated because there are not only increasing but also decreasing trends expected. Moreover, the climate model discrepancies are much more pronounced and a relatively small amount of variance can be explained by statistics. Figure 14 shows, as a very preliminary result, a statistical best fit study for the European 30°-35°N latitude zone using additionally (in comparison to Figs. 7-9) the North Atlantic Oscillation forcing and autoregressive terms in Eq. (1). The analysis is not yet finished. (For some preliminary results see Birrong and Schönwiese, 1988; Schönwiese and Birrong, 1990).

5. CONCLUSION AND OUTLOOK

Multivariate statistical signal assessments contribute to the projections and identifications of greenhouse-gas-induced climatic change and are of particular interest in comparison
Figure 14. Observed European 30°-35°N spring precipitation, see Fig. 3b, 10-year low-pass filtered and multivariate statistical reproductions. The dashed line (annual data $r_m = 0.46$) implies SVI*, SRN, ENW, NAO (North Atlantic pressure oscillation) and CO$_2$ forcing (without phase shifts), the dotted line implies 8 additional autoregressive terms (annual data $r_m = 0.60$; all computations based on unfiltered seasonal data, outcome 10-year low-pass filtered).

with corresponding GCM experiments. In future the analysis of temperature and sea level data should be extended to other climatic elements where the regional and seasonal assessments (signal and signal-to-noise) are very important. From the statistical point of view, autocorrelation, non-stability of correlation, and phase shifts remain to be the major problems. In order to enable some progress in the multivariate studies, not only GCM (CO$_2$ doubling etc.), but also simplified EBM and RCM experiments are necessary to simulate simultaneously the transient response of not only trace gas, but also natural forcing.

Nevertheless, the temperature assessments are most reliable at the moment and may be used to assess the time at which the anthropogenic signal in climate may be detected. Figure 15 indicates this will require an "equivalent" CO$_2$ concentration of about 445 ppm (exceeding the 99% confidence level in case of the minor response) which may take place approximately in the years 2000–2005.

ACKNOWLEDGEMENTS

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Figure 15. Temperature response (global-mean near surface, land areas, data only from Jones et al. (1982), so differing somewhat from the results listed in Table 1) to increasing greenhouse gases where the solid (CO₂ only) and dashed (CO₂ equivalent) lines delimit the standard deviations of the statistical assessments based on all regressions (similar to "±" in Table 1, without El Niño). A = "industrial" CO₂ signal, C = same but reduced due to equivalent computations, B = "equivalent industrial" signal (all equilibrium assessments) and D = "equivalent industrial" signal implying a 20-year phase shift (therefore appearing 20 years ahead). The three times standard deviation of observational data (corresponding approximately to a 99% confidence level, CL, or 0.01 error probability) may be exceeded, using the lower "equivalent" assessment, if the "equivalent" CO₂ concentration exceeds approximately 445 ppm (E). According to Tricot and Berger (1987) this may be between the years 2000 and 2005.

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ABSTRACT. Some broad issues arising in the statistical comparison of the output of climate models with the corresponding climate data are reviewed. Particular attention is paid to the question of detecting climate change.

1. INTRODUCTION

The purpose of this paper is to review some statistical approaches to the comparison of the output of climate models with climate data. There are many statistical issues arising in such a comparison. I will focus on some of the broader issues, although some specific methodological questions will arise along the way. One important potential application of the approaches discussed in this paper is the detection of climate change. Although much of the discussion will be fairly general, I will try to point out the appropriate connections to the detection question.

I have made no attempt to review the climatological literature on this issue. An exception is the recent paper by Barnett and Schlesinger (1987).

2. TESTING ONE CLIMATE MODEL

I will assume that climate model output and climate data are available at discrete locations in space and for discrete time periods. Let \( Y_{it} \) be the observed climate at location \( i \) (\( i = 1, ..., p \)) in period \( t \) (\( t = 1, ..., n \)). In general, \( Y_{it} \) is a multivariate observation, that is, more than one climatological variable is measured. This complicates the statistical methodology somewhat. However, the broad principles underlying the comparison of climate model output and climate data remain the same. For this reason I will proceed as if \( Y_{it} \) is univariate.
A general model for $Y_{it}$ is

$$Y_{it} = m_{it} + e_{it}, \quad i = 1, \ldots, p; \ t = 1, \ldots, n,$$

(1)

where $m_{it}$ is a systematic component and $e_{it}$ is a zero-mean random component. I will make the further assumption that the random component is normally distributed (perhaps after transformation). Non-normality of the random component introduces methodological complications, but again the broad principles remain the same.

A climate model produces output $a_{it}$ ($i = 1, \ldots, p; \ t = 1, \ldots, n$). I assume that there is a one-to-one correspondence between the values produced by the climate model and the climate observations. This is an approximation, as it is not possible in general to assign the output of a climate model to a particular location. I also assume that the climate model output is non-random. This means, for example, that the climate model should give exactly the same values of $a_{it}$ each time it is run under the same conditions. If the climate model output contains a random component, then I will suppose that it represents the average of the output values for location $i$ and period $t$ over a large number of runs.

The first question that we may wish to ask is whether the climate model has successfully predicted the behavior of the data. In statistical terms this means that we wish to test the null hypothesis

$$H_0: \ m_{it} = a_{it}, \ i = 1, \ldots, p; \ t = 1, \ldots, n$$

against the general alternative hypothesis

$$H_1: \ m_{it} \neq a_{it} \text{ for at least one } i \text{ and } t.$$ 

Under $H_0$ the distribution of

$$\chi^2 = (Y - a)'C^{-1}(Y - a)$$

(2)

is approximately chi-squared with pn degrees of freedom, where $Y$ and $a$ are the pn-vectors of observations and output, respectively, and $C$ is the pn-by-pn variance matrix of the $e_{it}$. The quantity $\chi^2$ represents a scaled version of the residual sum of squares. The scaling arises from the covariance structure of the random component. A test of $H_0$ can be made by calculating $\chi^2$ and comparing it to its null distribution. Note that this assumes that the climate model output is not "fit" to the data.

To apply this test it is necessary to know $C$, that is, it is necessary to know the covariance between $e_{ih}$ and $e_{is}$ ($h, i = 1, \ldots, p; \ s, t = 1, \ldots, n$). These covariances are not known, but they can be estimated from the data. In general, there are $[pn \ (pn + 1)]/2$ covariances to estimate. However, the estimation effort can be reduced greatly if the structure of $C$ can be simplified. For example, it may be reasonable to assume that the variance of $e_{ih}$ does not depend on $t$ (i.e., $\text{Var}(e_{ih}) = v_i$). It may also be reasonable to assume that the pattern of spatial correlation remains stable through time (i.e., $\text{Corr}(e_{ih}, e_{is}) = r_{hi}$).

Once the structure of $C$ is chosen, one way to estimate the elements of $C$ is through a version of the following iterative procedure:

1. Form an initial estimate of $m_{it}$ ($i = 1, \ldots, p; \ t = 1, \ldots, n$) and form the residuals $Y_{it} - m_{it}^*$ ($i = 1, \ldots, p; \ t = 1, \ldots, n$), where $m_{it}^*$ is the estimate of $m_{it}$.
2. Under the assumed covariance structure, estimate the elements of $C$ from the residuals.
(3) If necessary, improve the estimation in (1) by taking account of the estimated covariance structure from (2).

Note that the estimation of $m_{it}$ ($i = 1, \ldots, p; t = 1, \ldots, n$) should (and can) be done without imposing unrealistic parametric structure on the form of the systematic component. A number of non-parametric methods are available to do this (e.g., Cleveland, 1979). Solow (1988) applied this procedure to estimate the behavior through time of the variance of a single time series under the assumption that, if the variance changed through time, it did so smoothly. Interesting methodological issues arise in the estimation of the systematic component of spatial-temporal data. Note, too, that all available data – and not just that subset which is being compared to climate model output – can be used in estimating the elements of $C$.

How might this approach be used to detect climate change? Suppose that a climate model is used to portray the systematic behavior of climate without climate change. The climate model output can be compared to the corresponding climate data via $\chi^2$. Suppose that the null hypothesis is rejected. Taking account of the probability of false rejection, what can we conclude? Two possible conclusions can be drawn: either climate change has occurred or climate change has not occurred, but the unchanged climate is different from that portrayed by the climate model.

These conclusions are quite different, and the results of the test cannot distinguish between the two. A climatologist might try to choose between these two conclusions by examining the way in which the data depart from the climate model output. This is formalized in the next section where I consider testing the null hypothesis against alternatives in the direction of climate change.

Suppose now that $H_0$ is not rejected. Taking account of the probability of false acceptance, what can we conclude? Two conclusions are again possible: either climate change has not occurred or climate change has occurred, but it is not a very big change. These conclusions are again rather different, and again a climatologist might try to choose between them by examining what differences, if any, exist between the climate model output and the corresponding data. This problem can also be avoided to some extent by sharpening the alternative hypothesis.

In passing, note that a useful, informal way to explore the differences between climate model output and climate data is to estimate (nonparametrically) the systematic component of the differences $(Y_{it} - a_{it})$, $i = 1, \ldots, p; t = 1, \ldots, n$.

### 3. TESTING ONE MODEL AGAINST ANOTHER

Suppose that a climate model is run under two sets of conditions. Let $a_{it}$ ($i = 1, \ldots, p; t = 1, \ldots, n$) be the climate model output for the first set of conditions and $b_{it}$ ($i = 1, \ldots, p; t = 1, \ldots, n$) be the output for the second set of conditions. We wish to test the null hypothesis that the first set of conditions holds

$$H_0: \quad m_{it} = a_{it}, \quad i = 1, \ldots, p; t = 1, \ldots, n$$

against alternatives in the direction of the hypothesis that the second set of conditions holds

$$H_1: \quad m_{it} = b_{it}, \quad i = 1, \ldots, p; t = 1, \ldots, n$$

By extending the results of Hoel (1947) – see also Atkinson (1970) – a likelihood ratio test of $H_0$ against alternatives in the direction of $H_1$ can be based on the statistic.
\[ \chi^2 = (Y - a)'C^{-1}(Y - a) - (Y - c)'C^{-1}(Y - c) , \]  

where \( c \) is the \( pn \)-vector of values from the combined model

\[ c_i = w a_i + (1 - w) b_i, \quad i = 1, ..., p; \quad t = 1, ..., n , \]

with the weight \( w \) chosen to minimize \( (Y - c)'C^{-1}(Y - c) \). Under \( H_0 \), \( \chi^2 \) has an approximate chi-squared distribution with one degree of freedom. The test of \( H_0 \) can be made by calculating \( \chi^2 \) and comparing it to its null distribution. To use this test to detect climate change, a climate model could be run first "without" climate change and then "with" climate change, and the null hypothesis that climate change has not occurred could be tested against alternatives in the direction of the second model run.

By restricting the alternatives to lie in the direction of \( H_1 \), this test is preferable to that described in the previous section. It is important to note, however, that rejection of \( H_0 \) need not imply that \( H_1 \) is the better model, only that a linear combination of the two is better than both. With regard to the detection question, we must then decide whether this linear combination represents "no climate change" or "climate change."

In any case, this test restricts attention to models that can be expressed as linear combinations of the two basic models. This is a strong restriction, but it – or a similar restriction – is needed to allow formal testing. If the true model does not have this form, at least approximately, then the test is not valid.

A different problem with this test is that it treats the null and alternative hypotheses asymmetrically. This asymmetry may be justified in certain situations. For example, suppose that the null model has been performing well, but that we wish to decide whether it should be improved by adding a component in the direction of \( H_1 \). In order to make this change we may need to be convinced that the null model is no longer performing well. On the other hand, if we are faced with a fresh choice between two models, then this asymmetry is not appropriate. The situation with regard to detection is not altogether clear. The approach outlined in the next section is symmetric in the competing models.

4. DISCRIMINATING BETWEEN TWO CLIMATE MODELS

The statistical issues arising in the comparison of models whose parameter spaces bear no relation to each other was first discussed by Cox (1961, 1962). In such a comparison, Cox distinguished between hypothesis testing and discrimination. In hypothesis testing, the goal is to determine whether the null model is reasonable in light of a competing model. In discrimination, the goal is to assess the extent to which each of the models provides a reasonable fit to the data. In discrimination, the two models are on an equal footing.

In the discrimination problem, two competing climate models with output \( a_t \) and \( b_t, (i = 1, ..., p; t = 1, ..., n) \), respectively, are considered. A statistic that is useful in discriminating between these two models is the likelihood ratio

\[ L = \frac{(Y - a)'C^{-1}(Y - a)}{(Y - b)'C^{-1}(Y - b) } \]

or its logarithm. The distribution of \( L \) can be found under each model. Let these distributions be denoted by \( F_a \) and \( F_b \). The problem is then identical to the standard discrimination problem (Eand, 1981), that is, by comparing the observed value of \( L \) to \( F_a \) and \( F_b \) we can choose between four alternative conclusions: the data are consistent with the first model, the data are consistent with the second model, the data are consistent with both models,
or the data are not consistent with either model. Moreover, for a given discrimination rule we can use $F_a$ and $F_b$ to make statements about various misclassification probabilities.

To apply this approach it is necessary to know $F_a$ and $F_b$. Cox worked out these distributions using asymptotic results. Convergence to the asymptotic distributions appears to be slow. To overcome this, Williams (1970) suggested estimating $F_a$ and $F_b$ by simulation. This involves:

1. simulating a realization of the random component
2. forming the quantities

$$L_a = \frac{e^*C^{-1}e^*}{(a - b + e^*)(a - b + e^*)}$$

and

$$L_b = \frac{e^*C^{-1}e^*}{(b - a + e^*)(b - a + e^*)}$$

where $e^*$ is the pn-vector of the simulated random component;

3. repeating (1) and (2) many times and estimating $F_a$ and $F_b$ from the empirical distributions of $L_a$ and $L_b$.

One problem with this overall approach to discrimination is that a lack of fit to both models need not produce a value of $L$ that is unlikely under both $F_a$ and $F_b$. To avoid this problem, the approach can be extended in the following way. Instead of performing univariate discrimination based on $L$, we can perform bivariate discrimination based on the quantities

$$D_a = (Y - a)'C^{-1}(Y - a)$$

and

$$D_b = (Y - b)'C^{-1}(Y - b)$$

The bivariate distribution of $D_a$ and $D_b$ under each model can be estimated by simulation in a simple extension of the method outlined above. In this way the discrimination procedure can take account of absolute fit, not just relative fit.

The application of this approach to the detection question is obvious. The approach based on discrimination has several attractive features. It involves two alternatives that are treated symmetrically. Provision is made for rejecting both models or neither model. In addition, the approach can be extended to discriminate among more than two models. Finally, it is straightforward to take misclassification costs into account in devising the discrimination rule.

5. CONCLUSIONS

The purpose of this paper has been to review some of the broader statistical issues arising in the comparison of climate model output and climate data. The methods that I have described in this paper may seem overly formal, and less formal methods are certainly available. While formality is not necessarily a virtue, it does impose a kind of logic on the detection question. Without this kind of logic, it is easy to devise and apply a detection
method that is fundamentally flawed. Perhaps the main conclusion of this paper is that a little thought should go into the choice of a particular method of comparison.

Clearly, in order for this approach to be useful in addressing the detection question, climate models need to be reasonably accurate in representing the possible states of climate. Otherwise, we will be continually faced with data that do not match any set of climate model predictions. This poses a more serious problem for the hypothesis testing approach – which assumes that the true model lies either under the null hypothesis or the alternative hypothesis – than for discrimination – which allows for the possibility that neither model is (or both models are) correct. In any case, it seems sensible to base the comparison on climate variables that are expected to be accurately represented by climate models. Although these variables should also be expected to respond to climate change, they need not coincide with those for which the climate model predicts (perhaps incorrectly) maximal response under climate change.

Finally, although I have referred to situations in which a climate model is run “with” and “without” climate change, it seems to me that this concept is problematic. Climate change is not controlled by a switch that is either on or off. A climate model could be run with and without a specified change in atmospheric composition. However, inferences based on comparing the output to data would refer strictly to the question of whether or not atmospheric composition has changed. Since we are able to monitor atmospheric composition directly, this test is unnecessary and may well lead to absurd results (e.g., we may reject the hypothesis that atmospheric composition has changed when, in fact, we know that it has).

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Multivariate Methods for the Detection of Greenhouse-Gas-Induced Climate Change

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ABSTRACT. This investigation considers whether observed changes in surface air temperature are consistent with GCM equilibrium response predictions for a doubling of atmospheric CO₂. The model considered is a version of the Oregon State University (OSU) atmospheric general circulation model (AGCM). The study consists of three stages. In the first stage we examine the spatial structure of changes in the annual mean and annual cycle for surface air temperature, mean sea-level pressure (SLP) and precipitation rate. Signal-to-noise (S/N) ratios or equivalent test statistics are then computed (using the 1×CO₂ and 2×CO₂ data) in order to identify variables most useful for detection purposes. Changes in both means and variances are considered as possible detection parameters. The highest S/N ratios are obtained for annual-mean and winter surface air temperature, and the lowest S/N ratios are obtained for SLP. There are significant increases in the temporal and spatial variability of precipitation, and significant decreases in the temporal and spatial variability of surface air temperature.
In the second stage we examine the spatial structure of observed changes in surface air temperature and determine their statistical significance. Significance is assessed using a permutation procedure and the statistics applied in the S/N analysis. The results indicate that there have been significant observed changes in the annual mean and the means for individual months. Observed changes in temporal and spatial variability are generally in the same direction as the model-predicted results.

The final stage of the investigation addresses the question of whether the model-predicted surface air temperature signal is present in the observed data. Anomaly fields are computed for the observed (1977–86 minus 1947–56) and simulated (2×CO₂ minus 1×CO₂) data. The simulated anomalies are scaled using results from a one-dimensional model. Global tests of the mean indicate that there are highly significant overall differences between the observed changes and the scaled simulated changes in all tests performed. These results are strongly influenced by the low spatial variance of the simulated anomaly fields, and are not sensitive to the applied scaling or to the selection of data for defining observed changes. The spatial patterns of the observed and simulated temperature changes are significantly different in all cases except February.

Two possible explanations for the low degree of correspondence between observed and simulated patterns of temperature change are considered. The first is that the observed temperature signal is still too small to be detected against the background noise of natural variability. The second explanation is that the model signal may be erroneous due to model deficiencies and/or inherent differences in equilibrium and transient patterns of temperature change. Both explanations are likely to be valid.

1. INTRODUCTION

The problem of detecting the climatic effects of increasing CO₂ and trace-gas concentrations was initially considered by Madden and Ramanathan (1980) and Wigley and Jones (1981). These early studies used signal-to-noise analysis techniques in which only one variable was examined (e.g., zonally averaged surface air temperature). Univariate studies such as these may be able to detect a change in climate, but they are not well suited to the more relevant problem of attributing such a change, in whole or in part, to the enhanced greenhouse effect. To facilitate attribution, more recent investigations have addressed the problem of detection using the “fingerprint” strategy (MacCracken and Moses 1982; MacCracken and Luther, 1985) in which a number of different signal attributes are examined simultaneously, for example, tropospheric/stratospheric temperature-change contrasts (Epstein, 1982; Parker, 1985; Karoly, 1987, 1989). Recent fingerprint studies by Barnett (1986), Barnett and Schlesinger (1987) and Barnett et al. (1991) have attempted to identify optimum detection variables (i.e., those with high signal-to-noise (S/N) ratios in the model data) and have used a variety of statistics for comparing model output with observed data.

Almost all of these investigations employed results from GCM equilibrium-response experiments in order to define the climate-change signal associated with doubled or quadrupled levels of atmospheric CO₂. While only GCMs can provide a picture of the signal's regional and seasonal details, this does not mean that a state-of-the-art GCM can supply the correct picture of the possible atmospheric response to doubled atmospheric CO₂. There are large uncertainties in the modelling of cloud cover, height and radiative properties, the representation of the ocean, the coupling of atmosphere and ocean models, and in the parameterization of subgrid-scale physical processes. Such problems are common to all GCMs. The implication of these uncertainties is that current climate models are unlikely to incorporate correctly all the physical processes and feedback mechanisms that are necessary in order to simulate the climate impact of greenhouse-gas (GHG) forcing reliably.

