XXIII General Assembly
of the European Seismological Commission

ACTIVITY REPORT 1990 - 1992
and
PROCEEDINGS

Volume II

Prague, Czechoslovakia, 7 - 12 September 1992

Geophysical Institute, Czechoslovak Academy of Sciences
XXIII General Assembly
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INTRODUCTION

Doublets are seismic events characterized by very similar waveforms, so they are interpreted as events generated by the same source segment. Normally two kind of doublets are considered: spatial and temporal doublets. The formers are earthquakes that occur in a very restricted area and in a short lapse of time; in the second case the same dislocation source is affected, so that events occur in the same point in a time window of years.

Spatial doublets could be used to study the source time evolution, while temporal doublets could be usefully utilized to obtain information on time variation of seismic waves velocities (Got et al., 1990). Indeed it is possible to take advantage of the similarity characteristics of such seismic signals to increase strongly the precision of arrival time definition of different seismic phases that can reach few milliseconds.

In this paper a case of spatial doublets, recorded by the seismic network of the Genoa University (Cattaneo et al., 1990) that occurred in two days in an area of the Maritime Alps (Italy), has been analyzed. The analysis method has also been tested using synthetic seismograms and signals generated by controlled sources.

ANALYSIS METHOD

In the doublets analysis the relative location techniques are normally applied considering one of them as a reference event or a "master event". The similarity characteristics of the signals are used to determine the relative arrival times of seismic phases.

In the usual analysis a maximum precision comparable with the sampling rate can be achieved, but, applying time and frequency domain techniques, it is possible to define with higher precision the arrival times also in traces strongly noised. In the time domain analysis, for every pair of events, a very limited part (few seconds) of signal around the seismic phase on approval (P, S, PmP, and so on) is considered and analyzed evaluating the cross-correlation between the signals for every station.

In order to analyze the whole seismogram, the signal is shifted, step by step, with respect to the master one and the maximum of the correlation function defines the delay of the examined traces (Deichmann et al., 1992). The use of the normalized cross-correlation function has given the best definition of the peaks correlation allowing a more careful evaluation of the phase delays (Augliera, 1992). This is very important when the station is close to the epicentre, so that the S phase is contaminated by P-waves coda. The phase delay is usually determined with a precision comparable with the sampling rate. In order to overcome this restriction, an interpolation between the cross-correlation values is carried out. Tests performed using a parabolic function, cubic splines and sinc function have shown that, even if heavier
from a computational point of view, the sinc function interpolates in a better way the data and it gives also a less biased peak estimation.

The analysis in the frequency domain is carried out computing, over a time window, the cross-spectrum of the examined traces. The time delay between the signals is defined by the slope of the linear relationship between the frequency and the cross-spectrum phase lag (Poupinet et al., 1984). Through the cross-spectrum, the coherency, that represents the signal similarity for every frequency, is computed and used as weight function: accepting for example, only values of the spectrum with coherency greater than 90 per cent. Generally all data analyzed in the frequency range 1-16 Hz have shown better coherency values.

These techniques have been tested using synthetic traces delayed of a given amount (some fraction of the sampling rate). Besides a random noise has been added in order to evaluate its effect on the delay definition. For events without noise or with a weak noise, operating both in the time and frequency domain, the lag determination can reach a precision of less than a millisecond. Generally also for very noised traces, the difference between imposed and calculated delay is less than 5 milliseconds.

When we compare the results obtained using the two methods and the different interpolation functions, we must emphasize that: both sinc interpolation (in the time domain) and cross-spectrum method furnish very good results for low noised events; increasing the noise the time domain analysis (with the sinc interpolation) give the most stable results. In this last case, the parabolic interpolation introduces a bias of the order of 3 ms or more.

To test the location of seismic events with the "doublets method" some explosions performed on land for refraction crustal experiments have been considered. The relative distance between shot points was about 100 m. The low magnitude of the events (nearly M = 2), the distance of 30 km from the nearest station and the eccentricity of the network with respect to shot point, make the test very critical. Nevertheless, eliminating stations with an unfavorable signal to noise ratio, using both methods, the maximum error in the relative epicentral location is about few tens of meters (Augliera, 1992).

Another crucial point is to define the distance limit for which a pair of events could be still considered a doublet. To give a contribution to solve this problem, a set of data, obtained by air-gun shots performed in the Ligurian Sea during a wide angle reflection experiment, is used. In spite of their low energy, the events have been recorded in digital form by the Genoa’s network to a distance of 120 km. The cross-correlation coefficient between pair of signals coming from increasing distances (the shots are performed every 60 m) along a profile of about 3 km, has been evaluated. The cross-correlation coefficient determined between the master event (the nearest event) and the following ones, by increasing the distance between the pairs, has been compared with the coefficient evaluated between pairs at constant distance.

The comparison demonstrates that, whereas the correlation is nearly constant for equispaced pairs, the coefficient decreases exponentially with the distance. So that a good waveforms stability (correlation coefficient greater 0.90) is observed for distances less 250 m between the shots and it degenerated for increasing distances over this limit. Even if, in this case, the
form of the signals might be influenced by the geological and topographic condition close to the shot point, it is convenient to consider the previous limit as the distance limit for the existence of very coherent doublets.

THE SEQUENCE IN THE MARITIME ALPS

As application of the doublets analysis on real seismic events, a microseismic sequence, occurred in a very narrow area of the Maritime Alps, is considered. The sequence was composed of about 40 events 16 of which with magnitude ranging between 1.4 and 2.2.

Using the standard methods for phase readings, foci do not show any clear evidence of clustering even if, considering the waveforms of signals, it is possible to distinguish at least three families of earthquakes. Each family is characterized by very similar shape of waveform in all considered station.

Considering as the master event the greater one of each family, we have performed the localization by means of the arrival times defined by the cross-correlation computed, in the time domain, with sinc function interpolation. The cross-correlation times (namely the differences of events pair) are added on the time of master event for every station and every phase.

To localize the events we have utilized an improved version of the hypoellipse program (Lahr, 1979).

In Fig. 1 it is possible to show that the events are clusterized in three different zones (F1, F2, F3) placed at distance between 1 and 3 km, and in each source zone the distance between pair events is some tens of meters. To give a better picture of the spatial distribution of foci of the family F1, two perpendicular cross-section (oriented E-W and S-N) have been performed. From the cross-section it is possible to recognize that foci are distributed along a sub-vertical plane oriented in the East-West direction. The focal mechanism of the main shock of this family (Fig. 2), evaluated with few but very consistent data, shows as a most probable solution a dextral trascurrent fault with one of the nodal plane oriented E-W in agreement with the distribution of doublets.

CONCLUSIONS

The utilized methodologies, in doublets analysis, can provide seismic phase reading with a precision of the order of milliseconds. The use of the normalized cross-correlation simplifies the recognition of peak of the cross-correlation; the sinc reconstruction of this function is preferable in comparison with parabolic interpolation.

Applying relative location techniques it’s possible to determine the geometrical shape of seismogenetic structures (sensibility in the displacements of the order of meters).

It’s possible to discriminate the main fault plane from the auxiliary one of focal mechanism and to recognize the different clusters of seismic activity.

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ENERGY OF SEISMIC WAVES

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1. Introduction
At the turn of years 1985/86 there occurred an earthquake swarm in the region of Western-Bohemia. A lot of seismograms were recorded during the swarm and a lot of scientific work was done over the material. I determined the energy of seismic waves for selected events of the swarm.

I use the method suggested by Boatwright (1980). The method is a modern modification of Golicyn's approach which was published in 1915. The method allows the seismic energy to be determined from observations at one station. I determined the energy from seismograms of two nearest stations VAC and TIS.

2. Calculation of the energy
The determination of the energy of seismic waves (i.e. seismic energy) is based on time integration of segments of (digital) velocity seismograms. The seismic energy is determined by the following formula (Boatwright, 1980)

\[ E_c = \frac{4\pi}{G^2 A^2} \frac{\langle (F_c)^2 \rangle}{\langle (F_c(\theta, \varphi))^2 \rangle} \frac{(v(x,t))^2}{(\kappa_c)^2} \int_{t_1}^{t_2} dt, \]

where \( c \) depicts P or S wave and \( v_c \) is the velocity of their propagation, \( F_c(\theta, \varphi) \) is the value of the radiation pattern for azimuth (\( \varphi \)) and take off angle (\( \theta \)) of the ray which connects the source and the receiver, \( \langle (F_c)^2 \rangle \) is the R.M.S. value of \( (F_c(\theta, \varphi))^2 \) over the whole sphere, \( \rho \) is density, \( G \) describes the influence of media (i.e. geometrical spreading, interfaces, etc.), \( A \) is absorption, \( \kappa_c \) the conversion coefficient of the free surface, \( v(x,t) \) is particle velocity at the point of observation \( x \), times \( t_1 \) and \( t_2 \) determine the segment of direct P or S waves on the record.

The influence of the source type and its radiation patterns and also the influence of the medium (including the free surface effect, absorption, etc.) are corrected during the calculation.

Total seismic energy is given as

\[ E = E^P + E^S = E^P (1 + q), \]

where \( q = E^S/E^P \).

By the described method, the seismic energy can be determined from observations only at one station; a smaller number of observations has to be compensated for the knowledge of other required data. We must know the structure of the medium, actually of the
model, which is its quite good approximation, then we must know the localization of the
event, the source type and the source mechanism. And, of course, the digital record of
processed event must be available.

Records of 138 processed events of the Western-Bohemia swarm are fairy good, but
the quality of the knowledge of other required data is not sufficient. The source mechanisms
were determined only for a few events from the processed set. Moreover, the found solutions
are not exact and unambiguous because of the paucity of input data. But for the calculation of
seismic energy by the described method it is necessary to know the mechanism. First, I
supposed that due to swarm relationship of processed events, all of them could have the same
mechanism as the main shock of the swarm. The mechanism of the main shock was
determined quite reliably from a large number of observations (strike = 171°, dip = 75°,
rake = -30° - see e.g. Antonini, 1988; the main shock is not included in the processed set of
events). The obtained results, however, were rather unrealistic and also unstable. Numerical
tests showed that the source mechanism was decisive in final seismic energy. It was therefore
necessary to determine the source mechanisms of processed events to obtain more realistic
results. To find the required mechanisms, I have developed a method, which determines the
source mechanism from observation at two stations (VAC and TIS).

3. Method of determining source mechanism

The source mechanisms were determined by following method: I supposed that all
quantities in relation [1] were exact, except the three parameters describing the source
mechanism (strike, dip, rake), and found such a source mechanism, which satisfied best the
criteria

\[ [E_S/E_P]_{VAC} - [E_S/E_P]_{TIS} \] + 4^2 \left[ (E_S^{VAC})/(E_S^{TIS}) - 1 \right]^2
+ 0.2^2 \left[ (E_S/E_P)_{VAC} - q \right]^2 + 0.2^2 \left[ (E_S/E_P)_{TIS} - q \right]^2 = \text{min}.

Criteria [3] can be described as follows: seismic energy determined from both stations is
required to be equal, ratios of energy \( E_S/E_P \) to be equal and to be approximately 20 (i.e. their
theoretical value - see e.g. Boatwright, 1980). Expression [3] was constructed intuitively and
then modified into the present form according to the results of test calculations.

4. Selection of reliable solutions

The mechanisms of processed events found by the method described above are
certainly not determined with equal reliability. The quality of the solution can be
characterised by the value of the minimum from condition [3]. From the set of all solutions
(138 events) I therefore discarded those which had a relatively great value of the minimum
(then 60 events remained).
The selected solutions are plotted in Fig. 1. Average values of the parameters of mechanisms of 60 selected events (strike = 201°, dip = 76°, rake = -23°) approximately agree with the mechanism of the main shock (see above). I suppose, that these facts indicate that the determined mechanisms can be regarded as physically realistic. From the results given above it follows that the source mechanisms of selected events are not in principle different from the main shock mechanism and can be considered as its variations.

Fig. 1 Selected solutions.
Mechanisms of 60 events are plotted. Coordinates of points correspond to parameters strike and dip, rotation of the line segment corresponds to parameter rake.

5. Characterization of the method of determining the source mechanism

In my opinion the advantage of the method is that it works with seismic energy, i.e. with a quantity which is integral characteristic of the record interval that corresponds to the direct wave. For the determination of the source mechanism in standard way only signs of the first motion and/or maximum amplitudes of P waves are used (see e.g. Zahradník et al., 1988); i.e. only one value selected from all the record. The method should therefore work better especially for multiple events which cannot be realistically characterized by a single simple read value. But for multiple events it is usually more difficult to determine seismic energy since it is not easy to interpret properly the interval of the direct wave.

The performed tests showed that from the global point of view the results (both source mechanisms and seismic energy) are not strongly dependent on the starting values of the mechanism and on the model of a structure - e.g. the stability of solutions is fairly good. The conclusion is important because only a very approximate model of structure is available for the region under study.

Another advantage of the method is that the source mechanism can be determined from low numbers of observations (in an extreme case only from two).
As the seismic energy is determined from the square of velocity, the information on the orientation of the motion is lost, and therefore, for parameter rake we can determine its direction only, not its orientation - so its value can differ by 180°. I think, the orientation of rake can be established according to additional information (e.g. geological).

Another disadvantage of the method is that it is necessary to know a lot of additional information (about the medium structure, localization of the event, etc.).

It has to be stressed that the use of the source mechanism determined by the method described above for the calculation of seismic energy implies that seismic energy of one event determined from records of stations VAC and TIS is not independent, and neither is energy $E^P$ and $E^S$ determined from the record at one station.

6. Results and conclusion

The new method of determining the source mechanism was developed because the paucity of data did not allow the application of the standard determination. I suppose that the method can be useful as an additional one to standard ways of determining the source mechanism from signs of the first motion and/or amplitudes of P waves. Source mechanisms determined by the method described above have been used for calculations of seismic energy of processed events. The performed tests have shown that the stability of the results is quite good.

Seismic energy of 60 selected events was calculated from seismograms recorded at stations VAC and TIS - Kolář (1992). The reliability of obtained results is not very high because of the bad quality of the required additional knowledge (knowledge the structure model, localization, source mechanism, etc.).

References


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ANALYSIS OF CHANGE OF THE ROCK MASSIF STRESS STATE BETWEEN THE EARTHQUAKE INITIAL RUPTURE AND AFTERSHOKES ON EXAMPLE OF THE SPITAK SOURCE ZONE.

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The solution of inverse kinematic problems has a first priority among other seismic problems since no solution common for them all has so far been found. The existing methods such as the method of trials and mistakes as well as method of inversion etc. Have certain shortcomings and therefore the solution of inverse seismic problems cannot be simple. The typical example is cola super deep hole section where the study of the core refuted the composition of the rocks and structure of area of drilling assumed from seismic data.

Mean quantities of elastic waves velocities based on some geological strata having different quantity of some types of the rocks in its composition calculated according to the additivity rule well agree with the seismic data that as shown during the work with the samples from the super deep hole (1990). Taking into account the abovementioned information an attempt has been made in the present study to solve the cuverse kinematic problem on the example of the spitak earthquake zone based on the data of deeping geological cuts and the times of first arrival of longitudinal waves at the nearest from the epicenter s\station, i.e. Stepanavan and Bogdanovka.

It is known that the initial rupture of the spitak earthquake was at a depth about 3 km (1989). There are real geological cuts of the concrete description of the rocks composition which from zones from the epicenter of earthquake to abovementioned stations. The cuts from the place of the initial rupture in the spitak earthquake focus to the stations Stepanavan and Bogdanovka which fixed the time of first arroval of longitudinal waves are presented in fig.1.

There are the data from the catalogue (1988) for the determination of the distances from the stations to the focus and travel times of the elastic longitudinal waves and elastic characteristics of real rocks in the conditions of occurence is taken from reference book (1978).Extremal values of the longitudinal waves velocities for each type of the rocks are included into the calculation, that is the effects of the anisitropy of the velocities of the elastic waves in rocks are excluded.

The results of the calculations show that the mean quantity of the longitudinal waves velocities for the Stepanavan s\station was 4.59 km/s, but the same velocity from the catalogue was 6.29 km/s. For the Bogdanovka s\station, the opposite results were obtained-based on the calculation is 5.21 km/s, from the catalogue is 4.48 km/s. These results conform the conclusions drown in works (1990,1989) about strong change of deformation field at the moment of the initial rupture of the focus. Probably the zone of the
compression is directed towards the Stepanavan s/station, and the zone of stretching under the 90 deg. breiches towards the Bogdanovka s/station.

Analogous calculations were made for the aftershocks of the Spitak earthquake in during to 7.12.1988 - 11.12.1988, registred at the same s/station. The results of the calculation show on fig.2. The solid line correspondes the middle calculated values of the longitudinal waves velosities taken from the laboratory data and geological cuts for s/stations Stepanavan (a) and Bogdanovka (b). The points show values of the longitudinal waves velosities, calculated according to the catalogue data. All points under the solid line show that zone of stretching finds between focus of the earthquake or the aftershocks and concrete s/station. The points above the solid line show that zone of the compression finds between focus of earthquake or the aftershocks and concrete s/station. As you see from fig.2 the zone of stretching is directed towards the Bogdanovka station and the zone of the strong compression take place around Stepanavan station.

Therefore on example of the Spitak earthquake and its aftershokes is shown that if we have a real geological cut from the epicenter to concrete station forming the network of the nearest observations we can estimated a degree of the deformation of the near zones at the moment of the initial rupture of the focus of the earthquake and its aftershokes.

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Fig. 1. Scheme of the geological deep cut between Spital: earthquake focus (+ - the place of initial rupture) and a Stepanavan s/station, b- Bogdanovka s/station (1- andezite-dacite lava; 2- sandstone, clay, slate, limestone; 3- clay slate; 4- lava, tuff; 5- crystall slate).
Fig. 2. The values of the longitudinal waves velocities of the aftershocks of the Spitak earthquake: a— for s/station Stepanavan, b— for s/station Bogdanovka.
RELATIONS BETWEEN SEISMICITY AND FRACTURATION IN THE BETIC CORDILLERA

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INTRODUCTION

The Betic Cordillera (BC), together with the Rif, corresponds to the western termination of the Eurasian Alpine chains. To the north of the BC is the Iberian Massif, which forms the nucleus of the Iberian plate, and to the south of the Rif is the Moroccan Meseta and, further south and east, the Atlas Cordillera, which separates Northwestern Africa from the more stable part of the African plate.

The BC is basically made up of two zones: the External Zone, which was formerly the southern and southeastern border of the Iberian Massif, submerged during most of the Mesozoic and Tertiary, and the Internal Zone, which was displaced westwards mainly during the Lower Miocene (Andrieux et al., 1971, Sanz de Galdeano, 1990, etc.) from its original position approximately between Sardinia and Algeria. The Alboran Sea lies between the Rif and the BC, and consists of thinned continental crust, representing the western extension of the important stretching and thinning of the Algerian Basin.

Figure 1: Fracture network of the BC and situation of the epicentres of I ≥ VI earthquakes.

SEISMICITY AND SEISMIC SOURCES

There is moderate seismic activity, as regards the magnitudes obtained, in the BC, Alboran Sea, Gulf of Cadiz and Morocco. Most of this seismicity is superficial, but in certain zones
intermediate earthquakes occur at depths of 40 to 180 km. Surprisingly, three have been recorded south of Granada at depths of over 600 km.

The BC was divided into several types of seismic sources in order to study the superficial seismicity. Thus, those established by Martín Martín (1983) were based only on epicentre distribution. Sanz de Galdeano & López Casado (1988) and Muñoz et al. (1992) carried out divisions of seismic sources in which the main geological features were taken into account, particularly those of recent fracturation. The shape of the sources is thus based on that of those fault sets linked to earthquake groupings. Although this is useful at a certain scale, this procedure has its defects. The lack of homogeneity of the sources means that their seismogenetic characterisation (parameters a and b of the Gutenberg-Richter relation) is imprecise, since in this region there are several densely intertwined fault sets (Fig. 1) with different seismogenetic behaviour.

ACTIVE FAULT SEGMENTS

In order to avoid these defects, we have begun the study of the fault segments that can truly be considered active, correlating them with earthquake epicentres and neotectonic data. After relocation of the earthquakes on 1:200,000 scale maps, the data were fed into a computer together with the information on the known fault network. The positions of I≥VI earthquakes, and also those of ≥3 magnitude, when known, was correlated with the fault network. In some cases the relation was very clear and we marked that fault segment as active from a seismic point of view. Other cases are doubtful and in some the earthquakes cannot be related to any known fault. Part of the results are shown in Fig. 2.

The determination of active fault segments from the neotectonic point of view was mainly carried out on the basis of field data, with special consideration of the faults affecting the Quaternary and Pliocene. Geomorphological and other criteria were also taken into account. Fig.3 gives a graphic representation of the fault segments considered to be active.

We are aware of the difficulties involved in the location of earthquakes and of the incomplete state of knowledge of the fault network. Some important earthquakes cannot therefore be
Figure 3: Highly simplified general scheme showing the main fault segments considered to be active (indicated by thicker lines) on the basis of both neotectonic and seismic data.

explained at present as they were located far from the known faults. Moreover, the data obtained must be complemented with data on microseismicity, geophysical profiles, etc. For this reason the scheme shown in Fig. 3 is not complete as regards both scale and the data themselves. These maps must therefore be progressively improved, but nonetheless they represent a first real basis for calculation of seismic danger by the zonified method, as they allow determination of coherent seismic sources linked to concrete active faults.

THE INTERMEDIATE AND DEEP EARTHQUAKES

The distribution of the intermediate earthquakes can be seen in Fig. 4. The important concentration in the Granada-Malaga-Marocco, with indication of intermediate and deep seismicity and the main faults in the Atlas and the Azores.
al. (1988) by the possible existence of incipient subduction of the Alboran-Algerian crustal segment. However, the earthquakes of the Atlas region are not easily accounted for in this way. Moreover, their position coincides very well with that of the main faults in this area, which allowed the outflow of volcanic material during (and before) the Quaternary. We therefore interpret the earthquakes of the Atlas to be produced by faults affecting the entire lithosphere (Sanz de Galdeano & López Casado, 1988). The same may also be true in the Granada-Malaga region, although the geological context there is certainly different. Similarly, we interpret the intermediate earthquakes in the Azores-Gibraltar area as the result of movement of lithospheric (transformant) faults separating the Iberian plate from the African plate there.

Finally, the three deep earthquakes whose epicentre were located approximately to the south of Granada, are generally explained as being the result of the action of a fossil Wadati-Benioff slab.

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REFERENCES
INTRODUCTION

In general, deformation along the strike of the western Alps is quite regular: P-axes show an orientation roughly perpendicular to the Alpine chain and transcurrent faults prevail in the northern sector of the western Alps. The picture is more chaotic between the Argentera, Dora Maira and Pelvoux Massifs and in the Penninic domain between the Aar-Gotthard and Mont Blanc Massifs. Here, orientations of P- and T-axes published so far are very variable and in part inconsistent (e.g. Eva et al. 1990). In order to check whether these divergent directions of crustal deformation are real or due to poorly constrained data, we have reevaluated focal mechanisms of several earthquakes in the southern Valais and in the border region between Italy, France and Switzerland.

DATA AND METHOD

The analyzed events, with magnitudes from 2.7 to 3.6, occurred between 1985 and 1988. The available data includes seismograms from Swiss, Italian, French and German networks, much of it in digital form. Because of the shallow seismicity and large epicenter-station distances, focal depths of the studied earthquakes are poorly constrained by the usual location methods. Consequently, due also to the complex crustal structure, phase identification and take-off angles of the rays at the source are subject to large uncertainties.

This fact can lead to serious errors in the fault-plane solutions, particularly for mechanisms with significant components of normal or reverse slip. In order to obtain more reliable estimates of the possible focal-depth range of each event and of the consequent effects on take-off angles, we have resorted to 2-D ray-trace modelling. Based on P-wave velocities and Moho depths from previous earthquake and seismic refraction studies in the Alps and northern Alpine foreland, for each event we constructed seismogram sections and corresponding crustal profiles in three different directions: north into the Swiss Molasse Basin, Jura Mountains and Black Forest, as well as ENE and SSW along the strike of the Alps. In the usual location procedures, the origin time is as poorly constrained as the focal depth, and errors in origin time cause a vertical offset that is equal for all seismograms. Therefore, it is not possible to model absolute travel times but only relative arrival times. Focal depths were constrained by modelling arrival-time differences between the direct ray (Pg) and the reflection from the Moho (PMP) or the refraction in the upper mantle (Pn), as well as the cross-over distance of the Pn (fig.1). Because of the updipping Moho and the high station density north of the Alps, these phases were recorded particularly well along the northern profile (Eva,
1992; Deichmann et al., 1992). Having established a possible depth range, the other two profiles were used only to determine the corresponding take-off angles. Due to the remaining uncertainties of the epicentral coordinates, of the crustal velocities and of the Moho depth, a discrepancy of up to 0.3 s between observed and calculated arrivals was allowed in determining the acceptable focal-depth range.

