PAST CLIMATE CHANGES DERIVED FROM ISOTOPE MEASUREMENTS IN POLAR ICE CORES

XA0300363

J. BEER, R.MUSCHELER, G. WAGNER
Swiss Federal Institute of Environmental Science and Technology (EAWAG), Dübendorf, Switzerland

P.K. KUBIK
Paul Scherrer Institute, Zürich, Switzerland

Abstract. Measurements of stable and radioactive isotopes in polar ice cores provide a wealth of information on the climate conditions of the past. Stable isotopes ($\delta^{18}O$, δD) reflect mainly the temperature, whereas $\delta^{18}O$ of oxygen in air bubbles reveals predominantly the global ice volume and the biospheric activity. Cosmic ray produced radioisotopes (cosmogenic nuclides) such as ^{10}Be and ^{36}Cl record information on the solar variability and possibly also on the solar irradiance. If the flux of a cosmogenic nuclide into the ice is known the accumulation rate can be derived from the measured concentration. The comparison of ^{10}Be from ice with ^{14}C from tree rings allows deciding whether observed ^{14}C variations are caused by production or system effects. Finally, isotope measurements are very useful for establishing and improving time scales. The $^{10}Be/^{36}Cl$ ratio changes with an apparent half-life of 376,000 years and is therefore well suited to date old ice. Significant abrupt changes in the records of ^{10}Be, ^{36}Cl from ice and of $\delta^{18}O$ from atmospheric oxygen representing global signals can be used to synchronize ice and sediment cores.

1. INTRODUCTION

Stable and radioactive isotopes are useful tools to study present and past environmental changes. Due to progress in analytical techniques, very small samples can now be analyzed with high precision. This opens up the possibility of reconstructing past environmental changes with high temporal resolution in natural archives and of obtaining a comprehensive picture of the spatial and temporal variability of our climate system. For several reasons, ice cores merit special attention. They are the only archive, which stores not only all particular constituents removed from the atmosphere but also air samples. This unique property allows for instance to reconstruct the history of greenhouse gas concentrations as well as isotopic changes in atmospheric oxygen, a proxy for the global ice volume. During the last few decades, several ice cores were successfully retrieved from Greenland and Antarctica containing complete and continuous ice records of the last 100,000 or more years. Recently, the information derived from polar ice cores has been complemented by new data from an increasing number of low-latitude ice cores drilled into high-altitude ice sheets. The latter are crucial to the understanding of the regional patterns of climate changes. In this short overview, we will concentrate mainly on isotopes from polar ice cores.

1.1 Stable isotopes

Usually, an ice core is first analyzed for $\delta^{18}O$ and/or δD. The isotopic fractionation that takes place in the inversion layer during the condensation of water vapor to raindrops or snowflakes is strongly temperature dependent. The calibration of the relationship between $\delta^{18}O$ and temperature is usually made by applying linear regression to a set of annual data from different sites, where $\delta^{18}O$ and/or δD and temperature have been measured. In Fig. 1a, the δD record from Vostok, Antarctica is shown [1]. It covers more than 4 glacial cycles. Spectral analysis of the Vostok record reveals the Milankovich cycles caused by changes in the orbital parameters of the Earth. Periodicities around 100 kyrs (eccentricity), 40 kyrs (obliquity), and 20 kyrs (precession) were found. A detailed view of climate changes during the younger part of the last glacial is depicted in Fig. 2 showing the GRIP $\delta^{18}O$ record from Greenland [2]. Being representative for the North Atlantic temperature the GRIP $\delta^{18}O$ record is dominated by a series of very rapid fluctuations, the so-called Dansgaard-Oeschger events (DOE). Typically, $\delta^{18}O$ increases by about 5‰ within decades, then begins to decline slowly and subsequently drops rapidly to the low value it had at the beginning. The general explanation for this behavior is the switching on and off of the deep-water formation in the North Atlantic [3]. Melt water from the Laurentide ice sheet surging into the North Atlantic leads to a drop in the density of the surface water and shuts down the thermohaline circulation. As a consequence, the heat transport from low latitudes to Greenland is interrupted and the temperature decreases dramatically. After some time, the thermohaline circulation sets in again and the temperature rises. As we will see below, this process can be studied on a global scale using radioisotopes.