A further problem in detection studies is that most of our knowledge concerning the regional and seasonal details of the GHG signal has been obtained from equilibrium-response experiments – that is, experiments which consider the statistically steady-state
response of the model's climate to a step-function change in atmospheric CO$_2$ (e.g., Washington and Meehl, 1984; Hansen et al., 1984; Wetherald and Manabe, 1986; Wilson and Mitchell, 1987; Schlesinger and Zhao, 1989). An optimum detection strategy should make use of the time dependence of an evolving greenhouse-gas-induced climate signal. Furthermore, recent evidence from the few GCM experiments that have been performed with time-dependent GHG forcing suggests that there are important differences between the equilibrium and transient response (Hansen et al., 1988; Washington and Meehl, 1989). For example, some transient experiments with ocean general circulation models (OGCMs) show interhemispheric asymmetry in the surface temperature response, with equator-to-pole amplification of the surface air temperature change in the Northern Hemisphere but not in the Southern Hemisphere (Bryan et al., 1988; Stouffer et al., 1989). Some of these experiments have generated warming minima, or even cooling, in areas of deep-water production (Mikolajewicz et al., 1990). These features are absent in equilibrium-response experiments performed with mixed-layer ocean models. Only one previous study (Barnett, 1991) has attempted to detect a CO$_2$ signal in results from GCM transient experiments.

Given uncertainties in the models themselves and in knowing whether important features of the equilibrium and transient responses are likely to be significantly different, is it meaningful to use equilibrium-response results for GHG detection purposes? The answer is yes, since there is also evidence that other large-scale spatial details of equilibrium experiments, such as land-sea contrasts in surface temperature changes, are similar to those obtained by transient runs (Bretherton et al., 1990). Furthermore, equilibrium results can be of considerable value for testing detection methods. Nevertheless, we stress that any study which uses imperfect models and a potentially unrealistic pattern of climate change cannot give unequivocal results regarding the detection of the enhanced greenhouse effect.

2. MODEL SIGNAL-TO-NOISE ANALYSIS

2.1. Simulated Data

In this study we have used monthly mean surface air temperature, sea-level pressure (SLP) and precipitation data from the final 10 years of extended 1xCO$_2$ and 2xCO$_2$ integrations performed with the OSU two-layer AGCM. The OSU AGCM has a horizontal resolution of 4° latitude x 5° longitude and was coupled with a 60 m mixed-layer ocean. Details of model physics, integration procedure, control run performance and 2xCO$_2$ versus 1xCO$_2$ comparisons are given in Schlesinger and Zhao (1989).

2.2. Geographical Response: Annual Mean and Annual Cycle

The aim of the first stage of the detection study is to determine the strength of the model-predicted CO$_2$ signal in the model data. As a prelude to model signal-to-noise analysis, we require some basic information on the spatial structure of the signal in different variables. This information is obtained by analysing the geographical distributions of 2xCO$_2$ minus 1xCO$_2$ changes in surface air temperature, SLP and precipitation rate for the annual mean and the amplitude of the annual cycle. The amplitude of the annual cycle is calculated using temporal Fourier analysis (cf. Shea, 1986).

The motivation for considering changes in the amplitude of the annual cycle is that equilibrium-response experiments concur in their prediction of a decrease in seasonality from 1xCO$_2$ to 2xCO$_2$; that is, the predicted warming is greater in winter than in summer, particularly at high latitudes (Schlesinger and Mitchell, 1985, 1987). We are interested
in examining the spatial structure of changes in seasonality for surface air temperature and other variables in order to determine whether the geographical distributions provide a coherent signal that might be useful in detection studies.

2.2.1. Surface Air Temperature

The geographical distribution of changes in annual-mean surface air temperature is given in Fig. 1. The largest temperature changes occur along the Antarctic coast (> 5°C) and the smallest changes are in the tropics (less than 3°C). Temperature changes show equator-to-pole amplification, a feature common to other AGCMs coupled to mixed-layer oceans (Schlesinger and Mitchell, 1987). The model response is statistically significant at all gridpoints (see Schlesinger and Zhao, 1989).

In order to interpret model-predicted changes in the amplitude of the annual cycle, it is useful to consider first how successfully the OSU AGCM describes the observed annual cycle. In the observations, the annual cycle of surface air temperature is larger over land masses than over the oceans, and the amplitude reaches maximum values over Alaska (> 22°C) and Siberia (> 30°C; see Shea, 1986). The model simulates this pattern well in the control run (Fig. 2a), but the observed amplitude maxima are underestimated by 6-8°C.

Figure 2b shows the geographical distribution of 2×CO2 minus 1×CO2 changes in the amplitude of the annual surface temperature cycle. The amplitude changes are spatially coherent, with an overall tendency towards decrease. The largest decreases (4°C) occur at locations where sea ice retreats from 1×CO2 to 2×CO2. This is due to feedback mechanisms involving snow and ice cover, which generate a larger warming in winter than in summer [see Robock (1983), and Im et al. (1989), for a detailed explanation]. The largest zonally averaged decreases in amplitude occur close to the North Pole in the Northern Hemisphere and between 65-75°S in the Southern Hemisphere.

It is interesting that the OSU model has areas where there are slight increases (1°C) in the amplitude of the annual surface temperature cycle, particularly in the tropics and in the vicinity of major subtropical highs. Similar increases are found in other models,
Figure 2a. Amplitude of the annual surface air temperature cycle (°C) simulated by the OSU two-layer AGCM for 1×CO₂.

Figure 2b. Change in the amplitude of the annual surface air temperature cycle (°C) for 2×CO₂ minus 1×CO₂ as simulated by the OSU two-layer AGCM. Shading indicates a decrease in amplitude. The largest decreases in amplitude are at locations where sea ice retreats from 1×CO₂ to 2×CO₂.

and may be due to reductions in cloud cover and summer soil moisture (J. F. B. Mitchell, personal communication) or to latent heat releases associated with increased precipitation. For the OSU model there is some evidence that areas of precipitation increase correspond with areas where the amplitude of the surface temperature cycle increases (compare Figs. 2b and 5). Univariate t-tests (not shown) indicate that the significant changes in the
amplitude are confined to high latitudes, so some of the areas of amplitude increase may be noise.

The zonal-mean surface air temperature changes presented in Schlesinger and Mitchell (1987) for equilibrium-response experiments performed with the GISS and NCAR GCMs have seasonal-cycle changes that are similar to those presented here for OSU.

2.2.2. Sea-Level Pressure (SLP)

Changes in annual-mean SLP are given in Fig. 3. The OSU model simulates pressure increases of 1-2 mb in the vicinity of the Northern Hemisphere subtropical highs and in a broad band slightly northward of the circumpolar trough. In all other areas there are small pressure decreases (1-2 mb), except over Antarctica, where the decreases are large (2-5 mb). Some of these changes are likely to be related to interhemispheric transports of atmospheric mass and reductions in sea-ice coverage (see Wilson and Mitchell, 1987). Analysis of the time-mean 1×CO₂ and 2×CO₂ SLP fields reveals no evidence of any substantial latitudinal shifts of the major pressure systems. Univariate t-tests (not shown) indicate that the significant SLP response is generally confined to low latitudes, despite the large high-latitude perturbations in the surface air temperature field.

For the annual cycle of SLP there is good qualitative and quantitative agreement between the observed amplitude pattern (see Shea, 1986) and the pattern simulated in the model control run (Fig. 4a). Maximum amplitudes in both observed and simulated data are closely associated with the locations of semi-permanent highs and lows. The model overestimates the observed amplitude maximum in the vicinity of the Aleutian Low by about 4 mb.

Changes in the amplitude of the annual SLP cycle (Fig. 4b) are generally small (± 1 mb) and are significant at only a few gridpoints. The anomaly pattern shows far less spatial coherence than for surface air temperature.

Figure 3. Geographical distribution of the annual-mean sea level pressure change (mb) for 2×CO₂ minus 1×CO₂ as simulated by the OSU two-layer AGCM. Shading indicates a decrease in pressure.
2.2.3. Precipitation Rate

The largest increases and decreases in annual-mean precipitation rate occur between 30°N and 30°S (+0.5 to -1 mm/day; Fig. 5). Poleward of 30° latitude, precipitation changes are smaller and generally positive. This large-scale pattern of precipitation change is similar to that obtained in equilibrium-response experiments performed with other models (Schlesinger and Mitchell, 1987), although there is little intermodel agreement in the positions of maximum and minimum change. The changes in precipitation rate are only locally significant.

For precipitation rate the amplitude of the observed annual cycle is greatest in the tropics, with major regional maxima near the southwest coast of West Africa, the west coast of India and west of Thailand (Shea, 1986; Lau and Sheu, 1988). The OSU model fails to resolve these maxima, and control-run amplitudes are both qualitatively and quantitatively different from the observations (Fig. 6a). Schlesinger and Zhao (1989) have shown that the OSU AGCM underestimates the amplitude of the zonally averaged annual cycle by as much as 2 mm/day in the tropics, and that the model's errors in zonal-mean precipitation rate are predominantly annual cycle errors.

For precipitation, the 2×CO₂ minus 1×CO₂ changes in the amplitude of the annual cycle (Fig. 6b) are less spatially coherent than the annual-cycle changes for either surface air temperature or SLP (Figs. 2b and 4b, respectively), and are only locally significant. Areas where there are large increases (decreases) in the amplitude of the annual cycle tend to correspond to areas where there are large increases (decreases) in annual-mean precipitation rate.

This indicates that part of the simulated change in amplitude is simply related to changes in the total amount of precipitation. However, as Schlesinger and Zhao (1989) have shown, there are also noticeable interseasonal differences in the patterns of precipitation change which the OSU model generates, and such interseasonal differences are manifest in the annual-cycle changes. Other models also have season-to-season differences in their precipitation-change patterns, and in the case of the UKMO model these are related to shifts in the location of the ITCZ (Wilson and Mitchell, 1987).

2.3. Multivariate Significance of Simulated Changes

2.3.1. General Considerations

In this study we compare simulated and observed changes in means, temporal variances and spatial variances. This is the first time that the variability of a climate element has been considered as a possible detection parameter. For both first-order and second-order moments, one can identify optimum detection variables using some form of signal-to-noise ratio (S/N; Barnett and Schlesinger, 1987; Barnett et al., 1991). In general, a S/N ratio is an indicator of the statistical significance of a signal, so any measure of the significance level of a predicted change may be used to characterize S/N. By employing certain of the statistics introduced by Preisendorfer and Barnett (1983; henceforth PB) and Wigley and Santer (1990; henceforth WS), we are able to generalize the S/N concept to cover various measures of differences in the means and changes in higher-order moments.

1 For example, if the 1×CO₂ precipitation rate at a gridpoint is zero at any month in the annual cycle, a constant relative decrease must decrease the amplitude of the annual cycle.
Figure 4a. Amplitude of the annual cycle for mean sea-level pressure (mb) as simulated by the OSU two-layer AGCM for $1\times CO_2$. Maximum amplitudes are closely associated with the locations of semi-permanent highs and lows.

Figure 4b. Geographical distribution of the $2\times CO_2$ minus $1\times CO_2$ change in the amplitude of the annual cycle for mean sea-level pressure (mb) as simulated by the OSU two-layer AGCM. Shading indicates a decrease in amplitude. Changes in the amplitude of the annual cycle show less spatial coherence than for surface air temperature.

For changes in overall means, the PB SITES statistic and the WS T1 statistic are used as measures of the S/N ratio. These are, respectively, normalized non-directional and directional statistics in which the noise level is estimated using both the $1\times CO_2$ and $2\times CO_2$ data. In other studies the model noise level has been defined using $1\times CO_2$ data alone, which makes the implicit assumption that the true noise level does not change in
response to GHG forcing (Barnett et al., 1991). For changes in temporal and spatial variances we employ the SPRET1 and SPREX1 statistics as S/N indicators. Details of these statistics and their application in climate-change and model validation studies are given in WS and Santer and Wigley (1990).

We determine the significance of differences in means and variances with the PB Pool-Permutation Procedure (PPP). Permutation procedures such as PPP are useful in test situations where the reference distribution is unknown and the number of time samples is small. Significance levels calculated with PPP are sensitive to the effects of temporal autocorrelation (Zwiers, 1990). This is not a problem here.

Significance tests are performed in the following way. Let $T_{(1)}$ and $T_{(2)}$ represent the two space-time fields to be compared, where the subscripts in parentheses denote output from $1 \times CO_2$ and $2 \times CO_2$ integrations, respectively. In the simplest case we assume that $T_{(1)}$ and $T_{(2)}$ consist of one variable only (such as January surface air temperature) and that the spatial information is ordered sequentially with a one-to-one correspondence between the two data sets so that $T_{(1)}$ and $T_{(2)}$ are two-dimensional ($x, t$) arrays. We can write the elements of $T_{(1)}$ and $T_{(2)}$ as $T_{(1)xt}$ and $T_{(2)xt}$, where $x = 1, n_x$ and $t = 1, n_t$. Here $n_x$ is the number of gridpoints on the OSU global grid (3312) and $n_t$ is the number of time samples (10). We perform 45 tests of $1 \times CO_2$ versus $2 \times CO_2$ data. The reference distributions are generated with PPP using 1000 randomizations of the time ordering of $T_{(1)}$ and $T_{(2)}$. Further details of the significance testing procedure are given in PB and WS.

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2 3 variables (surface air temperature, SLP, precipitation) x (12 months + 3 harmonics ($k = 0, 1, 2$; annual mean, annual cycle and semi-annual cycle)).
Figure 6a. Amplitude of the annual cycle for precipitation rate (mm/day) as simulated by the OSU two-layer AGCM for 1×CO₂. The model amplitudes are both qualitatively and quantitatively different from the observations.

Figure 6b. Geographical distribution of the 2×CO₂ minus 1×CO₂ change in the amplitude of the annual cycle for precipitation rate (mm/day) as simulated by the OSU two-layer AGCM. Shading indicates a decrease in amplitude. Changes in the amplitude of the annual cycle are partly related to shifts in the location of the ITCZ.
2.3.2. Results

SITES

Actual statistic values for tests of $1\times CO_2$ versus $2\times CO_2$ data performed with SITES indicate that the S/N ratio for surface air temperature is much larger than for SLP or precipitation rate (Fig. 7a). This is consistent with results from S/N analyses for the OSU model performed by Barnett and Schlesinger (1987) and Barnett et al. (1991). The largest values of SITES are for changes in annual-mean surface air temperature and for temperature changes in winter months. For all three variables, the annual mean ($k = 0$) always has a higher S/N ratio than in individual months.

The significance levels (p-values) show that all SITES results for changes in surface air temperature and precipitation rate are significant at the 1% level or better (the p-value gives the probability that the null hypothesis of no signal is correct). For SLP, the null hypothesis of no signal cannot be rejected in 12 out of 15 tests. Although the p-values provide a better indicator of relative S/N, we have concentrated on the interpretation of actual test statistic values because many of the results are so highly significant that their p-values are zero (i.e., the actual test statistic value fell outside the range of values generated by PPP for the sampling distribution).

T1

The T1 statistic is a directional measure of overall differences in means. T1 values are given in Fig. 7b. Note that large but compensating differences in the $1\times CO_2$ and $2\times CO_2$ data sets can yield a non-significant result for T1 but a significant result for SITES (see WS).

The T1 values indicate that annual-mean and monthly mean surface air temperature and precipitation rate are consistently (and significantly) higher in the $2\times CO_2$ data than in the $1\times CO_2$ data. The reverse is true for SLP. There is an overall decrease in the amplitude of the annual surface air temperature cycle (see Fig. 2b). The amplitudes of the annual- and semi-annual cycles increase for precipitation rate and are close to zero for SLP. The p-values indicate that most differences in overall means are highly significant, with the exception of the semi-annual cycle for surface air temperature and 7 of the 15 tests involving SLP.

SPRET1

SPRET1 is an indicator of S/N ratio for changes in temporal variability. It is the ratio of spatially-averaged temporal variances in two data sets (see WS). Test statistic values greater than unity indicate that the overall temporal variance in the $2\times CO_2$ data is larger than in the $1\times CO_2$ data (Fig. 7c).

The most interesting results for SPRET1 are the consistent increase in the overall interannual variability of precipitation rate (9 of 15 results highly significant) and the decrease in the interannual variability of surface air temperature (7 of 15 results highly significant). Why should interannual variability show such changes from $1\times CO_2$ to $2\times CO_2$?

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3 Note that the data were not cosine weighted.
Figure 7. Tests of 1xCO₂ versus 2xCO₂ surface air temperature, SLP and precipitation rate data for the OSU two-layer AGCM. Results are the actual statistic values for tests of differences in overall means (SITES and T1; Figs. 7a,b), overall temporal variances (SPRET1; Fig. 7c) and overall spatial variances (SPREX1; Figure 7d). The statistics are measures of signal-to-noise behavior in the model-predicted data. The box at the right of each diagram shows results for tests of the annual mean, annual cycle and semi-annual cycle (k = 0,1,2, respectively). Results that achieve overall significance at the 1% level are indicated by solid symbols. Significance testing was performed using the Pool-Permutation Procedure (PPP) with 1000 randomizations of the 1xCO₂ and 2xCO₂ data sets. Both SITES and T1 show that surface air temperature has higher signal-to-noise ratios (in all months and for all harmonics) than SLP or precipitation rate (see Section 2.3). Statistic values greater than unity for SPRET1 and SPREX1 indicate higher overall variances in the 2xCO₂ data.
The spatial pattern of local variance changes for annual-mean precipitation (Fig. 8a) does not provide a clear physical explanation for the overall variance changes indicated by SPRET1. Variance changes for precipitation are extremely noisy and show no clear association with land/sea distribution or preferential latitudinal location. Possible explanations for the changes in temporal variability of precipitation include:

- Changes in the mean state. Since $1\times CO_2$ precipitation rates are small or close to zero at high latitudes, even a small increase in the number of high-latitude precipitation events can lead to local increases in temporal variance.

- An increase in spatial variability (see results for SPREX1).

- Shifts in the location of major storm tracks.

For surface air temperature there are large increases in temporal variability along the equator, and large decreases to the west of Chile, along the Antarctic coast and in the region of the Aleutian Low (Fig. 8b). Locally, decreases may be related to such factors as the retreat of sea ice, the weakening of major lows (e.g., pressure increases in the vicinity of the Aleutian Low; see Fig. 3) and large-scale changes in pressure gradients (weakening of the circumpolar trough and reduction of pressure over Antarctica; see Fig. 3). The increases in temporal variability along the equator require further investigation.

SPREX1

SPREX1, the spatial analogue of SPRET1, provides a S/N measure for changes in spatial variability (see WS). Statistic values greater than unity indicate higher overall spatial variances in the $2\times CO_2$ data (Fig. 7d). The spatial variability of surface temperature shows significant decreases in all months as a result of the reduced equator-to-pole temperature gradient. The largest reductions are in Northern Hemisphere winter.

The spatial variance of precipitation increases in 5 out of 12 months and for the annual and semi-annual cycles. Since SPREX1 is sensitive to outliers (Santer and Wigley, 1990), part of the change in spatial variance is related to large local increases and decreases in precipitation rate between 30°N and 30°S.

The spatial variance of SLP decreases in 9 out of 12 months and for the annual mean and annual and semi-annual cycles, reflecting overall decreases in large-scale pressure gradients, particularly in the vicinity of the Antarctic circumpolar trough (see Fig. 3). Only 2 of the 15 test results are significant.

3. OBSERVED CHANGES IN SURFACE AIR TEMPERATURE

Having determined the spatial structure and strength of the model-predicted signal in various fields, we are now interested in comparing simulated changes with observed changes in climate. How should the observed changes be defined? In order to address this question, we consider the spatial structure and significance of observed changes in surface air temperature.

3.1. Observed Data

The observed data consist of monthly mean land-based surface air temperatures and sea surface temperatures (SST) from the combined land/ocean data set described by Jones
The logarithm of the $2x\text{CO}_2/1x\text{CO}_2$ variance ratio is plotted in order to identify areas where $2x\text{CO}_2$ temporal variance is larger or smaller (shaded) than $1x\text{CO}_2$ temporal variance. For precipitation, variance changes are extremely noisy and show no clear association with land/sea distribution or preferential latitudinal location. For temperature, decreases in variability may be related to the retreat of sea ice, the weakening of major lows and large-scale changes in pressure gradients.

et al. (1991). The data are in the form of anomalies (relative to 1950-79) on a $5^\circ \times 5^\circ$ latitude/longitude grid, and are for the period 1854-1986. Coverage is nominally from $85^\circ\text{N}-90^\circ\text{S}$, but there are almost no data available south of $65^\circ\text{S}$ prior to 1958. To generate a data set of absolute temperatures (necessary for testing observed versus
observed changes in $k = 1, 2$), we added time-mean absolute values to the Jones et al. anomaly data.

Here we compare surface air temperature for a recent decade (1977–86) with data from decades which in global-average terms were among the coldest (1904–13) and warmest (1930–39) on record (see Jones and Wigley, 1990). The choice of these decades maximizes and minimizes the observed global-average temperature change. We also use a post-World War II decade (1947–56) with higher spatial coverage than either 1904–13 or 1930–39.