RESULTS

The results of this work show that the studied earthquakes are located in the upper 10 - 12 km of the crust (Table 1.), in accord with the cut-off depth of seismicity at about 15 km, generally observed below the Alps (Deichmann and Baer, 1990; Eva et al., 1990; Roth et al., 1992). The information about focal depths and take-off angles obtained by ray-tracing significantly improved the consistency of the fault-plane solutions. They are, in fact, well constrained by a sufficient number of reliable first motions with good azimuthal distribution (fig.2). Moreover, these solutions are often constrained by refracted arrivals that are not very sensitive to variations of seismic velocities and focal depth. Focal mechanisms of the Mauvoisin and Zermatt events, based on 1-D velocity models, have already been published previously (Nicolas et al., 1990). In both cases, our focal depths are significantly different. Whereas the fault-plane solution of the Zermatt event is quite similar, for the Mauvoisin event we obtain a strike-slip mechanism with a normal component of motion instead of a thrust mechanism. The Vissoie event with an almost pure normal faulting mechanism is located very close to two events (1976.01.29 11:39 and 1982.07.03 04:49) for which Mayer-Rosa and Pavoni (1977) and Jimenez and Pavoni (1984) obtained very similar mechanisms, with a nearly identical orientation of the T-axes. Our results give a very consistent picture of the contemporary deformation within the Penninic nappes of the Valais. The earthquakes in this region (the first six events in Table 1) are characterized by transcurrent or normal faulting mechanisms, with a nearly N-S trending average orientation of the T-axes.

DISCUSSION

The general trend of the Alps in the Penninic domain of the southern Valais is oriented ENE-WSW and thus forms an angle of 60 to 70 degrees with the average direction of the observed T-axes. It is therefore evident that the compressional structures of the Penninic nappes, that were formed during the Alpine orogeny, are presently undergoing extensional deformation and that a significant component of this extension is perpendicular to the Alpine arc. A similar picture, although with a more oblique orientation of the T-axes, can be observed also in the Penninic domain of the eastern part of the Swiss Alps, where normal faulting mechanisms predominate (Roth et al., 1992). In the northern Alpine foreland, on the other hand, we can observe a NNW-SSE direction of maximum crustal shortening and corresponding ENE-WSW oriented extension (Pavoni, 1980; Deichmann, 1992), which is consistent with the large scale stress field expected from the convergence of the African and European plates. This orientation is also confirmed by the fault-plane solution of the Ivrea event, which occurred along the Insubric Line at the border of the Po
Plain. Our observations of extensional deformation across the crest of the Alps is in good agreement with the physical model of the evolution of mountain ranges by Molnar and Lyon-Caen (1988).

This model, based on considerations of mechanic equilibrium only, predicts extensional deformation with normal faulting across the crests of the ranges, while the flanks and lowlands continue to undergo crustal shortening. Thus, the existence of crustal extension in a high mountain range and consequent variations in the deformation style across it do not imply necessarily changes in the regional horizontal stress field. In this sense, our observations of crustal extension across the alpine arc are not in contradiction with the large-scale compressive field that characterizes the Alpine region.

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<td>7.638</td>
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<td>345/21</td>
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Table 1. Focal parameters. Time = GMT; M = magnitude; Z = focal depth range (km b.s.l.); P, T = azimuth and plunge of P- and T- axes.

**fig.1:** Ray-trace model: focal depth is 6 km. Travel-times are reduced with 6 km/s, (+ calculated data; o observed data).

**fig.2:** Scatter plot of the fault-plane solutions (equal area lower hemisphere projection), solid circles: compression, open circles: dilatation.
SEISMOGEOLOGICAL PATTERN OF A TRANSITION AREA BETWEEN THE EAST ALPS AND THE WEST CARPATHIANS

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ABSTRACT and CONCLUSION

All available seismological data (earthquake epicentres, macroseismic observations) and the available structure-tectonical, geomorphological and geodetical data for the transition area of two fundamental orogenetic units of Central Europe - the East Alps and the West Carpathians are analyzed with the aim to find seismogenic zones of this area and to assess their possible earthquake activity. The relationship of small local geological structures to the fundamental European tectonic blocks is discussed from the viewpoint of seismotectonic processes expected for the transition area, especially in the Vienna basin and in the Little Carpathians. The special attention is paid to the inhabited area of the town Bratislava which is situated in the territory under study.

On the basis of geological and geophysical data it is possible to single out several faults or boundaries in the West Carpathians, on which a relatively intensive horizontal displacement can be observed. It concerns the following ones: Verona-Semmering-Little Carpathians-Pováží, Raab-Muráň, Periadriatic boundary-Balaton, and fault zone of central Hungary. As indicated by geophysical data, microstructure measurements, palaeogeographic maps, satellite images of the Earth, etc., these faults governed and influenced the fundamental movement of crustal masses in the domain of the West Carpathians in the direction to NE or N. The faults demarcate the extraordinarily mobile "West Carpathian block". In their southern parts the movements along the faults have mostly dextral character while in the northern parts the movements are sinistral. In the upper part of the Earth's crust, composed of rigid masses of the Inner Carpathian nappes, secondary systems developed gradually. They obviously generated or participated in the origin of Neogene basins of the "pull apart" type, in the segmentation of basement blocks. The whole territory of the Little Carpathians forms a system of differently uplifted or subsided blocks, separated by individual faults. It is not clear whether the whole formation of the Little Carpathians is not affected by movements on the original thrust planes as it can be observed at present on seismic profiles.

The paper will be submitted for publication on TECTONOPHYSICS
A PROPOSAL FOR THE SEISMOTECTONIC ZONATION OF THE HEPERIAN MASSIF

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Most seismotectonic studies aim at giving either a broad view of the Iberian Peninsula and its seismotectonic surroundings, or a close-up of one of the most seismically active zones. This report offers a proposal for the seismotectonic zonation of the Spanish sector of the Hesperian Massif, the Peninsula's basement.

Seismic, geological and geophysical data has been gathered, compared and interpreted. Finally, a set of twelve differentiable zones, each with its own characteristic seismotectonic traits, has been defined and established.

SEISMICITY

Menzuca and Solares' Database (1983) served as a starting joint for our seismological study. The periodical catalogues published by the Instituto Geográfico Nacional (IGN) since 1960 have also proved useful.

The seismological study included:
- doing an earthquake inventory, considering the date (historical and instrumental seisms), intensity and/or magnitude, depth and location.
- epicenter mapping.
- analysis of the general traits of the seismicity.
- interpreting the relationship between the seismicity of the Hesperian Massif and the Peninsula as a whole.

All seisms from the year 1393 to 1989, totalling 281 in all, were considered in this research.

The epicenter map clearly charts the study area's seismicity, roughly dividing the Spanish sector of the Hesperian Massif into three sections: Northern, Central and Southern. The Central Section has scarce seismic density as compared with the other areas, especially the South. As a whole, the study area shows medium-low seismicity, with somewhat greater seismic activity at the southern edge. The distribution of epicenters is directly related to the maximum intensities registered in each of the three zones (Fig. 1).

GEOLOGICAL CONTEXT

From Precambrian to Pennsian sediments can be found in the Hesperian Massif. The materials from the Cadomian cycle are generally represented by very monotonous siliciclastic series accompanied by igneous activity, currently in narrow gneissified bands, with a structure deeply obliterated by the Hercinian Orogeny.

The Hercinian cycle's sedimentation is rather homogenous throughout the Massif and is characteristic of a stable continental platform with slight tilting. Throughout the Upper Devonian and the Lower Carboniferous, a first phase of Hercinian deformation was produced, during which the current accretions occurred. The latter stages of deformation of this Orogeny and the important, mainly plutonic, igneous activity that went along with it, complete the Hesperian framework.

The Alpine Orogenic and neotectonic activity are difficult to detect because of the lack of sedimentary register. During this period, the Hesperian Massif acts as a rigid basement. Its most important structures are large vertical dip-slip faults characteristic of superficial-block tectonics. Its most notable feature is the formation of large tertiary continental basins (Duero, Tajo and Guadiana River Basins).

RECENT TECTONIC ACTIVITY

Since the work area is basically stable, data is scarce and difficult to evaluate. All available information has been considered in this report. The National Neotectonics Map (now being published) has proven quite valuable as a chart of accidents which acted or may have acted in the Neotectonic period, geomorphological lineal and areal anomalies, subsidence or uplift zones, tilting, geothermal springs and quaternary volcanic focal points.

270
HESPERIAN MASSIF MAXIMUM REGISTERED INTENSITIES

LEGEND
- --- LIMIT OF THE STUDIED ZONE
   " --- Topographic subdivision (1:200,000)

Fig. 1

ZONAL SUBDIVISION OF THE HESPERIAN MASSIF

Fig. 2
GEOPHYSICAL DATA

To complete the structural data, and especially to evaluate its in-depth evolution, the available work-scale geophysical data has been analyzed. Gravimetric data has been quite useful, granting the surface structural features a coherence by defining and clearly outlining like-acting bodies.

SEISMOEVTORIC

Superimposing the aforementioned seismic and structural data led to the division of the Hesperian Massif in twelve zones of marked seismotectonic characteristics (Fig. 2).

Zone 1: Fosa Blastomilomita. - This zone coincides with an allochtone tecto-sedimentary unit with important granite intrusions. It is structured by large Hercinian N-S fractures. Mainly located in the south, an active N-S fault family represents the neotectonic activity. Seismic activity is moderate and is distributed throughout the zone, with a greater concentration of seisms in the South.

Zone 2: Northwestern Complex. - This zone is made up of metasedimentary and granitic materials which come into tectonic contact with the large Hesperian zones adjacent to it. The neotectonic activity is represented by two NW-SE faults. There is scant seismic activity; what exists is focused around the Northwestern extreme and may be related to the NW edges of one of the aforementioned faults.

Zone 3: Orense-Alcanices. - This area is surrounded by geologically well-defined zones, though it is both lithologically and structurally varied. The structures are predominantly Hercinian, directed between E-W and SE-NW. There is scant neotectonic activity, manifested in small faults and tilting. An important concentration of geothermic springs to the south of Orense is noteworthy.

Zone 4: Asturoccidental-Leonesa. - This zone corresponds with the Hesperian Massif zonation. Hercinian structure is most conspicuous with the peculiar twist of the "rodilla asturica" (Asturian roller). The Upper Hercinian and Alpine tectonics (E-W and NW-SE) are superimposed on it, with associated small tertiary basins. Neotectonic activity is scarce. Only tilting of the Cantabric coast and small neotectonic faults are detected (Bierzo Basin), caused by the reactivation of large alpine faults. Seismicity is moderate to low, and the distribution of epicenters indicates a Hercinian related activity.

Zone 5: Duero Basin. - This area corresponds with the western and southern edge of the Duero River Basin. It is a great depression filled with mesozoic tertiary and quaternary continental deposits. The block cover tectonics is controlled by the reactivation of the Upper Hercinian fracturization during the Alpine Orogeny. The neotectonic activity is detected on the edges of the basin, with small fractures, fissures and tilting. Seismic data is very scarce. This is one of the Iberian Peninsula's most stable areas.

Zone 6: Zamora-Salamanca Granitoids. - This area is mainly made up of the syn-orogenic and post-orogenic granitic pluton bodies of the Zamora-Salamanca plateau. Neotectonic activity manifests itself in a series of N-S dip-slip faults with metric displacement near the western border of the Duero River Basin. Seismic activity is low in both number and intensity of earthquakes registered.

Zone 7: Northern Centroiberic. - In this zone, the Central System's granitoids and complex metamorphics outcrop. Block tectonics condition the structure, outlined by large vertical Upper Hercinian fractures. Neotectonic activity is reflected in a single N-S fracture located in the center of the zone. Quantified date taken at small neotectonic faults indicate compression near N-S. Seismic activity is very scarce. No epicenters are located in the zone.

Zone 8: Tajo Basin. - This zone reaches from the Tajo River Basin to the Guadiana River Basin. For the most part, quaternary and tertiary river basin deposits make up the zone's materials. No neotectonic activity has been recorded. Seismic activity is very scarce, and is only focused around some epicenters in the vicinity of Madrid and Ciudad Real.

Zone 9: Southern Centroiberic Area. - Thick siliciclastic series from Upper Proterozoic to Mid-Palaeozoic outcrop in this zone with some granitic plutons and small tertiary basins. Large Hercinian Orogeny folds are the predominant tectonics. Available data indicates scant neotectonic and seismic activity. The latter appears in the south, and is due to the intense activity found at the northern edge of the Guadalquivir River Basin.
Zone 10: Badajoz-Pedroches. - The magnetic alinement of Pedroches is the zone's most evident lithologic feature. Here granitoids intrude in paleozoic materials and are covered by a thin layer of neogenous deposits. All the structures and igneous bodies are elongated and concordantly arranged NW-SE, parallel to the regional shear band. Neotectonic activity is scarce. It is limited to two NW-SE faults, provoked by reactivated Hercinian fractures. Seismicity is scant. A maximum density of 2 earthquakes per 20 km occurs.

Zone 11: Costa Morena. - Outcrop materials in this zone are very varied and of a high structural complexity because of the various stages of deformation ranging from Cadomian to Hercinian Orogeny. The main tectonic is a southward movement which is intensely masked by the later stages of fracturization. Neotectonic activity is weak, limited to small dip-slip faults associated with Hercinian fractures. Seismicity is moderate. Like in the previous zones, the southeastern edge is affected by the seismicity of the active northern edge of the Guadalquivir Basin. As many as 7 earthquakes per 20 km occur at the border between this zone and the South Portuguese zone, in a site of high gravimetric anomalies (batholith of La Granada de Rio Tinto). This border is considered an old plate edge from the Hercinian Orogeny.

Zone 12: South Portuguese. - This zone coincides with the borders of the traditional zonation of the Hesperian Massif and in recent geodynamic models is interpreted as an amalgam of the Hesperian Massif. Its deposits are thick Devonian-Carboniferous vulcan-sedimentary series with an intense structurization in nappe-like southward thrust. Neotectonic activity is limited to the extreme south along the Guadalquivir Basin with dip-slip faults and even gentle folds.

CONCLUSIONS

Using geological, structural and seismological data to define the seismotectonic zones, the relationship between the seismicity and the associated geological phenomena can be determined for each zone, including neotectonic activity, pre-neotectonic structures, isostasy, geothermalmism, and areas of subsidence and uplift.

The study zone, according to any criteria, represents slight seismic activity and can be included among the Iberian Peninsula's most stable zones.

The thesis that the Hesperian Massif is a stable block is supported by the fact that the greatest seismicity is registered in the northwestern and southeastern borders, adjacent to the seismically active zones the western border of the Iberian continental crust and the southern border of the Iberian subplate, respectively.

REFERENCES

ON THE SUBDUCTION PROCESS IN THE ZONE PRESENTLY OCCUPIED BY THE CARPATHIANS

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The present age of an element P (Fig.1) of a lithosphere fragment subducted in a given zone is (Shiono and Sugi, 1985)

\[ A = A_c + t_{sb} \]  \hspace{1cm} (1)

where \( A_c \) is the cooling time of the lithosphere that is going to subduct, and

\[ t_{sb} = \frac{x_{sb}}{V_{sb}} \]  \hspace{1cm} (2)

is the subduction time; \( V_{sb} \) is the subduction velocity.

Taking into account a large number of data referring to the world's main subduction zones and the corresponding spreading zones, an empirical relation between the times \( t_{sb} \) and \( A_c \) was obtained as follows (Shiono and Sugi, 1985)

\[ t_{sb} = A_c / 9 + 2. \]  \hspace{1cm} (3)

From (1) and (3) one obtain

\[ A_c = 9/10 (A - 2). \]  \hspace{1cm} (4)

Of course, the above relations only hold good for the case of a still active subduction, uninterrupted by collision stages, mountain formation etc. Consequently, the way these relations were used above and the way graphs were used by Enescu et al., (1989) do not hold for the zone under investigation, since the oceanic lithosphere subduction ended in the Carpathians almost simultaneously with the beginning of the collision and mountain formation stage, then probably some lithosphere relict fragments (such as the Vrancea one) went further down as an effect of gravitation. In this conception, relations (1) and (2) give the age (A) and the subduction time (\( t_{sb} \)), respectively, for the lowest (i.e. oldest) element of the lithosphere fragment (Fig.1) at the moment where the collision process started halting subduction, for a time at least. So the age of the oldest (lowest) element of the lithosphere fragment at the moment when the collision started (and subduction ceased) would now (in present) have been

\[ A' = A + t_{pc} = A_c + t_{sb} + t_{pc} \]  \hspace{1cm} (5)

where \( t_{pc} \) is the postcollision time, i.e. the time interval from the collision beginning up to now. From data given by Bleahu (1989) we estimated

\[ t_{pc} = 80 \cdot 10^6 \text{ years}. \]  \hspace{1cm} (6)

Considering the new situation, relation (4) becomes

\[ A_c = 9/10 (A - 2) = 9/10(A' - t_{pc} - 2) \]  \hspace{1cm} (7)

which together with the relation

\[ A' = A_c + t_{pc} + \frac{x_{sb}}{V_{sb}} \]  \hspace{1cm} (8)

makes up a two-equation system whose unknowns are \( A_c \) and \( t_{sb} \).

The lithosphere which was subducted in the zone presently oc-
cupied by the Carpathians was associated, from the geotectonical point of view, with the distension period in the Carpathians' alpine evolution. The opening of the Tethys ocean, for the Carpathians, can be dated in the Middle Triassic (Șandulescu, 1984) and began by rifting processes, which then evolved in spreading processes. It is considered (Șandulescu, 1984) that there were two distension paroxysms. The first one took place in the Middle and Upper Triassic, and the second one since half of the Middle Jurassic to early Tithonian (the Upper Jurassic). If the oceanic lithosphere which subducted in the zone presently occupied by the Carpathians had really been generated by rifting during the second distension paroxysm, according to an assumption accepted by Enescu et al., (1989), it follows that the present age of this lithosphere would have been \( A' = (140−160) \times 10^6 \) years. From (6) - (9) it results that:

\[
A = (52.2−70.2) \times 10^6 \text{ years} ; \quad t_{sb} = (7.8−9.8) \times 10^6 \text{ years}.
\]

As we don't know either the length \( X_{sb} \) of the lithosphere fragment subducted up to the beginning of collision, or the subduction velocity \( V_{sb} \) of that fragment, but only their ratio, i.e. the time \( t_{sb} \), we shall further discuss two of the most likely cases. Thus, in the first case, we take a very extended length, \( X_{sb} = 700 \text{ Km} \). Taking into account (9), we get \( V_{sb} = 7.1−9.0 \text{ cm/year} \). Knowing that \( A = (60−80) \times 10^6 \) years, we have

\[
V_{sb}/A = 0.089−0.150.
\]

Using the values (10) in the graphs on Figure 2 (see Kostoglodov, 1988; Enescu et al., 1989), we find that, in this case, the maximum possible magnitude of earthquakes (occurred until the time when collision and mountain formation started) in the zone presently occupied by the Carpathians, would have been \( M_W = 8.3−9.1 \) i.e. far more strong earthquakes than those which occur in Vrancea region and comparable instead with present Aleutians and Kamchatka earthquakes. In the second case, we take \( X_{sb} = 350 \text{ Km} \), a length comparable with that in the case of the arc facing the Tyrrhenian Sea. Taking into account (9), we obtain \( V_{sb} = 3.6−4.5 \text{ cm/year} \). Considering the range \( A = (60−80) \times 10^6 \) years, it follows

\[
V_{sb}/A = 0.045−0.075.
\]

Account taken of (11), the graphs in figure 2 show that the highest magnitude of earthquakes preceding the collision stage would have been \( M_W = 7.5−8.1 \), values which are comparable to those of major earthquakes both in Vrancea and in other subduction zones. It follows from the above that for both cases, i.e. \( X_{sb} = 350 \text{ Km} \) and \( X_{sb} = 700 \text{ Km} \), as well as for any other particular value of \( X_{sb} \), the graphs in Figure 2 clearly demonstrate that the subduction zone preceding the collision and Carpathians formation stage must have been an island-arc type zone. A more significant illustration is given in Figure 3 which, besides data from Kostoglodov (1988), also includes several data from Smith (1989). Every possible value of the parameters \( A \) and \( V_{sb} \) (corresponding to the ranges \( A = 60−80 \) million years and \( X_{sb} = 350−700 \text{ Km} \)) for the zone presently occupied by the Carpathians, lies within the trapezium shown in figure 3. We note that here again the zone under investigation proves to fall within the island-arc type category (Fig.3). This conclusion, now based on a correct procedure, is confirming the view that alpine mountain arcs, which are in fact continental
Fig. 1.

- Island arc-type seismic zone
- Continental arc-type seismic zone

M = 3.10 log \( \frac{V_{sb}}{A} \) + 11.59

M = 3.20 log \( \frac{V_{sb}}{A} \) + 11.75

ISLAND ARC-TYPE

CONTINENTAL ARC-TYPE

Fig. 2.

\[ \log \left( \frac{V_{sb} \text{ [cm/yr]}}{A \text{ [Ma]}} \right) \]
arcs, are the result of a collision between island arcs and continents (Dewey, 1970).

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A NEW MODEL REGARDING THE SUBDUCTION PROCESS IN THE VRAICEA ZONE

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From the isobath map of the lower limit of the lithosphere, as calculated by Enescu (1992), the lithosphere’s thickness in the Carpathians area is found to be of 80-85 Km. This bears testimony that the lithosphere fragments, once subducted in the zone presently occupied by the Carpathians have fully (or nearly fully) vanished as a result of their assimilation and at least partial melting into the asthenosphere over the approximately 80 million years that have elapsed since the subduction process was halted almost simultaneously with the beginning of the collision and Carpathians formation process. One naturally wonders how come that Vrancea foci are found to occur up to as deep 200 Km (or even 220 Km), or, in other words, how come the lithosphere is present as far below as 200 Km in the sole Vrancea zone. An attempt will be made in present paper to answer this, assuming that a subduction process, started at a relatively recent date, i.e several million years ago, may still be under way in Vrancea. Thus, what we would have in Vrancea would be an active subduction area of continental arc type, a subduction which may have started after the Carpathians arc were formed and after relics of the old subduction had been assimilated and melted into the asthenosphere. Another argument, perhaps the most convincing one, supporting the assumption of an active subduction in the Vrancea zone, consists of data related to earthquake location and focal mechanism. Figure 1 shows the epicenters of several Romanian earthquakes whose foci lie deeper than the terrestrial crust (under Mohorovičić discontinuity). Subcrustal earthquakes may have expected to confine themselves just to the Vrancea zone (featured as a rectangle in Fig.1), yet seisms whose foci lie below the terrestrial crust are noticed to occur outside this area as well. Whereas the entire territory of Romania has been considered, only that part of it in which subcrustal foci are reported, is shown in Figure 1. Subcrustal foci outside the Vrancea zone are seen (Fig.1) to lie along the Intramoesian fault and on the strip between that fault and the Peceneaga – Camena fault (its north–western extension is included). Several subcrustal foci are evidenced close to the north-western border of the rectangle marking the Vrancea zone of intermediate-depth earthquakes. The findings based on Figure 1 are an important argument supporting the assumption that the lithosphere strip between the Intramoesian and the Peceneaga – Camena faults shifts slowly to the north–west resulting in a subduction and underthrust process in the Vrancea zone, where intermediate earthquakes are located. As this north–western movement involves the lithosphere (especially its lower part) between the two major faults, the earthquakes that occur may be not only crustal ones (along the faults streaking the area), but also subcrustal ones both along the Intramoesian fault and within the strip between the two lithospherical faults (Fig.1). Figure 2 shows the focal mechanism solutions for crustal earthquakes in this area. Mechanism solutions for the Vrancea intermediate earthquakes have not been included in Figure 2, for their being of little use at this stage, in
Zone of the intermediate Vrancea earthquakes.

Fig. 1.
proving the north-western movements of the lithosphere strip between the Intramoesian and Peceneaga-Camera faults. Focal mechanism solutions for the subcrustal earthquakes outside Vrancea zone have not been obtained due to a scarcity of observation data. According to Figure 2, the mechanism solutions for earthquakes occurring along the Intramoesian fault are nearly uniform to indicate a nodal plane that would be parallel to this large fault, suggesting that the rupture in the foci of these earthquakes occurred along the Intramoesian fault and that this fault is active with a seismic activity caused by the slow north-western movement of the block comprised between this fault and Peceneaga-Camera fault. The rest of the mechanism solutions in Figure 2 correspond to crustal earthquakes genetically associated to the fault field in the area, with the activity of these faults more or less involving the slow north-western movement of the lithosphere strip between the Intramoesian and Peceneaga-Camera faults. The two major faults are seen to extent (Fig.3 - Sândulescu, 1984) to the south-east, reaching close to the Anatolian fault. Moreover, the location of several earthquakes (occurred in only a couple of years) along the south-eastern fragments of the Intramoesian and Peceneaga-Camera faults proves once again these faults as being active. The slow north-western movement of the lithosphere strip between the two major faults is likely to be caused by the Anatolian subplate thrusting on the Black Sea subplate, even as the former is pressed in turn by the Arabian plate. An obvious effect of the Anatolian subplate's pressure on the Black Sea subplate is the slow north-western movement of the lithosphere strip between the Peceneaga-Camera and Intramoesian faults. The north-western movement of the lower part of the lithosphere and, implicitly, the subduction of this have an influence on the crustal lithosphere which undergoes underthrusting motions of much lower velocity. Several observation data confirm underthrusting of the terrestrial crust in the Eastern Carpathians bend area, following a south-east to north-west direction, at a rate of 0.3-1.0 cm/year (Lăzărescu and Popescu, 1983). As for the subduction velocity, it can be estimated to be $V_{ob} = 5.0$ cm/year (Enescu, 1985). Knowing the length of the subducted lithosphere fragment, the subduction can be assessed as having started some 2 to 3 million years ago. As a result of stresses associated to subduction process, subcrustal earthquakes occur not only in the subducted lithosphere fragment, but also in a surrounding area in the microplate under which the subduction takes place (Fig.1).