However, one should bear in mind that the simple linear relationships between $\delta^{18}O$ or δD and the temperature have been established based on a spatial network of climatically different stations. These relationships depend also on other parameters such as the source area and the transport of water vapor and the seasonal distribution of precipitation. In fact, using these simple relationships leads to a temperature change of approx. 10°C for central Greenland between the Holocene and the last glacial maximum. The analyses of borhole temperatures on the other hand points to a much larger change (20°C) [5] indicating that besides temperature changes other changes were involved as well.

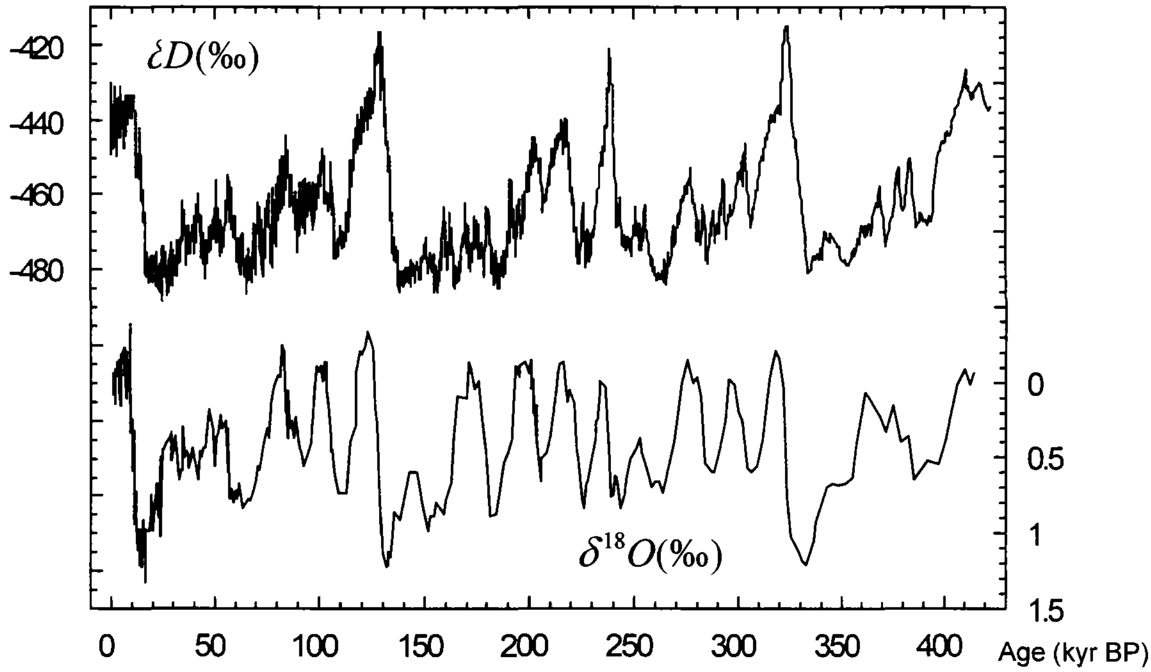

FIG. 1. Records of stable isotopes from the Vostok ice core over the last 440 kyrs, a) δD measured in the ice reflects the temperature [1], (b) $\delta^{18}O$ measured in air bubbles provides information on the global ice volume and the biospheric activity [4].