3.2. Spatial Structure of Observed Surface Air Temperature Changes

Figures 9a,b show the spatial structure of observed changes in annual-mean temperature for 1977–86 minus 1930–39 and for 1977–86 minus 1947–56. The corresponding figure for 1977–86 minus 1904–13 is not shown due to low spatial coverage in the earlier decade. It is interesting to compare these results with the corresponding simulated changes for 2×CO$_2$ minus 1×CO$_2$ (Fig. 1). At this stage we are concerned only with comparing the pattern and not the magnitude of the changes, since the real world can have experienced only a small fraction of the climate change simulated in response to CO$_2$ doubling.

Relative to both the “cold” (in global-average terms) decade 1904–13 and the “warm” decade 1930–39, the recent decade has undergone spatially coherent warming and cooling. Warming is largest relative to 1904–13, with maxima of 2°C over northwest Canada and the southeast coast of China. In the 1920s and 1930s the spatial characteristics of warming were more similar to model-predicted results than the more recent warming in the 1970s and 1980s (Wigley and Jones, 1981; Jones and Kelly, 1983). The 1930s were characterized by high-latitude warming in the Northern Hemisphere, whereas large-scale cooling occurred in the North Atlantic in the 1970s and early 1980s, possibly associated with changes in the rate of North Atlantic deep-water formation (Wigley and Raper, 1987; Mikolajewicz et al., 1990). These differences are evident in Fig. 9a. Relative to 1930–39, cooling of 1°–2°C has occurred in the North Atlantic and over large areas of the United States, Scandinavia, Europe and Africa.

All three of the observed patterns of temperature change are unlike the equilibrium-change pattern simulated by the OSU model, which projects warming at all gridpoints and equator-to-pole amplification of the temperature change (Fig. 1).

3.3. Significance of Observed Temperature Changes

Significance testing was performed as for the S/N analysis in Section 2.3 using the same test statistics and significance testing procedure. Surface air temperatures for the three selected decades (1904–13, 1930–39 and 1947–56) were tested against data for 1977–86 in order to determine the significance of overall differences in means and variances. Gridpoints with missing data were excluded from the analysis so that the number of gridpoints used in each test ($n_x$) varies from month to month and between the three decade-versus-decade tests. The number of valid gridpoints varies from 340 (for tests involving the annual mean, annual cycle and semi-annual cycle, 1904–13 vs. 1977–86) to 747 (for tests involving November temperatures, 1947–56 vs. 1977–86).

3.3.1. SITES and T1

As expected on the basis of the global-average temperature differences between the decades tested, the highest SITES values are for 1904–13 versus 1977–86 (Fig. 10a), indicating that differences in overall means are highest for these two decades. The lowest
Figure 9. Spatial pattern of observed changes in annual-mean surface air temperature (°C) for a recent decade (1977-86) minus an earlier “warm” decade (1930-39; Fig. 9a) and a decade with good spatial coverage (1947-56; Fig. 9b). The observed data are monthly mean land surface air temperatures and sea surface temperatures from a combined land/ocean data set (Jones et al., 1991). Shading denotes higher temperatures in the earlier decade. Note that spatially coherent warming and cooling have occurred in both cases. For plotting purposes only, data south of the equator are excluded and linear interpolation has been performed in order to fill in areas with missing data. The contour interval is 0.5°C.

Actual statistic values are for tests involving 1945-56. The results for all three sets of decade-versus-decade tests show that there have been overall changes in annual-mean and individual monthly temperatures that are highly field significant (p = 0.01). For all three decades the highest actual statistic values are obtained for tests of the annual mean.
The results for T1 indicate the direction of overall changes in surface air temperature (Fig. 10b). Positive T1 values denote higher overall mean temperatures in 1977–86. Overall, monthly and annual-mean temperatures in 1977–86 are consistently higher than
in 1904–13, and are higher than in 1930–39 and 1947–56 in all months except July, August and September. The direction of these changes is therefore broadly consistent with the model predictions (Fig. 7b). As for SITES, the largest and most significant changes are for 1977–86 versus the “cold” decade 1904–13. Although the model results predict a significant overall decrease in the amplitude of the annual cycle (Fig. 7b), the corresponding observed changes are not significant.

3.3.2. SPRET1 and SPREX1

While SITES and T1 reveal that there have been significant changes in overall mean temperature between different observed decades, the SPRET1 and SPREX1 statistics show that only a few of the observed changes in temporal and spatial variability are significant (Figs. 10c,d). However, the observed changes in individual months are generally in the same direction as the model-predicted results (Figs. 7c,d), that is, towards overall decreases in the interannual variability and spatial variance of surface air temperature.

4. STRENGTH OF THE MODEL-PREDICTED SIGNAL IN THE OBSERVED DATA

We are interested in comparing an observed change in climate, \( \Delta O \), with a simulated change in climate, \( \Delta P \), in order to see whether the surface air temperature signal predicted by the OSU model is present in the observed data. SLP and precipitation are excluded from the analysis since both variables have low S/N ratios in the model-predicted data (see Section 2.3).

4.1. Observed Data

Here we define \( \Delta O \) as the surface air temperature anomaly for 1977–86 relative to the time-average for 1947–56. We selected these decades in order to obtain reasonable spatial coverage, and because data before 1946 are less reliable over the oceans due to problems relating to the long-term homogeneity of SST. Other definitions of \( \Delta O \) are possible. For example, a previous study by Barnett and Schlesinger (1987) defined observed changes relative to an individual year rather than a reference-period average.

4.2. Simulated Data and Scaling Procedure

We define the \( 2\times\text{CO}_2 \) minus \( 1\times\text{CO}_2 \) change in the simulated climate, \( \Delta P_{\text{x}} \), as

\[
\Delta P_{\text{x}} = T_{(2)\text{x}} - T_{(1)\text{x}},
\]

where \( T_{(1)\text{x}} \) is the gridpoint time-average for \( 1\times\text{CO}_2 \). The surface air temperature data used for computing \( \Delta P \) are the same as those used for the model signal-to-noise analysis in Section 2.

The real world can have experienced only a fraction of the equilibrium temperature change predicted by the OSU model, partly because the equivalent \( \text{CO}_2 \) change over the period is much less than \( 2\times\text{CO}_2 \) minus \( 1\times\text{CO}_2 \), and partly because the observed signal is damped by oceanic thermal inertia. Before the observed and simulated anomaly fields can be compared, therefore, some form of scaling must be performed.

\footnote{In this notation, a time average is denoted by an overbar as well as a dot replacing the \( t \) in the index subscript.}
The modelled changes were scaled down by a factor $\xi$ defined as

$$\xi = \frac{\Delta T_{eq}}{\Delta T_{tr}},$$

where $\Delta T_{eq}$ is the global-mean equilibrium change for the OSU GCM (2.84°C) and $\Delta T_{tr}$ is the modelled transient global-mean temperature change between the mid-points of the decades 1947–56 and 1977–86. $\Delta T_{tr}$ values were calculated using observed greenhouse-gas forcing changes and the one-dimensional upwelling-diffusion model of Wigley and Raper (1987). $\Delta T_{tr}$ values depend critically on the climate sensitivity $\Delta T_{eq}$. Here, $\Delta T_{tr} = 0.39^\circ$C and $\xi = 7.29$ for the OSU model.

Changes are scaled by $\xi$ at all gridpoints, that is,

$$\Delta P'_x = \Delta P_x / \xi.$$

While we can use such a procedure to scale the model-predicted changes in the mean state, our information concerning the timing and magnitude of model-predicted variance changes is inadequate to justify any scaling of the variances of $\Delta P$. Therefore, no significance tests were possible for observed and simulated changes in temporal and spatial variances. For second-order moments, we are restricted to comparing the direction of model and real-world variance changes (see Section 3.3.2). Here, model-versus-observed significance tests are performed only for changes in means (which can be scaled) and changes in spatial patterns (which do not depend on $\xi$). We use the SITES and T1 statistics to assess the significance of differences in means and the $r$ statistic to assess the significance of differences in spatial patterns. The denominators (the pooled variance terms) in the expressions for SITES and T1 are unchanged and only the $\Delta P$ terms in the numerators are scaled (see WS)\footnote{Note that this scaling applies to the PPP-generated reference distribution as well as to the actual test statistic value.}.

In order to detect the modelled changes we need to compare the $\Delta O$ and $\Delta P'$ fields. Prior to statistical analysis, results from $\Delta P'$ are transformed from the model grid to the $5^\circ \times 5^\circ$ observed grid. Grid-points with missing data in $\Delta O$ and corresponding gridpoints in $\Delta P'$ are excluded from the analysis, so that $n_x$ varies from 802 (annual mean, annual and semi-annual cycles) to 1094 (November).

### 4.3. Observed Minus Scaled Simulated Difference Fields

The difference field for the observed annual-mean temperature changes minus the scaled simulated changes ($\Delta O$ minus $\Delta P'$) shows where the real world has experienced more warming or less warming than the scaled model projections (Fig. 11). If there were an exact correspondence between the observed and the scaled model changes, $\Delta O$ minus $\Delta P'$ would be zero at all gridpoints. This is clearly not the case. Since $\xi$ is large and since $\Delta P_x'$ is positive for virtually all values of $x_t$, the values of $\Delta P'$ are small and positive at all gridpoints. The $\Delta O$ minus $\Delta P$ patterns are therefore dominated by the observed changes.

### 4.4. Multivariate Significance

#### 4.4.1. General Considerations

Before discussing the results of significance tests for observed versus simulated changes, some general issues must be considered. First, detection and significance testing are
This must make detection difficult. Factors other than the enhanced greenhouse effect (solar, volcanic aerosols). These factors related in the following way. The tests performed have the null hypothesis that $\Delta O$ and $\Delta P$ are drawn from the same population. Rejection of the null hypothesis means that the predicted field cannot be identified in the observed data.

Second, we note that significant differences between $\Delta O$ and $\Delta P$ are to be expected a priori. The reason for this is that the scaled, predicted global-mean change of $0.39^\circ C$ is much larger than the observed global-mean change (ca. $0.15^\circ C$ over the decades considered). There are two alternative explanations for this discrepancy. First, it may be due to an error in the assumed climate sensitivity. However, a very low sensitivity (much less than $2.8^\circ C$) would be required to give a warming of only $0.15^\circ C$. Second, the global-mean warming over 1947–56 to 1977–86 has almost certainly been affected by forcing factors other than the enhanced greenhouse effect (solar, volcanic aerosols). These factors, together with internally generated natural variability (see, e.g., Wigley and Raper, 1990, 1991), constitute the noise against which the greenhouse signal must be detected. Since this noise level is apparently very substantial at the global-mean level, it is likely to be even more substantial at the smaller spatial scales being considered by this study. This must make detection difficult.

Finally, the results for tests of the means could be sensitive to the choice of the scaling factor $\xi$. An incorrect choice of $\xi$ could destroy a real similarity between the observed and simulated changes. As will be seen below, however, even if $\xi$ had been.

Figure 11. Observed minus scaled simulated changes in annual-mean surface air temperature ($^\circ C$). Observed changes are for the decade 1977–86 minus the 1947–56 average. Changes simulated by the OSU AGCM are for $2\times CO_2$ minus the $1\times CO_2$ average. The simulated anomaly fields have been scaled since the real world has experienced only a fraction of the equilibrium temperature change predicted by the OSU model (see Section 4.2). The difference field is dominated by the observed changes (see Fig. 9b). Shading denotes areas where the real world has experienced less warming than the scaled model projections. For plotting purposes only, data south of the equator are excluded and linear interpolation has been performed in order to fill in areas with missing data. The contour interval is $0.5^\circ C$.  

\[ a \text{ priori} \]
adjusted to produce the correct observed global-mean temperature change in the model
data, the results of our tests would be unchanged in terms of overall significance.

4.4.2. Results

The p-values for SITES indicate that there are highly field-significant (p < 0.01) overall
differences between the observed surface air temperature changes and the scaled simulated
changes (Fig. 12a). In all 15 cases (12 months + annual mean, annual cycle and semi-
annual cycle) we reject the hypothesis that the overall observed temperature changes are
consistent with the OSU model’s equilibrium response predictions.

Twelve of the 15 results for the T1 statistic also show a significant difference (Fig.
12b). For the annual mean and all individual months except March, the overall scaled
simulated changes are significantly larger than the observed changes. Thus, while the
observed warming in some regions (e.g., over Alaska) exceeds the scaled model-predicted
warming, overall surface air temperature changes show the opposite tendency (see Fig.
11).

The results for SITES and T1 are dictated by the previously noted differences in
the spatial coherence of the observed and simulated changes (see Section 4.3). The fact
that the observed changes are large and both positive and negative, while the simulated
changes are positive everywhere, guarantees a significant result for SITES6 no matter
how the scaling of AP is performed.

Based on the pattern correlation statistic (r), the time-mean spatial patterns of ob-
served and simulated temperature changes are significantly different in all cases except
February (Fig. 12c). For this month alone we cannot reject the null hypothesis that the
observed spatial pattern changes are consistent with the OSU results. The spatial pat-
tern results reflect the large differences in the spatial coherence of warming. The pattern
correlations are independent of the magnitude of the scaling factor $\xi$.

We can draw several general conclusions on the basis of the $\Delta O$ versus $\Delta P$ signifi-
cance test results.

First, since there are large differences in the spatial coherence of observed and model-
predicted surface air temperature changes, tests of differences in overall mean (using
statistics such as SITES, Hotelling’s $T^2$, or cumulative gridpoint t-tests) are relatively
insensitive to the value of the scaling factor $\xi$. The most likely explanation for model
versus real-world differences in the spatial coherence of warming is that the observed
changes are dominated by natural variability, which swamps any possible greenhouse
signal.

Second, these large differences in (spatial) noise levels do not depend on how the
observed temperature changes are defined. Similar significance levels are obtained for
tests of $\Delta O$ vs $\Delta P$ overall differences in means if decades other than 1947–56 and 1977–
86 are used to define the observed temperature changes.

Third, we expect that tests involving spatial patterns of change are sensitive to how
the observed changes are defined. Several studies, including the present one (see Section
3 and Fig. 9), have shown that the spatial patterns of observed surface air temperature
change are not stable with time, and that the spatial characteristics of the early twentieth
century warming were more similar to model-predicted results than more recent warming
(Wigley and Jones, 1981; Jones and Kelly, 1983). Actual statistic values and significance
levels for $r$ may therefore depend critically on how $\Delta O$ is defined.

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6 But not for T1 if $\xi$ is adjusted to scale the simulated changes to the observed global-
mean change.
Finally, we have concentrated here on differences in the spatial coherence of warming in order to explain why the temperature changes simulated by the OSU AGCM are not consistent with observed changes. We note that model versus real world differences in the year-to-year variability of temperature changes are also large. If $\xi$ is set to 1.0 (so that the model temporal variances are not scaled), SPRET1 results show that the overall observed
changes in interannual variability (for 1977–86 minus 1947–56) are significantly larger than the simulated variance changes in 14 out of 15 cases. Since temporal variance enters into the definition of both SITES and T1, biases in interannual variability will also affect the results of tests of the overall mean state.

5. DISCUSSION AND CONCLUSIONS

There is no evidence that the equilibrium CO$_2$ signal predicted by the OSU model exists in the observed surface air temperature data. We would have obtained similar results if we had used the equilibrium-response projections from other GCMs with mixed-layer oceans, namely, GFDL, GISS, NCAR and UKMO. These models also show warming at virtually all gridpoints and times, and equator-to-pole amplification of surface air temperature changes in the winter hemisphere (Schlesinger and Mitchell, 1987). As in the case of the OSU model, the spatial and temporal variances of AO would be much larger than for AP for all published equilibrium response results. This virtually guarantees significant differences between the means and spatial patterns of AO and AP, no matter how the scaling of AP is performed.

There are two possible explanations for the inconsistencies between the observed temperature changes and the equilibrium-response predictions: either the observed surface air temperature signal is still too small to be detected against the background noise of natural variability, and/or we have been trying to detect the wrong signal (i.e., the model results are wrong).

The first explanation for our failure to detect a GHG signal (observed noise level too high) is clearly correct. Furthermore, it is evident that it will be some time before the spatial details of the (basically coherent) signal rises above the (spatially highly variable) noise, unless a large part of the observed patterns of change can be removed deterministically.

Because this study concentrates on the spatial patterns of GHG-induced climate change, the second explanation (erroneous model results) is also almost certainly correct. Current GCM deficiencies are well-known (Schlesinger and Mitchell, 1987), and their inability to simulate reliably the spatial details of present climate is one of the main motives behind the need to improve them. As noted earlier, this problem is exacerbated by our comparison of GCM equilibrium-response results with observed (i.e., transient) changes. The spatial patterns of change are almost certain to vary with time in the transition from 1×CO$_2$ to 2×CO$_2$, and the simple linear scaling used here is no more than an unsatisfactory compromise pending the availability of output from coupled models with a dynamic ocean and transient forcing.

Nevertheless, there are other model “errors” which it is impossible to remove, namely, effects related to model neglect of external forcings (solar, volcanic aerosols) and to the unpredictable effects of internal variability within the climate system (see Wigley and Raper, 1990, 1991). There is no way that any model can simulate these changes in any deterministic sense, so, for model-reality tests involving the mean state, they must remain as noise. Models can, however, be improved in the way they reproduce the statistics of temporal and spatial variability (see Santer and Wigley, 1990). An obvious example is the simulation of spatially coherent changes in ocean circulation, such as El Niño. El Niño is an important component of observed year-to-year and spatial variability which cannot be simulated adequately in AGCMs coupled to a mixed-layer ocean. Preliminary results from coupled models with fully dynamic oceans suggest that both spatial and temporal variability characteristics are simulated more realistically (U. Cubasch, personal communication).

There are a number of ways in which this study could be improved:
• By the use of results from transient experiments with realistic time-dependent GHG forcing, thus obviating the need for any scaling operations.
• By the use of data-compression techniques as a means of noise reduction prior to significance testing (e.g., zonal means, spherical harmonics, EOFs).
• By the use of tests that involve the simultaneous testing of a number of variables with high S/N ratios in the model data.

The primary aims of this study have been to demonstrate methodology and to consider the usefulness of equilibrium-response results for GHG detection purposes. Even if we had found evidence for the existence of the model signal in the observed data, we would have had no justification for claiming detection of the CO2 signal. The problem of unambiguously attributing observed climate changes to GHG forcing has not been considered here. Future detection studies must either define the climatic fingerprints of competing forcing mechanisms, or eliminate these as possible explanations in some other way before we can attribute an observed climate change to an increase in GHG concentrations.

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On Greenhouse Gas Signal Detection Strategies

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ABSTRACT. Important elements of a greenhouse gas signal detection strategy are discussed and demonstrated with both model and observed data. The analysis also demonstrates the high level of unexplained interdecadal variability that occurs naturally in the climate system and how this “noise” will greatly complicate any detection strategy.

1. INTRODUCTION

The indisputable detection of a greenhouse gas signal in the global climate system will be a true decision point for mankind. Yet, at this stage of knowledge, the details of an appropriate detection strategy are only beginning to be developed. Two key elements in the eventual strategy are the subject of this report: (1) what variables should be monitored in a detection program, and (2) the comparison of the equilibrium versus transient climate system response. Subsequent sections consider these items in turn, while a final section summarizes the main conclusion of the study.
2. OPTIMAL FIELDS FOR CO$_2$ DETECTION

2.1. Estimating Signal Strength in Model Simulations

The Oregon State University (OSU) atmospheric GCM (AGCM) was coupled with a mixed-layer ocean model and run for two 20-year periods: one with "present-day" CO$_2$ levels (326 ppmv, control = 1x) and another with twice today's level (2x). The last ten years of each integration were averaged to obtain mean fields for a variety of variables, and these averages were integrated spatially over the globe to obtain a single number for each run and variable. Denote these values, which approximate the time mean state of selected global fields under 1xCO$_2$ and 2xCO$_2$ concentrations, by $S_{1x}$ and $S_{2x}$, respectively. The last ten years of the 1x-run were used to estimate a "global" standard deviation for each variable, $\sigma_c$. Among the various possible definitions we chose to characterize the strength of the greenhouse gas effect with a signal/noise (S/N) ratio defined by

$$\frac{S}{N} = \frac{|S_{2x} - S_{1x}|}{\sigma_c} .$$

(1)

The S/N values for different model variables are given in Table 1. As one might have guessed, the atmospheric and sea surface temperature (SST) fields in the model experience the largest greenhouse gas effects. The planetary boundary layer (PBL) mixing ratio and ground temperature show the next largest relative effect, but it is doubtful whether observations of these latter variables can be made adequately on a global basis.

The large signal found in the atmospheric moisture content was unexpected and clearly represents a possible variable to monitor, particularly in view of the 20+ years of data on moisture content available from the upper-air observing net. This point will be addressed in the conclusion section of this report.

Many of the variables shown in Table 1 have S/N ratios less than 5. Given the nature of the numerical experiments, we feel these variables deserve less attention in any scheme to detect greenhouse gas effects. Clearly, the model results suggest the variables listed above the heavy line in Table 1 are the leading candidates for monitoring in a CO$_2$ detection system.