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EXPLOSION-GENERATED SHORT PERIOD SURFACE WAVE DISPERSION: NOISE STUDIES AND Q ALONG SEISMIC PROFILES IN SOUTHERN SWEDEN

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ABSTRACT

The short period Rayleigh wave dispersion was determined along linear seismic arrays in southern Sweden, using data from the FENNOLORA and the EUGENO-S seismic refraction project on the Baltic Shield. The investigated area consists of crystalline and metamorphic igneous rocks of precambrian age and has essentially no sedimentary cover. The dispersion was grouped into dispersion regions and the regional shear velocity structure was determined down to 2-3 km, in one case to 6 km. Noise studies were undertaken in order to separate the effects of the heterogeneities from the experimental uncertainties. It was shown that the estimates of the shear velocity resolution could be significantly improved in comparison to conventional estimates. Q-values were determined from the SP Rayleigh waves in each dispersion region. In a broad sense the same Q-values were found in the largest part of the area, with lower values in the north-western corner only. Here the lowest shear velocities were also found. The Q-values were inverted to preliminary Qg-structures. The $Q_g$-value of the uppermost km of the crust was 110 in the dominating part of the area and 55 in the north-western corner. The lowest Q-values could be modelled with the same $Q_g$-structure below 1 km as in the other parts of the area.

Contribution as a whole has been sent to Tectonophysics for publication.
SEISMIC VELOCITIES AND Q-FACTORS IN THE UPPERMOST CRUST ALONG THE SVEKA PROFILE IN FINLAND *)

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The deep seismic sounding profile SVEKA located in central Finland was performed to study the deep structure of the Fennoscandian shield in the contact zone of Archaean and Svecokarelian provinces. Besides of the P- and S-waves well developed Rayleigh surface waves with periods of 0.5-1.5 s were recorded from all shot points. At the filter band 0.5-2.5 Hz the R-waves were observed clearly until the end of 320 km long profile. 2-D interpretation of data by Grad and Luosto (1987) was used as a base for this study. Amplitudes of P-, S- and R-waves were corrected for charge size. Attenuation coefficients and Q-factors were determined from the records of body waves using the dynamic ray tracing method (Červený and Pšeničk, 1983) and from the records of R-waves by calculating synthetic seismograms for 1-D velocity models using the reflectivity method (Kind, 1978). In the uppermost 1 km the quality factors Q_p and Q_s vary from 50 to 120, reaching higher value 140 in the marginal zone of the Archaean basement. For the depth range of 1-2 km Q-values of 80-400 were found. For the depth range 2-6 km Q-values of 300-800 were found for the Svecokarelian province and 200-400 for Archaean. The zones of strong attenuation of R-waves coincide with fault and schist zones, which are well known from surface geology.

![Graph](image)

**Figure.** Compilation of Q-values in uppermost crust beneath the SVEKA profile. Dotted area - data from 2-D modeling of body waves, thick solid lines - data from 1-D modeling of R-waves using the reflectivity method.

**References:**

*) **Contribution as a whole has been sent to Tectonophysics for publication**
Investigations of surface wave dispersion along several profiles in Central and Eastern Europe have been described in Proskuryakova et al. (1981) and Novotný et al. (1980). The present contribution deals with similar investigations for the profile Uppsala-Prague.

The profile Uppsala-Prague is situated on three large geological units: Baltic Pre-Cambrian Shield, Polish Palaeozoic Platform, and Bohemian Massif. The length of the profile is approximately 1110 km.

Seismograms from long-period seismometers at the stations Uppsala and Prague have been used for studying surface wave dispersion between these stations. Many earthquakes were selected for this purpose, but the best results were obtained for the following Italian earthquakes: Central Italy 19.9.1979, Sicily 28.5.1980, Southern Italy 7.5.1984 and Southern Italy 11.5.1984. The corresponding experimental values of phase velocities of Rayleigh and Love waves for this profile are shown in Fig. 1 by the isolated points; the usual graphic method has been used for the determination of these dispersion curves.

In interpreting dispersion curves the medium between the seismic stations is usually approximated by a layered, horizontally homogeneous model. However, here we use a more complicated model, composed of several layered blocks. Using results of previous seismic investigations we have divided the structure along the profile Uppsala-Prague into the following blocks: Uppsala-Oskarshamn, Oskarshamn-Karlskrona, Baltic Sea (northern and southern blocks), Tornquist-Teisseyre fault zone, Polish Palaeozoic Platform, and Bohemian Massif. Here we shall not give the parameters of the individual blocks. The theoretical dispersion curves of Rayleigh and Love waves for these blocks have been computed by standard matrix methods. Then the theoretical dispersion curves for the whole profile Uppsala-Prague have been constructed (see the procedure in Novotný et al. (1980)). These curves are shown in Fig. 1 by the solid lines. A good agreement between the experimental and theoretical dispersion curves is seen in this figure. Therefore, we can conclude that the investigations of surface waves confirm the other seismic results in this region.

The full text of this contribution will be submitted for publication in Tectonophysics.

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Fig. 1. Dispersion curves of Rayleigh waves (R) and Love waves (L) for the profile Uppsala-Prague.

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TOMOGRAPHIC DETERMINATION OF THE UPPER AND MIDDLE CRUST STRUCTURE IN THE LUANXIAN EARTHQUAKE REGION OF NORTH CHINA

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ABSTRACT

The three-dimensional velocity structure of the upper and middle crust in the Luangxian earthquake region of North China was obtained by the joint inversion of travel-time residuals of P waves generated by explosions and earthquakes. A temporary seismic station array, consisting of 88 analogue and digital recorders with three components, was arranged on a 30X40 Km² region for receiving the PmP and SmS waves reflected from the Moho at the critical distances generated by 6 shots from the different azimuths. Local earthquake data recorded by 12 stations located on the same region were also used. A joint inversion technique without blocks of explosion and earthquake data was developed to obtain the tomographic determination of the P velocity structures under the seismic array from depth of 1Km until the depth of 9Km and to modify the hypocenter determination of the earthquakes simultaneously. The results reveal that the velocity structures of the upper and middle crust in this region are laterally heterogeneous. The strip-shaped low-velocity and high-velocity distributions exist in the upper and middle crust. It seems that they are related to the surface tectonic features. The resulting seismic images also reveal that this local heterogeneity of velocity structures is possibly associated with the earthquake distribution in this region.

Contribution as a whole has been sent to Tectonophysics for Publication.
NEW LITHOSPHERE INFORMATION FROM SEISMO - GRAVIMETRIC MODELLING IN THE
MOESIAN PLATFORM (ROMANIA)

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ABSTRACT *)

A line with a NE - SW orientation has been recorded across the Moesian Platform and the Getic Depression using the reflection seismic method. This line is placed westward from C.aiova city and has a total length of 116 km. The gravity image of the Carpathian foredeep between Danube and Olt rivers, such as appears on the Bouguer anomaly map consists of a minimum zone. This minimum represents the important mass deficit, that consists of sediments of different ages, being juxtaposed, in a important contrast of density with the crystalline scaffolding of the Carpathians (in the northern part) and the mesozoic and palaeozoic deposits of the Moesian Platform (in the southern part). The sedimentary domain was separated in three sequences of higher reflectivity: Neogene, Mesozoic and Palaeozoic sequences. The base of the sedimentary layer could not be precisely drawn due to the advanced alteration of the crystalline basement surface. This limit has been established between 3 and 4s by correlation with other geological data. The Conrad limit was drawn taking into consideration some short and low intensity reflections at about 17 km depth. The structure of the Moho discontinuity appears to be a monocline tilting from SW to NE, having 31 - 34 km depth. A bidimensional gravimetric modelling has been carried out in order to verify the former crustal section through a comparison between the observed gravimetric anomaly and the computed one. The following equation:

\[ \rho = 2.18 + 0.07 V_p \]

has been established for the sedimentary cover. The values of the density have been adapted for the crustal layers as follows: upper crustal layer 2.67 g/cm³; lower crustal layer 2.87 g/cm³; upper mantle - 3.17 g/cm³. The programming peculiarities led to a reduced model with nine formations having different densities that permitted some simplifications of the geological section. As we found in the model obtaining process, the geological formations, have a reduced amplitude effect as support.

The information on the deep seismic reflection line showed the image of a lithospheric structural model in this zone. The thickening of the crust in the platform zone to the orogenic domain has been verified. The generally reduced thickness of this type of continental crust (30 - 34 km) is characteristic for the platforms and pre-orogenic depression. Our model has been confirmed by the gravimetric modelling after few iterations, the first operation being a gravimetric uncover that led to the effect evaluation of the geological section used as an initial model. A good reproduction of the observed gravimetric curve has been obtained accepting a high density for the metamorphosed rocks belonging to the Danube Domain.

*) Contribution as a whole has been sent to Tectonophysics for publication.
LITHOSPHERE STRUCTURE IN ROMANIA. I. LITHOSPHERE THICKNESS AND AVERAGE VELOCITIES OF SEISMIC WAVES P AND S. COMPARISON WITH OTHER GEOPHYSICAL DATA.

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ABSTRACT

On the basis of the seismic events, which occurred in Romania and in a few neighbouring areas in the period 1982-1985 (at depths of $C \leq h \leq 10$ Km) and were recorded by telemetered seismic stations of the Romanian network, the travel-time curves for seismic waves $P$ and $S$ reflected on the boundary $L$ between the lithosphere and the asthenosphere were obtained. These travel-time curves allowed to determine the average velocities of the $P$ and $S$ waves ($V_{\text{p}}$ and $V_{\text{s}}$) in the lithosphere and to estimate the lithosphere average thickness $h_{\text{l}}$ in different zones of Romania. The results obtained are the following:

<table>
<thead>
<tr>
<th>Nr.</th>
<th>Zone</th>
<th>$V_{\text{ml}}$(Km/s)</th>
<th>$V_{\text{mS}}$(Km/s)</th>
<th>$h_{\text{lm}}$(Km)</th>
</tr>
</thead>
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<td>1</td>
<td>Apuseni Mountains</td>
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<td>3.95</td>
<td>76</td>
</tr>
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<td>2</td>
<td>Transylvanian Depression</td>
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<td>3</td>
<td>Carpathians and Subcarpathians</td>
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<td>4</td>
<td>Central and Western Moesian Platform</td>
<td>7.05</td>
<td>4.00</td>
<td>72</td>
</tr>
<tr>
<td>5</td>
<td>Area between the Peceneaga - Camena fault and Intramoesian fault</td>
<td>6.70-7.00</td>
<td>3.80-3.97</td>
<td>68-80</td>
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<tr>
<td>6</td>
<td>Moldavian Platform and Northern Dobrudja</td>
<td>7.15</td>
<td>4.10</td>
<td>100-105</td>
</tr>
</tbody>
</table>

In the Eastern zone of the Pannonian Depression, $h_{\text{l}}$ = 60 Km. Knowing the distribution of depth values of the boundary $L$ as calculated for each reflection point, the isobaths of this boundary could be plotted approximately. A most significant result of this lithosphere thickness determination is its providing evidence of the Peceneaga - Camena fault (including its north-west extension) and of Intramoesian fault. A very important conclusion we can draw is that the two faults are lithospheric faults rather than crustal ones. Lithosphere thickness in the zone of intermediate Vrancea earthquakes is 90-200 Km, which has been adopted on account of earlier researches, which assumed an lithosphere fragment subducted in the Northern Carpathians bend area. Differences in lithosphere thickness between the zone of the Moldavian Platform (including the Northern Dobrudja) and the neighbouring ones are relatively large which clearly underscores the presence of the Peceneaga - Camena fault along with its north-west extension, and define it undoubtedly as a lithospheric fault. In addition, such differences in depth bear evidence of differences in point of physical and geological characteristics between the old platform zone and the Carpathian orogene zone. The results were compared with those obtained from heat flow and magnetotelluric sounding data.

x) Contribution as a whole has been sent to Tectonophysics for publication.
The same as in the first part of the study (Paper I), recordings of the seismic events occurred in Romania and in a few neighbouring areas, as obtained at telemetered stations in the Romanian network, have been used. The results given in this paper were obtained from data related to the quasi-continuously refracted waves \( P_c \) and \( S_c \), the refracted waves \( P \) and \( S \) and the \( P_p \) and \( S_p \) waves reflected on the Mohorovičić discontinuity; data on the \( P_p \) and \( S_p \) waves were used in the paper I. The travel-time curves of the \( P_c \) and \( S_c \) waves allowed to determine the following average velocities \( V_{p_c} \) and \( V_{s_c} \) of these waves and the average depth \( h_{Mn} \) of the Mohorovičić discontinuity:

<table>
<thead>
<tr>
<th>Nr.</th>
<th>Zone</th>
<th>( V_{p_c} ) (Km/s)</th>
<th>( V_{s_c} ) (Km/s)</th>
<th>( h_{Mn} ) (Km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Apuseni Mountains and Transylvanian Depression</td>
<td>6.11</td>
<td>3.52</td>
<td>33</td>
</tr>
<tr>
<td>2</td>
<td>Carpathians and Subcarpathians</td>
<td>6.18</td>
<td>3.51</td>
<td>45</td>
</tr>
<tr>
<td>3</td>
<td>Moesian Platform</td>
<td>6.21</td>
<td>3.50</td>
<td>34</td>
</tr>
<tr>
<td>4</td>
<td>Moldavian Platform and Northern Dobrudja</td>
<td>6.22</td>
<td>3.55</td>
<td>43</td>
</tr>
<tr>
<td>5</td>
<td>Eastern zone of the Pannonian Depression</td>
<td>-</td>
<td>-</td>
<td>25-30</td>
</tr>
</tbody>
</table>

Knowing the distribution of depth values of the Mohorovičić boundary as calculated for each reflection point, the isobaths of this boundary could be plotted approximately. The Peceneaga-Camena fault (including its north-western extension) is just as clearly revealed by a leap of 7-8 Km. The Intramoesian fault is marked at the \( M \) discontinuity level by a slight inflexion of the isobaths. The travel time curves of the \( P_n \) and \( S_n \) waves were plotted and the propagation velocity immediately under the \( M \) discontinuity was evaluated. Results of this assessment show that the values of this velocity in all zones investigated to lie around 7.8 Km/s for the longitudinal wave and 4.4 Km/s for the transverse one. The travel-time curves of the \( P_c \) and \( S_c \) waves were plotted and the Wiechert-Herglotz method was applied to determine the laws governing the variation with depth of the propagation velocities \( V_p(Z) \) and \( V_s(Z) \) across the Earth's crust. The distribution of average velocities \( V_{p_c}(Z) \) and \( V_{s_c}(Z) \) in depth was calculated using the known relation between these velocities and the \( V_p(Z) \) and \( V_s(Z) \) velocities. It was possible also to determine the variation laws of propagation velocities of the waves \( P \) and \( S \) in the subcrustal lithosphere (under the Mohorovičić discontinuity). Knowing the velocities of the \( P \) and \( S \) waves made it possible to calculate the Poisson coefficient, whose values range between 0.24 and 0.28 for the crustal lithosphere and from 0.25 to 0.27 for the subcrustal one.

\* Contribution as a whole has been sent to Tectonophysics for publication.
DEEP STRUCTURE OF THE VRANCEA ZONE

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ABSTRACT*^)

The observation data used in this paper are those which were recorded at Vrancea station from 1980 to 1987 and referred to 292 Vrancea earthquakes whose epicenters were located in an area between 45°.7 - 46°.1 northern latitude and 26°.5 - 26°.9 eastern longitude, at depths ranging from 63 Km to 160 Km. The area was selected so that distances from epicenters to Vrancea station do not exceed 0°.2 or 22 Km. The propagation times of the waves P and S are plotted in terms of focus depth h which is relatively large compared with epicentral distances. These hodographs t(h) allowed to determine the average velocities of the waves P and S below the Mohorovičić boundary. The results are: \( V_{2mp} = 8.19 \text{ km/s} \), \( V_{2ms} = 4.46 \text{ km/s} \), \( K_m = V_{2mp}/V_{2ms} = 1.836 \), \( \sigma_m = 0.302 \). Using the same hodographs t(h) it was possible to calculate the curves \( V_{m(h)} \) and \( V_{mS(h)} \), for \( 63 \leq h \leq 160 \text{ km} \), where \( V_{mP(h)} \) and \( V_{mS(h)} \) are average velocities which characterise the medium between Earth surface \((h=0)\) and any depth within the range \( 63 \leq h \leq 160 \text{ km} \). The distribution of the real and average velocities in the Earth crust and the thickness of the crust in Vrancea region was determined on the base of results obtained by Enescu et al. (Paper II). For depths larger than the depth \((h=50 \text{ km})\) of the Mohorovičić discontinuity we employed the ratio \( K \) whose value had been known beforehand, as well as several data on \( V_p \) provided by Cekunov and which allowed us to obtain the curves of the real velocities \( V_p(h-50) \) and \( V_S(h-50) \). We notice that the lithosphere part comprised between \( h = 50 \text{ km} \) and \( h = 160 \text{ km} \) is characterised by average velocities of around 8.2 \text{ km/s} and 4.5 \text{ km/s}, respectively. These values are in excellent agreement, practically identical, with those above expressed which were derived from different data through an entirely different procedure. Consequently, the results from the \( V_p(h-50) \) and \( V_S(h-50) \) are typical of the subcrustal lithosphere. It is also worth mentioning that the propagation velocities immediately below the Mohorovičić discontinuity are of about 7.8 \text{ km/s} for the P wave and about 4.3 \text{ km/s} for the S wave, values which are practically identical to those obtained from different data and through a different method in Paper II. The Mohorovičić discontinuity is seen to be marked by velocity contrast of about 0.8 \text{ km/s} for the P wave and around 0.2 - 0.3 \text{ km/s} for the S wave. These contrasts are, however, lower than expected. The finding might prove that the Mohorovičić discontinuity is not a clearcut separation, but rather a transition zone whose thickness is of order of kilometers and which the methods we applied failed to reveal distinctly.

*^) Contribution as a whole has been sent to Tectonophysics for publication.
NEW SEISMIC REFRACTION DATA IN THE EASTERN ROMANIAN TERRITORY

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ABSTRACT *)

Some new data concerning the terrestrial crust structure in the eastern part of the Romanian territory are presented. The seismic records have been carried out on four refraction seismic profiles, two of them being situated in the Moldavian Platform and the others in Dobrudja. Especially, the reflected wave groups from the bottom of the crystalline basement and from Conrad limit were put into evidence. The main contribution of these studies consists in the configuration of the Moho surface. Thus, the crust becomes thinner in the Moldavian Platform from West to East, having a thickness of 50 Km in the Eastern Carpathian Orogen and 39 Km in the southern part of Iași city, near the Prut River. On a North-South direction, the crust is thicker towards South, from 40 Km thickness on the Roman-Iași line, to 47 Km North of Galați. The Northern Dobrudja has a crust of 41 - 43 Km thickness, while in the southern part of the Peceneaga-Camena fault, the crust becomes thinner, from 38 Km to 30 Km, in the southern part of the Cernavoda-Constanta line.

The structural sketch shows a new model of the Earth crust in front of the Eastern Carpathians Arc Bend. It underlines three sectors of the crust with different evolutions which give a lifted block character to the sector between the parallels 45° and 46° N and the meridians 27° and 28° E in respect with an East-West direction.

*) Contribution as a whole has been sent to Tectonophysics for publication
CHARACTERISTICS OF ROMANIAN LITHOSPHERE FROM DEEP SEISMIC REFLECTION PROFILING

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ABSTRACT *

The seismic reflection studies for the knowledge of lithosphere structure started in Romania in 1976 in connection with prospecting for hydrocarbons. For this reason, the reflection lines have been recorded with an acquisition technology proper for the sedimentary domain study (short spreads and lines, reduced dynamite charges, etc.). In addition, the recorded time has been lengthened to 10 - 17 s TWT. The lines are uniformly distributed on the Romanian territory, ones of the major tectonic units being better represented.

Tectonically, the Romanian territory consists of prealpine cratons and alpine orogenic regions. The cratonic domain contains platform-type units (the Moldavian and Moesian Platforms) that are characterized by an old folded basement with a sedimentary cover above it and folded cratonic units (the Central Dobrudja and North Dobrudjan Orogen).

The alpine orogenic structures belong to the central and south-eastern European alpine area, containing more sectors with relative individual evolutions (the Eastern and Southern Carpathians and Apuseni Mountains). The alpine movements caused a strong subsiding of the platforms under the Carpathian Orogen. The Transylvanian and Pannonian Depressions are two internal basins formed in the Late-Jurassic period.

The Moesian Platform and the Carpathian Foredeep generally show a reflective sedimentary cover and a relatively transparent upper and median crust with some dipping events that evidence a brittle medium. The reflectivity increases with depth in the south-western and central part, suggesting an alpine tectono-thermal influence in this area. Moreover, a crust-mantle transition zone with 2-3 distinct and prominent reflections (at 9-11 s TWT), confirms this hypothesis. In the northern part of the central and eastern sectors of these tectonic units, the reflectivity decreases with depth, Moho being poorly distinguished. This zone has not been probably affected by the tectono-thermal effects of the alpine orogeny. Some evident subcrustal reflections in the 13-15 s TWT are also evidenced.

The transition area between the Eastern Carpathians and their molasse presents a weak reflective upper crust including dipping events that suggest possible nappes inside the crystalline basement on a brittle background. The median crust is well individualized through short and dipping westward reflections that could be generated by a shear zone. The lower crust is especially reflective on the margin of the Moldavian Platform domain, suggesting the tectono-thermal influence of the alpine orogeny. The subcrustal reflectivity is also present.

The Transylvanian Depression and the Romanian sector of the Pannonian Depression show a thin and reflective sedimentary cover (1-4 km). The upper crust is weak reflective and frequently presents dipping events. The median crust appears to be reflective in the Pannonian Depression and less reflective in the Transylvanian Basin. The lower crust does not show prominent reflections, this fact being in accordance with the low values of the heat flow (35-40 mW/m²) in the Transylvanian Depression but in a disagreement with the high values of the heat flow (~70 mW/m²) in the Pannonian Depression.

*) Contribution as a whole has been sent to Tectonophysics for publication
POSSIBLE LINK OF REFLECTORS IN THE VICINITY OF
THE URALSKAYA SUPERDEEP DRILLING SITE WITH STRESS FIELD

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The Uralskaya Superdeep drillhole is planned to reach 15 km depth. It is being drilled in highly deformed volcanogenic mainly mafic formations. There are two principle difficulties for application of seismic methods for the study of deep structures in this region.

The first problem is related to the character of the recorded wavefield which presents total effect from various reflecting elements often located aside of the profile. Reflections obtained at about the same recording time interfere destructively with each other and make the use of traditional seismic survey impossible for processing seismic section and analyzing the elastic waves velocities.

The second problem is that the volcanogenic formations of mafic composition with high velocities are present on the surface. In such media the velocity analyses requires long-offset data with maximum offset on the order of the depth of objective.

Seismic studies by controlled directional excitation in the vicinity of the Superdeep drilling site were carried out taking into account both problems and were based on the experience of preceding regional and detailed seismic works. Such approach realized during field observation and processing opens new opportunities for the study of the borehole areas by seismometric methods.

Deep reflected waves separated according to the approach directions are much more informative and travel time curves are much longer than in the initial interfering wave field. The system of direct and reverse observations with multiple coverage used along the profile allows not only to trace the main reflections under the surface but to calculate the seismic velocities in the drilled horizon.

The interpretation is made by comparison of seismic section with geological data on the surface and within the drillhole, different kinds of logging, acoustic emission and local seismicity. This interpretation shows that the strong reflectors in the vicinity of the Uralskaya Superdeep drilling site can not be explained only by tectonics and lithology. High reflectivity zones are also the images of highly stressed zones.

REFERENCES


MINERAL PHYSICS AND SEISMOLOGY

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INTRODUCTION

The purpose of this paper is to discuss the symbiotic relationship between mineral physics and seismology and to emphasize the need for more integrated and coordinated studies to achieve progress in our understanding of the Earth's interior.

Mineral physicists study the fundamental physical and chemical properties of minerals and rocks using a variety of experimental and theoretical techniques, some of which are imported from condensed matter physics, solid state chemistry and materials science. We focus on the silicates, oxides, metals and fluids which are major constituents in the Earth and on properties such as elasticity, equations of state, rheology, thermal and electrical conductivity, crystallography, phase equilibria, and thermodynamics. Mineral physics also has strong relationships with the fields of experimental petrology and geochemistry. We apply these studies to Earth problems from the atomic to the global scale; thus, we must consider pressures from 1 bar to more than 3.5 Megabar (350 gigapascals), temperatures from 300 to more than 4000 K, and densities from 1 to almost 15 g/cc.