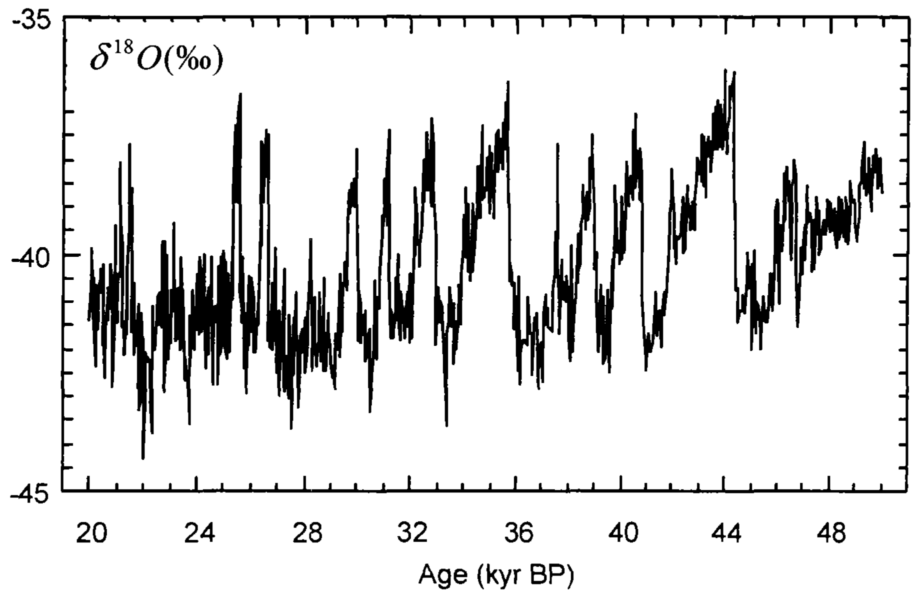

FIG. 2. Section of the $\delta^{18}O$ record from that section of the GRIP ice core, which is characterized by abrupt climate changes, the so-called Dansgaard-Oeschger events [2].

Ice cores offer the unique opportunity to measure $\delta^{18}O$ not only in solid precipitation (snow and ice) but also in gaseous O_2 trapped in air bubbles. $\delta^{18}O$ of atmospheric oxygen reflects the $\delta^{18}O$ signal of sea water after modification by hydrological (transpiration, evapotranspiration) and biochemical (photosynthesis, respiration) processes. [6]. The $\delta^{18}O$ difference between sea water and atmospheric oxygen (Dole effect) is mainly due fractionation during respiration. Hence $\delta^{18}O$ in an ice core provides information about the local temperature and the global ice volume and the biospheric activity at the same time. As shown in Fig. 1 the overall agreement between the δD of ice and the $\delta^{18}O$ in air bubbles in the Vostok core is good [4]. However, small differences indicate that some of the involved processes were not constant.

1.2. Radioisotopes

Ice sheets and glaciers are ideal archives for all radioisotopes, which are transported through the atmosphere and removed by wet or dry deposition processes. In the following discussion, we consider only so-called cosmogenic radionuclides. Galactic cosmic ray particles interact in high-energy nuclear reactions with atmospheric nitrogen, oxygen and argon. As a result, a cascade of secondary particles develops until all the energy of the primary particle is dissipated. Among the secondary particles, mainly neutrons and protons are responsible for the reactions leading to cosmogenic nuclides. For some of them the main properties such as half-life, main target elements and mean global production rate are given in Table I. The flux of the galactic cosmic rays is modulated by magnetic fields, which deflect mainly the low-energetic particles. The Earth's magnetic dipole field prevents particles of a rigidity below a geomagnetic-latitude dependent cut-off energy from penetrating into the atmosphere. The second magnetic modulation effect originates from the sun. Solar wind carrying frozen-in magnetic fields fills the vicinity of the solar system and also acts as a magnetic shield. During periods of high solar activity, the solar wind intensity increases enhancing the shielding effect and therefore reducing the production rate of cosmogenic nuclides and vice-versa. After production, the fate of cosmogenic nuclides depends strongly on their geochemical properties. Some of them, such as ^{10}Be, become attached to aerosols and follow their pathways. After a mean residence time of 1 to 2 years, they are removed from the atmosphere mainly by wet precipitation. ^{14}C, on the other hand, becomes oxidized to $^{14}CO_2$ and exchanges between the carbon reservoirs atmosphere, biosphere and ocean. The production rate of cosmogenic nuclides can be calculated as a function of solar and geomagnetic modulation [7]. In the following, we will discuss some of the most important applications of cosmogenic nuclides for climate change research of the past.

Table I. Some properties of cosmogenic nuclides

Nuclide	$T_{1/2}$ (y)	Target	Production rate (atoms cm^{-2} s^{-1})
^{10}Be	$1.5\ 10^6$	N, O	0.018
^{14}C	5730	N, O	2.02
^{36}Cl	$3.01\ 10^5$	Ar	0.0019