2.2. Estimating Signal Strength in Observations

Natural variability in the global climate system has a variety of spatial signatures. Ideally, this "natural" noise should be as distinctly different as possible from the expected greenhouse signal to maximize the "natural S/N." A measure of dissimilarity is provided by a pattern correlation function, that is,

$$r = \frac{\langle u(x)v(x) \rangle_x}{\langle u^2(x) \rangle_x^{1/2} \langle v^2(x) \rangle_x^{1/2}} ,$$

(2)

where $u(x) = S_{2x} - S_{1x}$, $v(x) = $ EOF1 of a variable listed in Table 1, and it's understood that $\langle u \rangle_x = \langle v \rangle_x = 0$ and that both $u$ and $v$ are defined on a common spatial grid. Since the spatially largest coherent pattern of variation in a variable is associated with its first EOF, it seems logical...
Table 1. CO₂ S/N Values: Model versus Observations.

<table>
<thead>
<tr>
<th>Field</th>
<th>Model S/N</th>
<th>Natural vs. CO₂ Pattern Similarity</th>
</tr>
</thead>
<tbody>
<tr>
<td>Temperature (400 mb)</td>
<td>13.24</td>
<td>0.60</td>
</tr>
<tr>
<td>Temperature (850 mb)</td>
<td>12.57</td>
<td>approx. as above</td>
</tr>
<tr>
<td>Temperature (Surface)</td>
<td>11.78</td>
<td>approx. as above</td>
</tr>
<tr>
<td>Sea surface temperature</td>
<td>11.31</td>
<td>0.04</td>
</tr>
<tr>
<td>Water vapor mixing ratio (PBL)</td>
<td>11.28</td>
<td>*</td>
</tr>
<tr>
<td>Water vapor mixing ratio (850 mb)</td>
<td>10.30</td>
<td>0.28†</td>
</tr>
<tr>
<td>Ground temperature</td>
<td>10.19</td>
<td>*</td>
</tr>
<tr>
<td>Water vapor mixing ratio (400 mb)</td>
<td>9.68</td>
<td>0.56†</td>
</tr>
<tr>
<td>Relative humidity (400 mb)</td>
<td>5.22</td>
<td>+</td>
</tr>
<tr>
<td>Surface evaporation</td>
<td>3.21</td>
<td>+</td>
</tr>
<tr>
<td>Relative humidity (850 mb)</td>
<td>3.17</td>
<td>+</td>
</tr>
<tr>
<td>Cloud cover</td>
<td>2.26</td>
<td>*</td>
</tr>
<tr>
<td>Snow melt</td>
<td>1.77</td>
<td>*</td>
</tr>
<tr>
<td>Relative humidity (surface)</td>
<td>1.74</td>
<td>+</td>
</tr>
<tr>
<td>Snow fall</td>
<td>1.52</td>
<td>*</td>
</tr>
<tr>
<td>Sea level pressure</td>
<td>1.39</td>
<td>**</td>
</tr>
<tr>
<td>Total precipitation</td>
<td>1.38</td>
<td>+</td>
</tr>
<tr>
<td>Ground wetness</td>
<td>1.29</td>
<td>*</td>
</tr>
</tbody>
</table>

* Development of these fields from observations is not likely.
+ Development of these fields from observations would require a large effort.
† Values based on limited data.
** Values not computed due to low model S/N.

to take this function as a measure of natural variability; that is, natural variation on time scales short relative to the time scale of a significant greenhouse gas signal. Thus if the expected greenhouse gas signal "looks" like natural, short-term variability, then \( r \approx 1 \). If the two signals are very different (orthogonal), then \( r \approx 0. \)

We found that the air temperature field was not a particularly good place to search for a greenhouse signal since \( r \approx 0.6 \), hence the short-term natural variation looks like the theoretically predicted signal (cf. Table 1). On the other hand, the SST field appears an excellent place to search for the expected signal. A further discussion of these results is given in Barnett and Schlesinger (1987).

The important new information in Table 1 from the model is the large S/N for atmospheric moisture content (denoted hereafter as Q). We obtained Q data for 9 stations in the tropical Pacific and computed the r values shown in Table 1. These show the natural variation in the Q field, especially at 850 mb, is rather dissimilar from the model-predicted CO₂ signal. Apparently the Q field also will be a good place to search for the greenhouse signal.

The above point is driven home by showing an annual average of the Q data in an atmospheric column between 700–500 mb (Fig. 1). The apparent change since the mid-1960s represents 30–40% of the mean.\(^1\) Other near-equatorial stations show the same

\(^1\) Additional work has shown that roughly 50% of this change is due to a change in instrumentation that took place in 1971–72.
Figure 1. Observed changes in column moisture content.

trend. Those stations away from the equator do not demonstrate such a remarkable signal. The signal at the surface (Fig. 1), while in the same sense as that at altitude, represents only about 5% of the mean and so will likely have a smaller S/N ratio.

3. EQUILIBRIUM VERSUS TRANSIENT SIGNALS

Much of the work done to date on projected greenhouse gas effects has been obtained from "equilibrium" runs, for example, viewing patterns from \((2x-1x)\) experiments. At the time we initiated this work we hoped to obtain results from a numerical "transient" run from the GISS model to investigate the difference between transient and equilibrium
runs. We were initially unable to obtain these data and this has limited our effort. But we have been able to use the transient parts of prior equilibrium experiments to obtain first-order estimates of the distortion present in equilibrium experiments and these are presented below. We will refer to these as "pseudo-transient" runs.

3.1. Response Pattern Stability

The key element in the greenhouse signal detection strategy proposed by Barnett (1986) was to search for a theoretically predicted greenhouse response pattern in the observational data set. While the original theory was general enough to handle time-dependent response patterns, the only model-predicted signals available to demonstrate the method at that time were those from equilibrium model runs. If the spatial response remains fairly static as the world progresses from a $1\times CO_2$ to $2\times CO_2$ concentration, then the spatial signals predicted by equilibrium runs are what we should be searching for in real detection studies. But if the spatial patterns change with time, then this fact needs to be taken into account when performing detection analysis.

How stable are the response patterns? To answer this question we used data from a 20-year model integration made with the OSU coupled ocean-atmosphere general circulation model. Denote a variable from the $2\times CO_2$ run as $U_{2x}(\overline{x},t)$, $t = 1, 2, \ldots, 20$ with $\overline{x}$ a global grid, and the $1\times$ run as $U_{1x}(\overline{x},t)$. We concentrate on the difference field $\delta(\overline{x},t) = U_{2x} - U_{1x}$. Some noise reduction is obtained by averaging four consecutive years of model data together, thereby reducing the number of temporal realizations to 5 in total, that is, $\delta_i(\overline{x})$ where $i = 1, 2, \ldots, 5$ and $i = 1$ refers to the average of years 1-4 inclusive, etc. The pattern correlation now measures the similarity between successive 4-year blocks of the "pseudo-transient" response,

$$r_{ij} = \frac{\left< \delta_i(\overline{x})\delta_j(\overline{x}) \right>_x}{\sqrt{\left< \delta_i^2 \right>_x \left< \delta_j^2 \right>_x}},$$

where $\left< \delta \right>_x \equiv 0$.

Thus if $r \approx 1$, there is little change in the spatial response of the model (the greenhouse signal) as the model transit from the $1\times CO_2$ to the $2\times CO_2$ regime. If $r \approx$ (small), the spatial patterns are different and the full time dependence of the spatial pattern needs to be taken into account in a detection study.

The values of $r_{ij}$ for various 4-year blocks are given in Table 2. The upper triangular element of the matrix refers to values obtained from the full 2-D boundary layer temperature field, while the lower triangular element refers to correlations obtained from zonally averaged boundary layer temperatures. The following conclusions are apparent:

(a) The detailed 2-D pattern rapidly approaches its "final" form (year 17-20). So to first order one could use the equilibrium response pattern in a detection study. However, the initial and final patterns are correlated ($r_{15}$) at only 0.59, so such an assumption is obviously omitting some potentially important information. Best to use the time-dependent response patterns if they are available.

(b) If one decides to use zonally averaged data in a detection study, then the equilibrium response is all that is needed because the first and last groups of 4-year averages are correlated at 0.83.
Similar estimates of $r_{ij}$ for the zonally averaged response were obtained for ground temperature, temperature and mixing ratio at 400 mb and 850 mb, precipitable water and PBL mixing ratio. Conclusion (b) above held for all these variables, except perhaps $T_{850}$, for which the assumption of "constant" response was marginal ($r_{15} \sim 0.58$). In summary, it is clear that use of the full 2- or 3-dimensional greenhouse gas signal will be a more discriminating detector than use of a zonally averaged signal.

Table 2. Response pattern correlation for "Pseudo-Transient" runs: Boundary layer temperature.

<table>
<thead>
<tr>
<th>Year Block</th>
<th>1–4 (1)</th>
<th>5–8 (2)</th>
<th>9–12 (3)</th>
<th>13–16 (4)</th>
<th>17–20 (5)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1–4 (1)</td>
<td>1.00</td>
<td>0.62</td>
<td>0.53</td>
<td>0.57</td>
<td>0.59</td>
</tr>
<tr>
<td>5–8 (2)</td>
<td>0.93</td>
<td>1.00</td>
<td>0.73</td>
<td>0.64</td>
<td>0.68</td>
</tr>
<tr>
<td>9–12 (3)</td>
<td>0.91</td>
<td>0.93</td>
<td>1.00</td>
<td>0.74</td>
<td>0.71</td>
</tr>
<tr>
<td>13–16 (4)</td>
<td>0.84</td>
<td>0.81</td>
<td>0.87</td>
<td>1.00</td>
<td>0.79</td>
</tr>
<tr>
<td>17–20 (5)</td>
<td>0.83</td>
<td>0.89</td>
<td>0.93</td>
<td>0.92</td>
<td>1.00</td>
</tr>
</tbody>
</table>

Zonal-Average Grid

3.2. Temporal Pseudo-Transient Response

An estimate of greenhouse-gas-induced surface air temperature change between 1880 and 1988 was constructed from the convolution integral

$$T(\lambda, \phi, t) - T(\lambda, \phi, t_0) = \int_{t_0}^{t} G \frac{dQ}{dn} F_c(\lambda, \phi, t - n) \, dn,$$  

(4)

where $t =$ time; $t_0 =$ initial time (1880); $(\phi, \lambda) =$ latitude, longitude; $T =$ annual average surface air temperature (°C); $G =$ "gain" of the climate system, taken to be 0.724°C/Wm$^{-2}$ based on 20-year 1 xCO$_2$ and 2 xCO$_2$ simulations with the coupled O/A GCM (Schlesinger et al., 1985); $Q(t) =$ time-dependent forcing taken from Wigley (1984) and shown in Fig. 2, which includes the effects of CO$_2$, CH$_4$, N$_2$O and CFC's; and $F_c =$ climate response function.

The main problem was to estimate $F_c$. This was done as follows. The $\Delta T(\lambda, \phi, t)$ from the 1x- and 2x-coupled O/A GCM simulation was normalized by the equilibrium $\Delta T(\lambda, \phi)$ from the AGCM/mixed-layer ocean model simulations for 1 xCO$_2$ and 2 xCO$_2$ simulation described by Schlesinger and Zhao (1989); the result is shown in Fig. 3.

The data were fit at each of the model's 3312 grid points by

$$F_c(\lambda, \phi, t) \approx 1 - \exp\{-[a(\lambda, \phi) + t/\tau(\lambda, \phi)]\} ,$$  

(5)

where the e-folding time $\tau(\lambda, \phi)$ and the parameter $a(\lambda, \phi)$ were obtained by least squares fit of

$$a(\lambda, \phi) + \frac{t}{\tau(\lambda, \phi)} = -\ln[1 - F_c(\lambda, \phi, t)] .$$  

(6)
Thermal forcing due to increase of different greenhouse gases

The zonal-mean value of the e-folding time is shown in Fig. 4, the global-mean of which gives a value of 28.6 years. Figure 5 gives the geographical distribution of the e-folding time. The negative values obtained in certain regions indicate a negative surface air temperature trend rather than a warming. These negative trends are illustrated in Fig. 6 for selected points. Figures 7a and 7b show the zonal mean and geographical distribution of the parameter $a(\lambda, \phi)$, respectively. The approximated zonal-mean values of $F_c(\lambda, \phi, t)$ are best fit by functions that either increase or decrease monotonically with time and so do not reproduce all the features of Fig. 3.

Both the actual global-mean value of $\tau(\lambda, \phi) = 28.6$ years and a value selected to be 60 years for contrast were used to calculate the temperature change expressed in Eq. (4). The resultant global means of $T(\lambda, \phi, t) - T(\lambda, \phi, t_0)$ are compared in Fig. 8 with the observed temperature changes from 1880-1986. The observations of Hansen and Lebedeff (1987, 1988), Jones et al. (1986) and Vinnikov et al. (1990) are shown with their zero points (reference periods of 1951-1980, 1951-75, and 1951-1975, respectively) shifted to be equal to the 1880-1884 average value of the average of the three observations. The case with $\tau(\lambda, \phi) = 0$ is also shown for comparison; this case represents the instantaneous equilibrium temperature change corresponding to the forcing $Q(t)$. From Fig. 8 it is evident that the global-mean e-folding time likely lies between 0 and 60 years.
Interestingly, if this analysis is correct, it means either that the 1880-1940 warming (actually an abrupt change in the early 1920s) was not due to greenhouse gases or: (i) the lag of the system would have to be near zero, (ii) the sensitivity to a doubling would have to be somewhat larger than the 2.84°C of the present model (a possibility), and (iii) there would have to be “something else” forcing a large cooling from 1940 to the present to offset the otherwise large greenhouse-gas-induced warming we should have seen by now. In either case, the agreement between the greenhouse-gas-induced warming and the observations is sufficiently poor that one should not state as Hansen (1988) that “Global warming has reached a level such that we can ascribe with a high degree of confidence a cause and effect relationship between the greenhouse effect and the observed warming.”

The resultant global distributions of $T(\lambda, \phi, t) - T(\lambda, \phi, t_0)$, $t_0 = 1880$, are shown in Figs. 9a through 9f for the actual, unscaled $\tau(\lambda, \phi)$ for the years 1900, 1920, 1940, 1960, 1980 and 2000 respectively. It is seen from these figures that in certain regions the warming is very small. These regions correspond to the areas where $\tau(\lambda, \phi) < 0$ and $F_c(\lambda, \phi, t)$ monotonically decreases and can become negative after a sufficiently long time. In such a case $F_c(\lambda, \phi, t)$ is set equal to zero in Eq. (4). Similarly, if $F_c(\lambda, \phi, t) > 1$ occurs, $F_c(\lambda, \phi, t)$ is set equal to unity in Eq. (4). In summary, the expected greenhouse-gas signal is not homogeneous over the earth. This in turn means attempting to detect such a signal using
Figure 4. Zonal mean of e-folding time of climate response function \((F_c)\) obtained by least squares fit.

global-mean data is a less-rigorous, less-discerning approach to the detection problem than an approach based on the full 2- or 3-D character of the signal.

4. MULTI-DIMENSIONAL DETECTION

In this section we summarize briefly several measures of signal detection and their interpretation. The approach is that of Barnett (1986) and could easily be expanded to a multivariate or "fingerprint" strategy as demonstrated in Barnett and Schlesinger (1987). Let us denote the predicted signal by \(\partial P(\bar{x},t,t_o)\) which is given by Eq. (4). The mean of \(P\) is \(\overline{\partial P}(t,t_o)\) given by

\[
\overline{\partial P}(t,t_o) = \langle \partial P(\bar{x},t,t_o) \rangle_x ,
\]

while biased and unbiased estimates of the variance of \(\partial P\) are, respectively,

\[
\sigma^2_p(t,t_o) = \langle \partial P^2(\bar{x},t,t_o) \rangle_x
\]

and

\[
\sigma^2_p(t,t_o) = \langle (\partial P(\bar{x},t,t_o) - \overline{\partial P}(t,t_o))^2 \rangle_x
\]
The exact same quantities may be defined for the observed variations of the predicted signal. We denote these by $\delta T$, $\sigma T$, $\sigma^2$ and $\sigma_v^2$ which are computed as in Eqs. (7) and (8). The expected change, $\delta T$, is computed directly from the observations relative to a reference time $t_0$. In the following examples we have:

(i) Set $t_0 = 1920$ to obtain adequate data coverage. The data extend through 1985.
(ii) Used as a variable the “surface” air temperature field.
(iii) Selected 32 geographic locations where continuous observations were available and extracted theoretical values of predicted temperature change for these locations (Fig. 10). Twenty-seven of the stations are in the Northern Hemisphere and five are in the Southern Hemisphere.

We now use various measures to determine if the predicted temperature change at the 32 locations since 1920 is observable in the observations from those measurement sites. Effects of urban heating and other contaminants are neglected so conclusions may error
in the direction of concluding the predicted signal has been observed (but see summary section).

4.1. Measure 1: "Global Means"

The areal-average mean ("global") changes since 1920 (Eq. (7)) are shown in Fig. 11 by decade, for example, the difference 1930–1920 and 1940–1920. The observed data were smoothed about the decadal boundary to reduce noise, that is, the average of 1928 through 1932 inclusive was assigned to 1930. The theoretical values were smooth enough and changed slowly so this procedure was not used on them.

This measure of observed comparative change shows much the same result as seen in Fig. 8: An abrupt warming (~0.2°C) between 1920–1930; little change thereafter until 1960; a sharp temperature drop to near 1920 levels between 1960–1970; a sharp temperature increase of about 0.4°C between 1970–1980. By contrast, the theoretical curve shows a slow monotonic rise such that 1980 is approximately 0.6°C warmer than 1920. The observed data show a warming of 0.5°C between 1920–1980. But clearly the observed data exhibit a large interdecadal change and do not have the time dependence.
Figure 7. The zonal mean (a) and geographic distribution (b) of the parameter a.
of the theoretical signal. In our view, the two curves are in no way equivalent except that they fortuitously have about the same ending point in 1980, and that observed ending value was realized in 10 years, not 60 years required for the predicted signal.

4.2. Measure 2: Raw Pattern Similarity

Let us denote the "raw pattern" similarity by $C_1(t)$ as in Barnett (1986), that is,

$$C_1(t) = \frac{\left< \frac{\partial P(x,t)}{\partial T(x,t)} \right> \cdot x}{\sigma_P \sigma_T} ,$$  

(9)
Figure 9. Temperature change due to increase in greenhouse gases for 1900, 1920, 1940, 1960, 1980 and 2000. Units are in °C.
Temperature change due to increase of greenhouse gases, 1940

Temperature change due to increase of greenhouse gases, 1960

Figure 9. Continued
Temperature change due to increase of greenhouse gases, 1980

Temperature change due to increase of greenhouse gases, 2000

Figure 9. Continued
where \( t_0 \) is suppressed but understood to be 1920. Note this measure looks like a pattern correlation but the means remain in the data. Nonetheless, it should be clear that \( C_1(t) = 1 \), independent of time, if \( \delta P(x,t) = \delta T(x,t) \).

The temporal evolution of \( C_1 \) (Fig. 12) has typical values of 0.5 up to 1960. The small value of \( C_1 \) was for the decade of the 1970's (relative to 1920) and shows the decadal change bore no resemblance to those expected. The sharp increase in \( C_1 \) between 1970 to 1980 is dominated by the behavior of the mean fields (see above). Unfortunately, the interdecadal variation demonstrated by \( C_1 (\sigma \approx 0.2) \) suggests the greenhouse-gas-signal will have to be significantly larger than the values predicted here before it will dominate the high-frequency "noise".

4.3. Measure 3: Pattern Correlation

Consider a traditional measure of pattern similarity given by

\[
C_2(t) = \frac{\langle (\delta P - \overline{\delta P})(\delta T - \overline{\delta T}) \rangle}{\sigma P \sigma T},
\]

where the \( (x,t) \) dependence on the right-hand side has been suppressed.

The mean change relative to \( t_0 \) has been removed from Eq. (9) to give Eq. (10). This mean change is a large part of the expected signal particularly as \( (t - t_0) \) increases. We saw from Measure 1 that this can be a noisy component of the signal. Yet the expected signal does have spatial structure (cf. Fig. 5) and it is that element that Eq. (10) will key on, that is, the relative changes in temperature between different parts of the globe.