Fig. 1. Preliminary reference Earth model (PREM) of Dziewonski and Anderson (1981) showing variation of P and S wave velocities and density with depth for (a) entire Earth and (b) upper mantle and transition zone.
Seismology provides Earth models in terms of velocity-depth and density-depth profiles (e.g., Fig. 1a). Mineral physics provides the code via which we hope to interpret these seismic models in terms of variation of pressure, temperature, mineralogy and/or chemical composition. Of particular interest to global geodynamics is the transition region between the upper and the lower mantle, which is characterized by discontinuities near 410, 520, and 660 km and/or high gradients of velocity and density (Fig. 1b). A key question in this regard is whether these discontinuities and gradients can be explained by adiabatic compression and isochemical phase transformations in upper mantle mineral assemblages; if this is not so, then we must conclude that the upper and lower mantle are chemically distinct now and, presumably, for some considerable period in geological history.

Our approach is to measure the elasticity of mantle minerals in the laboratory as a function of pressure and temperature from which data we can construct profiles of velocity and density vs. depth to compare with those observed from seismology. In this paper, we will focus on recent efforts to determine the pressure derivatives of the high-pressure phases of olivine and to apply these data to interpretations of the transition zone. Both seismologists and mineral physicists have used an olivine mantle to model this important region of the Earth (e.g., Revenaugh and Jordan, 1991; Akaogi et al., 1989).

MINERAL PHYSICS EXPERIMENTS

Two recent technological advances in our laboratories have enabled major progress in this field. Firstly, at Stony Brook we have developed techniques to hot-press polycrystalline specimens at pressures of 20 GPa and temperatures of 2000°C in a 2000-ton uniaxial split-sphere apparatus (Liebermann and Wang, 1992). This multi-anvil apparatus is complimentary to the diamond-anvil cell and is distinguished by the ability to produce large-volume specimens in an environment which permits adjustment and control of the sample environment (e.g., T and P and their gradients, deviatoric stress and state of strain, and oxygen fugacity).

Cylindrical specimens (millimeter dimensions) of the beta and spinel phases of Mg$_2$SiO$_4$ have been synthesized in their stability fields in runs of 1-4 hr duration and recovered at ambient conditions by simultaneously decompressing and cooling along a computer-controlled P-T path designed to preserve the high-pressure phase and to relax intergranular stress in the polycrystalline aggregate (Gwanmesia et al., 1990a; Gwanmesia and Liebermann, 1992). These specimens were characterized by X-ray, optical, scanning and transmission electron microscopy, and ultrasonic techniques, and found to be single-phased, fine-grained (<5 micron), free of microcracks and preferred orientation, and to have bulk densities greater...
than 99% of X-ray density. Grains are uniform in size throughout the specimens and have boundaries that are well-sintered with no observable pores or cracks at the TEM scale. Ultrasonic measurements of the P-wave and S-wave velocities at atmospheric pressure demonstrate that these polycrystals are elastically isotropic and exhibit velocities within 2% of the Hashin-Shtrikman averages calculated from the single-crystal elastic moduli.

The second technological advance has been made at the Australian National Observatory with the development of a buffer rod technique for measuring elastic wave travel times at ultrasonic frequencies using the phase comparison method on jacketed polycrystalline specimens (Rigden et al., 1988, 1992a; Niesler and Jackson, 1989). Application of this technique to the hot-pressed specimens of the high-pressure polymorphs of Mg$_2$SiO$_4$ in a liquid-medium piston cylinder apparatus at room temperature and pressures up to 3 GPa yields pressure derivatives of the velocities and elastic moduli (Gwanmesia et al., 1990b; Rigden et al., 1991). These measurements clearly demonstrate that these polycrystalline specimens are not only elastically isotropic at room conditions but also at elevated pressures (Rigden et al., 1992a). We can, therefore, use these data with confidence in comparisons with seismic models of the mantle.

APPLICATONS TO EARTH'S INTERIOR

410 km discontinuity

The new data for the pressure dependence of the elastic moduli for the beta phase provide a firmer basis for a discussion of the magnitude of the 410-km velocity discontinuity than has hitherto been possible (Gwanmesia et al., 1990b). A number of recent upper mantle body wave studies indicate that the velocity contrast across the discontinuity near 400 km depth is 3.8 to 4.9% for P waves and 4 to 5% for S waves (Nolet and Wortel, 1989; see also Kennett, 1991). There is thus no clear requirement that the velocity contrasts for P and S waves to be different; therefore, we choose for illustration the results from two models based on synthetic body-wave studies with velocity contrasts of 4.6% for both P and S waves (S-25 for P waves from Le Fevre and Helmberger, 1989; and TNA for S waves from Grand and Helmberger, 1984).

At ambient conditions, the olivine to beta phase transformation results in an increase of ~13% in velocity for both P and S waves (Fig. 2). If the magnitude of this contrast were preserved at high pressure and temperature, matching the size of 410-km discontinuity would require ~35% olivine in the mantle. However our experimental data demonstrate that this velocity contrast decreases with pressure with the consequence that an olivine content of 42 to 51% can be inferred for the model mantle with \( \Delta V_P = \Delta V_S = 4.6\% \) at 410 km (Gwanmesia et al., 1990b).
Fig. 2. Percentage change in velocity across the olivine-beta and beta-spinel phase transformations in Mg$_2$SiO$_4$ for both P and S waves as a function of pressure at ambient temperatures.

For a more complete analysis of this problem, the temperature derivatives of the elastic bulk (K) and shear (G) modult are also required. Because these are yet to be measured and can probably not be estimated with sufficient precision, the approach we adopt is to determine plausible combinations of $dK/dT$ and $dG/dT$ for the beta phase and various olivine contents of the model mantle that provide a quantitative explanation of the magnitude of the observed discontinuity near 400 km, under the assumption that this discontinuity is due solely to an isochemical transformation from olivine to the beta phase. Allowable combinations of olivine content (%), $dK/dT$ and $dG/dT$ (both in GPa/K) for the observed seismic contrasts include (65%, for -0.018, -0.020), (55%, for -0.015, -0.018), and (45%, for -0.012, -0.016).

520 km discontinuity

Suggestions have also been made that the higher pressure beta to spinel phase transformation could be responsible for the seismic discontinuity which is sometimes observed at ~520 km depth. Calculation of the pressure dependence of P and S wave velocities for an olivine mantle to pressures equivalent to this depth in the Earth (Fig. 2) indicates that the effect of pressure alone is to decrease slightly (to below 1%) the small velocity contrast that exists between the beta and spinel phases at ambient laboratory conditions (Rigden et al., 1991).

In view of the uncertainty surrounding the temperature derivatives for the beta and spinel phases, calculation of the amplitude of the velocity
discontinuity in the vicinity of 520 km depth from the beta-spinel transformation was carried out using a range of plausible temperature derivatives in combination with the measure temperature derivatives (Rigden et al., 1991, 1992a). At 18 GPa, discontinuities associated with beta-spinel transformation in the olivine component are predicted to be: \( \Delta V_P = 1-2\% \), \( \Delta V_S = 0.8-1.5\% \), and \( \Delta \rho = 2.5-3\% \). The magnitudes of these discontinuities can be changed dramatically only if the temperature derivatives are substantially (>10\%) different for the beta and spinel phases. These values will be reduced by the dilution with non-olivine components amounting typically to 40-60\%. Furthermore, recent experimental work on the system \((\text{Mg,Fe})_2\text{SiO}_4\) (Akaogi et al., 1989; Katsura and Ito, 1985) suggests that the transition occurs over a depth interval of ~50 km for the \((\text{Mg}_{0.9}\text{Fe}_{0.1}\text{SiO}_4)\) composition appropriate for the mantle.

Although the velocity discontinuities expected at this depth from the beta-spinel transformation are small, the effect of the density contrast is to produce impedance contrasts of 2.5-5\% for P waves and 3-4.5\% for S waves. In many cases, there is no compelling observational evidence for a velocity discontinuity near 520 km depth from short-period and refraction data; see, for example, the seismic models S25 and TNA in Fig. 3.

![Fig. 3. Comparison of velocity models from the Earth's surface to 600 km depth derived from synthetic body-wave studies and mineral physics data for an olivine mantle which transforms successively to the beta and spinel phases. The velocity dependence on pressure is calculated from the data of Gwanmesia et al. (1990b) and Rigden et al. (1991) and on temperature is assumed to be identical in all three phases.](image)
However, the inference of such a discontinuity from near-vertical incidence long-period records can be reconciled with the dominant contribution of density to the impedance contrast (Shearer, 1990). Moreover, numerical modelling of a discontinuity dominated by a density contrast showed that although there was no evidence for a triplication caused by the discontinuity at normal epicentral distances for either short or long periods, near-vertical reflections appear from this discontinuity (Jones et al., 1992). Thus, the beta-spinel transition cannot be ruled out as the cause of the 520 km discontinuity although the velocity jump may be too small to be generally observable, particularly with short-period waves.

**Velocity gradients in transition zone**

Rigden et al. (1991, 1992a) examined expected transition zone velocity gradients for the olivine component for a range of plausible values of \(\frac{dK}{dT}\) and \(\frac{dG}{dT}\) for the high-pressure phases. Although the size of the calculated 410 km discontinuity is strongly influence by the choice of temperature derivatives for the beta phase (as discussed above), the velocity gradients are relatively insensitive to the choice of temperature derivatives. From Fig. 3, it is clear that adiabatic compression of the beta and spinel phases is not sufficient to explain the observed seismic gradients for either P or S waves in the transition zone. This discrepancy has been attributed to neglect of other mineral components with higher velocity gradients, to the occurrence of other phase transitions within the transition zone, and to steepening of the S wave velocity gradient as a consequence of the gradual recovery of a shear modulus deficit associated with anelastic relaxation (see discussion and references in Rigden et al., 1991, 1992a). Moreover, as Kennett (1991) has pointed out, neither the velocity gradients nor the velocity contrasts at the main discontinuities in the transition zone (at 410 and 660 km) are so tightly constrained as to rule out many petrological models.

**SUMMARY AND FUTURE DIRECTIONS**

It should be clear from the above discussion that future progress in the interpretation of seismic models in terms of mineralogy and chemical composition requires integrated and coordinated investigations in the fields of both mineral physics and seismology. From the mineral physics vantage point, work is underway in our laboratories to determine the temperature derivatives of the velocities of the high-pressure phases of mantle minerals, to extend these measurements to other high-pressure phases (e.g., stishovite-Li et al., 1992; majorite-Rigden et al., 1992b), and ultimately to measure the velocities of the high-pressure phases within their stability fields at simultaneously elevated pressures and temperatures.
REFERENCES


This paper is based on the results of a collaborative research program between the Stony Brook High Pressure Laboratory and the ultrasonics laboratory of I. Jackson and S. M. Rigden at the Australian National University; this program has been primarily energized by the innovative work of Rigden and G. D. Gwanmesia. This research was supported by the National Science Foundation Division of Earth Sciences as part of the NSF Science and Technology Center for High Pressure Research (EAR 89-20239 and 89-17563) and by the U. S.-Australia Cooperative Research Program of the NSF (INT-89-13363) and the Australian Department of Industry, Technology and Commerce and NSF grant (EAR 91-04563).
ABSTRACT
Surface waves dispersion curves have been measured using the single station method for the paths in the Atlantic Ocean between Azores and West Iberia Margin. The vertical and horizontal components of Rayleigh waves have been analyzed. In order to assure that this study is concerning only oceanic structures, the pure path method has been applied to separate continental from oceanic propagation. The pure oceanic Rayleigh wave phase velocities have shown to be higher in the northern vicinity of the Iberian Peninsula. Pure phase velocities have been inverted using an isotropic model, to propose S-wave velocity models of the crust and upper mantle in the sampled regions.

INTRODUCTION
From March 1988 to March 1989, the NARS Broad-Band network has been installed in the Iberian Peninsula, under the sponsorship of the ILIHA project. Its configuration (figure 1) was especially designed to allow the study of the lithospheric anisotropy beneath the Peninsula. As a side effect, the existence of a few records, from earthquakes with epicenter on and around the Azores Islands suggest the study of the Azores-West Iberia lithosphere and upper mantle structure using the NARS data.

SURFACE WAVES DISPERSION MEASUREMENTS
The location of the selected earthquakes, their focal mechanisms and the surface waves paths reaching the NARS network, can be seen in figure 1. Their identification and main source characteristics are presented in table I. The crossed structures in the "open" Atlantic, at the depths reached by surface waves, are more or less homogeneous, then we may hope that any asymmetry between the northern and southern lithosphere should be caused by heterogeneities in the vicinity of Iberia.

As a preliminary analysis, we have obtained group velocity dispersion curves using the multiple filter techniques [Dziewonski and Hales, 1972]. Residual dispersion analysis, with amplitude correction [Cara, 1973], have also been applied regarding phase measurements of both vertical and longitudinal components of Rayleigh waves.

Phase velocities have then been calculated using the single station method [Dziewonski and Hales, 1972]. In order to obtain the propagation phase shift, the instrumental and source phase components have been removed. The first one has been performed altogether the residual dispersion analysis. To calculate the source phase we have used a theoretical model from Nishimura and Forsyth (1989) for a 0-4 m.y. old ocean and the centroid focal mechanisms from the ISC reports [table I]. The measured phase velocities have been grouped in three different sets, according to the crossed structures in the Atlantic (figure 1): 1) Azores - Northwest Iberia, 2) Azores - Southwest Iberia and 3) North Atlantic Ridge - Iberia, hereafter referred as AZNI, AZSI and NARI, respectively.

The longitudinal Rayleigh wave phase velocities have been much more difficult to evaluate due to a relatively poor horizontal cemponants data quality. Technical problems related to the trigger, malfunctioning of some horizontal sensors, and noise level higher
Figure 1 - Epicenter location of the analyzed earthquakes

TABLE I - Event Source Characteristics

<table>
<thead>
<tr>
<th>ID</th>
<th>Origin Time</th>
<th>Location</th>
<th>m</th>
<th>F.ult Parameters (T,P)</th>
<th>Long.</th>
<th>Lat.</th>
<th>Plg</th>
<th>Azm</th>
<th>Plg</th>
<th>Azm</th>
<th>Hd</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1</td>
<td>880722 21:16:04</td>
<td>-29.59</td>
<td>39.87</td>
<td>5.0</td>
<td>3</td>
<td>290</td>
<td>81</td>
<td>178</td>
<td>2.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>A2</td>
<td>881016 06:15:29</td>
<td>-25.33</td>
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than the vertical component one also reduce the amount of data. Nevertheless, the computed longitudinal Rayleigh wave phase velocities have shown to be identical to the corresponding vertical ones, thus confirming that actually the measured dispersion correspond to Rayleigh wave phase velocities.

For the paths AZNI and AZSI, the phase velocities have been regionalized in order to separate the continental path from the oceanic path. We have used the "pure path method" presented by Forsyth (1975) assuming an isotropic structure, since the data are insufficient for evidencing anisotropy.

The obtained pure phase velocities can be seen in figure 2, together with their error bars. The first point to notice is that for the periods between 20 and 50 seconds the phase velocities for AZNI are significantly higher than for AZSI.

This difference tends to disappear for higher periods. Beyond 100 seconds, phase velocities in the paths AZNI seem to be once again higher than in the AZSI. Nevertheless, we must remember that the NARS velocity instrument response is flat only until 100 seconds, decaying beyond that value. Therefore, the signal to noise ratio...
becomes very low for periods higher that 100 seconds, unabling the interpretation in that range.

In the North Atlantic Ridge - Iberia set, and due to the lack of data, the pure path method has not been applied and no separation between the Northern and Southern vicinity of the Iberian Peninsula has been tried. Nevertheless, and as seen on figure 1, the proportion of continental path over the total path is very small. Furthermore, the individual dispersion curves are less scattered than the ones obtained on AZNI and AZSI. For this reason, the measured phase velocities for NARI, which are shown on figure 3, may be analyzed as a whole. It is interesting to notice that, in general, the phase velocities seem to be higher than those obtained for the paths AZSI, and similar to those measured for AZNI, probably reflecting some slight anisotropic behaviour.

RESULTS OF PURE PATH INVERSION

Only the pure phase velocities have been inverted for an isotropic model using the Tarantola and Valette algorithm (1982), in order to obtain S-wave velocities as a function of depth. The a priori covariance matrices for data and model parameters have been built as described in Lévéque et al. (1991). The parameter a priori variance (i.e., the S-wave a priori variance) was taken equal to 0.2 kms\(^{-1}\) with a correlation length of 20 km. The Nishimura and Forsyth (1989) oceanic model for an ocean 20-52 m.y.old was used as the a priori model. The matrix of the partial derivatives with respect to the model parameters has been computed using the Saito software package (1988).

The obtained S-wave velocity models, for the paths AZNI and AZSI, together with the S-wave initial model (dashed line) and the a posteriori error bars, can be seen in figure 4. These error bars presented in the final model must be interpreted keeping in mind the chosen correlation length. The most valuable information we can infer from the data concerns the depths between 15 and 150 km, where error bars have been significantly
reduced. At the bottom of that figure, the differences between the data final fit (solid line with the error bars) and the initial fit (dashed line) are represented. It must be noticed that the final model fits the data well inside their error bars, with some exceptions in the short range periods.

CONCLUSIONS
When comparing the final models obtained for the northern and southern regions, sets AZNI and AZSI, respectively, we found that the S-wave velocities for shallow depths are higher for the AZNI. This difference tends to disappear with depth, and after 70-80 Km the two models exhibit similar behaviour. This same discrepancy between the northern and southern structures in the vicinity of the Iberian Peninsula has been suggested by Marillier (1982). Also, the dispersion curves obtained for group velocity in the AZSI agree with those obtained by Cara (1976). According to this author, the reason for the lower velocities may be the existence of a thick sediment layer in the Tagus Basin.

REFERENCES
A SMALL ARRAY STUDY OF S-WAVE ANISOTROPY OF THE DEEP LITHOSPHERE ACROSS THE PROTOGENE ZONE IN SOUTHERN SWEDEN

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INTRODUCTION

It has been recognized during the last decade that seismic anisotropy is an important property of the deep continental lithosphere (Babuška et al., 1984; 1988a; Anderson, 1989). However, at the present time, we do not dispose of sufficient knowledge of this parameter to derive satisfactorily the 3-D anisotropic structure of the upper mantle. A passive seismic experiment was organized in southern Sweden for the study of the deep lithospheric structure and focused mainly on the differences in velocity anisotropy on both sides of the Protogene Zone interpreted as a suture resulting from the continental collision of the Swedish and Norwegian plates.

During the experiment, conducted in the period May-August 1991, twenty digital seismic stations were deployed on 3 profiles orientated in the VSV-ENE direction perpendicularly to main tectonic zones of the Värmland region (Fig. 1) to record teleseismic P, SKS and S waves. The network covered an area of 140 x 60 km with a centre at 13.5°E and 60.0°N. The experiment was directed toward studying lateral variations of mantle anisotropy as was indicated in a preliminary P-residual study (Babuška et al., 1988b) based on teleseismic observations of permanent seismological observatories in Fennoscandia. This contribution reports on the results of a new study of SKS/S waves recorded during the experiment.

METHOD

Anisotropic studies of the upper mantle based on shear-wave polarizations, so far performed in several regions (e.g. Kind et al., 1985; Vinnik et al., 1989; Makeeva et al., 1990; Silver and Chan, 1991; Savage and Silver, 1992) deal with an azimuthal anisotropy and a-priori suggest that the symmetry axis of the hexagonal medium used in the models is horizontal and thus, in fact, restrict the complete 3-D solution to a search for only the horizontal component of the upper mantle anisotropy (Savage and Silver, 1992). The horizontal component of anisotropy dominates in some parts of the upper mantle, mainly in the asthenosphere, but this is not necessarily true for the lithospheric part of the mantle.

To analyze S wave polarization, we use a method by Silver and Chan (1991). The method estimates the splitting parameters for the anisotropy (an orientation of the fast S polarization and time delay $\delta t$ between the two split S waves) that mostly returns the split
signal in the N-E-Z or L-Q-T coordinate systems (L is a ray direction, Q and T are perpendicular to L in the vertical and horizontal planes, respectively) to an isotropic form in a new X-Y-Z coordinate system. For S waves, this is done by minimizing the least eigenvalue of the covariance matrix of the corrected particle motion. In the case of SKS or SKKS, the corrected transverse component is minimized. We solve the problem as a fully 3-D one, i.e., we search for an arbitrarily inclined plane, where the fast and slow S are polarized. In such a case, the fast S polarizations observed at a station depend on azimuths and incidence angles. Therefore, it is impossible to make the fast S polarizations identical with the fast velocity directions. To determine the orientation of a plunged symmetry axis that coincides either with the fast- or slow-velocity direction of the anisotropic medium, we have to invert the S polarizations.

Digital seismic stations - Värmland experiment 1991

1-component short-period stations
1-component medium-period stations
1-component medium- and long-period st.
3-component short-period stations
3-component medium-period stations
3-component medium- and long-period stations
University of Uppsala, Sweden

Polarization analysis of S, SKS and SKKS showed elliptical polarizations at many stations of the experimental array. The dense spacing of stations allowed us to study lateral variations of the orientation of the fast S polarizations and changes of the time

Fig. 1. Tectonic sketch-map of southern Sweden and the location of experimental stations (stippled area in the left).

OBSERVATION

Particle motion analysis of S, SKS and SKKS showed elliptical polarizations at many stations of the experimental array. The dense spacing of stations allowed us to study lateral variations of the orientation of the fast S polarizations and changes of the time.
Fig. 2. Lateral variations of SKS particle motion across the array: (I) only slight elliptical polarization westward of the PZ; (II) linear polarization in the central part, near the approximate position of the PZ; and (III) distinct elliptical polarization eastward of the PZ.

Fig. 3. SKS waveform rotated into the L-Q-T coordinate system (a) with the Q component delayed relative to the T component and splitting parameters achieved by a correction for the anisotropy (b).
delay between the two split S waves. Figure 2 demonstrates changes of the particle motion of SKS of one event recorded at different stations. Distinct elliptical polarization of SKS waves, typical of their propagation through anisotropic structures, is observed mainly at stations situated in the Swedish Plate (stations C, SV13, SV14), eastward of the Protogene Zone. Such a polarization pattern disappears at stations deployed near the approximate position of the Protogene Zone (stations B and SV12), where SKS waves are polarized linearly in a direction close to the back azimuth. Weak elliptical polarization appears again at stations situated in the Norwegian Plate (station A), westward of the Protogene Zone. Pronounced lateral variations of the polarization pattern across the array that are related to the Protogene Zone are typical for all shear waves of various events (see another example in Fig. 4).

Undisturbed SKS waves propagating through isotropic media exhibit linear polarizations whereas in anisotropic media their polarizations are elliptical unless they propagate in the direction along the symmetry axis coinciding with fast- or slow-velocity direction. Straight line in the polarization diagram of station C (Fig. 2, lower right-hand corner) constructed from the original signal in the N-E-Z coordinate system (Fig. 2, upper left-hand corner), may correspond to the fast S wave. An estimate of the time delay between the fast and slow S waves from the diagram is 1 s.

When the same SKS wave recorded at station SV13 with the broadest ellipticity, is rotated into the L-Q-T coordinate system (Fig. 3a) a clear shift between the Q and T components corresponding to the two split S waves is evident. The minimum (MQ) of the waveform on the Q component is delayed 0.7 s relative to that (MT) on the T component. Figure 3b presents an estimate of the splitting parameters obtained by a correction for the anisotropy.

![Diagram of the particle motion of S wave.](image)

Fig. 4. Lateral variations of the particle motion of S wave.
To show how the polarizations depend on a dip of the symmetry axes of anisotropic structures, we computed the fast and slow S polarizations of waves propagating through an anisotropic medium with hexagonal symmetry, whose symmetry axis coincides with a low-velocity direction (axis b, Babuška et al., 1992). The diagrams show (Fig. 5) that the polarizations are independent of the back azimuths and incidence angles only if the symmetry axis is horizontal. The right-hand column of the figure shows orientations of the fast S polarizations found for stations at the Norwegian (A) and the Swedish Plates (C and SV14). These preliminary results indicate different tendencies in orientation of the fast S polarizations at stations situated on both sides of the Protogene Zone and surely demonstrate their dependence on the azimuths and incidences.

Fig. 5. Stereographic projections of S polarizations showing their dependence on the back azimuths and incidence angles (up to 30°) if the symmetry axis of the anisotropic medium with hexagonal symmetry, coinciding with the low-velocity direction (b), plunges. Examples of the fast S polarizations observed at stations A, C and SV14 indicate the difference between anisotropic structures of the Norwegian (A) and Swedish Plates (C and SV14). Dashed crosses are used in the cases when the SKS wave was polarized linearly, i.e. the wave may propagate in one of the prominent directions.
CONCLUSIONS

1) Polarization analysis of SKS waves that exhibit elliptical polarizations caused by split fast and slow S, which occur in anisotropic media, proved the existence of velocity anisotropy in the upper mantle as indicated by the teleseismic P-residual study (Babuška et al., 1988b).