2. SOLAR VARIABILITY

The sun is the fundamental source of energy that drives the climate system. Changes in solar radiation inevitably lead to adjustments in the radiation balance of the earth and to modifications of the complex climate system. Direct measurements of solar irradiance (solar constant) with satellite-based radiometers revealed variations in phase with the 11-year Schwabe cycle over the last two decades [8]. The yearly averaged amplitude of this cycle is too small (0.1%) and the changes are too fast to induce relevant climate changes. However, this does not mean that over longer time scales, the solar irradiance did not undergo much larger changes, which may have significantly affect the climate [9]. In fact, there are clear indications that solar irradiance has increased since the Maunder minimum (1645-1715 AD) by about 0.24% [10]. The Maunder minimum period is famous for its almost complete lack of sunspots pointing to a different mode of the solar magnetic dynamo. At the same time, this period is known for much cooler temperatures (Little Ice Age). Evidence for a much larger potential of solar variability than observed in our sun so far comes from other solar type stars [11]. A long-term study of many such stars suggests that changes in radiation of up to 1% can occur. To investigate, whether the sun ever made use of this potential it becomes necessary to learn more about the solar activity of the past. Direct, detailed and reliable observations of solar variability are limited to the telescopic era, which goes back to about 1600 AD. To extend this period, one has to rely on indirect solar parameters such as cosmogenic nuclides. The analyses of ^{14}C records in tree rings and ^{10}Be data from polar ice cores [12] reveals periodicities around 11 years (Schwabe cycle) and 90 years (Gleissberg cycle), as expected from the sunspot records. The 205 De Vries cycle could be found during the Holocene and between 25 and 50 kyrs B.P. [13]. This suggests that the solar dynamo is characterized by several periodicities which appear to be active and rather constant averaged over long times, although the amplitudes may change with time.

Special features attributed to reduced solar activity are the so-called "grand minimum" periods like the Maunder minimum. Such minima occur irregularly with different durations throughout the Holocene. In Fig. 3, the $\Delta^{14}C$ (of the atmospheric $^{14}C/^{12}C$ ratio from a standard in permil) derived from tree rings [14] is shown for the last 7 kyrs after subtraction of the long-term trend. The solar minimum periods, which correspond to production maxima, are indicated with arrows. It is interesting to note that several of these minima in solar activity can be related to deteriorations in the climate [15] underlining that cosmogenic nuclides play a key role in establishing a causal relationship between solar variability and climate change [16].

3. ACCUMULATION RATE

Probably the most fundamental parameter of climate change is the precipitation rate. In theory, an ice core represents a complete collection of all the past precipitation and measuring annual layer thicknesses can therefore be used to reconstruct the history of precipitation. However, in practice, there are two problems that complicate the situation considerably. Firstly, ice flows and this leads to a thinning of the annual layer thickness with increasing depth. To correct for this effect one has to apply ice flow models. Secondly, the discrimination between annual layers becomes impossible at larger depths. Again, cosmogenic nuclides provide a solution. It is based on the following two assumptions: 1. Averaged over a large area, the flux of e.g. ^{10}Be from the atmosphere into the ice is proportional to the production rate in the atmosphere: $F = \alpha * P$. 2. Averaged over millennia the production rate is mainly determined by the geomagnetic field intensity.
F can also be expressed by:

$$F = \rho * c * acc,$$

the product of ice density ρ, radionuclide concentration c and accumulation rate acc which leads to