The results of comparing observed and predicted changes in temperature since 1920 (Fig. 13) are illuminating. For instance, the observed difference between 1920–1940 is just about orthogonal to the pattern of predicted changes (\( C_2 \approx 0 \)). But one decade later, the observed and predicted patterns of change have their highest similarity, \( C_2 \approx 0.32 \). It seems clear that interdecadal variability is apt to dominate this measure and so its value in a greenhouse-gas-signal detection strategy is questionable (but see summary section).
4.4. Measure 4: Skill Score

In measuring and evaluating forecast models it is traditional to use a measure of "the variance accounted for" by the model, for example,

$$C_3(t) = 1 - \frac{\langle (\partial P - \partial T)^2 \rangle_x}{\sigma_T^2}.$$  \hspace{1cm} (11)

This measure, which is similar to Eq. (9), will equal unity if $\partial P$ and $\partial T$ track exactly. A variant of this measure replaces $\sigma_T$ in the denominator with $\sigma_T$, but the results of doing that in this case affect the results little. Note the numerator requires both the spatial mean and variation of $\partial P$ to equal those of $\partial T$ if a high score is to be obtained.

The skill score, which would be unity, independent of $t$, if the theoretical predictions were exact, shows the same interdecadal variation seen in the other measures (Fig. 14).
Figure 12. Raw pattern similarity ($C_1$) for temperature changes since 1920. $C_1 = 1$ if the theoretically predicted and the observed temperature change are equal. See Fig. 11 legend regarding time axis.

Apparently, it will be exceedingly difficult to say a CO$_2$ signal has been detected based on data increments of one decade of observations.

5. SUMMARY

Several elements of a CO$_2$ detection strategy have been investigated. The principal results of these studies are as follows:

(i) The most effective variables to monitor in a greenhouse gas detection program are sea surface temperature, lower-middle tropospheric moisture content, and free air temperature, respectively. This conclusion is based on a variety of considerations such as availability of the observations, $S/N$ values, and the similarity of natural and greenhouse gas signals.

(ii) Detection studies should concentrate on the full N-dimensional structure of the predicted signal. Failure to do so results in disregarding valuable data and doing so makes detection statements less vulnerable to error.
(iii) If one attempts to detect a full N-dimensional signal then the temporal evolution of the signal should be included in the detection study. If lower dimensional components of the signal must be used in detection studies, then it seems likely that results from equilibrium runs, that is, $1\times$CO$_2$ and $2\times$CO$_2$, will adequately represent the predicted signal.

(iv) Four different measures were used to see if a theoretically predicted increase in surface air temperature due to all greenhouse gases since 1920 could be detected. It could not. Although each measure focused on different aspects of the predicted signal, all were swamped by interdecadal "noise," with the last two decades providing much of the problem. It seems likely, subject to the reservations noted below, that the greenhouse gas signal will not be detectable with high certainty in the air temperature field until it is quite large.

Our study had many deficiencies. We ignored them in the text because we wanted mainly to demonstrate techniques and methods. But it is important to list them here to remind the reader that earlier conclusions must be considered tentative. Major shortcomings are:
(i) The results of sections 2 and 3 are likely to be model dependent. For instance, the results of Table 1 could change from model to model and hence affect the design of a greenhouse gas monitoring program.

(ii) The transient versus equilibrium nature of the greenhouse gas signal needs to be investigated with results from a fully transient O/A GCM run. Such data were not available to us at the time of this study so results such as in Table 2 must be considered tentative.

(iii) The theoretically predicted signal developed in section 3 and used in detection measures (section 4) implicitly took account of thermal lags introduced by the oceans and changes in cloud cover. But it is likely not as accurate as a full transient integration. Likewise, the observation set used in section 4 was spatially limited and subject to bias (urban warming). Only air temperature data were used and this likely isn't the best variable to use in a detection study. A different data set, more densely sampled and smoothed to avoid interdecadal variability, would be more appropriate for use in a detection study. In spite of these shortcomings, we feel the method demonstrated here will, when properly implemented, be a key element of a greenhouse gas detection strategy.

Figure 14. As in Fig. 12, but for the skill score, C₃.
ACKNOWLEDGEMENTS

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An Attempt to Detect the Greenhouse-Gas Signal in a Transient GCM Simulation

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ABSTRACT. Results from the GISS model forced by transient greenhouse-gas (GHG) increases are used to demonstrate methods of detecting the theoretically predicted GHG signal. The signal predicted to occur in the surface temperature of the world's ocean since 1958 is not found in the observations but this is not surprising since the signal was small in the first place. The main result of the study is to demonstrate many of the key issues/difficulties that attend the detection problem.

1. INTRODUCTION

Prior discussions of the greenhouse gas (GHG) signal-detection problem have generally focused on the use of GHG signals from GCM equilibrium runs (Barnett, 1986; Barnett and Schlesinger, 1987) or from "pseudo-transient" simulations (e.g., Barnett et al., 1991). In those studies no true transient simulations were available. That situation has changed and we present here some preliminary detection methodologies using the GISS transient simulations (Hansen et al., 1988).

Subsequent sections of this paper describe the data sets and mathematical/statistical detection strategies. Results of applying the methods to the GISS runs are described next. A final section summarizes the results of the study.

2. DATA

2.1. Model Data

The globally gridded fields from the GISS-A and GISS-B transient GHG simulations were provided courtesy of Roy Jenne (NCAR). The integrations begin in 1958 and extend to
2062 (A scenario) or 2029 (B scenario). The model fields were annually averaged on a grid with intervals of 7.83° latitude and 10.0° longitude. The model, the runs themselves and descriptions of scenarios are given in Hansen et al. (1984, 1988). In this study we will use the sea surface temperature (SST) data from these runs since this appears to be a particularly good variable to use in a “real world” search for the GHG signal (Barnett and Schlesinger, 1987; Barnett et al., 1991). However, the GISS model’s representation of the ocean leaves much to be desired so, from the model’s point of view, another variable might be more desirable for detection purposes.

2.2. Observational Data

The updated COADS data set (1950-86) gives gridded estimates of SST at the monthly average time scale. The data are obtained from ship reports, edited and projected onto a 2°x 2° grid. We further averaged the data annually and projected them onto the GISS grid. Latitudes above 60° were generally ignored as the data were too sparse in those areas. Further, we considered only grid points that had less than 10% of the possible data missing. The minor data gaps were filled by linear interpolation in space and time. A total of 331 grid points were available for comparison with the model SST fields.

2.3. Crude Comparison

Averaging the B-scenario model SST over all grid points and all years between 1958-86 gave a mean global ocean SST of 21.449°C (no area weighting). The comparable number for the observations was 21.265°C. The difference was 0.184°C. The agreement for a comparable measure, air temperature, was 0.059°C. These apparently excellent agreements are likely due in large part to the flux correction scheme applied to the model (cf. Hansen et al., 1984). (cf. Discussion section).

When viewed geographically the model performance is less spectacular. The difference between model and observed SST averaged over 1958-86 at each grid point is shown in Fig. 1. Large differences (2-4°C) of both signs are seen over the high latitude oceans, with the southern oceans containing the largest differences. Differences in

![SST DIFFERENCE FIELD OF LONG TERM MEANS (GISS-B-OBS)](image)

**Figure 1.** Difference between 29-year averages (1958-86) of sea surface temperature from GISS B-scenario transient greenhouse-gas simulation and the observed sea surface temperature. Units are degrees Celsius times 10. Negative values are stippled.
the tropical strip are generally less than 1°C. In all regions the differences are spatially coherent. Their sum over all oceans may be near zero, but regional differences may be significant with regard to model response.

3. METHODOLOGY

3.1. Selecting a GHG Metric

The model-predicted GHG signal in the surface air temperature for the “A” scenario is shown in Fig. 2. This “signal” was obtained by subtracting the first decade of model integration from the third decade of integration. We used the air temperature for this illustration since there is little difference between the air temperature and SST signals over the oceans. The full global field demonstrates that the model predicted signal has substantial spatial structure. Use of a global-average number for detection would ignore this structure.

A metric that will test the spatial structure of the signal, as well as the temporal evolution, is a pattern correlation metric. Such a measure is described by Preisendorfer and Barnett (1983). For present purposes we work with a measure that does not depend on time (but easily could; Barnett, 1986; see also Wigley and Raper, 1991.) The appropriate metric is

$$p(t_0) = \left< \frac{(\Delta_D - \overline{\Delta_D}) \cdot (\Delta_M - \overline{\Delta_M})}{\sigma_D \sigma_M} \right>_t,$$

where

$$\Delta_M(\mathbf{x}, t, t_0) = M(\mathbf{x}, t) - M(\mathbf{x}, t_0)$$

$$\Delta_D(\mathbf{x}, t, t_0) = D(\mathbf{x}, t) - D(\mathbf{x}, t_0)$$

$$\overline{\Delta(\cdot)} = \left< \Delta(\cdot) \right>_t$$

$$\sigma^2(\cdot) = \left< (\Delta(\cdot) - \overline{\Delta(\cdot)})^2 \right>_t,$$

with (D,M) the (observed, model) data fields and $t_0$ the initial year, in this case 1958. Thus, $p(t_0)$ describes the level of agreement between the way the model evolves the GHG-forced SST signal in space and time since 1958 and the way the actual SST field evolved. If $p=1$, then the model prediction would be fully realized in the observations. If $p=0$, the model predictions and observations would be orthogonal, that is, the evolutions of model and observed SST field over the oceans between 1958-86 bear no relation to each other.

Prior to computing $p(t_0)$ one ought to filter the (D,M) fields to concentrate on the spatial scales for which the model performance is deemed adequate. For example, expand
Figure 2a. Increase in decadal surface temperature from the GISS A-scenario greenhouse-gas simulation. The difference is computed by subtracting the grid-point average temperature for the period 1958-67 from a similar average for the period 1975-86. Units are 1/10 of a degree Celsius. The zero contour is indicated by the heavy line. Negative values are lightly stippled. A contour for ±0.5°C is also included.

Figure 2b. As above, except for the GISS B-scenario greenhouse-gas simulation.
AN ATTEMPT TO DETECT THE GREENHOUSE-GAS SIGNAL

M(\bar{x},t) as a set of EOFs\(^1\), \(e_{n}(\bar{x})\),

\[ M(\bar{x},t) = \sum_{n=1}^{N_c} A_n(t) e_n(\bar{x}) + \sum_{N_c+1}^{n_{\text{max}}} A_n(t) e_n(\bar{x}) . \]

Using some (undefined) objective scheme, we say that the M and D fields for, say, the annual cycle, agree for spatial scales equal to or larger than those associated with EOF mode No. For smaller scales associated with modes greater than No, the model and data disagree to the point that the model simulations are unreliable. This prefiltering allows one to use the most reliable portions of the model simulations in a detection scheme. It also limits the discussion to scales on which the model is valid and eliminates noise at high spatial wavenumbers that has nothing to do with the GHG signal. In the current work we are concentrating on detection methodology, not model validation, and so perform no prefiltering of the (D,M) fields. This should be remembered when viewing the results.

3.2. Significance Testing

A significance testing method called the Pooled Permutation Procedure (PPP) is used to determine if \(p(t_o)\) is meaningful (see Preisendorfer and Barnett, 1983). The PPP method essentially computes the statistic \(p\) for the original model/observation data sets. Call this value \(p_o\). The original (D,M) are then randomly interchanged (in time only; the spatial ordering is preserved) and \(p\) is recomputed. Many realizations of the interchange are carried out and a probability distribution function for \(p\) is constructed. If \(p_o\) falls within the mid-range of this distribution, we say that M and D are equal and the model predictions are verified. If \(p_o\) falls at or beyond the lower end of the distribution, we say the model predictions are not found in the observations.

The PPP method has both a major strength and weakness. It is nonparametric and automatically accounts for spatial/temporal correlation in the data fields. One need not worry about estimating degrees of freedom, field significance, etc. The weakness of the PPP method is that it has very high statistical power – too high for reasonable use if the number of time values in (D,M) exceed, say, ten. In a small sample environment, for which it was invented, PPP works fine. In the tests to be described below, we tried to create a small sample environment by computing \(p\) from non-overlapping three-year averages of (D,M). The averaging left 9 time-terms in both data sets and covered the model prediction for the period 1958-1984 inclusive.

4. RESULTS

4.1. Perfect Data

The globally integrated model/data SST are shown in Fig. 3. Based on this sample curve, one might be tempted to conclude the GHG signal is present in the observations. Removal of the 3-year average in the early 1980s, that includes a huge ENSO event, might dampen one's enthusiasm for the level of agreement. But perhaps not. A more rigorous approach to the hypothesis testing was obtained by following the procedures of Section 3.2.

\(^1\) Obviously any other basis set could be used in this decomposition.
GLOBAL AVERAGE SST CHANGE RELATIVE TO 1958-60

Figure 3. Globally averaged sea surface temperature difference relative to the tri-annual average 1958-60. Each point represents an average of 3 years of data. Symbol A refers to results from GISS-A run, symbol B refers to data from the GISS-B run, and the circles refer to observations.

The cumulative distribution functions (cdf) for \( p \) are shown in Fig. 4 along with the value of \( p_o \). For both the A and B scenarios the value of \( p_o \) is at the extreme low end of the cdf and has magnitude less than 0.2. Indeed, this end point of the cdf was the result of the random permutation selecting exactly the original (D,M) data. Thus any mixing of the original data sets gives higher pattern correlation than the original unrandomized (D,M) sets. It is concluded that the model-predicted behavior of the global SST field in response to GHG forcing was not present in the observations with confidence 100% (based on 100 permutations of D,M).

4.2. Nonperfect Data

Power tests suggested that reduction of (D,M) to nine equivalent time points still endowed the PPP test with a stringency that was somewhat unreasonable given that neither D nor M are error-free. This problem was partially overcome by softening, or detuning, the test as follows. The model was assumed to have an error in its mean field identical to
Figure 4. Cumulative distribution functions for the GISS-A and GISS-B runs. The variable under consideration is sea surface temperature. The detection metric is the pattern correlation over 27 years of common data. The heavy X's indicate the value of p obtained from the original model/data set (p₀). The pattern correlation values are given, as is the number of times the original D/M set was selected in the permutation process (in parentheses). It is concluded that the pattern of SST change predicted over the world's oceans by the model is not present in the observations.
the difference \( D(\bar{x}, t_0) - M(\bar{x}, t_0) \), that is, the tuning error in the flux correction scheme (cf. Fig. 1 and Hansen et al., 1984). This error was subtracted from all \( M(\bar{x}, t) \). It was further assumed that the yearly averages had a random rms error of order 0.5°C at each grid point. The observed data at each grid point were assumed to be unbiased but with a random rms error of 0.5°C.

New realizations of the original (D,M) data were constructed by incorporating the above error structure into the original (D,M), each realization being obtained by prescribing a new set of random errors. Each realization was then submitted to the PPP and a new realization of the cdf derived. The new values of \( p_0 \) were compared with the ensemble of cdf's to determine if the (D,M) were equivalent under the assumed error structure. Again, it was concluded that the model predictions were not present in the observations with confidence 98%. This statement is conditional on the assumed error structure.

Clearly, we could increase the magnitude of the assumed errors until the model and data agreed. But the size of the assumed errors is already generous; perhaps unsupportably so. In fact, one can "see" the statistical result in the observations – compare Fig. 5 with Figs. 1 and 2.

![Figure 5](image-url)

**Figure 5.** As in Fig. 2, except for observed sea surface temperature. Some results from upper-air stations appear over the land masses.

### 5. DISCUSSION

The results show rather clearly that the model-predicted changes in SST have not been observed. But obtaining that result raised some fundamental points regarding GHG detection strategies. These are:

(i) The use of a global average (anything) is a very poor metric for detection purposes. It neglects fundamental model predictions, is poorly able to discern between competing models, and apt to be biased by large, short-lived climate events (e.g., ENSO). Further good observations of truly globally averaged quantities are virtually impossible to obtain reliably except perhaps in the last several decades.
(ii) The error structure of the model/observations should be known or estimated so that the statistical testing procedures can be tuned to the appropriate power. Otherwise one is apt to use a test that is impossibly stringent or too lenient. It would also seem wise to evaluate the accuracy of the model as a function of spatial scale length. For instance, if the model is valid only for scales $\geq 1000$ km, then the model and observational data should be filtered accordingly prior to application of the detection test. That was not done in this work.

(iii) The GISS model was tuned to a climatological data set of unknown length. Such a procedure is a currently acceptable way to enhance model performance. But it also means that the tuning period may not be useful in detection studies (although it was used here!). Thus if the model were tuned to the mean SST, air temperature, etc. for the period 1950-1980, say, then one could begin GHG signal detection only for the post-1980 period. Presumably any GHG signal present in the 1950-80 period would affect the mean and hence be partially accounted for in the tuning.

(iv) Suppose the (D,M) sets had been termed “identical” by the PPP. This does not mean the GHG effect is responsible for the result. There may be other mechanisms that could cause a similar signal. This is the well-known problem of “attribution,” a problem that will not be easy to solve.

6. CONCLUSIONS

A particular GHG detection methodology has been applied to the transient GHG simulations from the GISS atmospheric GCM/ocean model. The model-predicted signals were not found in the observations of global SST. In making this statement, it is important to realize that the detection strategy was not “all encompassing” and was subject to shortcomings. But the main purpose of this study has been to demonstrate some of the issues and problems that will attend any attempt to attribute climate change to GHG, or any other, forcing. Hopefully the problems noted in the text will give those anxious to confirm or deny a GHG effect pause for thought.

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2 If the magnitude of the “flux correction” term is small relative to the GHG forcing, then the “tuned” portion of the simulation may still be useful for detection purposes. In the present case we have no way of determining if this constraint is met.


Part 6:

Working Group Reports
1. THE NEED FOR OBSERVATIONS

With the recognition that climate changes are inevitable as a result of the observed increase in greenhouse gases in the atmosphere, there is a great need for better observational data bases and analyses to document the past and current climate, and how it has changed with time. The need is for both regional and global evidence, to the extent that it exists, for many climatological variables.

The available variables for documenting change associated with the increasing greenhouses gases, roughly in order of importance appear to be surface air temperature, sea

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1 In the sense that reflects both the availability of useful data and the signal from models associated with increased carbon dioxide compared with the noise of natural variations.
surface temperature, temperature throughout the free atmosphere, moisture content of the
atmosphere, precipitation and other hydrological cycle parameters, sea level pressure,
wind, sea ice and glacier extent, snow cover, soil moisture, subsurface ocean tempera-
tures and sea level. Past information on several of these variables is quite sketchy. The
instrumental record is limited, with upper air observations only after about World War
II, and there are very few surface observations outside of Europe prior to about 1880.
For the more distant past it is necessary to rely on proxy data of various kinds such as
historical documents, tree ring widths, times of harvest, phenological records, last or first
snow falls, freeze-up or ice breakup in harbors and glacier extent. Most of the important
variables are discussed in more detail in Section 3. In addition, it is desirable to obtain
more reliable and comprehensive observations of quantities that have generally not been
available in the past. Examples include cloud cover, subsurface ocean temperatures, soil
moisture, biomass, sea ice thickness, and satellite-derived quantities, especially the Earth
radiation budget components.

While there are numerous problems with all the observational data sets, certain
characteristics of the climate changes have clearly emerged and a near-global record
exists of surface temperatures for the twentieth century. Average near-global surface
air temperatures appear to have risen over the past century by as much as half a degree
Celsius, although the warming has been far from uniform in either time or space. Because
of the natural variations in temperatures evident in the climate record, it is not possible to
attribute this warming to the observed increases in greenhouse gases, although the latter
is likely to have been a factor.

The generic uncertainties in the record result from changes in instrumentation, ex-
posure, and measurement techniques, changes in station location and observation times,
variations in the spatial distribution and areal coverage of the observations, and changes
in analysis methods; the latter are discussed in Section 2. Most observations taken in the
past that are now used to reconstruct the climate record were taken for weather forecasting
purposes. It is vital in the context of Global Change to ensure that future observations are
not beset with the same problems that have been identified with the past data. Suggestions
on how to redress these issues are given in Section 4.

In addition to documenting change, observational data sets are invaluable for val-
idating models, developing understanding through empirical studies of climate, and ex-
amining processes or mechanisms. Analysis of individual observations into global fields
for climate purposes has often used monthly mean data, but there is also a need for daily
data so that the time series character of the record can be examined.

Increasingly, climatologies of variables must include not only the spatial distributions
of means, but also variances and covariances or correlations, frequency or probability
distributions and extremes. In addition, time series of variables are needed along with
their spectral character, autocorrelations and trends. Wherever complete fields are not
available, it is desirable to devise simplified, reliable indices for monitoring. In either
case, it is unlikely that the full global distribution of variables can be maintained without
some inhomogeneities and, for reference purposes, it is therefore essential to designate a
more stable, homogeneous and reliable subset of stations as baseline or reference stations,
and to ensure they are maintained.