2) The dense spacing of stations allowed us to study variations in the orientation of the fast S polarizations and in the time delay between the two split S waves. Distinct lateral changes of polarizations across the array at stations associated with the Protoene Zone indicate changes in the anisotropy within the small region related to the prominent suture. This suggests that the main part of the anisotropic signal originates in the uppermost mantle beneath the stations, most likely in the subcrustal lithosphere.

3) Observed S polarizations seem to depend on azimuths and incidence angles of arriving waves. Therefore, the observed anisotropy is not compatible with an anisotropic medium with hexagonal symmetry, whose symmetry axis coinciding either with fast- (a) or slow-velocity (b) directions lies in the horizontal plane. We suggest an anisotropic model of the subcrustal lithosphere, where the symmetry axis plunges as indicated by the P residual spheres. To find the model, mainly as regards an inclination of its symmetry axis, we will perform an anisotropic inversion of the S polarizations observed at individual stations.

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REFERENCES

THE DISTRIBUTION OF Pn VELOCITIES FOR ALBANIA BASED ON TIME-TERM METHOD FOR NEAR EARTHQUAKES

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INTRODUCTION

As is well-known, the Pn phase is a diffracted one travelling along the Mohorovicic discontinuity. This phase begins to distinguish itself in the near earthquake records from 130-150 km and up to almost 1000 km where it is the first phase reaching to the seismological stations happening to be in this distance range from the earthquake foci.

The velocity of Pn phase depends from the crustal thickness as well as from the crust/mantle density.

Using a modified version of the time-term method in this paper we attempt to give a picture of Pn velocity distribution for Albanian territory.

THE USED METHOD

The time-term method is a well-known one which has been used in refraction seismology (Scheideger, Willmore, 1957) and later also in crustal and earthquake studies (Willmore, Bancroft, 1960; Berry, West, 1966; Haines, 1979; Haines, 1980; Liaw, Yeh, 1983; Hearn, 1984; Kayal, De, 1987). According to this method, the travel time of Pn phase from the earthquake foci to the seismological station would be given:

\[ T_i = T_0 + d_0 + \left( D_i - (\delta_o + \delta_i) \right) / V + d_i \]

where \( T_0 \) - earthquake origin time; \( D_i \) - epicentral distance; \( \delta_o \), \( \delta_i \) - offset distances at the focus and the recording site; \( V \) - apparent velocity for a path below the refractor between offset points \( \delta_o \) and \( \delta_i \) (Fig.1); \( d_0 \) and \( d_i \) - are called the time-terms for the focus and recording site. They express the travel times above the refractor plus allowances for the differences between the true offsets and their estimates. Assuming that they depend only on the structure between the focus and the refractor and the refractor and the recording point respectively, one can consider \( d_0 \) as being the same for all observations of the phase coming from the focus and \( d_i \) as being the same for all the recordings of the phase at one seismological station.

Some authors were trying to solve the velocity comprising into the calculation even the time-terms (Berry, West, 1966; Liaw, Yeh, 1983; Hearn, 1984). Some others, have simplified the problem solving the velocity in the border crust/mantle using the pairs of stations (Haines, 1979; Haines, 1980; Kayal, Smith, 1984; Kayal, De, 1987). Modifying the method given by Scheideger and Willmore (1957) and Willmore, Bancroft (1960), Haines (1979) used only the cases when the pairs of stations lay at roughly the same azimuth from the epicenter. Doing so, it is possible to avoid from the calculations the time terms. In this case, we can write the travel time difference of Pn wave for two stations:

\[ T_1 - T_2 = (D_1 - D_2) / V_{1,2} + (d_2 - d_1) \]

where \( D_1 \) and \( D_2 \) are epicentral distances, \( d_1 \) and \( d_2 \) are the time terms for the recording site (Fig.1), \( V_{1,2} \) the Pn wave velocity. Assuming that \( d_0 = d_i \) for a horizontal refractor and minimum observational error, the above equation can simplify more: \( \Delta T = \Delta D/V \) from which knowing the difference on the travel time of Pn at two seismological stations with the same azimuth from the
earthquake epicentre, and the distance between those two stations, we can obtain the Pn velocity for the crust/mantle discontinuity limited between two above seismological stations.

RESULTS AND DISCUSSION

This study utilises the data of ISC Bulletins for the period 1980-1984. By a computer program, in the Earthquake Research Institute of University of Tokyo, Japan, there were selected only these events when the azimuth difference epicentre-station for each pair of seismological station does not pass over 2°. There are taken into consideration the stations of Albanian Seismological Network. Additional Pn arrival-time information was obtained from neighbouring seismological stations of TTG, OHR, JAN, ULC.

The epicentral distances of selected cases are in the range 150-700 km. We have left apart the data with anomal values of residuals as well as the earthquakes with depths over 40 km. For 29 combinations there were used about 250 events and keeping in mind that one earthquake may be used for more than one pair of stations, the total of our data is 275. In some cases, when it was possible, reversed Pn arrivals for some station pairs were taken into consideration too.

In the Tab. 1 are given the obtained results and the Fig. 2 embodies these results according to the directions of each station pair. One can see that the high values of Pn velocity are near the shore part of the country, in the direction TIR-PUK, TIR-SDA, TIR-TTG, SDA-TTG, SRN-BERA, SRN-VLO. These values vary from 8.1 to 8.9 km/sec. The highest value, 8.92 km/sec is noted for the station

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pair TIR-PUK, with 14 data. In the inner part of Albania, generally result lower value of Pn velocity. It is true both in transversal and longitudinal directions.

As we can see from the Table 1, the Pn velocity values obtained from the data analysis of time-term method, are in the range 7.0-9.0 km/sec which is a considerable difference for a small territory like Albania. In this variation influence the crust thickness, the rock density in the crust/mantle discontinuity and the nonhorizontal being of refractor for particular directions.

On the other hand, the tectonic setting of a region influences so much on Pn velocity. The high values of Pn velocity there are reported also by other investigators for different areas of the world (Kosminskaya et al., 1972; Ansorge et al., 1979; Fuch et al., 1979; Hirn, 1977). For the South Adriatic Sea, where is included also Albanian Adriatic shore, results a crustal thickness of 41-47 km (Calcagnile et al., 1982). This region is characterised by strongly positive Pn residuals (Panagiotopoulos et al., 1985). Going to the northwest, for the Montenegro shore, the Pn velocity results over 8.0 km/sec and the Moho discontinuity depth is over 40 km (Glavatovic, 1989).

So, if we attribute the high values of Pn velocity to the increasing of the crustal thickness from inner part towards the seaside, we can suggest that this crustal thickening is caused by the collision of Adria and orogene in this part of Balcan peninsula. This collision is also demonstrated by focal mechanism solutions (Muco, 1992). It is well confirmed that the outer part of Albania is dominated by compressional stress regime while the tensional stresses are more present in the inner part (McKenzie, 1972; Aliaj, 1988; Sulstarova, 1986; Muco, 1992). From this study one can see that the zone of highest Pn velocities coincides with that one where the compressional regime is dominant.

In spite of limitations imposed by this method and the earthquake location and observational errors, it is clearly revealed a considerable variation in the Pn velocity values between the inner and the outer part of Albania.

CONCLUSIONS
1. It is provided a picture of Pn velocity distribution for Albania using a modified version of time-term method.
2. A distinct variation between the Pn velocity values for the inner and outer part of Albania is noted confirming that crustal thickening goes towards the sea.

ACKNOWLEDGEMENT
This work was largely carried out while the author was visiting the Earthquake Research Institute, Tokyo, Japan and he would like to express his deep gratitude and his thanks to M. Mizoue and all the members of his Department for the generous provision of resources.
technical facilities and for their invaluable support.

REFERENCES
INVESTIGATIONS OF THE CONNECTION BETWEEN SEISMICITY AND $^{222}\text{Rn}-\text{CO}_2$ CONTENT IN SPRING WATERS AT THE VOGLAND AREA (GERMANY): FIRST RESULTS

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INTRODUCTION

The Saxon State Spa Bad Brambach is situated in the south west of Saxonia, in the western part of the Ore Mountains. It is well-known because of its mineral waters with high natural $^{222}\text{Rn}$ and $\text{CO}_2$ concentrations. The location has a temperate-humid climate with an average precipitation of 656 mm·a$^{-1}$ and an average temperature of 4.7 °C.

Our investigation area is located almost exactly on the boundary between the so called Fichtelgebirgs granite in the south and the Vogtland mica slate/phyllite massiv in the north (Fig. 1). The region is tectonically disturbed in a wide range. It is characterized by the effects of Tertiary volcanism, having been also the reason for the recent seismic activity in this area (swarm-, micro-earthquakes).

The seismic events are registered by a seismological network, which is pursued by the Geo Research Center Potsdam/Jena. The typical Vogtland micro-earthquakes have magnitudes $< 4$ but a relatively high frequency of the events.

The mineral springs of Bad Brambach are very useful for the investigation of earthquake prediction methods on geochemical basis because of their high mineral, $^{222}\text{Rn}$, and $\text{CO}_2$ concentrations (Fig. 2) and their location near the most frequent epicenter of the microquakes. Furthermore the area has been explored very thoroughly by means of several modern techniques during the last 15 years (geophysics, isotope hydrology) /7,8,11,18,6/.

MEASUREMENTS

In 1989 we started a continuous measurement programme of radon in spring water and soil air as well as hydrological and meteorological parameters (Fig. 3). After a three years measuring period it could be found that the soil air is not suitable very well to investigate the connection between radon behaviour in the ground and seismic events. Radon in soil air shows a high sensitivity against atmospheric pressure, temperature, soil moisture and groundwater level fluctuations.

The measurements in the spring water are carried through in the 'Radonquelle' of Bad Brambach ($^{222}\text{Rn}$ activity: 25 kBq·l$^{-1}$). This high activity makes it possible to detect the gamma activity of the radon daughter products as a radon equivalent /5/.

The measuring values are statistically checked, for example by auto correlation method. Furthermore values with an amount above the one sigma level are interpreted as anomalies only.
spring parameters | Radonquelle | Eisenquelle | units
--- | --- | --- | ---
Discharge | 150 | 230 | l/h
Type | Na-Ca-HCO₃-SO₄ - acidulous with Fe + Rn | | |
Temperature | 7 - 9,5 | 7 - 9,5 | °C
Depth of the spring capture | 6 | 5 | m
Content of radon 222 | 25 000 | 1 900 | Bq/l
Content of radium 226 | 1,8 | 0,06 | Bq/l
Content of uranium 238 | 2,5 | 1,5 | ppm
U 234 / U 238 activity ratio | 3,1 | 2,8 | |
Content of CO₂ | 2540 | 2740 | mg/l
Content of δ 13C | - 3,5 | - 3,7 | %
Mean residence time (tritium) | 19 - 150 | 1 - 13,5 | a
Content of "young" water | 12 - 19 | 34 - 48 | %
Content of He in the spring gas | 0, 001 | 0,001 | Vol.%
Electric conductivity | 1,65 | 1,22 | mS/cm

Fig. 2 Hydrological, chemical and isotope data of the mineral springs 'Radonquelle' and 'Eisenquelle'.
RESULTS AND DISCUSSION

The results of the radon measurements in the spring water over a period of three years demonstrate some significant anomalies in the radon activity before, during, or after microseismic events in the region (Fig. 4, 5). The measuring interval for calculating the daily average value is < 1 hour. Fig. 5 shows an example for an anomaly caused by an earthquake of higher epicentral distance (about 450 km NW of the Vogtland area).

Former investigations in the area had shown that the origin of the very high activity of the 'Radonquelle' - an uranium/radium deposit - is situated in a relatively low depth (<100 m) which is not influenced by the tectonic stress field directly /3,7,8,11, 12,18/. Thus another 'transmitter' must be responsible for the signal from the ground due to a seismic event. Such a transmitter could be the CO₂ as a typical component of all the mineral waters in area (concentrations 1200-2800 mg·l⁻¹). Particularly tritium and uranium analyses show that the spring waters consist of two main water components - an old mineralized part and a younger one (10-40 %, Fig. 2) /6,7/. The CO₂ enriched old water is from greater depth because the δ¹³C values of the spring gas indicate a magmatic CO₂ origin (-3.5 to -3.7 ‰) /3/. So it may be assumed that carbon dioxide reflects the changes in the stress field built up in the seismically active faults (see also /15/).

Fig. 6 shows an example for the influence of CO₂ in 3 springs of Central Italy /13/. The springs belong to different hydrological circuits. The apparent similitude in the reaction of the BORRA and PETRIOLO springs is not related to meteorological effects. But probably it is due to geodynamic stress being able to trigger strong CO₂ activities.

Therefore an improved measuring system registers CO₂-, Rn-concentration, electric conductivity, and water flow rate continuously at two mineral springs of Bad Brambach ('Radon'--,'Eisenquelle') since April 1992. The measuring interval is 15 minutes. The values are recorded by a personal computer.

The results show a good correlation between the parameters CO₂ and radon in an interval measured during a seismically quiet period (Fig. 7). There can be seen simultaneous fluctuations in both curves due to anthropogenic influences (pumping effect, taking out of spring water).

As a result of the lapse of the curves we can assume that earlier observed radon anomalies in the temporal vicinity of the earthquakes could be caused by anomalies in CO₂ concentration.

According to Hurtig and Oelsner (1979) as well as Čermak (1979) a deep source of heat has been identified in the investigated area. Kampf at al. (1989) demonstrate that this heat source could be related to a low velocity zone in the crust at about 10-15 km depth. In this zone strong CO₂ degassing activity occurs, and deep originated gases escape towards the ground surface. The macroscopic effect of the degassing activity has been identified in the CO₂ bubbling which strongly affects many springs in the Bohemian Massif and in the Vogtland. The isotopic ratio δ¹³C measured in CO₂ bubbling in the Bad Brambach springs is consistent with a deep metamorphic or magmatic origin /3,1/.
BASIC INVESTIGATIONS AT BAD BRAMBACH

soil air

80 cm depth

springs

* Radon-Quelle* 25 kBq/l
* Eisen-Quelle* 1.9 kBq/l

* temperature
* air pressure
* precipitation (humidity)
* discharge
* conductivity
* radon
* CO2
* groundwater level

Fig. 3 Schematic overview of the measuring principle and the measured parameters.

Fig. 4 Example for typical radon anomalies in spring water of the Radonquelle in connection to micro-earthquakes.

Fig. 5 Example for radon anomalies in spring water caused by a micro-earthquake with Vogtland origin and an earthquake of higher epicentral distance (about 450 km NNW of Bad Brambach).

---

rel. radon content in the Radon spring

00000 seismic events (epicentral dist. < 50 km)

---

rel. radon content in the Radon spring (counts/0.001 min)

---

discharge [l/h]

---

00000 seismic events (epicentral dist. < 50 km)

---

00000 seismic events (epicentral dist. approx. 450 km)

---

m=1.9/1.1

6.0

NRW

1.9/1

Swarm

00

00

0

---

time in days (measuring interval Jan.-June 92)
Radon concentration in the liquid phase at different sites

Fig. 6 Example for the radon behaviour of three springs belonging to different hydrological circuits, Central Italy (see text).

Fig. 7 First results of continuous measurements of $^{222}\text{Rn}$ together with CO$_2$ in the mineral spring 'Eisenguelle', Bad Brambach. Obviously there is a good correlation between $^{222}\text{Rn}$ and CO$_2$ concentration.
The earthquake occurrence triggers the CO$_2$ generation, and the deep originated gases escape towards the surface and act as gas carrier for radon. $^{222}$Rn is subjected to natural radioactive decay but the gas carriers are continuously supplied of radon originated from relatively shallow accumulations also. This can explain the very close correlation between CO$_2$ and $^{222}$Rn.

Varhegyi et al. (1992) experimentally confirmed the ability of CO$_2$ to cause such a process. Toutain et al. (1992) reported about the similar phenomenon after having monitored volcanic gas emanations.

The spike-like form of the observed radon anomalies in the 'Radonquelle' of Bad Brambach reflects the above discussed model conception very well. It confirms the assumption of an outburst-like CO$_2$ release from the depth within a very short temporal interval.

CONCLUSIONS

The results of continuously radon measurements in the Bad Brambach area have shown that - in opposite to the soil air - radon in spring water is a suitable tracer to obtain precursor informations in connection with micro earthquakes also. They suggest that radon alone is not able to penetrate the water from the deep located geodynamic stress field up to the surface within a time being necessary for inducing earthquake precursors. For such a short-time reaction an additional gas carrier has to be responsible. This is CO$_2$ at our measuring site. In rare cases it may be methane also /13,14/.

Irwin and Barnes (1980) found that most of the great seismic areas of the earth are characterized by a lot of springs with CO$_2$ of magmatic origin. Thus it should be useful for earthquake prediction research to observe such typical CO$_2$ sources continuously with regard to variations in CO$_2$ concentration.

In this connection it should be mentioned that the costs for the installation of a continuously measuring device do not need to be very high. In our case they amount to less than 3,000 $.

ACKNOWLEDGEMENTS

The authors wish to thank the director of the Saxon State Spas Bad Brambach - Bad Elster, P. Scheler, for the long standing support of their work as well as Dr. H. Neunhöfer and Dr. K. Klinge, GFZ Potsdam/Jena/Moxa, for sending seismic data periodically.

REFERENCES


PHYSICAL BASIS AND EXPERIENCE OF USING COMPLEX OF PARAMETERS FOR THE ESTIMATION OF STRONG EARTHQUAKE APPEARANCE DANGER.

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INTRODUCTION

Development and deepening of ideas about failure process of rocks and geological medium as a self-similar and self-organizing system of blocks have created the physical and theoretical basis for analysis of failure process at different scales (Sadovsky et al., 1982).

An earthquake is an act of internal destruction of strained material of the Earth. In this connection the ideas about failure process and strength developing in solids failure physics, especially for bodies with different kinds defects of structure, which are the blocks of geophysical medium, have a great meaning for the earthquake source physics (Myachkin et al., 1975). Significant progress for understanding of physical origin of strength have carried out a study of its dependence from temperature and time and have resulted in creation of kinetic conception of strength of solids (Zhurkov, 1968, Zhurkov et al, 1977). It was found out that the failure of solids is not a single, instantaneous act, but a longterm process developing in time and space. This process proceeds in stages, progressing gradually from the lowest hierarchy rank of crack dimensions to the highest (submicro-, micro-, macro-cracks (fig.1) (Miachkin et al., 1975). At a given moment, the failure process, so far quasi-homogeneous in nature, becomes concentrated at certain points on the object under loading. The transition from one hierarchy rank of failure J (lower) to another hierarchy rank J+1 (higher) occurs when the crucial concentration of ruptures of appropriate dimensions (J-th rank) is reached in the failure source; the interaction between them results in the formation of a higher J+1 rank. behavior in earthquake source zone.

PRECURSORS AND INFORMATION DATA BASE

Based on the kinetic conception of strength of solid materials authors have made the image of anomaly behavior of different
seismological parameters before strong (M≥5.5) earthquakes. These are the density of seismogenic fractures Ksf, slope of magnitude-frequency graph (KSI-B value), number of earthquakes per time unit in a form of seismic quiescence (KSI-NQ) and activity increase (KSI-NA) and seismic energy released E*+f£/3 per a time unit also as quiescence (KSI-EQ) and activity increase (KSI-EA) (Sobolev et al., 1991). This list may be complemented with formal presentations of already known earthquakes precursors or with further advance in gaining about earthquake preparation process. Regional earthquake catalogs of Caucasus, Kopet-Dag, Southern California and North-East China were used as the information data base.

Table 1. Retrospective characteristics of prognostic parameters.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Size of elementary cells, km</th>
<th>Alarm level</th>
<th>Probability of detection</th>
<th>Probability of false alarm</th>
<th>Expectation time, years</th>
<th>Prediction effectiveness, years</th>
</tr>
</thead>
<tbody>
<tr>
<td>CAUCASUS</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ksf</td>
<td>50×50</td>
<td>13.9</td>
<td>0.3984</td>
<td>0.0837</td>
<td>2.8±2.7</td>
<td>4.31</td>
</tr>
<tr>
<td>KSI-B</td>
<td>100×100</td>
<td>+2.0</td>
<td>0.2591</td>
<td>0.1350</td>
<td>3.6±3.4</td>
<td>1.78</td>
</tr>
<tr>
<td>KSI-NQ</td>
<td>100×100</td>
<td>-2.0</td>
<td>0.1917</td>
<td>0.0275</td>
<td>4.3±3.1</td>
<td>4.00</td>
</tr>
<tr>
<td>KSI-NA</td>
<td>100×100</td>
<td>+1.5</td>
<td>0.1819</td>
<td>0.0292</td>
<td>4.3±4.7</td>
<td>3.87</td>
</tr>
<tr>
<td>KSI-EQ</td>
<td>100×100</td>
<td>-1.3</td>
<td>0.0881</td>
<td>0.0373</td>
<td>2.9±1.7</td>
<td>2.22</td>
</tr>
<tr>
<td>KSI-EA</td>
<td>100×100</td>
<td>+2.5</td>
<td>0.1152</td>
<td>0.0758</td>
<td>5.8±6.0</td>
<td>1.54</td>
</tr>
<tr>
<td>KOPET-DAG</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ksf</td>
<td>50×50</td>
<td>11.8</td>
<td>0.5441</td>
<td>0.0647</td>
<td>3.7±3.4</td>
<td>6.84</td>
</tr>
<tr>
<td>KSI-B</td>
<td>100×100</td>
<td>+2.0</td>
<td>0.2434</td>
<td>0.0591</td>
<td>3.2±2.8</td>
<td>3.27</td>
</tr>
<tr>
<td>KSI-NQ</td>
<td>100×100</td>
<td>-2.0</td>
<td>0.1018</td>
<td>0.0241</td>
<td>6.7±5.0</td>
<td>2.25</td>
</tr>
<tr>
<td>KSI-NA</td>
<td>100×100</td>
<td>+2.0</td>
<td>0.1726</td>
<td>0.0191</td>
<td>3.6±2.7</td>
<td>4.73</td>
</tr>
<tr>
<td>KSI-EQ</td>
<td>100×100</td>
<td>-1.3</td>
<td>0.1018</td>
<td>0.0367</td>
<td>3.9±2.4</td>
<td>2.36</td>
</tr>
<tr>
<td>KSI-EA</td>
<td>100×100</td>
<td>+2.5</td>
<td>0.1327</td>
<td>0.0731</td>
<td>5.0±5.2</td>
<td>1.80</td>
</tr>
<tr>
<td>SOUTHERN CALIFORNIA</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ksf</td>
<td>50×50</td>
<td>5.5</td>
<td>0.7778</td>
<td>0.1730</td>
<td>6.4±6.9</td>
<td>6.21</td>
</tr>
<tr>
<td>KSI-B</td>
<td>100×100</td>
<td>+3.0</td>
<td>0.5098</td>
<td>0.3620</td>
<td>4.8±4.2</td>
<td>1.98</td>
</tr>
<tr>
<td>KSI-NQ</td>
<td>100×100</td>
<td>-2.0</td>
<td>0.4902</td>
<td>0.2015</td>
<td>7.6±9.2</td>
<td>3.18</td>
</tr>
<tr>
<td>KSI-NA</td>
<td>100×100</td>
<td>+2.0</td>
<td>0.5894</td>
<td>0.2537</td>
<td>7.7±11.1</td>
<td>2.93</td>
</tr>
<tr>
<td>KSI-EQ</td>
<td>100×100</td>
<td>-1.1</td>
<td>0.1176</td>
<td>0.0219</td>
<td>4.0±2.9</td>
<td>5.08</td>
</tr>
<tr>
<td>KSI-EA</td>
<td>100×100</td>
<td>+2.0</td>
<td>0.3529</td>
<td>0.3730</td>
<td>7.1±9.8</td>
<td>1.68</td>
</tr>
<tr>
<td>NORTHEAST CHINA</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ksf</td>
<td>50×50</td>
<td>63.3</td>
<td>0.4211</td>
<td>0.1446</td>
<td>2.5±2.0</td>
<td>2.86</td>
</tr>
<tr>
<td>KSI-B</td>
<td>100×100</td>
<td>+2.0</td>
<td>0.2857</td>
<td>0.1285</td>
<td>2.6±2.0</td>
<td>2.18</td>
</tr>
<tr>
<td>KSI-NQ</td>
<td>100×100</td>
<td>-2.0</td>
<td>0.3636</td>
<td>0.0505</td>
<td>4.4±3.7</td>
<td>5.42</td>
</tr>
<tr>
<td>KSI-NA</td>
<td>100×100</td>
<td>+2.0</td>
<td>0.2392</td>
<td>0.0456</td>
<td>2.1±2.8</td>
<td>4.99</td>
</tr>
<tr>
<td>KSI-EQ</td>
<td>100×100</td>
<td>-2.0</td>
<td>0.2392</td>
<td>0.0456</td>
<td>2.1±2.8</td>
<td>4.99</td>
</tr>
<tr>
<td>KSI-EA</td>
<td>100×100</td>
<td>+2.0</td>
<td>0.2456</td>
<td>0.0458</td>
<td>1.9±1.9</td>
<td>5.15</td>
</tr>
</tbody>
</table>

RESULTS

Estimations of probabilities of detection P(KillD1), of false alarms P(KillD2), average time of strong earthquake expectation Tpr
were made for each prognostic parameter during the work (Table 1). Using the Bayesian approach maps of conditional probability distribution of strong earthquake occurrence $P(D_{11<k})$ were calculated (Sobolev et al., 1991). These maps were named as a maps of expected earthquakes (MEE) (e.g. fig.2.3). Their analysis have shown (Table 2) that in zones with 70% level of conditional probability occurred 56-72% of strong earthquakes. At that time the total square of 70% level zones was not above 38%.