$$acc = \alpha * P/(\rho * c).$$

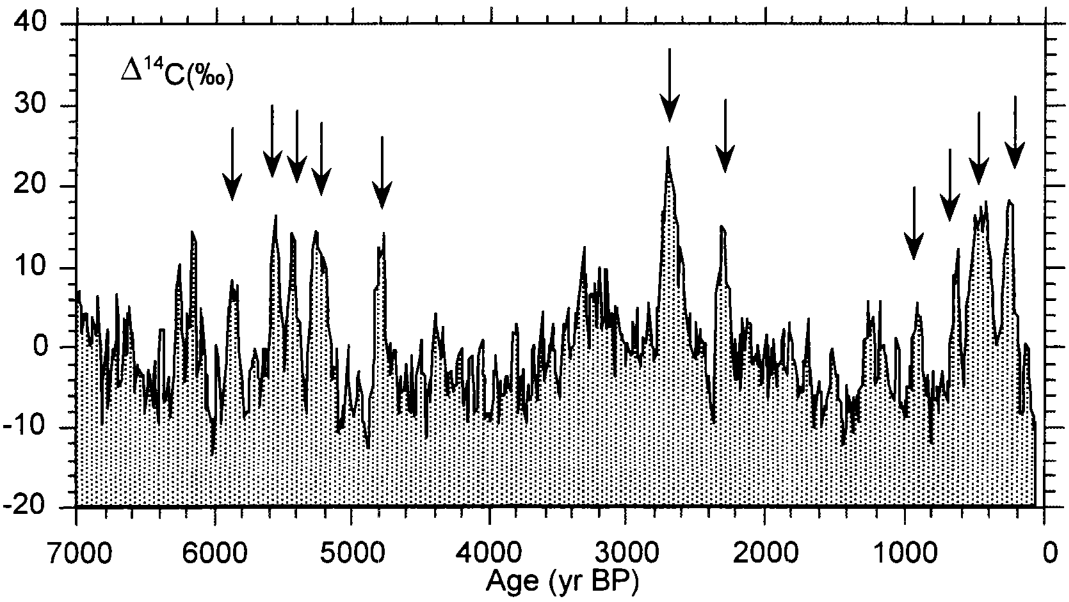

FIG. 3. $\Delta^{14}C$ record derived from tree rings [14] *after subtraction of the long-term trend. Periods of increased ^{14}C production, which are generally thought to coincide with periods of a quiet sun are indicated by arrows.*

Since ρ is virtually constant (0.92 g/cm^3), α can be determined for modern times and P can be derived from paleomagnetic measurements [17] the accumulation rate can be calculated. The results of this approach when using ^{10}Be in the GRIP ice core confirm Johnson's reconstruction which is based on an empirical relationship between $\delta^{18}O$ from ice and the annual layer thickness corrected for the thinning [18]. Fig. 4 shows that, in Greenland, the precipitation or accumulation rate changed dramatically during the last glacial in phase with the temperature.

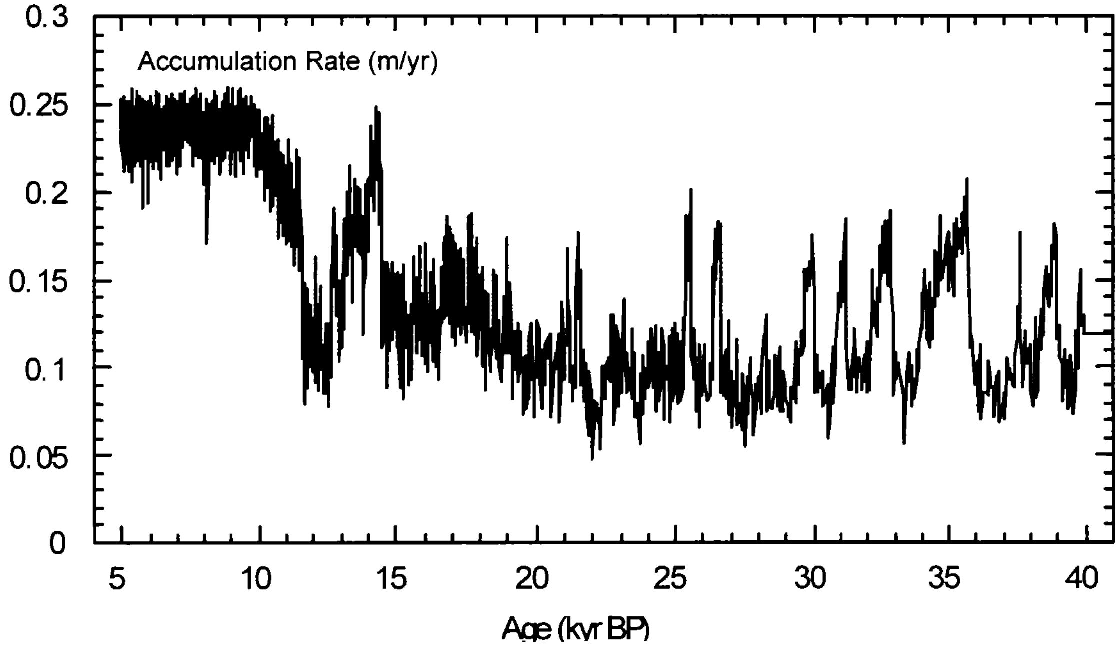

FIG. 4. Accumulation rate at Summit (Central Greenland) based on an empirical relationship between $\delta^{18}O$ [18] and the accumulation rate determined by layer counting and extrapolated to older ages (see text). This accumulation rate is confirmed by the accumulation rate derived from ^{10}Be measurements in the GRIP ice core.