2. ISSUES IN ESTABLISHING GLOBAL CLIMATE TRENDS

Examination of the observational data base reveals the following problems: (1) changes
in instrumentation, exposure and measurement techniques; (2) changes in station location,
or station discontinuance; (3) changes in observation times and methods of processing
individual observations into time (e.g., monthly) means; (4) changes in the spatial distribution and areal coverage of observations; and (5) variations in analysis methods and ways of estimating areal averages, both regional and global. All of these can lead to inhomogeneities in the record. Other issues arise in the interpretation of the results and the extent to which they represent the climate record. Thus: (6) changes in the station environment and effects of urbanization should be identified; and (7) other non-climatic changes can also be important in certain parameters.

Changes in the time of day of the observations, which are common, can result in the diurnal cycle being aliased onto long-term trends. Similarly, if there are missing data the annual cycle can be aliased onto interannual variability and long-term trends. This is especially an issue in the stratosphere since premature balloon bursts mean missing soundings are not uncommon. The difference in a monthly mean computed for a few days at the beginning of the month versus the end of the month can lead to large apparent "anomalies" in the monthly mean if it is simply computed as the average of the available data without proper regard for the nonstationary features in the records. Another example is a measurement near sea ice which ceases once ice cover prevents ships from entering the area, thus leading to biased sampling.

Spatial coverage is an issue for all parameters whenever global averages are required. There is no such thing as true global coverage, even today. Vast areas of the southern oceans, especially south of about 40°S, are inadequately observed by conventional means to provide reliable analyses of any quantity. Satellite estimates of sea surface temperatures are becoming available globally, but their absolute accuracy is uncertain; substantial corrections are found to be necessary where calibrated buoy observations are available to provide ground truth. Satellite observations, however, do provide useful information about patterns and gradients of sea surface temperatures. Over land, fairly complete observations now exist, although their continuity back in time is an issue. Over Antarctica useful climatological records are virtually nonexistent prior to about the year of the IGY, 1957, except at Orcadas. Similarly, much of the Greenland Ice Cap is not monitored. Elsewhere, observations over about 70% of the globe are reasonably complete after 1950, but prior to World War II the usable network is diminished. Less than 50% of the globe is covered prior to 1900. Some studies have suggested that global coverage is not essential because there are limited numbers of spatial degrees of freedom. Because of the absence of observations, it is not possible to assess whether the southern oceans, for instance, could be changing in some way different from the rest of the world. Standard errors of the mean and other estimates of uncertainty are desirable, but the few that do exist are greatly dependent on the assumptions made in computing them. The overall goal of having a useful measure of the uncertainty or representativeness of the results has not been achieved.

The above comments apply to all parameters and analyses to some degree, although certain aspects become more critical for some parameters, as discussed in the next section.

3. CLIMATE VARIABLES

3.1. Surface Air Temperature Over Land

Several compilations of surface air temperature variations for the land areas of the two hemispheres are available. All agree that, in global terms, a warming of 0.4 to 0.5°C has occurred since 1880. The character of the temperature rise, however, between the two hemispheres has differed. The Southern Hemisphere shows a gradual rise during the twentieth century while the Northern Hemisphere record is characterized by two
warming periods, from 1910 to 1940 and from 1975 to 1988, with cooling between. Both hemispheric records are averages and as such they are not representative of any region, country or continent.

Consensus among the data compilations, however, does not guarantee that they are correct because they are not independent. All analyses use the same basic set of station data, many of which are affected by a variety of problems. These include the problems outlined in Section 2. Despite careful analysis to minimize many of these problems, uncertainties remain about the magnitude of the implied temperature change that are significant relative to the overall trend.

Air temperature measurements, expressed as anomalies from a climatological mean, tend to be more coherent spatially than those of precipitation and can often be adjusted for changes in measurement techniques and site locations, although that is best achieved if there is an overlap period of the two measurements in parallel. Resolving some of these uncertainties may be possible by a thorough re-analysis of the station data on an international basis.

Several of the potential problems with the temperature data base are unlikely to seriously affect estimates of hemispheric-average temperatures because of the cancelling effects of random errors. The effects of changes in station environments, particularly with respect to urban growth, are unlikely to be so benign. For instance, there were many shifts of measurements from city centers to airports in the 1950s. But the process of building a city around the measurement site, and the evolution of the airport site from a location on the outskirts of town to one surrounded by urban activity and increased jet aircraft frequency, which produce urban heat island effects, are pervasive. The effect of urbanization over the United States has been assessed by comparing averages based on the dense Historical Climate Network data set with the sparser density of stations that is internationally exchanged. These studies have estimated the urbanization influence on United States temperatures of about 0.1°C, or sometimes more depending on the data set, during the present century. In order to generalize these results to the rest of the world, studies in other large countries such as the Soviet Union, Peoples' Republic of China, India, Canada, Australia and Argentina are much needed. Other systematic biases arise from changes in times of observations and in instrument shelters.

3.2. Marine Temperature

Marine sea surface and air temperatures, like the land temperatures, reveal long-term trends. Recent decades appear to have been somewhat warmer than any other time in the past 120 years or so, and the period from about 1890 to the 1920s was distinctly colder than average. After 1910 these fluctuations are quite similar to those found from the land data alone. Although uncertain, temperatures from about 1860 to 1890 may have been higher than those in the subsequent 40 years, but still not as high as in the post World War II era. The decadal variations range over about half a degree Celsius.

Most of the marine observations have been taken by ships of opportunity. Some of the computer-readable data date back to the middle of the last century, but over three-quarters have been taken since World War II. The ships of opportunity follow preferred routes and consequently leave vast areas of the oceans inadequately sampled. Over parts of the southern oceans, coverage was better during the last century, prior to the opening of the Suez and Panama Canals.

For sea surface temperatures there have been many changes in measurement techniques over the years, with several different kinds of buckets used to bring up the water and different (and often unknown) instructions for exposure of the thermometer to the water before registering the reading. The greatest inhomogeneity, however, arises from
the transition to measurements in the engine intake. More recently, the addition of satellite measurements has provided further questions about some results. Attempts have been made to derive adjustments to the SST record based on an assumed but unknown distribution of bucket-versus-intake measurements, or by making use of apparent changes in the annual cycle. Other adjustments have used coastal stations as a control, thereby potentially spreading over the oceans effects such as those due to urbanization. Adjustments are unnecessary after World War II.

There are still good prospects for improving the data coverage in both time and space by digitizing the known marine observations available in manuscript form. It is estimated that about 15 million such observations exist in the U.S., several million in Europe, over a million in Japan, and about a million in Maury's collection for the last century. These data need to be evaluated and considered for digitizing. The U.S. and European observations have the potential for filling in the big data gaps during both World Wars. The Japanese data cover the North Pacific from 1911 to 1935, a time and place only sparsely covered by currently available data.

The main two global marine data sets that presently exist, COADS (Comprehensive Ocean-Atmosphere Data Set) and the U.K. Met. Office collections, reveal differences that have not yet been resolved, and it is important to compare these and incorporate buoy, expendable bathythermograph and other data, and thereby gain greater confidence in the observations. In addition to sea surface and marine air temperatures, these data sets provide pressure and wind data which may be useful for circulation studies.

3.3. Precipitation

Reliable precipitation climatologies are not available over the oceans because, even where rare measurements from ships do exist, they are unreliable. Over land, the climatologies are undoubtedly useful, but still subject to inhomogeneities, and it is difficult to obtain good estimates of changes with time. The records that do exist exhibit strong variability on all time scales. Interannual variations, often associated with the El Niño-Southern Oscillation phenomenon, are strong, and there are also large decadal-scale variations that can be verified by such phenomena as changes in lake levels. In the United States record high levels of Lake Michigan and the Great Salt Lake in 1985 provide evidence of the persistently above-normal rainfalls from about 1970 to 1985. Some lower-frequency variations and trends also appear in several records, but their reality is often less certain. One example of a strong and persistent reduction in rainfall is in the Sahel region of Africa, with the result that Lake Chad in the central Sahel has fallen substantially in recent decades.

For precipitation the large spatial variability and inherent small scales make coverage a paramount consideration. Moreover, precipitation is not a normally distributed quantity, so that analysis and averaging methods are important considerations. Even small changes in station location and changes in exposure, such as trees growing nearby, commonly corrupt rainfall records, but are generally not detectable unless detailed station-history records are kept. Adjustment of rainfall records for inhomogeneities, using an adjacent nearby station comparison, requires a much denser network of stations than for temperature because of the small-scale spatial variability. The latter also means that many more stations are needed to adequately monitor rainfall than for temperature, especially in complex terrain areas.

In addition, changes in instrumentation through alterations in rain gauge design to address such things as the effects of wind, wetting losses, and evaporation that affect the collection efficiency and measurement of rain, and especially snow, introduce biases with time for which adjustments must be made. National practices for converting frozen
precipitation into liquid equivalent vary. In the past, discontinuities in isohyets across the borders of adjacent countries have often been noted. Corrections of the pervasive inhomogeneities in rainfall records with complete station records is essential if confidence is to be gained that rainfall is changing with time.

3.4. **Circulation**

Little attention has been devoted to integrating results of all quantities into a consistent picture. The atmospheric circulation forms a link among the changes in winds, precipitation and temperature. Although this is not a strong constraint on any parameter, internal consistency can add confidence to the results. Thus, analysis of the pressure field and the implied changes in northerly or southerly winds implies advective changes in temperatures and moisture flow. Changes in pressures can also indicate changes in cyclonic activity and, perhaps, give inferences about precipitation. But possibly changes in the variance of pressures, which are related more to storm tracks, may potentially be even more useful for checking consistency with precipitation. Above all, it will be necessary to reconstruct the circulation changes in order to improve our chances for understanding the reasons for any change.

Historical sea level pressure analyses north of 20°N are available from 1873, and regionally in Europe from about 1780, although prior to about 1924 they clearly suffer from some problems. Variations in pressure patterns occur on all time scales, and multi-decadal time-scale anomalies are not readily apparent. An exception perhaps is a marked change in circulation in the North Atlantic. The more westerly flow from about 1900 to the 1930s led to milder conditions in northern Europe and to an absence of very cold outbreaks. Other changes in the Icelandic Low around the 1920s are reflected as a fairly abrupt warming of the polar regions which is also seen in temperatures for the entire Northern Hemisphere.

3.5. **Upper Air Measurements**

Examination of the signal-to-noise ratio and pattern similitude in model experiments involving increased carbon dioxide concentrations indicate that tropospheric humidity and column air temperature are important quantities to monitor. Cooling of the stratosphere is expected to accompany the greenhouse warming of the troposphere, and tropospheric specific humidity is expected to increase by as much as one-third with a doubling of CO₂. Radiosonde observations, which provide data, have the advantage that they are relatively free of urban heating and other local problems. However, there have been changes in type and design of radiosondes, and there is an absence of a reliable station-by-station record of the changes with time. In particular, for humidity a complicating factor has been the use of poor types of sensors and changes in the type of humidity sensing element and, more important, changes in reporting procedures at low humidities. The latter mainly affects the upper troposphere values.

The upper air data do not extend back in time nearly as far as the surface data. Some upper air data exist from the 1930s onward, but they are limited in areal extent and of somewhat questionable quality. Nevertheless, it is desirable to digitize as much of these early data as possible. At the very least, digitization would provide some data for the period of the U.S. drought of the 1930s, data that are not now readily available.

A 63-station network of radiosondes has been used to monitor layer-mean temperature since 1958. Although sparse, the network appears to give reasonable estimates of temperature changes. The network shows the period from 1964 to 1976 to be cooler and the past decade to be warmer, than the long-term mean, by a few tenths of a degree C,
and there is a distinct recent cooling of the lower stratosphere. It is desirable to expand
this network in both space and time to compare with other temperature records.

Water vapor can be examined reliably only from radiosonde data below about 500mb
because humidity sensors are not reliable at low temperatures. After neglecting the water
vapor above 500 mb, which accounts for less than 10% of the total, a reasonable estimate
of the changes of water vapor should be obtainable from radiosondes. To construct a
reliable record since 1958 will require knowledge of station and instrument changes from
the countries conducting the observations. Although this information will be difficult to
obtain in many cases, the task should be undertaken.

A new and potentially valuable source of climate variations in the free atmosphere
is the twice-daily global operational analyses which provide complete and dynamically
consistent fields of temperature, humidity and wind on constant-pressure levels. These
have the great advantage of complete spatial and temporal coverage, although the values
assigned in data sparse areas depend greatly on the veracity of the analysis system. Four-
dimensional data assimilation systems, originally developed for numerical weather pre-
diction purposes, provide a means of integrating all the information by carrying forward
the information from past observations, and by assimilating different kinds of observa-
tions (satellite and ground-based) and different variables through a multivariate analysis
in which the expected error characteristics of each kind of observation can, in principle,
be accounted for properly.

Numerical weather prediction models and four-dimensional data assimilation meth-
ods, however, are continually evolving and being improved, and any such changes can
have a major impact on the climate record implied by the results. Operational analyses
are therefore of limited use for establishing the climate record. Development of addi-
tional stages in this process to produce a more stable product is essential. If, or when,
changes are introduced into such a system, painstaking efforts are required to ensure that
the effects of such changes are known. An alternative is re-analysis of the entire set of
observations using a data assimilation and analysis system that is held constant.

3.6. Historical Observations

Whatever anthropogenic impacts on climate occur in the future, they will be superimposed
on natural climatic variability which may mask or amplify such impacts. From the limited
data currently available it is apparent that the climate for much of the 16th to 19th
centuries was quite different from that of the 20th century. It is therefore important that
efforts be made to extend the record of interannual climatic variations back over past
centuries, to provide a longer-term perspective on contemporary climate.

Because the spatial coverage of instrumental data declines markedly back into the
early 19th century, in addition to long instrumental data sets it is necessary to incorporate
proxy records of climate (climatically sensitive natural phenomena) into the mix of data
used. These include written historical records, dendroclimatic and phenological data, ice
core and sedimentary data, glacier extent and permafrost data.

For reliable and meaningful climatic reconstructions, such proxy data need care-
ful calibration with contemporary instrumental data, and attention should be focused on
improving the quality of such calibrations. No single proxy provides a globally exten-
sive data set; a synthesis of multiple proxies is needed to provide the necessary global
coverage.
3.7. Subsurface Ocean Data

As a major part of the Earth's climate system, the world ocean will play an important role in the response of the climate system to the increased production of greenhouse gases. For example, the ocean will almost certainly delay the onset of the expected atmospheric warming. Additional subsurface data, in particular, temperature and salinity, need to be acquired for the purpose of monitoring, as well as for the study of past variations. Substantial numbers (several hundred thousand) of temperature and/or salinity profiles exist in manuscript form which can be used to increase the size of the historical archive of subsurface ocean data in digital form. In addition, subsurface data exist in analog form (e.g., mechanical bathythermograph slides and expendable bathythermograph strip charts) at oceanographic institutions which could be digitized to increase the size of the historical data base. The improved temporal and geographical distributions of data that these additional observations provide will allow scientists to increase their knowledge of the temporal variability of temperature and salinity in the world ocean, and help to understand the physics responsible for these changes.

Present knowledge of temporal variability of the subsurface temperature/salinity structure of the world ocean indicates that gyre- and basin-wide-scale variabilities do occur. For example, in the North Atlantic a well-documented cold, fresh anomaly occurred in the surface layer of the subarctic gyre beginning in the late 1960s. Convection in this water mass was responsible for creating a relatively cold, fresh anomaly at intermediate depths (500 m) within this gyre. In the subtropical gyre of the North Atlantic, an upward displacement of isopycnals (25–50 m) occurred between the late 1900s and the early 1970s. This led to substantial changes in temperature (0.5°C) and salinity (0.025%) in the 500–1200 m depth range throughout the gyre. Monitoring of the temperature and salinity structure of the world ocean is important for detecting temporal variability due to natural and anthropogenic causes.

3.8. Sea Level

Although individual station records of sea level can display considerable variations over short distances, there is a consistent pattern overall indicating a rise in sea level since about 1880 of 1.0 to 1.5 mm/year. For sea level, the biggest issue is the interpretation of the results in terms of the climate record because of contamination due to isostatic rebound of the Earth's crust and tectonic activity. Nor is there a global network. Nevertheless, the observed increase is consistent with alpine glacier melting and thermal expansion of the upper layers of the ocean. Future expansion of the global network of stations is desirable.

3.9. Cryosphere

Future greenhouse-gas-induced warming effects are expected to be amplified in high latitudes. However, the observational records of hemispheric snow cover and sea ice extent span barely two decades. Decadal fluctuations of the order of 10 percent in both hemispheric snow cover and sea ice are apparent, but there are no evident trends. Continued monitoring via satellite all-weather (passive microwave) remote sensing is essential. Efforts should also be made to assemble station observations of snow depth from national archives. Sea ice concentration and thickness can be determined from satellite remote sensing (passive microwave and SAR) and upward-looking sonar. Research to improve these techniques and make them routine needs to be accelerated. Ice concentration estimates for the 1980s derived from passive microwave data need rechecking to establish
whether or not trends exist. Space-time averages of sea ice thickness need to be prepared from available sonar tracks in the Arctic; no such data have been collected in the Antarctic.

Changes in land ice volume are a major contributor to sea level changes. The general retreat of mountain glaciers over the last 100 years is thought to be represented in the observed 20th century sea level rise, but the mass balance records from mountain areas are sparse, and there are some recent signs that glacier melting may have ceased. Continued support for national monitoring programs at selected representative glaciers is essential; such programs are being discontinued or under threat in several countries. Limited field surveys in Antarctica and comparison of radar altimetry from Seasat and Geosat over southern Greenland indicate these ice sheets may be thickening (which would tend to lower sea level). Radar and laser altimetry surveys from ERS and EOS-type satellites will be crucial to monitoring such trends.

Ground temperature profiles in Alaskan permafrost regions show a widespread 2°–4°C integrated warming of the surface over the last 75–100 years. To the extent that these kinds of data add to our overall information, circumpolar data should be collected from well logs, or obtained by installing new ground temperature sensors at additional sites in Eurasia.

3.10. Clouds

One of the major weaknesses of present climate models is the way that clouds are treated. Climatologies of cloud type, height and amount, and the extent of cloud-radiation feedbacks, are uncertain. The radiative effects of changes in low cloudiness are markedly different from changes in high clouds. Thus, detailed observations of the 3-dimensional cloud fields are needed to establish cloud climatologies and to compare with clouds in models. The cloud fields derived from satellite data under ISCCP should meet many of these needs. Continued studies using cloud observations from land stations and ships are also required.

3.11. Radiation

Top-of-the-atmosphere radiation components have now been produced from several different satellites, although it is only with the Earth Radiation Budget Experiment that comprehensive fields which include the diurnal cycle can be produced. As well as providing climatologies of albedo, outgoing longwave radiation, absorbed solar radiation and the net budget, recent advances allow the effects of cloud-radiation forcing to be examined. Even within the fairly short available record, the radiation observations have captured the huge regional perturbations associated with the 1982-83 El Niño events. The need to continually monitor the top-of-the-atmosphere radiation budget is readily apparent.

In addition, progress is being made in obtaining the surface radiation budget. The latter is very important as part of the overall surface heat balance. Furthermore, the correct partitioning of the surface radiation into solar and longwave components is important because of their different effects on the ocean. Preliminary algorithms exist for computing the solar component, and research is progressing for the longwave component. Stable and reliable estimates of these quantities should prove invaluable for interpreting changes in climate.
3.12. Aerosols

It has been estimated that the effect of a major volcanic eruption is to lower the global mean temperature by about 0.2°C in the following year, and perhaps by as much as 0.1°C over the following few years. The effect varies with the amount of small particles and gases injected into the stratosphere, as well as with their chemical composition. Accordingly, extensive volcanism can complicate interpretation of the climate record. Improved measures, or perhaps indices of volcanic activity, based on historical information, ice core data, turbidity measurements, early Smithsonian data and aircraft and satellite measurements are needed. In the future more quantitative-estimates of aerosol loading and its vertical distribution and spectral characteristics are needed over a wide range of locations.

4. FUTURE CONSIDERATIONS AND RECOMMENDATIONS

In spite of the intense interest in climate trends and global change and the clear evidence that Man's activity is changing the concentrations of greenhouse gases in the atmosphere, there is no coherent national or international program to ensure that problems experienced with past data will not also occur in the future. Collection of new surface data is increasingly being automated so that, for instance, mercury-bulb thermometers are being replaced with thermisters. Stations are being closed for economic reasons and commercial companies are taking observations, but making a profit is of more concern than ensuring long-term quality control of the measurements. New sources of remotely sensed data, especially from satellites, are now available.

It is considered of the utmost importance that the relevant agencies of the world meteorological community devote the effort and resources necessary to ensure the proper maintenance of national observing networks. This means taking necessary action to identify the best and longest climate records, to make sure that they are continued in the future, and to ensure that the data are available promptly to the international meteorological community through the various WMO-sponsored international protocols. In the past, observations from pre-existing stations with long periods of record have been terminated under economy measures, or the instruments have been relocated or changes in observation methodology made with little regard to their impact on climate-trends monitoring efforts. Comprehensive station history files are needed to enable the user to apply any necessary corrections.