Table 2. Results of analysis of maps of expected earthquakes.

<table>
<thead>
<tr>
<th>Region</th>
<th>Caucasus</th>
<th>Kopet-Dag</th>
<th>S. California</th>
<th>N-E China</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zones with $P(D_{11&lt;k})$</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>70%</td>
<td>23-26</td>
<td>15-30</td>
<td>8-18</td>
<td>12-22</td>
</tr>
<tr>
<td>90%</td>
<td>12-25</td>
<td>10-26</td>
<td>2-3</td>
<td>6-13</td>
</tr>
<tr>
<td>in % to the whole observation area</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Number of predicted earthquakes at zones with $P(D_{11&lt;k})$ levels</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>70%</td>
<td>72</td>
<td>59</td>
<td>56</td>
<td>60</td>
</tr>
<tr>
<td>90%</td>
<td>40</td>
<td>54</td>
<td>15</td>
<td>20</td>
</tr>
<tr>
<td>in % to the total number of target earthquakes</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total number of target earthquakes</td>
<td>25</td>
<td>39</td>
<td>27</td>
<td>5</td>
</tr>
</tbody>
</table>

CONCLUSION

Results of the research show that the proposed method of MEE construction and the maps of expected earthquakes may be recommended to take preventive measures in order to mitigate damage from the future strong earthquakes. Authors see further development of this method in the expansion of number of using prognostic parameters and accumulation of an experience of it testing in various seismoactive regions with different geological and geophysical characteristics.

REFERENCES

Fig. 2: Map of expected earthquakes in the Caucasus for 1987-1991 by data from an earthquake catalogue for 1963-1986. Earthquake epicenters with K12.5 (M26.5) for 1987-1991 are shown. 1 - earthquake epicenters, size of circles correspond to the size of earthquake source; 2 - isolines of conditional probability of occurrence of strong earthquakes 0.7; 3 - contour of a representative record of earthquakes with K28.5 (M22.5). Notations: GRO - Grozny, MAK - Makhachkala, LGD - Lagodekhi, BKR - Bakuriani, TIF - Tbilisi, STE - Stepanavan, ERE - Erevan, ERZ - Erzerum, VAN - Van, VAU - Vardenis, GRS - Goris, SHE - Shemakha, BAK - Baku. Along X and Y axes - distances in km.
Fig. 3 Map of expected earthquakes in the S. California for 1990-1997 by data from an earthquake catalogue for 1932-1989. Earthquake epicenters with M≥5.5 for 1-30-1992 are shown. 1 - earthquake epicenters, size of circles correspond to the size of earthquake source; 2 - isolines of conditional probability of occurrence of strong earthquakes 0.7; 3 - contour of a representative area of Caltech catalog of earthquakes. Notations: HO - Holister, FR - Fresno, PF - Parkfield, BF - Bakersfield, PD - Palmdale, LA - Los Angeles, SB - San Bernardino, SD - San Diego. Along X and Y axes - distances in km.
IS THE TIDAL A TRIGGERING FACTOR FOR Vrancea Intermediate Earthquakes?

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Figures 1 a and 1 c contain the representation in form of histograms of the distribution upon time that suggested the dependence between the number of Vrancea intermediate earthquakes $N$ and their position on sinusoid from Figure 1 b; Figure 1 a refers to the period between the 1st April, 1977 and the 1st April, 1986, and Figure 1 c refers to the period between the 1st April, 1977 and the 1st August, 1987 (Zugrăvescu et al., 1985, 1987). At first analysis, these representations suggested that Vrancea intermediate earthquakes could occur especially during a maximum or minimum of tidal curve, and also during the positive or negative inflexions of the tidal curve (Figure 1). The approximation of the gravity tide curve by the sinusoid in Figure 1 b made possible, like any model, some error introduction, in other words, it makes possible a wrong or forced assignation of the moments of earthquakes occurrence to some certain phases of the gravity tide curve (maximum, minimum, etc.). Let us suppose that the representations in Figure 1 do not contain errors (and neither exaggerations!) and let us see how much significant they are. For this, we calculated the probability $P_f$ that a Vrancea intermediate earthquake to occur during a maximum of the gravity tide, during a minimum, in one of the inflexion zones of this curve or in other phases (Table 1). Table 1 was made on the basis of data from Figure 1 c; the other notations have the following significance: $N_f$-number of earthquakes corresponding to one or more gravity tide phases; $N_t$-total number of Vrancea intermediate earthquakes recorded during the respective period of time.

<table>
<thead>
<tr>
<th>Phase</th>
<th>Number of events ($N_f$)</th>
<th>$P_f = N_f/N_t$</th>
</tr>
</thead>
<tbody>
<tr>
<td>maximum</td>
<td>155 [252]</td>
<td>0.16 [0.26]</td>
</tr>
<tr>
<td>minimum</td>
<td>142 [223]</td>
<td>0.15 [0.23]</td>
</tr>
<tr>
<td>inflexion</td>
<td>157 [221]</td>
<td>0.16 [0.23]</td>
</tr>
<tr>
<td>other phases</td>
<td>507 [265]</td>
<td>0.53 [0.28]</td>
</tr>
</tbody>
</table>

$N_t = 961$ earthquakes

The values that are not placed in brackets (Table 1) correspond to the case when for maximum and minimum phases, and to each of the two inflexion zones were admitted time intervals of one hour, and those in brackets were established allowing time intervals of three hours for all the mentioned phases. Because only 15-16% (in total 47%) of earthquakes (Table 1) occurred during the maximum, minimum and inflexion zones of the gravity tide, and 53% of earthquakes occurred in the course of the other phases of this curve, it is not possible to conclude for a certainty that Earth tides play the role of triggering factor for Vrancea intermediate earthquakes. Even if we admit for each maximum, minimum or inflexion phases a time interval of three hours (exaggerately too large!), that is even if we take the values in brackets (Table 1), the results are
Fig. 1. Position of earthquakes on the tidal curve.
Fig. 2

Fig. 3

Fig. 4

- \( N_0 \) - number of earthquakes triggered during the maximum.
- \( N_{10} \) - number of earthquakes triggered during the minimum.
- \( N_{P0} \) - number of earthquakes triggered during the positive inflexion.
- \( N_{12} \) - number of earthquakes triggered during the negative inflexion.
- \( N_{\infty} \) - number of uncorrelating earthquakes.

**Legend:**
- \( A \) - number of earthquakes triggered during the maximum.
- \( V \) - number of earthquakes triggered during the minimum.
- \( VV \) - number of earthquakes triggered during the positive inflexion.
- \( VVV \) - number of earthquakes triggered during the negative inflexion.
- \( VVVV \) - number of uncorrelating earthquakes.
not significant enough to demonstrate the existence of clear correlation between the moments of Vrancea intermediate earthquake occurrence and the gravity tides.

A set of 309 earthquakes (with the focal depth well known) was used to make a statistics of the distribution upon depth of the dependence between Vrancea intermediate earthquakes and their position on sinusoid from Figure 1 b. The results are presented in Figure 2. Figure 3 shows on basis of data from Figure 2 the curves of variation upon depth of the probability $P_f$ of triggering Vrancea intermediate earthquakes during the maximum (empty circles), minimum (crosses) and inflexion zones of tidal curve (empty triangles) and also during the other phases of this curve (full circles). Concerning the distribution upon depth, Figures 2 and 3 show that the order of size of $P_f$ probability is alike with that in Table 1. Consequently, as the data from table 1, those from Figures 2 and 3 do not allow to conclude for a certainty that Vrancea intermediate earthquakes are triggered by earth tides.

From practical point of view, the correlation between the moments of Vrancea intermediate earthquakes occurrence and the different characteristic phases of the Earth tides should be useful to contribute to the prediction of these earthquakes. It's obvious that this correlation can not contribute to seismic prediction because, first of all, we should have to know which is the day when the earthquake will occur (that is almost everything) and just after that we will be able to state that the earthquake will occur in only 16% probability during the maximum of tidal curve, with nearly 15% probability during the minimum of tidal curve, and with 16% probability during one of the inflexion zones of tidal curve (Table 1). On the other hand, the probability of earthquake occurrence in any other phases should be of 53% (see Table 1). The tidal might make the earthquake to occur some time before (a few days) the moment when tectonic stress should attain to the critical value (Figure 4 - see Curchin and Pennington, 1987), but the probability that this fact to take place is very small. To the knowledge of this fact a systematic study of Earth tides might play an important role, but because the phenomenon is much more complex than is shown in Figure 4, the chances of success are very little. On the other hand, knowing this preceding could not matter as long as it is not known to whom is applied this correction, that is as long as it's not solved the problem of earthquake prediction itself.

REFERENCES
VARIATIONS ON THE SHALLOW UNDERGROUND WATER LEVEL AND TEMPERATURE RELATED TO THE SEISMIC ACTIVITY

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1. INTRODUCTION

In Northern Greece close to Thessaloniki there is a seismic zone which is associated with the Servomacedonian geological massif. In this zone there is a basin which includes the lakes Lagada and Volvi. This is a very active seismic area in which many large earthquakes have occurred in the past.

In order to verify if the well known (Rikitake 1976,1981, Wakita 1982) pre- and post-seismic response of the underground water level and temperature manifest itself in the case of the shallow underground water, a shallow underground water level and temperature monitoring network was installed (Asteriadis and Contadakis 1991) in December of 1983 in this seismic active area and has been followed up ever since. The reason is obvious and has to do with the cheap instrumentation and the low running cost of such a monitoring network.

In this paper we present the work which has been done up today in the frame of this research as well as the respective results.

2. THE EXISTING NETWORK AND THE MEASUREMENTS

Four KLT-OTT type contact gauges were installed in 1983 by the Department of Geodesy and Surveying, University of Thessaloniki, at selected wells in the seismically active area of Northern Greece, between the latitudes 40°.35 and 41° N and longitudes 22°.7 and 23°.85 E.

Figure 1 shows the position of the four villages, Liti, Assiros, Melissourgos and Nymfopetra, where the six selected wells belong. Table 1 displays the position and the depth of the well stations of the network as well as the respective number of measurements which were performed in the time interval between 1983, December 20 and 1990, December 31.

The continuous daily measurement at the stations of Liti, Assiros 1&2, Melissourgos 1 and Nymfopetra have started on 1983 December 20 and at the station of Melissourgos 2 on 1984 April 16.

However it happened that some stations were out of work for some periods of time, mostly because the respective well have became dried due to the drought, as it is happened during the extremely dry year of 1990, where all the wells except that of the Assiros 1 and 2 stations have became dried by the end of May. It should be noted that the well of Liti remain dry until today and we were forced to replace the well in the network.

Two HELLMAN-type rain-gauges for the rainfall measurements were installed in the area in order to study for the influence of the rain in the changes of the underground water level.
Table 1. Position and depth of the wells of the network stations and the total number of the performed measurement.

<table>
<thead>
<tr>
<th>Monitoring wells</th>
<th>(\phi) [(^\circ)N]</th>
<th>(\lambda) [(^\circ)E]</th>
<th>height (m)</th>
<th>depth (m)</th>
<th>numb/meas</th>
</tr>
</thead>
<tbody>
<tr>
<td>LITI</td>
<td>40.75</td>
<td>22.98</td>
<td>170</td>
<td>11.0</td>
<td>1591</td>
</tr>
<tr>
<td>ASS1</td>
<td>40.82</td>
<td>23.03</td>
<td>200</td>
<td>4.2</td>
<td>2210</td>
</tr>
<tr>
<td>ASS2</td>
<td>40.82</td>
<td>23.03</td>
<td>200</td>
<td>2.2</td>
<td>2210</td>
</tr>
<tr>
<td>MEL1</td>
<td>40.60</td>
<td>23.42</td>
<td>110</td>
<td>4.1</td>
<td>1652</td>
</tr>
<tr>
<td>MEL2</td>
<td>40.60</td>
<td>23.42</td>
<td>110</td>
<td>5.1</td>
<td>1652</td>
</tr>
<tr>
<td>NYMF</td>
<td>40.69</td>
<td>23.34</td>
<td>90</td>
<td>10.0</td>
<td>1885</td>
</tr>
</tbody>
</table>

The seismic data have been selected from the monthly bulletin of the Seismological Institute of the National Observatory of Athens (SINOA) for the time period between 20.12.83 and 31.12.89, and from the files of the Seismological Institute of the University of Thessaloniki for the time period between 1.1.90 and 31.12.90. These data are for earthquakes with epicentres extending from \(\phi=40^\circ.3\) N to \(\phi=41^\circ.0\) N and from \(\lambda=22^\circ.7\) to \(\lambda=23^\circ.8\). During the above mentioned time period 496 earthquakes with epicentres in this area occurred, 66 from them had magnitudes \(M > 3.0\). Unfortunately the monthly bulletin of SINOA does not provide informations about the magnitudes if this is less than 3.0.R.

3. RESULTS AND DISCUSSION

During the seismic calm periods the overall behavior of the underground water level shows a gradual long term variation in water level causing by the rainfalls. This variation has a period of a year in accordance to the rainfall annual periodicity. The underground water temperature shows the well known seasonal, one year periodic variation with amplitudes ranging between 0.6 °C in the deepest well, NYMF (10.0m), up to 8 °C in the shallowest well, ASS2 (2.20m), and a phase lag with respect to the atmospheric temperature which depend on the depth of the well. These long term variations do not bother the detection of any sharp change caused by other reasons.

Sharp changes in the underground water level and temperature can be related to earthquakes with epicenters in this area as pre- and post- seismic phenomena (Asteriadis and Livieratos 1989 a, b, Asteriadis and Contadakis 1990). There is not a unique response pattern of these changes for all the wells, a fact which is also reported by others investigators (Wakita 1982, Ma Zongjin et al. 1990). The magnitudes of these changes vary between a few centimetre to 90 centimetre in the water level changes and from a few tenths up to one degree centigrade in the temperature changes. Figure 2 displays an example of the underground water level variations. Our results indicate that there is a correlation between the magnitude of the change in underground water level and temperature and the magnitude of the shock as well as the epicentral distance of the respective station. The nature of our data does not permit the expression of this correlation in figures but it can be easily detected from our data. Finally the water level and the temperature changes occur in a time interval up to seven days before or after the associated shock (see also Asteriadi and Zioutas 1990), except the disturbances in underground water level and temperature which correspond to the strongest shock observed during this investigation. This earthquake occurred in 1990 December 21, has a magnitude of 5R and the corresponding disturbances in underground water level and temperature started 20 days before. This fact is found in accordance with the suggestion of Zongjih et. al. (1990) that the precursory disturbances in underground water precede the event by 10 to 20 days, if the strength of it is 5R. In general we may say that the observed time lag of the precursory disturbances in underground water.
level and temperature are in favor of the suggestion of Zongjin et. al. (1990) that they are correlated with the magnitude of the associated event.

**Figure 2.** Underground water level variation at the Nymfopetra station in the period between January 1 and March 31, 1989. Dots indicate the earthquakes which have occurred the same period in magnitudes of the Richter scale which is given on the left y-axis.

Table 2 displays a brief statistics of the network response on the seismic activity of the region in the time interval between 1988, January 1 and 1989, August 31. In this time interval all the stations were in work. So we have an homogeneous sample. During this period 143 shocks with epicenters in the area of interest have occurred. In this table the number of the sudden changes in the underground water level (row named "W/L") as well as in the underground water temperature (row named "Temp.") at each well station within the above mentioned time interval are given. Those changes which can be attributed to a particular shock are given in the columns under the heading A while those changes which occur within an earthquake swarm or sequence are given in the columns under the heading B. The number as well as the percentage of the total number of the shocks which can be correlated with a sudden change in water level or in temperature are displayed in the last two respective columns.

Table 2. The observed sudden changes in underground water level and temperature and the earthquakes which have occurred in the above mentioned area within the time interval 1988, January 1 and 1989, August 31.

<table>
<thead>
<tr>
<th></th>
<th>Assiros</th>
<th>Liti</th>
<th>Melissourgos</th>
<th>Nymfopetra</th>
<th>Shock</th>
<th>%</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>A</td>
<td>B</td>
<td>A</td>
<td>B</td>
<td>A</td>
<td>B</td>
</tr>
<tr>
<td>W/L</td>
<td>26</td>
<td>3</td>
<td>40</td>
<td>5</td>
<td>30</td>
<td>4</td>
</tr>
<tr>
<td>Temp</td>
<td>19</td>
<td>5</td>
<td>38</td>
<td>8</td>
<td>21</td>
<td>6</td>
</tr>
</tbody>
</table>

During the above mentioned time interval only two water level changes can be hardly attributed to a shock (Nymfopetra and Liti station at JD 2447516), while no temperature disturbance had ever failed to be correlated with a respective shock. That is the probability that an event follow an observed sudden change in the underground water level or in temperature in a well of the network is practically 1. This means that the necessary condition is fulfilled.

On the other hand 40 shocks cannot be connected with any water level change and 52 shocks failed to be connected with any temperature variation. This means that the probability that an event will
produce an observed water level change in a well of our network is 0.72 and the probability that an event will produce an observed temperature change is 0.64. That is, the sufficient condition is fulfilled with a probability of 0.72 for the underground water level variation and 0.64 for the underground temperature variation. Asteriadis and Zioutas (1990) have found that these two conditions are fulfilled with probabilities 0.88 and 0.65 respectively for the underground water level variation in the period between 1983 December 20 and 1986 August 31. The relative low probability for the verification of the sufficient condition is to be expected in view of the different response pattern of the well stations in the shocks as well as on the dependence of the magnitude of the water level changes on the epicentral distance.

4. CONCLUDING REMARKS

The results of this investigation indicate that there is a correlation between the underground water level and temperature variation and the microseismic activity in the broader area of the network manifested by sharp changes of these quantities within 7 days before the shock. However the time interval between the 5.0R shock of 1990 December 21 and its associated disturbance in underground water level and temperature, which is 20 days, indicates that a correlation between the event magnitude and the time lag of the associated disturbance may exist.

Crustal movements, expansion or contraction of the ground cause changes in the pore pressure of the underground rocks and deformation of the underground water system. Therefore the seismogenic process is expected to cause such a deformation. The exact way that this deformation manifest itself depend on many factors e.g. the epicentral distance the magnitude of the shock and the tectonic environments of the well network. The last factor presumably is responsible for the different response pattern of the different wells of the network.

REFERENCES


RUPTURE ZONES OF STRONG EARTHQUAKES IN THE THESSALIA REGION,
CENTRAL GREECE

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INTRODUCTION

The Thessalia region, central Greece, has experienced three
strong earthquakes in the present century: 30 April 1954,
Karditsa, M=7.0; 8 March 1957, Magnesia, M=6.8; 9 July 1980,
Magnesia, M=6.5. This region has been selected as a European
earthquake prediction test-site. It is, therefore, of special
interest to study the faulting process by defining the lateral
extent of rupture zones of these earthquakes and examining the
previous seismic history of the region.

DATA, METHODS AND MAIN RESULTS

A large number of papers has been taken into account to
collect information about the earthquake activity and tectonic
features of the region. In a forthcoming full paper all these
studies will be cited. Here only a few papers will be referred.

Figure 1 shows the main neotectonic lines (sharp lines),
surface fault breaks (heavy dashed lines) associated with the
1954 and 1980 strong shocks, the epicentres of known strong
(M $\geq 6.0$) earthquakes from the ancient times up to the present
(circles; cross means that two events had the same epicentre,
solid circles= epicentres of the three strong events mentioned
above), the strongest aftershocks of the three last, strong
events, as well as the degree VIII isoseismals (sharp curves)
and the rupture zones (heavy curves) of the three last strong
events. The main results have as follows:
(1) Before 1621 there was a lack in the earthquake reporting.
(2) There is another lack during the 19th century.
(3) There is an apparent time clustering of strong events.
(4) The southern part of Thessalia has ruptured three times
in the last about 38 years after an apparent quiescence since
1773. The rupture zones practically abut and do not overlap.
(5) The northern part did not ruptured since 1781.

Figure 2 shows the G-R diagram of Thessalia: triangle=
cumulative frequency, $N$, of $M \geq 4.5$ for 1964-85 as integrated
to 1901-85, open circle= single frequency, $n$, square= cumulative
frequency, $N$, of $M \geq 6.0$ for present century (all data from
Comninakis and Papazachos,1986), solid circle= $N$, of $M \geq 6.0$ for
present century (data from Ambraseys and Jackson,1990). One
may observe that:
(1) There is a magnitude gap of $5.8 > M > 5.2$ in $N$. Extending
$N$, to lower magnitudes it results that the gap persists over
the whole century (magnitude error $\leq 0.3$). This means that
the magnitude range of about 5.8 - 7.0 is characteristic in
The asalia.

(2) Reliable b-value can be determined only for M $>$ 6.0. From Comninakis and Papazachos (1986) and Ambraseys and Jackson (1990) data we get $\log N = 6.23 - 0.88M$ and $\log N = 6.78 - 0.99M$, respectively, which imply mean return period of events of M $>$ 6.0 equal to 10.2 years and 13.2 years, respectively. For M $>$ 5.8 we get mean return periods equal to 6.8 years and 8.4 years, respectively.

Previous results indicate that should the seismicity pattern of the southern Thessalia area repeat in its northern part than a migration of the activity would be expected northwards in the near future.

REFERENCES


ACKNOWLEDGEMENTS

This is a preliminary paper of a seismological study of Thessalia region which has been undertaken with the financial support of the European Centre on Prevention and Forecasting of Earthquakes, Council of Europe, Athens, Greece.
ABSTRACT: We are carrying out since 1988 repeated relative gravity observations along the western part of the North Anatolian Fault Zone (NAFZ), Turkey. Analysis of seven observation periods shows gravity decrease at stations near the fault especially in the middle and at the eastern part of the network. All other stations show moderate, statistically insignificant gravity increase. Gravity changes/year are converted into height changes using a gradient $\frac{dg}{dh} = -2000 \text{ mm*sec}^{-2}/\text{m}$. These height changes are compared with vertical strain accumulations at particular stations, determined by other experiments. Obviously these strain components are valid only for the crust near the surface. Height changes as observed by gravity changes are probably caused by stress accumulations in the deeper earth crust and by mass movement within the earth crust.

INTRODUCTION

Within the so-called Turkish-German Earthquake research project which takes place at the western end of the North Anatolian fault zone, Turkey, (Berckhemer et al., 1991) we have established a gravity network to observe temporal gravity changes in the preparation phase of an awaited earthquake of magnitude $> 6$. The size and the geographical location of the network are shown in Fig. 1. More details about the network, measurements and adjustment technique are given in (Demirel and Gerstenecker, 1989, and Akin et al., 1991).

Fig. 1: Location of the Gravity Network at the North Anatolian Fault.
Since September 1988 we have carried seven observation campaigns, whereby we have collected altogether about 4995 observations. Analysis of the data have shown significant gravity changes, which we will correlate with some results of the other experiments.

**TEMPORAL GRAVITY CHANGES**

The gravity values of each station at any epoch is estimated by least square adjustment. Linear regression analysis of gravity values with time delivers the gradient "gravity changes/ year" \( \frac{\text{dg}}{\text{a} \cdot \text{year}} \), which we have plotted in Fig. 2. The isoline distance is 10 nm*sec\(^{-2}\)/year.

**Fig. 2 Turkey Gravity Changes / Year 1988-1992**

*Isoline distance: 10 nm*sec\(^{-2}\)/year*

The plot shows two major structures of significant gravity decrease:
- between the stations Abant and Mudurnu at the eastern part of the network
- between the station Dokurcun and Göynük.

At all other stations of the network more or less insignificant mainly positive temporal gravity changes are found. We have inverted the gravity changes in height changes using an average gradient \( \frac{\text{dg}}{\text{dh}} = -2000 \text{ nm*sec}^{-2}/\text{m} \) (1 nm = 1*10\(^{-9}\)m) (Atzbacher and Gerstenecker, 1992). Height changes/year dh/a at Dokurcun and Abant are listed in table 1.
TEMPORAL STRESS AND STRAIN ACCUMULATION

Height changes can be caused by elastic deformation of the earth crust due to stress accumulation during an earthquake cycle. Between Dokurcun and Abant strain changes were observed in the following experiments:

- deformations of a geodetic horizontal network (10km * 10km) near Taskesti
- local temporal modulation of the response of borehole tiltmeters due to tidal forces at six "multiparameter stations".