4. THERMOHALINE CIRCULATION

The thermohaline circulation is a basic process for the oceanic heat transport from equatorial to polar regions. Since most of the deepwater formation takes place in the North Atlantic, Greenland's climate is especially sensitive to modifications in thermohaline circulation. In fact, it is generally believed that the rapid climate fluctuations during glacial times (DOE) (Fig. 2) are due to changes in this transport system. Since the ocean is the main reservoir of atmospheric CO_2, and the mean oceanic residence time of CO_2 is about 1,000 years, the thermohaline circulation affects the atmospheric $^{14}C/^{12}C$ ratio. A slowdown of the thermohaline circulation leads to a reduction in the removal rate of ^{14}C from the atmosphere and therefore to an increase in atmospheric $\Delta^{14}C$. But an increase could also be caused by a higher ^{14}C production rate due to a reduction in solar activity or geomagnetic field intensity. To distinguish between these two possibilities (thermohaline circulation or production), additional information is necessary. This is provided by ^{10}Be, which is produced in a very similar way as ^{14}C but differs considerably in its geochemical behavior. As a consequence, production changes affect both nuclides, while changes in the global deep-water formation only affect ^{14}C. This new approach has been successfully applied to the $\Delta^{14}C$ peak during the Younger Dryas cold event (Fig. 5) [19]. By comparing the $\Delta^{14}C$ record with the ^{10}Be record from GISP 2 [20] it has been shown that only the combination of a production change and a reduction of the global deep-water formation is consistent with both the ^{14}C and ^{10}Be measurements. Should a high quality, detailed $\Delta^{14}C$ record become available for the glacial time period from 15,000 to 50,000 years, this technique can be applied to study the DOE events.

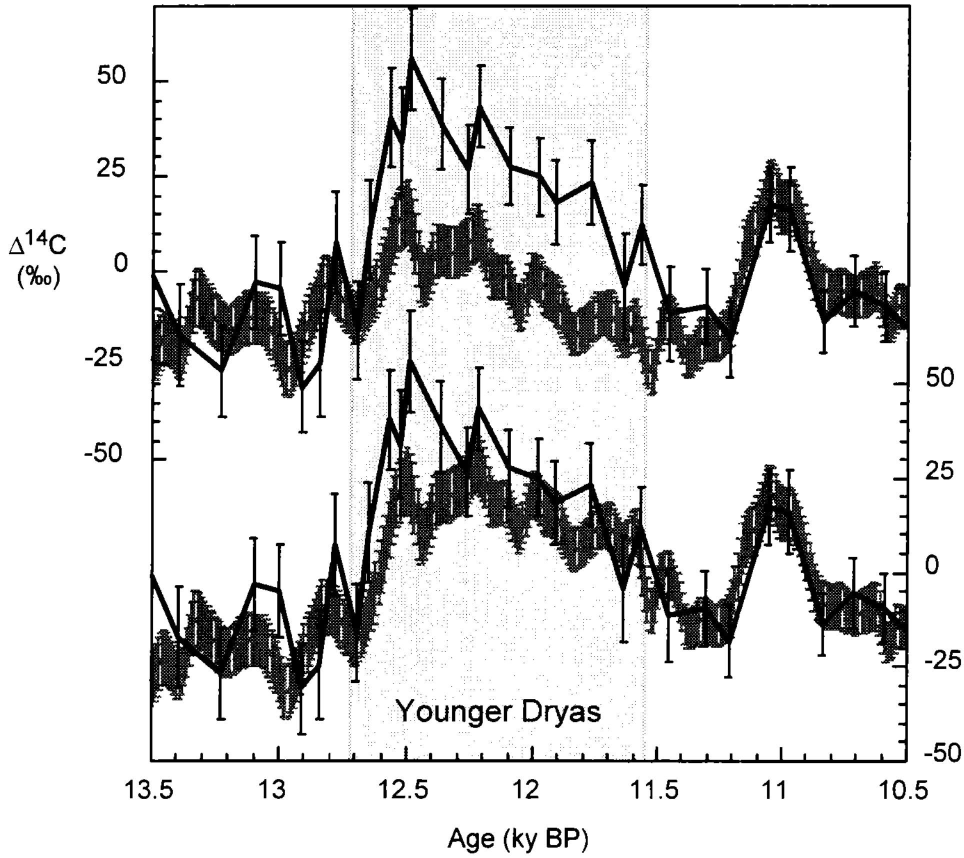

FIG. 5. Comparison of measured $\Delta^{14}C$ with $\Delta^{14}C$ calculated using the ^{10}Be data from the GISP ice core [20]. The discrepancy between the two curves during the Younger Dryas period indicates that the peak of the measured $\Delta^{14}C$ is not only due to a production change (upper panel). In fact, assuming in addition to the production increase a reduction of the global deep-water formation leads to a much better agreement (lower panel) [19].