There are no global networks specifically maintained for monitoring surface climate data. Current surface climate data are almost exclusively derived through the utilization of weather observations from operational weather sites. The Global Telecommunications System (GTS) operated for the World Meteorological Organization is the primary source of monthly surface data (temperature, precipitation, surface pressure and hours of sunshine). Recently, the spatial coverage of these surface climate data, often sent as CLIMAT reports, has been decreasing. Responsibility for collecting and transmitting the monthly data reports, whether through the GTS or sent in manuscript form to the principal archiving center at Asheville, rests with the individual countries. To the extent that political instability and economic difficulties in various countries make the regular collection of weather data an onerous task, problems associated with incomplete data coverage will persist. Long delays in receiving decadal updates from some countries have been a particularly serious problem in attempts to produce the regular upgrades to the World Weather Records archival data. Delays of 10 to 20 years in publication of update summaries for some world regions are occurring. For example, the World Weather Records update for the decade of the 1970s (1971–80) has been published only for the European region (in 1987). Publication of regional updates for North America and for South America are
expected in 1989 and 1990, respectively. Publication dates for the updates for Africa and Asia are presently unknown.

It is desirable that the archival data centers should not only be responsible for routine data collection, but should also pursue a more active role in the prompt acquisition of data from other countries, and undertake additional efforts to digitize the large backlog of historical records that are in storage in national archives in the U.S. and abroad. Critical gaps exist in the National Climate Data Center (NCDC) operations, and there is a need to set priorities and establish a framework for correcting such problems and for obtaining the appropriate level of long-term financing necessary to accomplish these priorities.

Advances in technology and budget considerations have fostered the development of state-of-the-art automated surface-weather observation systems. These automated systems will be implemented by the U.S. National Weather Service over the next five years. It is imperative that the instruments in these systems be extensively compared and calibrated with conventional observations, and that accurate station histories be maintained. Otherwise there is a real risk of a diminished capability to monitor climate change.

Operational satellite data are an underutilized resource to monitor and understand the climate system and may have an untapped potential to provide a basis for recognizing patterns of climate change, such as those expected with increasing greenhouse gases. To be useful, satellite data must be calibrated with ground truth on a continuing basis. In addition, a comprehensive climate data management system must be part of all operational satellite products. Such a system would include the capture, formatting and archival of satellite-related and supporting in situ climate data; these should include the Earth radiation budget, precipitation estimates, sea surface temperature, surface and upper air temperatures, water vapor, snow cover and water equivalent, soil moisture, sea ice and a vegetation index.

4.1. Establishment of a Global Benchmark Climate Monitoring Network

Questions regarding the reality of reported trends in climate will continue to arise as long as significant problems associated with the reliability of the climate records are not satisfactorily addressed. One way to at least partially overcome such problems is to establish an appropriate worldwide climate monitoring network which will eliminate, as far as possible, many of the sources of non-climatic noise. Recent WMO programs to revitalize and improve the spatial coverage of climate-related data transmitted over the GTS should be vigorously pursued. A proposal by the ICSU Commission for Climatology (CCL) seeks to improve the amount and quality of internationally exchanged climate data so as to provide more reliable information on climate trends and variability. Each member of WMO would be asked to assess the homogeneity of their records currently in the global temperature data base, and to amend records where necessary and add additional ones to ensure a distribution of specified density toward a goal of 10 stations per 500 x 500 km square. This increase could be accomplished through the efficient utilization of systems such as CLICOM at existing stations, and through implementation of the CCL-proposed Climate Change Detection Project. Such a re-analysis would achieve two basic objectives. First, it would produce a data set of known reliability and, second, it would constitute a network of stations collected principally for climate studies. Although such a network will require a commitment of resources, it is worth noting that the cost of such a benchmark network would be far less than that involved in monitoring for short-term forecasting purposes, yet the potential economic benefits far outweigh its cost.

One means of improving data quality and volume is to disseminate the results of hemispheric and global analyses back to member countries. All the world's countries are aware of global warming, but many do not realize that it is their data that form the basis
of the global temperature data base. Improvement to the network may be achieved if
countries see the use to which their data are put.

4.2. Data Management

The main scientific need for data management is to collect and prepare the basic ob-
servations and derived products into well-structured data sets that can be used by the
scientific community. There is a related need to know where data are located and the
changes in data coverage with time. It is important to keep a reasonable balance between
data management activities and data acquisition, with the main focus on quality assess-
ment and control, archival, distribution and analysis, and without an undue emphasis on
data set catalogs, hardware systems, communications and networking. Directories of data
holdings are desirable, but experience shows that monolithic “top-down” systems do not
satisfy research needs and are often expensive. Scientists should play a key role in de-
veloping procedures for data validation, quality control and re-analysis, and in evaluating
the degree to which the data centers are satisfying programmatic needs. The scientific
community and funding agencies have a responsibility to contribute to the maintenance
of a viable data management system.

Projects such as data “rescue”, that is, the digitization of unique manuscripts or other
hard-copy archives, have not been implemented as a result of shortfalls of funding. Thus
it is not yet possible to meet all of the identified requirements for Global Change research
program.
1. INTRODUCTION

The complexity of the climate system and the absence of definitive analogs to the evolving climatic situation force use of theoretical models to project the future climatic influence of the relatively rapid and on-going increase in the atmospheric concentrations of CO₂ and other trace gases. A wide variety of climate models has been developed to look at particular aspects of the problem and to vary the mix of complexity and resource requirements needed to study various aspects of the problem; all such models have contributed to our insights of the problem.
The most comprehensive of the available climate models are known as General Circulation Models (GCMs). These models represent what are believed to be the most important components of the atmosphere, ocean and cryosphere system relevant to determining the potential for an enhanced greenhouse effect. That biospheric interactions are at present largely omitted may be important; for example, their seasonal changes and role in forming cloud condensation nuclei may affect planetary albedo. The GCMs are being constantly modified to include the most up-to-date extent of our understanding of atmosphere, ocean and geophysics from analytic, laboratory and field experiment studies, consistent with limitations in computer resources, numerical methods and human resources required to update, test, and verify the new algorithms in the models. A variety of tests indicate that the GCMs, although not yet providing adequate representations of the climate system, nonetheless provide the most plausible way for projecting future changes in the climate. A wide range of improvements will be necessary, however, to permit development of increased confidence in the detailed regional and seasonal simulations of the models.

2. GENERAL CHARACTER OF CLIMATE MODEL RESPONSES TO INCREASED CONCENTRATIONS OF GREENHOUSE GASES

Two types of studies are performed to study the potential consequences of the increasing concentrations of greenhouse gases. In the first, the gas concentration (usually CO$_2$) is arbitrarily and instantaneously increased (usually doubled) and the model is run until a new equilibrium climate is reached, the difference between the perturbed and control results being a measure of the sensitivity of the climate to the imposed perturbation. This type of calculation is intended primarily to determine the equilibrium commitment to future climate change. In the second type of simulation, the gas concentration is slowly increased as time advances, sometimes, but not always, in a manner representing either the observed change in gas concentration over some past period, or as projected for some future period. Although this type of simulation is intended to provide a more realistic representation of climate change, it requires that the model realistically represent oceanic behavior. Quite often the results from the equilibrium simulations, which are much easier to perform because the effects of the deep ocean are neglected, are interpolated to develop estimates of present climate change; there is little likelihood, however, that this is a valid assumption on scales below global.

The general character of results from GCM simulations of the potential climate effects from the continuing increases in emissions of CO$_2$ and other trace gases are given in Table 1. Although results differ somewhat from model to model (results from about 5 modelling groups are generally available), and these differences are certainly cause for deeper investigation, there are also many similarities in the results of the various models. These results indicate that the enhancement of the greenhouse effect will raise global average surface air temperatures by as much as a few degrees Celsius by the middle decades of the next century, assuming that the concentrations of CO$_2$ and other greenhouse gases continue to rise at significant rates and that other forces influencing the climate system do not act in the opposite direction. The increasingly comprehensive calculations through the 1980s remain consistent with assessments and evaluations made over the last decade (e.g., NRC, 1982; Bolin et al., 1986) suggesting a temperature sensitivity to a doubling of the CO$_2$ concentration generally within the range of about 1 to 5°C. Such a climate response will make the change induced by CO$_2$ and other trace gases over the next several decades larger than changes that may be induced by other natural factors (solar, volcanoes, etc.) over time scales of a century and longer. The warming is expected to be somewhat larger in high latitudes during the transition season as the warm
snow-free and ice-free period lengthens and the cold winter period is shortened; in lower latitudes, the warming may be somewhat less than the global average.

Precipitation patterns are also expected to shift somewhat in season and location. The only distinctive change found so far in all models, however, is that high latitudes seem likely to receive increased precipitation. Warmer temperatures, coupled with modest precipitation changes, will increase evapotranspiration, especially over mid-latitude continental regions during the warm season, suggesting increased drying, soil moisture limitations, and water resource stresses.

Atmospheric warming will lead to increasing heat uptake by the oceans and enhanced melting of mountain glaciers, both of which will contribute to sea level rise. The actual rise in sea level will be determined by changes in the mass of polar ice sheets, which may experience either increased accumulation of ice as the very cold temperatures warm (but remain below freezing) or increased deterioration as decay of ice shelves allows an increased rate of glacial flow and calving. The potential exists for a diminution to significant enhancement of the rate of 5 to 20 cm/century rise in sea level experienced over the past century, such that sea level changes over the next century could range from slightly negative to increases of as much as a meter or more.

3. FACTORS CONTRIBUTING TO CONFIDENCE IN SIMULATION RESULTS

In their implementation of the fundamental conservation laws, climate models incorporate findings derived from analytical and experimental studies, but must also include a sizeable sprinkling of empirically based algorithms to try to overcome simplifying assumptions resulting from limitations in understanding and computer resources. That the models are largely theoretical necessitates that they be evaluated and tested as extensively as possible to provide a basis for developing confidence in their results. Such tests are being widely attempted and many of the tests provide important indications that the models can reasonably represent the large-scale, long-term aspects of the greenhouse effect and potential future changes of climate.

1. Climate models based on well-known laws governing the physics of radiation and convection have successfully simulated the relative intensities of the greenhouse effect on Venus, Mars, and the Earth—three planets with vastly different atmospheric compositions and surface temperatures. These models are also able to roughly explain the relative constancy of Earth's temperature throughout its geological history as solar intensity has increased and atmospheric composition has changed. More comprehensive models are now also able to simulate many of the differing aspects of circulation systems of these and other even more different planets, lending confidence that we understand the basic physics governing the climate system.

2. Algorithms representing specific processes in climate models can and are being tested against results from field and experimental studies. For example, the infrared radiative transfer schemes in climate models, which provide the initial forcing for climate change, have been intercompared among themselves, with more detailed radiative calculations, and with laboratory observations. The results of these intercomparisons indicate that the radiative fluxes calculated by current climate models are generally within several percent when compared to observations measured under a variety of atmospheric conditions.
Table 1. General character of results of model simulations by geographical domain

Global Domain

**Highly Likely**
- Equilibrium temperature sensitivity of 1 to 5°C for a CO₂ doubling (or equivalent).
- Acceleration of the rate of climatic warming with continuing increases in emissions.
- Increased rates of evaporation and precipitation.
- Stratospheric cooling.
- Equilibrium warming will be greatest in high latitudes and lowest in the tropics.

**Likely**
- Warming of land areas will lead warming of ocean areas.

**Possible**
- Increased intensity of precipitation.

High Latitudes

**Highly Likely**
- Warming initially greatest in late fall/winter and spring; as greenhouse gas concentrations continue to increase, warming will become greatest in mid-winter.
- Warming least in summer.
- Precipitation increase throughout the year that would increase soil moisture and runoff.

**Likely**
- Retreat of sea ice.

**Possible**
- Delay of warming as a result of ocean circulation changes.

Middle Latitudes

**Highly Likely**
- Increased summertime evaporation accompanying warming.

**Likely**
- Precipitation increases in winter and spring leading to increased soil moisture in spring.

**Possible**
- Soil moisture decreases in summer.

Low Latitudes

**Highly Likely**
- Temperature increase less than global average increase.

**Likely**
- Precipitation changes in the dry subtropics will generally be small with broad areas of decreased precipitation.
- Increased zonal mean precipitation in the tropics, but not uniformly distributed.
- Relatively uniform increase in temperature throughout the year.

**Possible**
- Increased frequency and/or intensity of tropical cyclones.
3. The models successfully simulate many features of the present climate. When used in a weather forecast mode, models essentially identical to the climate models represent the evolution of atmospheric dynamics, precipitation zones, storm development, and many other important features. The models also successfully simulate the latitudinal, seasonal, and land-ocean distributions and changes of climate. A second aspect of such testing has involved simulations in which the atmospheric (or oceanic) part of the model is forced by observed oceanic (or atmospheric) conditions. In such tests the freely varying component of the model is able to quite accurately emulate observed behavior.

4. Climate models have been used successfully to simulate a number of past climate changes, as well as important features of the present observed climate and interannual variability. Most notable among these are:
   a. the ability to simulate inferred large-scale changes in precipitation and soil moisture patterns in response to large-scale changes in seasonal radiative forcing over the past 18,000 years.
   b. the simulations of various types of climate models that do not overestimate the few-year, low-amplitude coolings in response to large, short-term volcanic perturbations.
   c. the ability of some coupled atmosphere-ocean general circulation models to simulate some features of El-Niño oscillations.

5. The first-order, large-scale estimates of the climatic response to a CO₂ doubling are consistent with our intuitive understanding of how the climate system operates. For example, the amplification of equilibrium warming at high latitudes is fully consistent with well-known processes involving ice and snow feedbacks and the surface inversion. The same can be said for the poleward shift in the midlatitude precipitation maximum associated with baroclinic processes, the overall intensification of the hydrological cycle, and the tendency for summer evaporation from soils to increase. The fact that the most important features of the model responses can be understood intuitively is not trivial, in that the models are highly complex and characterized by innumerable nonlinear couplings among the various model components.

The success of the models in dealing with this wide variety of situations gives us reasonably high confidence in some important aspects of model performance, particularly the short-term (< few years) and very long-term (millennial) physics. Unfortunately, however, there are few good tests on the decadal-to-century time scales for which the greenhouse gas changes are projected as the dominant influence.

4. FACTORS LIMITING CONFIDENCE IN MODEL RESULTS

Although many aspects of model behavior lead to increasing confidence, neither our understanding of atmospheric physics nor the model representations of our understanding are perfect. These limitations contribute to uncertainty and reduced confidence in model estimates of climate sensitivity and future climate change.

1. The representations of many important processes are absent or highly simplified in ways that may affect model sensitivity (whether positively or negatively is not known). These limitations arise due to both limitations in understanding and in resources (computer, staffing, etc.).
a. Resolution: With typical horizontal resolution of hundreds of kilometers, climate models are unable to resolve many aspects of atmospheric and oceanic processes, surface-atmosphere interactions, the intensity of storm and frontal activity (e.g., hurricanes), and the resulting climate changes.

b. Cloud-Radiation Interactions: With clouds having dimensions typically less than tens of kilometers and complex vertical structures, algorithms for estimating cloud cover, and hence radiative interactions, are necessarily highly simplified. Approximations are quite variable among models and have not been adequately validated against observations. Variations in cloud liquid water content and in cloud optical properties with climate change are only starting to be represented. Simulations using different cloud algorithms dramatically affect estimated sensitivity to doubled carbon dioxide and other perturbations.

c. Clouds and Convection: Convective algorithms are generally highly simplified and the ability to loft and condense water vapor is poorly tested. In that water vapor feedback is of great importance and that convection dries the atmosphere, even small shortcomings in the algorithm could have large consequences.

d. Surface Hydrology: The simple 15 cm buckets most often used to estimate soil moisture conditions do not allow for the role of subsurface recharge, evapotranspiration, soil crusting, etc.

e. Surface-air fluxes and the planetary boundary layer: Estimates of fluxes assume simple homogeneous terrain and do not adequately resolve the structure of the planetary boundary layer (PBL) or the transfer of moisture upward from the PBL to the free atmosphere.

f. Oceans: Models are only beginning to represent oceans with a variable depth mixed layer, interactive ocean currents, and couplings to the deep ocean.

g. Sea ice: Climate models do not yet represent sea ice transport, ridging, and other features strongly influencing extent and thickness. These models also do not represent brine rejection and its role in bottom water formation, ocean stratification, etc.

h. Chemistry: No climate models yet interactively represent ozone and related chemistry and few represent the full range of radiatively active gases and aerosols.

i. Biosphere: Interactive coupling of the biosphere into climate models is in the early stages; thus, changes in plant-related albedo, evapotranspiration and other properties with season and region are not adequately represented (if at all).

2. Models have generally been tuned in various ways to achieve improved agreement with the observed climate. Examples of adjustments include selection of cloud albedo and transmissivity, adjustment of the solar constant, setting of oceanic mixed layer depth, arbitrary determination and incorporation of oceanic fluxes, adjustment of precipitation fluxes to correct salinity, adjustments of the top-of-the-atmosphere radiative flux to account for kinetic energy dissipation and/or to achieve global energy conservation, etc.

3. Model verification against the present climate has been limited in part by limitations in observational data sets. Particularly for summer conditions over continental and smaller scales, verification of representations of the mean climate is quite poor. The models are also not adequately representing climatic variability (interannual, quasi-biennial oscillation, etc.)

4. Model simulations have generally had to be made with simplified oceans and instantaneously increased greenhouse gas concentrations rather than to perform the proper
transient calculations. Only recently have simulations with slowly increasing CO₂ concentrations shown that sensitivities derived from equilibrium calculations do not match those from transient calculations, either while the climate is changing (when differences in land and ocean heat capacities seem certain to affect the results) or perhaps in their final equilibrium state (or states). There is also no assurance that climate will change slowly and gradually (as reliance on interpolation of equilibrium results assumes) rather than in fits and jumps, nor that any equilibrium state of the ocean-atmosphere system is unique.

5. Neither data nor adequate understanding exist for models to represent the simultaneous presence of non-greenhouse forcing factors, including solar variations, aerosol injections and oceanic circulation changes.

6. Aspects of model simulations of CO₂-induced perturbations are in apparent conflict with observations.

   a. If we assume that greenhouse gases are the only cause of climate change for the past century, then the present results of model simulations (as interpolated from equilibrium calculations) suggest that a larger and more rapid warming than has been observed should have occurred (by about a factor of 2). The observations also show irregular variations, such as the Northern Hemisphere cooling from about 1940 to the 1970s and the changes in interhemispheric temperature anomalies, which are not shown by models with simple mixed-layer oceans. If, however, we assume that a substantial amount of this observed climate change is due to other causes, with changes in ocean circulation and in volcanic aerosol loading being the main contributors and perhaps solar variation, tropospheric aerosols, natural variability, ENSO variations and unremoved urban warming also contributing, then the models and observations may not be in conflict. Further, work to better determine the other components of climate change is required to assess the importance of these apparent conflicts.

   b. Analysis of climate changes for warm and cold periods over the past 100,000,000 years suggests a high degree of stability of tropical temperatures, whereas climate model estimates of CO₂-induced changes suggest a relatively high sensitivity.

The combined effects of these many limitations and shortcomings in the models suggest that: (1) the uncertainty implicit in the 1–5°C consensus estimate of the sensitivity of the climate to doubled carbon dioxide should not now be narrowed, and (2) there is not yet an adequate estimate of the magnitude and pattern of climate change that we should be expecting to see as the atmospheric composition is slowly changing in the simultaneous presence of changes in greenhouse gases and other forcing factors. Increased confidence in model results requires addressing all of the enumerated shortcomings.

5. PATHS AND OPPORTUNITIES FOR ENHANCING UNDERSTANDING AND CONFIDENCE

Research over the past several decades has significantly increased our understanding of the factors controlling the global climate. There remain an interconnected set of paths and opportunities for further progress in overcoming shortcomings and limitations in our understanding and in the model simulations.

1. Detailed studies and field experiments are needed to improve the level of understanding of individual processes, including clouds, oceans, hydrology, surface-atmospheric interactions, radiation and others.
2. Enhancement of the assembly of observational data sets is essential in order to significantly advance model verification activities. This will require support of networks and satellite systems and acquisition, assembly, and analysis of the data.

3. Detailed and systematic diagnosis and intercomparison of climate models, both of their individual processes and coupled behavior, can identify strengths and shortcomings requiring improvement.

4. The representations of atmospheric and surface processes must be significantly improved and better verified to assure adequate representation of the many complexities evident in observations of system behavior. Increased use of improvements made in weather forecast models should be made.

5. Expanding the domain of climate models to more fully include oceanic, cryospheric, chemical, and biospheric processes and interactions is essential in order to simulate recent climate variations.