(Zschau et al., 1991) have deduced from these observations principal strain components, showing the strain accumulation in the area between Dokurcun and Abant from 1985 -1989 (Table 1). The averaged shear strain accumulation is between 1 to 3*10^-6/year.

<table>
<thead>
<tr>
<th>Station</th>
<th>e_{11} 1*10^{-6}</th>
<th>e_{22} 1*10^{-6}</th>
<th>&amp;</th>
<th>dh/a mm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dokurcun</td>
<td>-7.8</td>
<td>14.5</td>
<td>33.2</td>
<td>12</td>
</tr>
<tr>
<td>Abant</td>
<td>-5.6</td>
<td>4.2</td>
<td>80.3</td>
<td>13</td>
</tr>
</tbody>
</table>

Assuming a homogeneous earth crust with a surface free of normal stress (\( \sigma_{33} = 0 \), Young-modulus = 7.25 N/m², Poisson-number = 0.25) we can estimate the vertical strain changes \( e_{33} \). The elastic parameters are taken from (Roth, 1983).

Results for the station Abant and Dokurcun are shown in Table 2.

<table>
<thead>
<tr>
<th>Station</th>
<th>( \sigma_{11} ) MPa/a</th>
<th>( \sigma_{22} ) MPa/a</th>
<th>( e_{33} ) 1*10^{-4}/a</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dokurcun</td>
<td>-0.06</td>
<td>0.20</td>
<td>-0.5</td>
</tr>
<tr>
<td>Abant</td>
<td>-0.07</td>
<td>0.04</td>
<td>0.1</td>
</tr>
</tbody>
</table>

The observed stress changes agree well with values obtained by (Roth, 1983), who modelled stress and strain changes during earthquake cycles at the eastern part of NAFZ.

Vertical compressional strain changes are found for Dokurcun. Stress accumulation at this station is connected with height decrease. We cannot explain the negative gravity changes at Dokurcun with horizontal stress accumulation of the earth crust near the surface. The reason for gravity changes are probably mass movements in the subsurface like groundwater. Dilatation in the deeper crust also can predominate the effects of vertical strain accumulation at the surface.

At Abant tensional strain changes will cause positive height changes. If height changes are caused only by the dilatation of the
homogeneous earth crust, the free air gradient \( \frac{dg}{dh} = -3090 \text{ nm*sec}^{-2}/\text{m} \) have to be used, to convert gravity changes to height changes. In this case the homogeneous horizontal stresses in table 2 has to be applied to a homogeneous earth crust thicker than 100 km. Such an idea is quite unrealistic. Probably also there inhomogeneous stress fields in the deeper earth crust as well as mass movements are responsible for the found gravity changes.

**CONCLUSION**

Observed strain- and stress accumulation along the western part of the NAFZ can be deduced by means of geodetic observations in horizontal and gravimetric networks as well as by borehole tiltmeter measurements. The results agree with models of stress and strain accumulation within an earthquake cycle developed for the eastern part of the NAFZ.

We are not able to explain height changes, which we have calculated from temporal gravity changes, by the observed accumulated horizontal surface stress components. The stress accumulation is too small (Abant) or has the wrong sign (Dokurcun). Stress components observed at the surface do not represent the real subsurface inhomogeneous stress field.

**REFERENCES**


1. INTRODUCTION

Strain analyses are frequently performed on the basis of repeated measured local or regional geodetic networks. Doing so, strain is usually considered homogeneous for the whole polygon or, at least, for parts of it. However, this approach does not always work in reality. This can be done by dissecting the area concerned into finite elements. The triangle was the finite element used for computations of the deformation parameters in Gruiu-Caldarusani geodynamic polygon (fig.1).

Taking into account tectonic and geodynamic considerations, the Gruiu-Caldarusani polygon is situated in the Northern part of the Moesian Platform, on the transition zone with the Pericarpathian Depression.

The crystalline basement is divided into different sectors, the Dobrudjan and Vallachian, separated by an important tectonic accident called the Intramoesian Fault which extends quite some distance, especially in the Southern part of the Black Sea area.

This one seems to be a tectonic line of major importance which divided the Platform into different domains with various sedimentary conditions, depending on the petrological constitution of the basement and the geothermal regime.

Geodetic measurements of triangulation and trilateration were performed in the polygon in the year 1980, 1984, 1989 (Ghitau et al, 1989).

For the periods mentioned above, the compensating coordinates were resulted after processing the initial data. The base data of the calculation for the deformation parameters are the displacement vectors x. The paper presents the deformation parameters for each finite element of the networks.
2. MATHEMATICAL TREATMENT.

a) Deduced displacement vectors and their error ellipses.

Displacement vectors \( \mathbf{x} \) are deduced from the new and old position vectors \( \mathbf{X}, \mathbf{X}' \):

\[
\mathbf{x} = \mathbf{X} - \mathbf{X}'
\]

(1)

Variance-covariance matrix \( \sum_{\mathbf{x}} \) of the displacement vectors can be obtained from variance-covariance matrix \( \sum_{\mathbf{X}}, \sum_{\mathbf{X}'} \) of new and old position vectors as follow:

\[
\sum_{\mathbf{x}} = \sum_{\mathbf{x}'} + \sum_{\mathbf{x}''}
\]

(2)

where:

\[
\sum_{\mathbf{x}} = \begin{pmatrix}
\sigma_{xx}^2 & \sigma_{xy}^2 \\
\sigma_{xy}^2 & \sigma_{yy}^2
\end{pmatrix}
\]

\[
\sum_{\mathbf{x}'} = \begin{pmatrix}
\sigma_{x'x}'^2 & \sigma_{x'y}'^2 \\
\sigma_{x'y}'^2 & \sigma_{y'y}'^2
\end{pmatrix}
\]

\[
\sum_{\mathbf{x}''} = \begin{pmatrix}
\sigma_{xx''}^2 & \sigma_{x'y''}^2 \\
\sigma_{x'y''}^2 & \sigma_{y''y''}^2
\end{pmatrix}
\]

\( \sigma_{xx}^2, \sigma_{yy}^2 \) is variance of \( x, y \) components of the determined displacement vectors respectively, and \( \sigma_{xy} \) is covariance. Similarly, \( \sigma_{x'x}'^2, \sigma_{y'y}'^2, \sigma_{x'y'}^2, \sigma_{x''y''}^2 \) are variances and covariances of new and old positions. Usually \( x \) is taken as the N-S and \( y \) as E-W directions in geodesy.

b) Strain analysis

Strain parameters are deduced from displacement vectors by the following method expressed in matrix form. Let \( \mathbf{x} \) be the displacement vector, \( \mathbf{U} \) the strain matrix, and \( \mathbf{A} \) the coefficient matrix:

\[
\mathbf{X} = \mathbf{A} \mathbf{U}
\]

(4)

where:

\[
\mathbf{X} = \begin{pmatrix}
x_1 \\
y_1
\end{pmatrix}
\]

\( i = 1,2,3 \)

(5)

\[
\mathbf{U} = \begin{pmatrix}
u_1 \\
u_2 \\
u_3 \\
u_4 \\
u_5 \\
u_6
\end{pmatrix}
\]

\[
\mathbf{X} = \begin{pmatrix}
x_0 \\
y_0
\end{pmatrix}
\]

\[
\varepsilon_{xx} \\
\varepsilon_{xy}
\]

(6)
\((x_0, y_0)\) represents shift. \(\varepsilon_{xx}, \varepsilon_{yy}\) are the strain components and \(\omega\) is the rotation. We represent the coordinates of three terminal stations of a triangle with respect to its centre:

\[
\begin{vmatrix}
\frac{dx_1}{dy_1} & \frac{dx_2}{dy_2} & \frac{dx_3}{dy_3} \\
\frac{dx_1}{dy_1} & \frac{dx_2}{dy_2} & \frac{dx_3}{dy_3} \\
\frac{dx_1}{dy_1} & \frac{dx_2}{dy_2} & \frac{dx_3}{dy_3}
\end{vmatrix}
\]

then:

\[
\begin{vmatrix}
1 & dy_1 & dy_2 & dy_3 \\
1 & 0 & 0 & 0 \\
1 & 0 & 0 & 0 \\
0 & 0 & 1 & dx_1 \\
0 & 0 & 1 & dx_2 \\
0 & 0 & 1 & dx_3
\end{vmatrix}
\]

Solving eq. 4 we obtain:

\[
U = A^{-1} X
\]

Dilatation \(\Delta\), maximum shear strain \(\gamma_{\text{max}}\), pure shear \(\gamma_{1}\), engineering shear \(\gamma_{2}\), principal strain \(\varepsilon_{1}\) and \(\varepsilon_{2}\) are:

\[
\begin{align*}
\Delta &= u_2 + u_6 = \varepsilon_{xx} + \varepsilon_{yy} \\
\gamma_{\text{max}} &= \sqrt{(u_2 - u_6)^2 + (u_3 + u_5)^2} = \sqrt{(\varepsilon_{xx} - \varepsilon_{yy})^2 + 4\varepsilon_{xy}^2} \\
\gamma_{1} &= \varepsilon_{xx} - \varepsilon_{yy} ; \gamma_{2} = 2\varepsilon_{xy} \\
\varepsilon_{1} &= (\Delta + \gamma_{\text{max}}) / 2 ; \varepsilon_{2} = (\Delta - \gamma_{\text{max}}) / 2
\end{align*}
\]

The authors made a program which is able to compute all the strain parameters above mentioned. The program is structured in three sections, as follows:

1. Creation and working with data files.
2. The computation of strain parameters as described in the [10] relations.
3. Graphic representation of the triangular network.


Compensation by indirect observations method has been used for the computation of the network benchmarks coordinates. After the compensation, the \(x, y\) coordinates that are denoted with \(X\) vector in the strain analysis are at the measurement time.

Putting into practice the above working method, we have calculated the deformation parameters for each finite element of the polygon in both periods of measurements.

The \(\mu\)-strain \(10^{-5}\) is the measure unit of the maximum principal strain \(\varepsilon_{1}\), and the minimum principal strain \(\varepsilon_{2}\). One can observe that, the maximum
principal strain values cover a limited domain, between $-0.31 \times 10^{-5}$ and $+3.40 \times 10^{-5}$, for the first period (1980 - 1989). Moreover, only four values from the eleven of them, which correspond to the finite elements are negative, being close to zero.

For the second period (1984-1989), only two values are negative, namely $-0.28 \times 10^{-5}$ (in the 4-3-6 triangle) and $-0.42 \times 10^{-5}$ (in the 4-5-7 triangle). The other values are grouped in the $[+ 0.06 \times 10^{-5}, + 2.15 \times 10^{-5}]$ interval. So, taking into account only the tendency of this area, it would be an extension one. But considering the evolution of the minimum principal strain $e_{2}$, in the 11 finite elements we can conclude that the general aspect of this area is one of compression. This idea relies on the minimum principal strain $e_{2}$ values which have negative values for both periods.

Thus, the minimum principal strain $e_{2}$ has values in the interval $[-0.06 \times 10^{-5}, -5.69 \times 10^{-5}]$, for the 1980-1984 period, and in the interval $[-0.15 \times 10^{-5}, -3.55 \times 10^{-5}]$, for the 1984-1989 period.

Figures 2 and 3 are a good proof for the above hypotheses.

Fig. 2 Earth’s strains in the Gruiu-Găldărușan network during 1980-1984. Magnitude and orientation of the principal axes.

Fig. 3 Earth’s strains in the Gruiu-Găldărușan network during 1984-1989. Magnitude and orientation of the principal axes.

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INTRODUCTION

The Vrancea region is an important seismic area of Europe being situated in the Romanian Carpathians where a sharp of direction of their folded structure takes place. Specific peculiarities of the seismic activity of this European region (earthquake energy, area of macroseismicity, character of the seismic foci and their surface distribution) are only compared with seismic foci from Hindukush Mts. and Bucaramanga region (Colombia).

The Romanian seismological studies showed that the Vrancea seismic region consists of an epicentral area of 2000 km$^2$ where intermediate earthquakes (depth of 70 - 200) were produced in the folded internal zone of the Carpathians and another area of 7000 km$^2$ in the external zone (placed in the Carpathian fore - deep) with crustal earthquakes (depth of 0 - 40 km) (Cornea, Lazarescu, 1980). The energy of these intermediate earthquakes is important, they causing major damages and human lives losses. The macroseismic area of the strong earthquakes is large, being extended between Moscow and the southern Greece (Fig. 1).
The geodynamic studies of this region have been carried out in the external area of the crustal earthquakes by repeated high precision levelling measurements. These works performed along three levelling lines in the 1937 - 1990 period (Fig. 1). The results of the geodetic surveys in the deep geological and geophysical context of the investigated region are discussed.

GEOLOGICAL AND GEOPHYSICAL CONSIDERATIONS

The levelling lines are placed in the Carpathian foredeep. Here, the sedimentary cover has a great thickness. The unconsolidated crust has thicknesses greater than 15 km in the Focșani - Odobești Depression that is situated in the eastern part of the Vrancea region.

On the levelling alignment Marașești - Focșani - Rimmicu Sarat - Buzau the Quaternary sediments are thick, being represented by the Cindești layers (gravels with rare insertions of sands) that have thicknesses of about 500 m and lie on a Pliocene complex (argillaceous marls, clays and argillaceous sands) with a 3000 m thickness (Mutihac). The mentioned area has been affected by an active subsidence process that began in Sarmatian and especially continued in Pliocene. Some general geodynamic characteristics from Quaternary, are showed in the Romanian neotectonic map (Brandabur et al., 1970).

The recent vertical crustal movements map (Popescu, Dragoescu, 1985) indicates some subsidence movements of the topography surface (with the maximum vertical velocity of 2.00 mm/y). These subsidence movements of the medium intensity in the western part of the Vrancea intermediate earthquakes region, extending than over the orogenic area of the Eastern Carpathians (in the southern part of the Trotuș Line).

After DSS data the regional structure of the crust is defined by the evolution of the Peceneaga - Camena and Capidava - Ovidiu crustal fractures. For example, the Peceneaga - Camena fault produces a 10 km jump of the Mohorovičić discontinuity, separating a northern crustal block with great thickness (47 - 48 km) (the North - Dobrudjan Orogen and its north-western prolongation) and a southern one with a smaller thickness (38 - 39 km) developed in the Moesian Platform. A depression zone with depths more than 50 km is separated in the Vrancea region on the crust thickness map of Romania.

GEODETICAL DATA

The performed works refer to the vertical displacements of the Earth's crust surface from repeated high precision levelling along three lines: Buzau - Focșani - Marașești, Marașești - Tecuci - Filești and Filești - Faurei - Buzau (Fig. 1).

Fig. 2 presents the vertical relative displacements along the Marașești - Filești line for the 1961 - 1986 period. In this time interval produced two strong earthquakes on March 4, 1977 (M=7.4) and August 30, 1986 (M=7.0). The entire area was affected by a constant subsidence movement. The correlation between this image and the deep geological structure indicates a different geodynamic behavior in the area of the crustal fault Peceneaga - Camena. Local intense upheaval and subsidence movements that are in corresponding with this important fault and with another superficial fracture of the sedimentary cover, appear here too.

The heights differences of the geodetic benchmarks on the Buzau - Marașești line showed same general features. These relative data are reported on the initial measurement from 1937 and in comparison with the Marașești reference point. The following measurements (1961, 1983, 1990) pointed out the general tendency of the benchmarks subsidence from north.
(Marașești) to south (Buzau). The upheaval tendencies of the benchmarks appeared in the middle part of the profile where the sedimentary has a maximum thickness (about 18km) and contains more faults. The most intensive tendencies have been pointed out in the northern part of Buzau city, where some Paleozoic magma circulated along an important fault, determining positive magnetic anomalies (ΔZ) at the surface.

![Diagram](image)

Fig. 2. Vertical relative displacements and crustal structure along Marașești-Filești profile
(1-sedimentary, 2-upper crustal layer, 3-lower crustal layer, 4-fault)

Strong seismic events with \( M \geq 7.0 \) have occurred in each period between the geodetic measurements (1937 - 1961; 1961 - 1983; 1983 - 1990). The ground subsidence had the maximum values in the 1990 of about 110 mm - after 53 years from the initial measurement in the northern part of the geodetic line (Buzau).

**GENERAL REMARKS**

The present dynamics of the crustal surface in the Carpathian Arc Bend (Carpathian fore-deep) is represented by a general subsidence movement. On this general tendency, local zones with modification of intensity and sense of these vertical movements have been pointed out. Some of them have been correlated with the major crustal faults of the Romanian territory (for example the Peceneaga-Camena fault), that produce important modification of the crustal thickness and internal structure. A fluid transfer between depth and surface took place along these tectonic lines, influencing the recent vertical movements.

The absence of the measurements immediately before and after the strong earthquakes (1940, 1977, 1986, 1990) determined the lack of
information about the effects of these seismic major events on the crust surface.

The levelling measurements along Rm. Sarat – Cioraști line (eastern from Rm. Sarat) before and after the Rm. Sarat crustal earthquake (M=3.9) (produced on September 1, 1991) have been performed in 1991. A relative subsidence have been observed at the most benchmarks. The largest subsidences were about 15 mm. But upheaval movements with maximum values of 4 mm appear too. The correlation between subsidence and the heights of the levelling benchmarks have been observed too; the reduced altitudes have large values for the subsidence.

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EVIDENCE FOR POSSIBLE RELATION BETWEEN LOCAL IRREGULARITIES IN THE HORIZONTAL DEFORMATION FIELD AND MICROSEISMIC ACTIVITY OF THE MYGDONIAN BASIN (NORTHERN GREECE)

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1. INTRODUCTION

Geodetic methods have been proved very useful tools for the detection and monitoring of the earth crustal deformations. The accuracy of the today surveying technics approaches the level of about 1 to 2x10^-6 which correspond to 1 or 2 cm for a 10 km side length of a triangulation. Furthermore the interpolation by a least square collocation method (see for instance Dermanis et al. 1982) or by other analytical methods enable the determination of the horizontal deformation field with an accuracy of a few tenths of the microstrain (\(-10^{-6}\)). This fact enable us to follow up the possible variations of the deformation field of a seismic active area over time intervals of one or two years with the help of repeated surveys.

Using the repeated measurements in the decade 1979-1990 of the 16-point trigonometric network, located in the seismic active area of lake Volvi in Northern Greece, a continuous strain field was determined for the area and for each epoch of the corresponding surveys.

In this paper we discuss the pattern as well as the local and time variations of the horizontal deformation field of the area in connection with the seismic activity in the area.

2. THE AREA UNDER STUDY, OBSERVATIONAL MATERIAL AND ANALYSIS

Figure 1 (Vlachos 1980) displays the map of the area of the lakes Lagada and Volvi which was the locus of the major seismic events of 1978. In this map one can see the epicentres of the main shock as well as the epicentres of the largest foreshock and aftershock as it is given by the Geophysical Laboratory, University of Thessaloniki. The trace of the surface faults which were observed after the seismic activity (Papazachos et al. 1979) are also shown. The 16-point trigonometric network is also displayed in this map. This network has been measured during seven 15-days surveys, which include direction observations and was performed first in 1979 July-August, and was repeated in 1980 April, 1981 July, 1982 April, 1983 April, 1989 May, and 1990 July.

The adjustment of the network was based on the free network methods. The standard de-
viation of the derived distances lie within the limits of 0.6 cm and 3.6 cm. The 2-D local reference frame has the x-axis parallel to the W-E direction (Dimanidis et al. 1983, Elemenoglou et al. 1989, Goutsi et al. 1990).

The analysis for the derivation of the displacement and velocity field (using as a reference epoch August of 1979) was performed by splitting up the observational data into a systematic and a random part and with development of the systematic part into a series of an appropriate system of functions using Legendre and Hermite polynomials (Asteriadis et al. 1988, Schwan & Asteriadis 1989) and subsequently the strain parameters and the strain history of the deformation field of the area were determined by analyzing the continuous displacement field (Asteriadis et al. 1992).

Finally, the seismic data for the earthquakes with epicenters in the area extending from \( \varphi = 40.3^\circ \) N to \( \varphi = 41.0^\circ \) N and from \( \lambda = 22.7^\circ \) E to \( \lambda = 23.8^\circ \) E, which include the area under study, were taken from Scordilis (1985) for the years 1981 to 1983, the bulletin of the Seismological Institute of the National Observatory of Athens for the years 1984 to 1988, and the archives of the Geophysical Department, University of Thessaloniki for the years 1989 and 1990.

### 3. RESULTS AND DISCUSSION

Our results show that there is local as well as time variations of the deformation field. Figure 2 displays an example of the deformation field expressed in terms of dilatation. In this Figure the maximum as well as the minimum normal strain are plotted together with the isodilatation lines. Continuous lines mean elongation and dashed lines contraction. The epicenters of the microearthquakes which occurred within six months before (filled cycles) as well as of those which occurred within the following six months (open cycles) the epoch of the survey are also displayed.

First of all it should be noted that the sites of absolutely minimum value in dilatation \( \Delta \) or in maximum shear strain \( \gamma_{\text{max}} \) coincide or fall very close to the trace of the surface faults which appeared after the great earthquake of 1978 June 20 (see Figure 1). This might be explained in terms of the rebound theory of earthquake generation according to which the zone along the fault is locked to pre-earthquake movements (Rikitake 1981). In addition, our results show that the absolute values of dilatation correlate with the microseismic activity in the region. This correlation may be expressed by the correlation coefficient between the mean absolute values of dilatation for the sites of the network stations and the earthquake frequencies for the corresponding to the epoch of the survey year, which is \( r = 0.76 \). The dilatation (in absolute values) for the years 1982 and 1989, years of minimum microseismic activity, is much smaller than that for the other years. The small value of the dilatation for the year 1981 for the sites of the stations 11, 13, 15, and 16 may be explained by a reverse post-seismic deformation, in an analogous way to that which has been observed after the earthquake of Kanto, in Japan (Rikitake 1976 and reference therein). As it is shown in Figure 2, most of the earthquakes, which have occurred in 1981, occurred before the period of the 1981 survey (filled circle in the Figure) and their epicenters fall in the area between or near the surface fault (see Figure 1), where the sites of the network stations 11, 13, 15, and 16 are situated. The strongest earthquake of this period is among these events. This kind of behavior is especially pronounced in the maximum shear strain values of the sites of these stations. The mean maximum shear strain for these sites are strongly anticorrelated with the frequencies of the earthquakes occurred within a year before the survey period (correlation coefficient \( r = -0.9 \)).

The deformation field of the area at the epoch of April of 1982, which was seismically a relatively quiescent year, shows a slight uniform N-S elongation with a normal strain of 2.4 \( \mu \)str, in accordance with the N-S elongation which is indicated by Geological and Seismological studies (Papazachos et al. 1986, Mercier et al. 1979, Carver and Bollinger 1981, Sutleris et al. 1982).
However, the deformation field of the area at the other epochs, which correspond to years with increased microseismic activity, presents regions of extension as well as regions of contraction.

The location of the contraction areas relative to the locus and time of intensive microseismic activity favors the explanation that these deformation fields were created by a post-seismic reverse deformation mechanism. As it is shown in Figure 2, in the dilatation field of April 1981, the contraction area coincides with the area of intense microseismic activity which preceded the survey period by one month. The elongation and contraction directions are exactly reversed to those at the regions of extension, which are the same to those found from Geological and Seismological studies of the region. This is what one should expect to happen by this mechanism if a uniform extension area before the period of the microseismic activity is assumed. Similarly, the contraction areas in the deformation field at the epochs of April 1983 and May 1990 are located at the northwest site in 1983 and at the northwest and northeast site in 1990, close to the locus of intense microseismic activity for these years. As far as the areas of extension concern, the preferable direction for the elongation is ENE-WSW in close accordance to the seismological results, except the areas near the existing surface faults where the direction of the elongation trend to E-W.

The existence of regions of contraction in this area which is dominated by extension is also confirmed by seismological evidences (Scordilis et al. 1989, Hatzfeld et al. 1987).

As a conclusion we can say that the deformation field of the area shows minimum strain values in sites which are close to the system of the surface faults, which have appeared after the strong earthquake of 1978 June 20. In addition, there is local as well as time variations of the strain field which are well correlated with the microseismic activity of the area. This microseismic activity has its origin at shallow depths. Our results are consistent with those of the Geological and Seismological studies in the area. The deformation field is dominated by a N-S extension superimposed by local and time irregularities due to the local seismogenic process which is expressed by the microseismic activity on the area.

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ON UNUSUAL HUMAN BEHAVIOR BEFORE EARTHQUAKES

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Earthquake prediction is the most important and spectacular field of research in seismology.

The methods implemented in elaborating predictions also rely on investigations of pre-events or foreseeing revelations associated with the occurrence of earthquakes. A large proportion of the foreseeing phenomena associated with earthquake occurrence can be grasped in different ways by human beings through their sensory and extrasensory abilities.

Such involvements can be detected with people in the following conditions of body and mind:
- healthy and sound people having entirely preserved their ancestral sensibility to excitations from the environmental medium;
- people having become more sensitive who display a high sensitivity than normal, healthy and sound people;
- people whose mind works at its best and display a perfect reflex-conditioned activity;
- people enjoying special parapsychological powers providing them with foreseeing abilities.