5. DATING

A classical but still very important application of cosmogenic nuclides is dating. Precise time information is crucial to a comparison of hemispheric or regional climate changes and to determine rates of changes. In the case of polar ice cores, there are two major problems using cosmogenic nuclides for dating. Firstly, the concentration of radionuclides such as ^{14}C is very low. Even when using accelerator mass spectrometry (AMS), the extraction and processing of the tiny amounts of CO_2 occluded in air bubbles is extremely difficult. In addition, there is still no reliable ^{14}C calibration curve available for the glacial part, where the uncertainties of other dating techniques are larger and ^{14}C would be useful. Secondly, in most cases the initial concentration of a radionuclide is not well enough known to calculate an age with a precision of a few percent. One way of circumventing this problem is to use isotopic ratios. Since the production rates of both nuclides fluctuate in phase, the production rate ratio is constant at first order and can be determined experimentally on modern samples. One example is the $^{10}Be/^{36}Cl$ ratio, which is ideal for dating very old ice. It has an apparent half-life of 376,000 years and is therefore well suited to date the older parts of deep-drilling ice cores. Unfortunately, the situation is complicated by processes such as changes in transport or loss of chlorine from the ice. The latter process mostly occurs in low accumulation areas with low dust content (Antarctica) [21]. In Greenland, however, this dating technique looks promising, as shown by work in progress.

The accumulation rate derived from measurements of the cosmogenic nuclides as described above can then be used in ice flow models to establish a time scale.

Significant abrupt changes in the production rate of cosmogenic nuclides provide global markers that can be used to synchronize time scales of different ice and sediment cores [22]. An example of such a marker is the Laschamp event around 40 kyrs BP when the magnetic field intensity almost vanished [23].

In a similar way, $\delta^{18}O$ derived from oxygen in air bubbles can be used to synchronize ice and sediment cores [6].

6. CONCLUSIONS

Stable and radioactive isotopes are extremely versatile tools to study various processes and to deepen our understanding of the complex environmental system and the couplings between its components. In natural archives such as ice cores, they allow us to reconstruct past climate conditions and to analyze the dynamics of processes on time scales from decades to millennia. A look back into the past is therefore crucial for the development of realistic models that may help to shed light on the future of the earth.

ACKNOWLEDGEMENTS

This work was supported by the Swiss National Science Foundation

REFERENCES

[1] PETIT, J. R., JOUZEL, J., RAYNAUD, D., BARKOV, N. I., BARNOLA, J.-M., BASILE, I., BENDER, M., CHAPELLAZ, J., DAVIS, M., DELAYGUE, G., DELMOTTE, M., V. KOTLYAKOV, M., LEGRAND, M., LIPENKOV, V. Y., LORIUS, C., PEPIN, L., RITZ, C., SALTZMAN, E., STIEVENARD, M. "Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica," Nature **399** (1999) 429-436.

[2] DANSGAARD, W., JOHNSEN, S. J., CLAUSEN, H. B., DAHL-JENSEN, D., GUNDESTRUP, N. S., HAMMER, C. U., HVIDBERG, C. S., STEFFENSEN, J. P., SVEINBJOERNSDOTTIR, A. E., JOUZEL, J., BOND, G., "Evidence for general instability of past climate from a 250-kyr ice-core record," Nature **364** (1993) 218-220.

[3] BROECKER, W. S., BOND, G., KLAS, M., BONANI, G., WOELFLI, W., "A Salt Oscillator in the Glacial Atlantic? 1. The Concept," Paleoceanography **5** (1990) 469-477.

[4] SOWERS, T., BENDER, M., RAYNAUD, D., KOROTKEVICH, Y. S., and ORCHARDO, J., "The $\delta^{18}O$ of atmospheric O_2 from air inclusions in the Vostok ice core: timing of CO_2 and ice volume changes during the penultimate deglaciation," Paleoceanography **6** (1991) 679-696.

[5] DAHL-JENSEN, D., MOSEGAARD, K., GUNDESTRUP, N., CLOW, G. D., JOHNSEN, S. J., HANSEN, A. W., BALLING, N., "Past temperatures directly from the Greenland ice sheet," Science **282** (1998) 268-271.

[6] LEUENBERGER, M., "Modeling the signal transfer of seawater $\delta^{18}O$ to the $\delta^{18}O$ of atmospheric oxygen using a diagnostic box model for the terrestrial and marine biosphere," J. Geophys. Res. Lett. **102** (1997) 26841-26850.