6. Pursuit of a hierarchy of climate models, ranging from conceptually simple phenomenological models to coupled ocean-atmosphere general circulation models, will continue to provide opportunities for identifying, interpreting and understanding the roles of mechanisms and feedbacks in amplifying and moderating climate change.

7. Climate models must be verified against a wider range of temporal and spatial conditions, including diurnal, synoptic (taking more advantage of coupling to the numerical weather prediction community), interannual (including ENSO, QBO, and other low-frequency variations), and decadal (e.g., 1980s vs. 1970s) periods, and for various external (e.g., volcanic, solar and oceanic) forcing factors.

8. The set of model sensitivity tests should be expanded, seeking to better understand the potential for natural climatic variations and changes, for example, in the North Atlantic Ocean circulation or by the locking in of atmospheric circulation patterns.

9. The increasing speed of computers offers the potential for more completely representing the climate-system domain, improving model physics, lengthening and broadening the set of model simulations, and many other advances. Significantly expanded computer resources are a vital need for improving understanding.

10. The level of effort to understand global climate change has been rather modest, considering the importance of the issue (e.g., each of the several CO₂ modeling groups is composed of only a few scientists). A significant increase in human resources is needed at all levels, from the senior research scientist to the upcoming graduate student.

11. Closer collaboration, coordination and communication among the various modeling groups can aid in more efficient allocation of available resources to issues and problems, although it must be recognized that it is differences among groups pursuing similar questions that have been an important driver of progress.

12. Development of a more coordinated modeling strategy is needed, including establishment of standard test simulations (e.g., of CO₂ and trace-gas-increase scenarios), verification tests, analyses, and algorithm testing in order to facilitate intercomparison and diagnosis of model results.

These various activities represent a combined and interacting program for progress. Although these efforts will require a sustained commitment of increased financial resources, the potential for significant and steady progress over the next decade and beyond is substantial.
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1. SUMMARY AND OVERVIEW

Quantitative efforts to detect the greenhouse-gas signal (GHG) in nature are in their infancy. The reasons for this state of affairs are numerous. It is only in the last few years that GCMs have advanced to the point where their simulations of GHG signals might be marginally believable. Without reasonably good a priori predictions of expected GHG signals from the models, the detection problem is moot. The observational data sets describing changes in the global climate system over the last 50-100 years needed for adequate detection studies have also only come into existence in the last five years. Finally, no coherent, generally-agreed-on detection strategy has been developed by the scientific community interested in the GHG problem. The lack of adequate model predictions and observational sets are largely responsible for this latter condition.

The rudimentary detection efforts that have been conducted have generally been based on recognizing the “fingerprint” of GHG signals in the oceans and atmosphere. GCM results for 1x and 2xCO₂ “equilibrium” runs have been used to search for GHG effects induced in tropospheric air and ocean surface temperature fields since the early 1900s. No significant effect has been found. But given the huge assumptions attending these studies, the results should not be taken seriously; they only demonstrated technique.
More recent attempts at detecting the GHG signals predicted by "transient" runs also failed to find the model-predicted signals. These recent studies, though much more powerful and improved over previous work, still have serious flaws and represent only the most meager start in the detection arena.

2. STATE OF THE RESEARCH

2.1. Overview

Because of the strong theoretical basis for greenhouse warming, scientists are concerned about the potential climatic effects that are likely to result from the current observed increases in CO$_2$ and other atmospheric gases. However, considering the many very significant uncertainties and inadequacies of both historical data and current computer models, there is a very strong reluctance for scientists working in this field to make the definitive statement: "Yes, we have now seen a greenhouse effect." In the data area, the limited number of variables available, the poor and changing quality of their observations, their short length of time, and the innate spatial and temporal variability of climate all contribute to this reluctance. Additionally, although GCMs have improved markedly over the past few decades, modellers keenly realize their many deficiencies. Particularly important to the GHG problem are the primitive treatment of oceans, poor parameterization of clouds and an overly simplistic view of the hydrological cycle.

2.2. Meeting Summary

Over the past century, mean Northern Hemisphere land temperatures have risen about 0.5° C. However, the limited data prior to 1881, the year from which hemispheric data were first systematically collected, indicate a cooling in that early period. This part of the observed warming may represent a natural recovery from the earlier cold epoch. It also appears that about 0.1° C of the 0.5° C warming indicated by the data set may be due to urban effects, that is, it is fictitious.

Southern Hemisphere land and ship observations have recently been collected and analyzed, and contrary to expectations, have indicated a more rapid warming than the Northern Hemisphere land data. They also show other unexpected and still unexplained variations, in particular, higher temperatures than now before 1880 and an abrupt cooling near 1903. However, due to their sparseness it is likely that these data are less reliable than the Northern Hemisphere land data in indicating past climate change.

The observed warming has not been steady in time nor has warming over this period occurred at every location for which observations are available. A major fraction of the Northern Hemisphere warming occurred in abrupt jumps around 1920 and again between 1976 and 1981 after a prolonged period of cooling between 1938 and 1964. While the warming up to 1938 was smallest in the low latitudes and greatest in the Arctic, the most recent phase of warming has been just the opposite: Strongest in low latitudes and accompanied by actual cooling in both the North Pacific and the North Atlantic. These recent observations are in general disagreement with all GCM predictions of climate change associated with enhanced GHG.

Our best climate models predict a mean global warming of 2.8 to 5.2° C for an equivalent doubling of CO$_2$. For the approximately 40% increase in equivalent GHG concentrations in the atmosphere since the 1880s, there should have been an equivalent warming of 1 to 2° C. Most scientists who have addressed this apparent discrepancy between models and observations have concluded that the large thermal inertia of the
oceans is delaying the appearance of the warming and that perhaps less than half of the equivalent warming can be expected to have manifested itself to date. An even more difficult part of the problem of trying to identify the observed warming with that predicted for increased GHGs is the differences in the latitudinal and vertical variations in the spatial patterns of the observed and predicted warmings.

In short, if one considers not "global means" but the actual change in air and ocean temperatures in time and/or space, then the model-predicted GHG signal cannot reliably be found. This result was amply demonstrated by the several speakers at the conference who described quantitative attempts to detect model predicted changes in the real world. Accordingly, it is the consensus of this meeting that the climatic warming of the past century cannot be identified as the warming predicted by our climate models as a result of the known increases in GHGs. But neither can we rule out the possibility that the models are wrong and some of the apparent warming is due to GHG effects. This is an unsatisfactory state of affairs. Due to uncertainties in the historical data, the natural variability of the climate, and the primitive nature of model transient runs, it seems unlikely that we will be able to detect the GHG signal with confidence during the next decade.

3. THE ELEMENTS OF A QUANTITATIVE DETECTION STRATEGY

In this section, we outline and briefly comment on what seems to us the key issues of a GHG-detection strategy. The discussion includes possible solutions in each area, but these are largely for illustrative purposes.

3.1. Selection of Variables for Detection Schemes

In this section we set out a strategy for selecting the variables that will be used in a GHG-detection program. This is necessary because the numerical models (GCMs) that are used to predict GHG effects offer numerous climate variables and physical quantities with which to attempt a GHG-signal detection. This is contrasted against the relatively few real world variables that have been observed well enough to be used in a detection study. There is also the problem that many of the variables output by the GCMs replicate nature so poorly that they should be excluded from use, even though they possess a large GHG signal.

The following strategy offers one quantitative methodology for sorting through the maze of possible choices and arriving at a few variables for detection work. Thus, we endorse a multivariate or "fingerprint" approach to the detection problem. The strategy is offered in three steps and assumes only that: (a) the model predictions will come from a GCM, and (b) the observed data are historical observations in quantitative format.

STEP I: Estimation Model S/N

The GHG signal (S) for a specific variable, say air temperature, from a transient or equilibrium model simulation is isolated. The model noise (N) is next estimated and the ratio (S/N) formed. This represents the relative strength of the GHG signal in the GCM simulation.
STEP II: Estimation of Model Validity

A measure of model "goodness" needs to be evaluated for the variable mentioned above from the GCM validation process. Many such measures are possible (and probably desirable).

STEP III: Estimation of Similitude

A measure of the degree of similitude between S (the model-predicted GHG signal) and the principal pattern of high-frequency variation in the observed data set needs to be estimated.

The candidate detection variables are then ranked by the three measures noted above, for example, select variables for detection work that have the strongest signals in the model's world, that are most accurately simulated by the model, and which do not appear similar to the signals found in high-frequency, natural variability.

3.2. Selection of Space and Time Scales for GHG Signal Detection

There are two aims in choosing specific space and time scales: (1) identification of a specific pattern or "fingerprint" of GHG-induced climate change which is specific and separable from climate change associated with any other mechanism; and (2) maximization of the signal-to-noise ratio for the GHG signal relative to all other climatic variations.

Complex climate models are unreliable on their smallest resolvable scales in space and time. Model simulations at single gridpoints usually show little or no relationship to single station or small region observations for the current climate and thus cannot be expected to provide reliable estimates of climate change on these very small scales. Spatial-averaging of model variables (larger-scale estimates) compare much better with similarly averaged estimates from observations of current climate. The models cannot represent the regional-scale variations of the observations because these small scales are not accurately represented (or parameterized) in the models.

Hence, only large spatial scales should be selected from models and observations for comparison and GHG-signal detection. This scale selection may be achieved by averaging in space or by spatial filters that pass only larger-scale (small wavenumber) components. The most severe scale selection is to use a global (horizontal) average similar to that used for surface temperature. This allows for simple intercomparison between a range of models (complex and simple) with observational data. However, it has the major drawback that it cannot separate possible mechanisms of climate change that may have a similar effect on global-mean temperature but different effects on spatial patterns.

There are specific aspects of the spatial structure of GHG-induced climate change in models that can be used as a guide to scale selection. Models show an enhanced warming at high latitudes in winter (associated with reductions in sea ice and snow amounts) relative to the tropics. The thermal response associated with increased amounts of GHGs in the troposphere is warming in the troposphere and cooling in the stratosphere. Thus the vertical and/or meridional structure of the zonal-mean temperature variations in models and observations should be compared. Model simulations also generally show larger warming over continents than oceans, so comparisons of full 2-D fields truncated in wavenumber space to retain only the largest-scale variations (zonal wavenumbers < 10) should be used. An additional benefit of this selection of only the largest spatial scales is the immense data compression that it provides, allowing easier computations for statistical testing.
Scale selection in time is intended to reduce the "noise" in model and observational data associated with processes other than GHGs. Low-pass time filters that select periods comparable to the time scale of significant variations of GHG amounts should be used. At the present rate of increase of GHGs (~ 0.4%/year for equivalent CO₂), variations on decadal time scales are still small. Thus, comparisons of variations of the model and observational data on time scales longer than a decade must be used, with longer time scales being even better. This filtering will eliminate the direct contribution of higher-frequency components in the data, such as those due to meteorological fluctuations or ENSO events, but it cannot remove the influence of other sources of very low frequency variations in the model or observations such as internal nonlinear dynamics, deep ocean circulation or external forcing (volcanic dust or solar irradiance) variations.

So far, attention has been concentrated on the direct thermal impact of GHGs, for which selection of large space and very long time scales is best. Different scale selection may be required for other parameters or to make best use of other data. Even for temperature, it may be sensible to select other specific scales. The larger sensitivity of the climate system at the sea-ice and snow boundary regions in winter to GHG warming should lead to a reduction in the annual cycle of temperature at high latitudes. This suggests selection of this time scale for comparison between model and observations. The sensitivity of the high-latitude regions and the sea-ice/snow boundaries (in winter) may indicate the selection of specific proxy data types for collection.

3.3. METRICS: How to Measure the Greenhouse-Gas Signal Quantitatively

Previous detection studies can be divided into three categories:

(1) Signal-to-noise analysis. This considers one variable only (generally global-mean surface temperature) and compares the signal with the noise level defined by natural variability.

(2) Noise reduction analysis. As in (1), this considers a single variable only, but attempts to reduce natural variability by identifying and factoring out the effects of forcings other than CO₂ (e.g., solar, volcanic).

(3) Fingerprint analysis (multivariate studies). These consider a number of 2-D or 3-D spatial fields, either individually or simultaneously.

Studies in (1) and (2) generally involve statistics which are used to test for the existence of a trend. Detection studies in (3) have employed a variety of statistics to test model-versus-observed changes in means, variances and spatial patterns. Here, only those statistics which are potentially useful in fingerprint analysis/multivariate studies are considered, since this later detection strategy appears to be more powerful in its abilities to discern a predicted GHG signal.

3.3.1. Statistics for First-Moment Tests

Several univariate/multivariate measures of the mean differences between two distributions of data, such as a GCM-predicted GHG signal and observations of that signal, have been employed in detection studies. These include:

(1) one-tailed/two-tailed t-tests

(2) "cumulative" t-tests
(3) Hotelling's $T^2$
(4) Mahalanobis $D^2$/Multivariate Recurrence Analysis
(5) SITES

The first two measures are commonly described in statistical text books and we defer the interested reader to such sources. The Hotelling's $T^2$ statistic (and variants thereof; see Hasselmann, 1979) has been used in model validation and predictability studies but has not been applied in a GHG detection context. An important constraint on the use of $T^2$ is the dimensionality problem, that is, the number of spatial points in tests involving GCM data is often several orders of magnitude larger than the number of time samples. This means that some form of data compression must be employed (e.g., spatial averaging, EOFs, spherical harmonics). The SITES statistic introduced by Preisendorfer and Barnett (1983) is a dimensionless measure of the separation of data set centroids. It has been used as a test of global differences in means in detection studies by Barnett et al. (1991) and Santer et al. (1991). The reference distribution of SITES is unknown, so its significance is judged using a permutation approach.

3.3.2. Statistics for Second-Moment Tests

(1) one-tailed/two-tailed F-tests
(2) "cumulative" F-tests
(3) SPRED, SPRET1
(4) SPREX1

Local F-tests supply spatial and directional information concerning differences in temporal variances. "Cumulative" F-tests are the variance ratio analog of cumulative $t$-tests. The Preisendorfer and Barnett (1983) SPRED statistic and the SPRET1 statistic (Wigley and Santer, 1989) are both measures of overall differences in temporal variance. SPRED is a dimensionless measure of the radius of data set "swarms," while SPRET1 is defined as the ratio of spatially averaged temporal variance in two data sets. Both give equivalent information (in terms of significance levels), but SPRET1 provides directional information which SPRED cannot supply. SPREX1 (Wigley and Santer, 1989) is the spatial analogue of SPRET1 and is defined as the ratio of the time-averaged spatial variance in two data sets. SPRED, SPRET1 and SPREX1 have been used in previous detection studies.

3.3.3. Tests of Patterns

(1) Spatial Field Correlation ($r$)
(2) SHAPE
(3) RBAR

Perhaps the most interesting category of statistics for detection purposes includes measures of the similarity in time-mean spatial pattern ($r$) and in the temporal evolution of spatial anomaly fields (SHAPE, RBAR). SHAPE (Preisendorfer and Barnett, 1983) and RBAR (Wigley and Santer, 1989) give equivalent significance information. The only advantage of RBAR is that the actual statistic value can be directly interpreted in terms of
common variance. SHAPE and $r$ have been used by Barnett (1991) in a detection study with the GISS transient response experiments, and by Santer et al. (1991) in a detection study involving the UKMO and OSU equilibrium response experiments.

3.4. Statistical Inference Strategies

In this section we consider the problem of determining whether the model-predicted GHG signals are found in the observations with a given confidence level. We assume it is possible to identify one or a few GHG indices that can be computed from simulated climates and currently available analyses to efficiently capture the GHG signal if it exists (see subsections 3.1 and 3.2 above). Our goal is then to make point and interval estimates of the GHG signal. Hypotheses testing, *per se*, should at best be a secondary concern.

There are some key issues which impact on estimation and inference procedures. A partial list of these items includes:

1. There is ONLY ONE observed record and that is likely non-stationary (chaotic?) on time scales of decades to centuries even in the absence of GHG signals (see Lorenz, 1991; Zebiak and Cane, 1991; Hansen et al. (1988, the control run); Zwiers (1987); amongst many others).
2. Other anthropogenic signals are likely confounded with the GHG signal. These include effects due to:
   a. Albedo changes due to land-use practices,
   b. Potential evaporation and soil hydrology changes due to land-use practices,
   c. Tropospheric aerosol loading, and
   d. Albedo changes of Arctic and Glacial Ice due to the fallout out troposphere aerosols.

The impacts of these issues (from an estimation/inference point of view) are:

1. Traditional estimation and inference procedures, both parametric and non-parametric, implicitly assume that samples consist of independent observations or perhaps represent a sample taken from a stationary time series. The fact that GHG index time series may well be non-stationary (even in the absence of a GHG signal) suggests that traditional inference techniques will ascribe too much confidence to estimates of GHG signals. It is likely that present estimates of the rate of increase of annual global-mean surface temperature suffer from this problem.
2. Statistical procedures will not be able to separate the GHG signal from other anthropogenic (and natural) signals unless a clear GHG-only "fingerprint" can be established (see subsections 3.1, 3.2 and 3.5).

3.4.1. One Estimation/Inference Strategy

Promising statistical tools for estimating the strength of the GHG signal in an appropriate index which has not yet been investigated by atmospheric scientists are the non-stationary, non-Gaussian, state-space modelling techniques recently described by Kitagawa (1987). The strategy would consist of the following components:

1. Develop a body of expertise in the atmospheric sciences on the use and application of state-space models and other "structural" models.
(2) Apply a state-space model to currently available data. In the absence of inhomogeneity introduced by the observing network and its "homogenization," the model should reveal the drift of the climate (due to natural and anthropogenic causes) together with overriding shorter duration spikes due to volcanism. Indeed, the detection of such spikes would serve as a useful means of validating these techniques.

In addition to the estimated drift, the model will also provide confidence limits indicating how well the current climate state is known. One would expect that the width of these intervals would vary with the density of the observing network and the (possibly state-dependent) short time scale internal variability of the climate system. One would also expect to see the secular warming trend superimposed on the background drift and shorter-term climate signals. It may not be possible to ascribe significance to the secular warming trend because the instrumental record is not long enough to assess the "natural" low-frequency variability of the climate state.

(3) The latter issue could perhaps be addressed by applying the same state-space models to coupled A/OGCMs.

(4) Analyze current transient experiments to determine the sensitivity of state-space methods.

3.5. Attribution

The fact that a model-predicted GHG signal* is found in the observations with high statistical confidence does not necessarily mean the GHG effect has been observed. There are other mechanisms that might produce a similar signal, for example, natural variability associated with chaos, tropospheric aerosols or variation in the solar constant. To attribute the observed changes predicted by the models to the GHG mechanism will require that other potential competing mechanisms be defined. Model simulations forced by these competitors then must be carried out and the resulting responses defined. If the response appears similar to the GHG signal then some means of critically distinguishing between two competing mechanisms must be found, or else attribution (defining cause and effect) becomes impossible.

4. RECOMMENDATIONS

The current state of our ability to assess the possible role that GHG has played in the recent history of our global climate is unsatisfactory. In fact, given the implication for humanity, the current situation is unacceptable. We offer the following suggestions that, if followed, will help ameliorate the present situation.

A. Detection Panel

A "Detection" Panel should be formed immediately. The chief function of this group would be to coordinate GHG research activities toward the goal of making definitive quantitative statements about the impacts of GHG on the environment of the planet. Among the immediate tasks of this group would be to:

* "Signal" here means the full 4-dimensional GHG "fingerprint."
(1) Define a rigorous, well-reasoned detection strategy, perhaps along the directions outlined in Section 3.

(2) Standardize and make readily available observational data sets for GHG work. The quantitative uncertainties in these observations should be defined. This task should be done ideally through a single "clearing" house directed by a "Data Tzar."

(3) As in (2), except for model simulations of GHG and other effects (see (5) below).

(4) Ensure that a multidisciplinary approach is taken to the detection problem. This means the use of the best scientific minds from meteorology, oceanography, chemistry, statistics, etc. in the development of a detection strategy.

(5) Commission new, specific GCM (transient) runs to address the problems of attribution and intransitivity as it affects GHG-signal detection.

The membership of the Panel should be drawn from the scientific community (not politicians) and rotate on/off the Panel to ensure continued broad perspective and objectivity.

B. Computer Resources

It is our view that current research on the GHG problem is seriously hampered by lack of computer resources. Transient O/A GCM runs are the heart of the GHG research program and they require massive resources which now exist but are not available to researchers. It is true that the problem of obtaining a stable coupled model is difficult, but coupling strategies can be tested only with long (10-20 year) simulations, and this requires large computer usage. And this must be done before any actual GHG scenarios or other "attribution" problems are run.

We strongly recommend that at least one CRAY 2-equivalent machine be dedicated to the necessary GCM simulations. Allocation of these resources should be carried out by the Detection Panel to ensure that the most critical simulations get the highest priority.

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