[7] MASARIK, J. BEER, J., "Simulation of particle fluxes and cosmogenic nuclide production in the Earth's atmosphere," J. Geophys. Res. **104** (1999)12,099 - 13,012.
[8] FROEHLICH, C., "Observations of irradiance variations," Space Science Reviews **94** (2000) 15-24.
[9] BEER, J., MENDE, W., STELLMACHER, R., "The role of the Sun in climate forcing," Quat. Sci. Rev. **19** (2000) 403-415.
[10] LEAN, J., BEER, J., BRADLEY, R., "Reconstruction of solar irradiance since 1610: implications for climate change," Geophysical Research Letters **22** (1995) 3195-3198.
[11] RADICK, R. R., LOCKWOOD, G. W., BALIUNAS, S. L., "Stellar activity and brightness variations: a glimpse at the sun's history," Science **247** (1990) 239-247.
[12] BEER, J., JOOS, C. F., LUKASCZYK, C., MENDE, W., SIEGENTHALER, U., STELLMACHER, R., SUTER, M., "^{10}Be as an indicator of solar variability and climate," in The solar engine and its influence on terrestrial atmosphere and climate, vol. I. 25, NATO ASI, E. Nesme-Ribes, Ed. Berlin: Springer-Verlag (1994) 221-233.
[13] WAGNER, G., BEER, J., KIBIK, P. W., LAJ, C., MASARIK, J., MENDE, W., MUSCHELER, R., RAISBECK, G. M., YIOU, F., "Presence of the solar de Vries cycle (205 years) during the last ice age," Geophys. Res. Lett.(2000 in press).
[14] STUIVER, M., REIMER, P. J., BARD, E., BECK, J. W., BURR, G. S., HUGHEN, K. A., KROMER, B., McCORMAC, G., VAN DER PLICHT, J., SPURK, M., "INTCAL98 Radiocarbon age calibration, 24,000-0 cal BP," Radiocarbon **40** (1998) 1041-1083.
[15] VAN GEEL, B., RASPOPOV, O. M., V. d. P. J., and RENUSSEN, H., "Solar forcing of abrupt climate change around 850 calendar years BC," in Natural catastrophes during the Bronze Age civilisations, BAR International Series 728, B. J. Peiser, T. Palmer, and M. E. Bailey, Eds.(1998) 162-168.
[16] BEER, J., "Long-term indirect indices of solar variability," Space Science Reviews. Nov **94** (2000) 53-66.
[17] MASARIK, J., BEER, J., "Monte Carlo simulation of particle fluxes and cosmogenic nuclide production in Earth's atmosphere," presented at ICRC 25, Durban (1997).
[18] JOHNSEN, S. J., DAHL-JENSEN, D., DANSGAARD, W., GUNDESTRUP, N., "Greenland palaeotemperatures derived from GRIP bore hole temperature and ice core isotope profiles," Tellus **47B** (1995) 624-629.
[19] MUSCHELER, R., BEER, J., WAGNER, G., FINKEL, R. C., "Changes in deep-water formation during the Younger Dryas cold period inferred from a comparison of ^{10}Be and ^{14}C records," Nature **408** (2000) 567-570.
[20] FINKEL, R. C., NISHIIZUMI, K., "Beryllium-10 concentrations in the Greenland ice sheet project 2 ice core from 3-40 ka," J. Geophys. Res. **102** (1997) 26699-26706.
[21] WAGNON, P., DELMAS, R. J., and LEGRAND, M., "Loss of volatile acid species from upper firn layers at Vostok. Antarctica," J. Geophys. Res. **104** (1999) 3423-3431.
[22] BEER,J., JOHNSEN, S. J., BONANI, G., FINKEL, R. C., LANGWAY, C. C., OESCHGER, H., STAUFFER, B., SUTER, M., WOELFLI, W., "^{10}Be peaks as time markers in polar ice cores," presented at Proc. Erice Nato Workshop. The last delegation: absolute and radiocarbon chronologies (1992).
[23] WAGNER, G., MASARIK, J., BEER, J., BAUMGARTNER, S., IMBODEN, D., KUBIK, P. W., SYNAL, H.-A., SUTER, M., "Reconstruction of the geomagnetic field between 20 and 60 ka BP from cosmogenic radionuclides in the GRIP ice core," Nucl. Instr. Meth. **172B** (2000) 597-604.