This paper presents some aspects relative to foreseeing phenomena grasped by certain human beings prior to the earthquake of magnitude Ms=7.2 occurred on March 4, 1977 in the Vrancea region (Romania), obtained from a survey covering Prahova, Dimbovita and Arges districts. A zone located on the external side of the Carpathians, in the Carpathians Fore-deep has been investigated. In this particular area, the highly tectonised structure and composition of the crystalline bed gives rise to gravimetric and magnetic anomalies (Dumitrescu and Sandulescu, 1974). The geological and structural features specific of the zone are due to the Peripherical Carpathians fault and the regional faults that were strongly activated during the earthquake resulting in major macroseismic effects extending over a large area and reaching a seismic intensity of VII-VIII (MSK-1964 scale) which proved the activated faults to be particularly deep.

The investigation was essentially aimed at studying, recording and quantizing the macroseismic effects caused by the March 4, 1977 earthquake. As part of the research, lots of people were interviewed to supply detailed information on damages caused to both civilian and industrial buildings. The queried often volunteered additional, uncommon data which somehow appeared as the most attractive and exciting side of the survey. Much of this information referred to the “unusual” behavior noticed in both animals and humans prior to the quake.

Odd animal behavior as described by the interviewed was largely similar to the one studied and reported by many scientists around the world, particularly in Japan and the U.S. (Rikitake, 1978; Simon, 1975; Logan, 1977; Reasenberg, 1978; Kolev et al., 1988; Radu et al., 1978). As for the unusual human behavior prior to the earthquake, this also resembled the one earlier depicted in the literature (Keeton, 1976; Hill, 1976; Buffe and Nanevicz, 1976). More surprising, though, was the account by several people of ominous dreams they had had 24h before the quake occurred. Despite the risks implied by a foray into the arcana of
According to Jung, dreams are a crucial, indispensable source of information which develops in terms of a symbolic language of the cause-effect type as a consequence of brain excitation by various internal and external stimuli. Moreover, as these stimuli succeed one another at different time spans, their rate and intensity are changed, which makes it possible to estimate time lengths and determine some complex, fully developed space-time connections.

Scientists generally contend that psychology is a realm too far apart from physics, thus concluding that prediction theory as developed in the latter field can hardly be of any good in the former.

Classical research, however, especially recent investigations on simple physical systems, has demonstrated that individual prediction is ineffective in many physical systems owing to their instability. By contrast, for numerous various reasons, biological systems are known to be more stable than simple physical ones. This structural stability, which makes prediction possible, is a striking peculiarity of biological systems. Under these circumstances, it is not at all unreasonable to assume that psychology will some time in the future emerge as having a higher prediction-making potential than physics.

To underscore the usefulness of psychology as a predictive factor does not mean this can be regarded as the one straightforward approach to earthquake prediction, but rather that alongside other methods it can help improve the chances of scientific forecast in the seismologic field.

REFERENCES
A model of induced anisotropy in crystalline rocks

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Commonly minerals in deformed rocks show crystallographic preferred orientations and also mechanical textures i.e. layering and cracks, which are genetically connected to each other. It is of great interest how external conditions and physical processes influence on development of the texture in an initially isotropic material. In this work we introduce a model medium, which under directed stress is transferred from the isotropic to an anisotropic state. It corresponds to a polycrystalline monomineralic quartzite under physical conditions, where structural phase transformations take place.

The considered model medium consists of three solid state components. At the initial stressless state each of them is an isotropic polycrystalline aggregate of quartz crystals. The first component is assumed to be α-quartz forming a rigid “matrix”. The second component named “inclusions” consists of β-quartz which exhibits either elastic or plastic behaviour depending on physical conditions and stress state. At least we introduce limited regions of plastic behaviour appearing only during deformation in the matrix as well as in the inclusions. This third component we call “plastic microzones”.

Each inhomogeneous medium under stress is characterized by stress differences and stress fluctuations at the component boundaries. This means, that in our model medium stress concentrations appear around the inclusions in the rigid matrix. Additionally a temperature gradient exists between inclusions and matrix – β and α quartz – resulting in a zone with anomalous mechanical properties along the boundary. This is based on the negative Poisson coefficient of quartz in the temperature range 524°C to 580°C, i.e. stretching results in dilatation and not in shortening perpendicular to the stretching direction. The increase of volume amplifies the local stress magnitude, which at least induces the generation of the plastic microzones mentioned above. Assuming initially spherical inclusions the resulting shape of the plastic microzones depends on the nonhydrostatically stress. Generally they will get the shape of an ellipsoid with its long axis oriented parallel to the direction of maximal stress component. The volume of the microzones depends on the stress magnitude, the axis ratios on differential stresses.
First we consider the problem how the macroscopic anisotropic properties, i.e. the components of the bulk elasticity tensor, depend on the symmetry of the external stress and the shape and the concentration of the microplastic zones. For calculation the statistical theory of elasticity of microinhomogeneous media was applied to our model for the plane case.

The effective stiffnesses of the microplastic zones are defined by the rule of small elastic-plastic deformation

\[ \mu^k_j = \mu_k \left( 1 - \omega(\langle \varepsilon^k_i \rangle) \right) \]  

where 
- \( k \) is the number of component deformed plastically;
- \( \mu_k \) is the shear modulus of \( k \)-component;
- \( \omega(\langle \varepsilon^k_i \rangle) \) is the Ilyushin's function.

We can write the equation formulated for the effective stiffness tensor \( \varepsilon^* \)

\[ (\varepsilon^* + b^*)^{-1} = \langle c + b^* \rangle \]  

Here \( b^* \) is the auxiliary tensor given by

\[ g(c^* + b^*) = -I. \]  

The dependence of the tensor \( g \) on the anisotropy parameter is defined by the expressions

\[ g_{ijkl} = a_{ij(k,l)j} \]  

\[ a_{ijkl} = \frac{1}{8\pi^3} \int \int \tilde{G}(r,\alpha) r^{-\alpha} d\alpha dr \]  

\[ \tilde{G} = \int G(\tilde{r}) e^{-i\alpha r} dr \]

The integration in (5), (6) must be done over the effective volume of the inhomogeneous zones which has the form of an ellipsoid. After the integration of (3) and the symmetrization of (2) we establish that the tensor \( g \) depends on the surface of the inhomogenity and expressed in terms of shape anisotropy \( k \) which is determined below.

Equation (4) can be transformed to the form

\[ g_{ijkl} = g^0_{ijkl} + k^2 \phi_{ijkl} \]  

where \( g^0_{ijkl} \) are analogous to the isotropic tensor \( g \).

The function \( \phi_{ijkl} \) is a measure of plastic anisotropy.
We found that in the case of uniaxial and biaxial stresses the tensor \( g_{ijkl} \) is characterized by five independent components

\[
\begin{align*}
    g_{1111} &= g^0_{1111} \left(1 + \frac{2}{7} k^2 \frac{4\lambda_c + 11\mu_c}{2\lambda_c + 7\lambda_c}\right) \\
    g_{3333} &= g^0_{3333} \left(1 + \frac{2}{7} k^2 \frac{6\lambda_c\mu_c}{2\lambda_c + 7\lambda_c}\right) \\
    g_{1122} &= g^0_{1122} \left(1 + \frac{2}{7} k^2\right) \\
    g_{1133} &= g^0_{1133} \left(1 + \frac{6}{7} k^2\right) \\
    g_{2323} &= g^0_{2323} \left(1 + \frac{2}{7} k^2 \frac{3\lambda_c - 2\mu_c}{3\lambda_c + 8\lambda_c}\right)
\end{align*}
\]  

where \( \lambda_c, \mu_c \) are Lame parameters of comparison body, \( k^2 \) is the shape anisotropy of the microplastic zone.

As a result we obtain five components of the tensor \( e^* \) which depend on the concentration of the microplastic zones, the loading and the extend of plastic deformation.

The concentration of the microplastic zones \( \Theta_1 \) is expressed by

\[
\Theta_1 = (1 - \Theta_2) \left(1 - \frac{\Phi(\sigma_{a1} - <\sigma^1_a>)}{D_{\sigma a}}\right)
\]  

with \( \Phi(u) = \frac{1}{\sqrt{2\pi}} \int_{-\infty}^{u} e^{-\frac{t^2}{2}} dt \),

\( D_{\sigma a} \) - dispersion of the stress intensity,

\( <\sigma^1_a> \) - mean stress intensity in \( \alpha \)-phase,

\( \Theta_2 \) - inclusion volume concentration.

Therefore, the concentration of the microplastic zones is a function of the mean stress and its dispersion.

According to the Ashelby model the shape anisotropy \( k \) is described by the formula

\[
k^2 = (\eta - 1) \left(1 - \frac{A}{B} \frac{\Theta_2}{\Theta} \frac{A_2(1) - A_2(2)}{A_1 A_2(1) (\chi - 1) + A_2(2) (\chi - \sigma_{a2})}\right)
\]  

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where \( A = \frac{2}{3} \left(1 - \frac{1}{5(1 - \nu_1)}\right), \)

\[
\chi = \frac{\sigma_i^0}{\sigma_{s1}},
\]

\[
\bar{\sigma}_{s2} = \frac{\sigma_{s2}}{\sigma_{s1}},
\]

\[
B = \frac{1}{5} \left( \frac{5}{1 - \nu_1} - \frac{1}{3} \right) \eta + 1 + \frac{12}{7(1 - \nu_1)}.
\]

Both mechanisms are in agreement and depend on the type of the stress-strain state of the model, the concentration of the microplastic zones, the deformation rate and the temperature.

The model proposed in this paper gives quantitative results for induced anisotropy of the elastic properties in inhomogeneous media. This model also explains the development of the texture under the action of hydrostatic pressure.

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The National Building Code of Canada has provided guidance for earthquake resistant design since 1953. Third generation seismic hazard maps were included into the Code in 1985. These maps used the Cornell-McGuire method to map peak horizontal ground velocity and acceleration values for a probability of 10% in 50 years. The Geological Survey of Canada has now been charged with producing new seismic hazard maps for prospective use in the year-2000 edition of the Code. The new model should incorporate the following advances: improved evaluation of earthquake magnitudes, maximum magnitudes, and seismicity location and rates as a result of the last 10 years of data collection and research; new ground motion relations; multiple models of earthquake distribution reflecting the incomplete understanding of seismogenesis and explicit recognition that the short historical record of earthquakes has not identified all potentially active faults; additional ground motion parameters to PGA and PGV including spectral estimates allowing presentation of uniform hazard response spectra. It is important that new hazard maps be presented for discussion to the affected user groups by the end of 1994, well in advance of the proposed date of inclusion in the Code in order to facilitate general acceptance and a smooth transition to the new provisions.
INTRODUCTION

Different probabilistic methods are being applied in the new seismic hazard assessment of the Italian territory (see GNDT Working Group on Hazard Assessment, this volume). They correspond to increasing levels of computing complexity, requested to take advantage from the available seismotectonic knowledge.

Shorty, we refer to a Gumbel type approach (Epstein and Lomnitz, 1966) when speaking of a simple statistical analysis, based on the extreme value theory. Its best application is performed on a list of felt intensities, experienced at a site: obviously this approach is usually limited to old, historically well documented cities. The second method is referred as Cornell (1968) type approach. Here some assumptions on the equiprobable occurrence of the seismicity in each seismogenic source, as well as the independence of earthquake in time are needed. The method is usually applied when a good seismotectonic background is available but no particular knowledge supports the sequence of earthquakes in a given zone to be interpreted as a non-stationary process. The Poisson process is adopted in modeling earthquake occurrence, while the general Gutenberg-Richter exponential distribution describes the earthquake size.

Renewal process (Grandori et al., 1984) introduces a step of memory maintaining the seismotectonic zoning but removing the correlation between magnitude (or intensity) and interoccurrence times. The interoccurrence time distribution can be modeled by a combination of functions (e.g. with two distributions) where the short inter-times represent earthquakes occurred during a seismic crisis, while the long interoccurrences are the times between different crises. In this way, each site is characterized by:

a) the intensity distribution, which is obtained by the contribution of all the seismogenic zones, can be modeled by a function chosen among simple exponential type and multiple parameter double exponential functions; in a similar way, the source magnitude distribution is not forced to follow the Gutenberg-Richter relation, but can be modeled using various functions;

b) the interoccurrence time distribution is obtained directly from the data of each site (experienced or computed from the epicenter by an attenuation relation); when many samples of interoccurrence times are available, combined distributions (e.g. a Weibull plus a Gamma function) substitute the simpler exponential type distribution.

A very simple application of the renewal time-dependent process to a site in Southern Italy is presented in the following, with comparing the result obtained by the memoryless Cornell approach.

APPLICATION TO SOUTHERN ITALY

An application of renewal process is tempted for the same Southern Italian sites analyzed by the TERESA Project (see Barbano et al., 1989), where remarkably different estimates of seismic hazard were obtained by different operators who started from similar input sets. To give an example, the renewal
process applied to the town of Benevento (one of the chosen test sites of the TERESA project) is here presented and compared to the result obtained following Cornell hypotheses (i.e. exponential type distribution for both intensity and interoccurrence times).

The seismotectonic zonig proposed by Scandone et al. (1992) has been adopted: the distribution function of intensity greater than VI MCS for each source has been computed using the last 300 years of the Italian earthquake catalogue (Postpischl, 1985), choosing the best fitting among three different kinds of exponential distributions. Not all the considered seismogenic zones suggest the same intensity distribution type to be the most adequate for fitting (in Fig. 1 an example of intensity distribution for two sources is given): in general it can be said that a 2 parameter double exponential function is preferable when high intensities are experienced in the source, while a simple exponential function (Gutenberg-Richter type) can be accounted for the other cases.

![Fig. 1: Intensity distributions for two seismogenic sources](image)

Figure 1: Intensity distributions for two seismogenic sources; dotted line for simple exponential distribution; continuous line for double exponential with 1 parameter; dashed line for double exponential with 2 parameter distribution.

Then, the intensity distribution at the site of Benevento has been computed by attenuating each source contribution with an ad-hoc developed non-circular attenuation relation. In Fig. 2a stars describe the intensity distribution obtained in this way; on the contrary squares indicate the intensity distribution at Benevento when the Gutenberg-Richter relation for seismicity is applied to the sources, like in the Cornell approach.

Considering interoccurrence times (Fig. 2b) a combined function of Weibull and Gamma distributions (continuous line) fits very well the samples experienced at Benevento; the simpler exponential distribution (dotted line), that corresponds to the Cornell hypothesis of independence in time of earthquakes, has a worse fit in the shortest times.

The differences between renewal and Cornell approaches are finally evaluated in term of return period for intensity greater than, or equal to, VIII MCS, giving
similar results; 17.56 years following Cornell, and 19.45 considering renewal process.

This agreement, that suggests the Poisson process to be only lightly the cautious one, is very important when evaluating seismic hazard for national seismic code purposes. Other sites, in fact, have given much different estimates in following the two approaches to seismic hazard assessment: therefore, the comparison of the results may enhance different characteristics of the sources, as well as of the sites, moving towards a more complex and detailed definition of the seismic process.

The great improvement in the renewal approach is that no "a priori" hypothesis on the seismic process is needed, and the data support by themselves the best distribution to be applied. The modelling of interoccurrence times can consider...
not only the major earthquakes but even the smaller ones, interpreted as elements of a seismic sequences. Obviously this detailed analysis is possible only using complete segments of the earthquake catalogues. So a kind of memory has been introduced in the seismic process, making dependent the hazard estimates to the time when the forecast is made. A further control on the comprehension of the seismic process is therefore possible, deleting the last ten-years seismicity and comparing the forecast with the reality.

REFERENCES
SEISMIC HAZARD IN GREECE BASED ON RADIATION AND ATTENUATION MODELS OF STRONG MOTION

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The focusing effects of seismic intensity due to the radiation of seismic waves have been recognized for quite some time. Recently the influence of the directivity effects in assessing seismic hazard has been investigated. Based on the geographical distribution of seismic sources of the shallow earthquakes in Greece, the azimuthal attenuation model and the anisotropic radiation model of seismic intensity were utilized. The elliptical shape of the attenuation models were adopted. Using the available macroseismic information of the area studied, the main axes of the elliptical attenuation models were calculated. Azimuthal attenuation relationships of each seismic source were estimated and the dependence of the seismic intensity on the source properties, geometrical spreading and the anelastic attenuation was taken into account. Incorporation of these improved attenuation models in seismic hazard assessment was carried out for the area of Greece. Simultaneously, comparison of the the results of seismic hazard analysis based on these improved attenuation models were accomplished. The comparison of the results of seismic hazard analysis using the azimuthal variation and anisotropic radiation showed a good agreement. On the other side the comparison of the results of seismic hazard analysis using the anisotropic radiation of seismic intensity with those which have been calculated relied on the mean attenuation model presented some differences. as was expected. This paper present the necessity for improved models of earthquake source and attenuation law in seismic hazard assessment.

TO BE PUBLISHED IN SPECIAL ISSUE IN THE JOURNAL NATURAL HAZARDS
SEISMIC HAZARD AND SEISMOTECTONIC PARAMETERS FOR SOUTHERN TURKEY

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Introduction

The analysis of earthquake recurrence in terms of magnitude, intensity, peak ground acceleration and strain energy release in time and space may be used in the construction of seismic hazard zoning maps. Dealing with one of these parameters alone can be taken into account as a measure of seismic hazard and ensuing risk, but may not explicitly reflect the effect of various factors contributing to damage potential seismic zones. For example, a region of large earthquake occurrence does not necessarily mean an area of high degree of seismic hazard. In this study, we apply several methods to earthquake data from 1900 to 1990 to assess earthquake hazard in southern Turkey. Our intention is to combine the spatial distribution of different hazard parameters in order to prepare an overall zoning map. A combination of different parameters for regionalisation may well highlight the zones where earthquake occurrence has meaningful and significant lateral variations in damage potential.

The data set is based on the CD-ROM earthquake catalogues—the Global Hypocenter Data Base (USGS/EDBS). Most of the data are extracted from the Mediterranean catalogue (MED) which is uniform for magnitude \( M_a \). Some data come from the HDS (NEIS-preliminary Determination of epicenters) and the European catalogue (EUR) in addition to the ISC bulletins which are used to update the catalogue for especially recent years.

Methods

The method of extreme value distributions is a particularly suitable method when the earthquake catalogue is extended back to macroseismic observations and is clearly incomplete. For such catalogues, the use of increasing time intervals to extract extreme magnitudes is a way to mitigate such catalogue deficiencies. Completeness analysis suggests that annual extremes may not be
used for statistical estimation of seismic hazard for the whole area studied. An optimum time interval for extracting extremes while maintaining reasonable spatial resolution (cells of 0.4° side) is found to be four years in which case surface wave magnitudes Ms lower than 5.0 are eliminated. Formulation and discussion of this method can be found in various studies (e.g., Gumbel, 1958; Burton, 1990).

Gumbel's first, $G^I$, and third, $G^{III}$, asymptotic distributions of extremes are used for comparison and results obtained for recurrence rates of magnitude, intensity and peak ground acceleration.

The probability of a specific acceleration for a given magnitude can be estimated as a function of distance (McGuire, 1978; Bath, 1979; Campbell, 1981; Makropoulos & Burton, 1985; Joyner & Boore, 1988). Since there is no attenuation relation developed using earthquakes in southern Turkey alone, we prefer to apply the attenuation model proposed for the Aegean region (Makropoulos & Burton, 1985) to the earthquake catalogue for the estimation of peak ground acceleration in the study area. An example of the spatial distribution of pre-

![Figure 1 Interpolated contour map of peak ground acceleration distribution for an average return period of 100 years.](image)

dicted peak ground acceleration for a specified period of time is given in Fig.
1. As seen in this plot, the hazard level increases towards the Burdur region in comparison to the surroundings. The maximum acceleration expected in the Rhodes and Burdur areas is about 30 percent of gravity for an average return period of 100 years.

Several empirical intensity attenuation relations proposed for different regions of Turkey are compared and discussed in Burton & Yilmaztürk (in prep.). The spatial distribution of intensity recurrence for a given time period is then also estimated using a relation based on isoseismals from the Burdur and Antakya (north-east of Cyprus) regions.

The variety of methods used to contribute to the final zoning methodology also incorporates non-probabilistic parameters. For example, the non-probabilistic parameter $M_3$, the upper bound for earthquake magnitude equivalent to the maximum possible strain energy release encompassed by the seismic cycle, is estimated through the energy-magnitude relation (Bath, 1973; Makropoulos & Burton, 1983, 1985). The distribution of $M_3$ based on strain energy release, shown in Fig. 2, agrees well with the expected patterns of seismicity corresponding to the most active regions. The hazard level in the Burdur region using this parameter again shows high compared to most of its surroundings.

![Figure 2 Distribution of $M_3$ values.](image)
Preliminary Zoning Results

A set of hazard maps has been produced similar to the examples illustrated in Figs. 1 and 2 and these will be described elsewhere (Burton & Yılmaztürk, in prep.). The next step in the process is the synthesis of these hazard maps, and the different contributary components to damage potential they represent, into one zoning map. The preliminary criteria for synthesis at this stage of the study are tabulated below:

Table 1 A comparison of seismic zones in terms of maximum expected intensity (MSK), magnitude, the upper bound $M_3$ and peak ground acceleration (PGA) for the average return period of 100 years.

<table>
<thead>
<tr>
<th>Zone</th>
<th>Magnitude</th>
<th>$M_3$</th>
<th>I (MSK)</th>
<th>PGA</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>$\leq 5.0$</td>
<td>$\leq 6.0$</td>
<td>$\leq 5.0$</td>
<td>$\leq 0.030$</td>
</tr>
<tr>
<td>II</td>
<td>5.0-6.0</td>
<td>6.0-6.6</td>
<td>5.0-7.8</td>
<td>0.030-0.075</td>
</tr>
<tr>
<td>III</td>
<td>5.0-6.0</td>
<td>6.4-7.0</td>
<td>5.0-7.8</td>
<td>0.075-0.120</td>
</tr>
<tr>
<td>IV</td>
<td>6.2-7.0</td>
<td>6.4-6.6</td>
<td>7.5-9.5</td>
<td>0.090-0.150</td>
</tr>
<tr>
<td>V</td>
<td>6.2-7.0</td>
<td>6.6-7.2</td>
<td>7.5-9.5</td>
<td>0.090-0.195</td>
</tr>
<tr>
<td>VI</td>
<td>6.6-7.0</td>
<td>6.8-7.2</td>
<td>8.5-9.0</td>
<td>0.165-0.195</td>
</tr>
<tr>
<td>VII</td>
<td>6.6-7.0</td>
<td>6.8-7.2</td>
<td>9.0-9.5</td>
<td>0.210-0.285</td>
</tr>
<tr>
<td>VIII</td>
<td>7.0-7.6</td>
<td>7.4-7.8</td>
<td>$\geq 9.5$</td>
<td>0.210-0.285</td>
</tr>
</tbody>
</table>

A preliminary seismic zoning map (Fig. 3) results from these criteria and individual hazard maps taking into account not only the size of earthquakes but also the effect of focal depths, distance, and other characteristics of the seismicity. The hazard maps suggest that the study area may be divided into eight seismic source zones, from negligible hazard to most hazard. Zone I (Konya zone) is quiescent at the present time. The Antalya zone (zone II) is characterised by earthquakes as deep as 100 km and is relatively less active in comparison to southern and western Cyprus. In contrast, zones VII and VIII contain areas where the most damaging earthquakes occur.
Figure 3 Preliminary zoning map derived from different hazard parameters.

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Burton, P. W and Yılmaztürk, A. Seismic Hazard and Seismotectonic parameters for southern Turkey, (in preparation).


SUMMARY

The earthquake seismic hazard in Jordan and its vicinity is assessed on the basis of probabilistic methods. For this purpose a new up-to-date earthquake catalogue is compiled which covers the period between 1-1989 A.D.. The earthquakes lie between latitudes 27.0°-35.5°N and longitudes 32.0°-39.0°E. Thirteen seismic zones are defined on a seismotectonic map presented for the area. Point-source and line-source models are used. The seismic hazard parameters, namely, $b$-parameter (of Gutenberg-Richter relationship), $m_1$ (the upper bound magnitude), and $\lambda_4$ (the annual rate of occurrence of earthquakes with local magnitude $M_L \geq 4.0$) are calculated for every zone using an approach which was used by many previous studies (e.g. the least-squares method) and a new method of Kijko and Sellevoll (1992). The results of the seismic hazard assessment are displayed in the form of iso-acceleration contours expected to be exceeded during typical economic life times of structures, i.e. 50 and 100 years. For every model, two seismic hazard maps are derived. In order to determine the importance of the SE-Mediterranean zone and the North part of the Red Sea zone from a seismic hazard point of view for Jordan, one seismic hazard map which correspond to 50 years economic life for every model, excluding the seismicity of these zones, is derived. A sensitivity analysis is carried out for the seismic hazard parameters with inherent uncertainties.
SEISMIC HAZARD ANALYSIS - AN IMPORTANT BASIS FOR PLANNING AND DESIGN

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INTRODUCTION

The objective of seismological studies and maps is to provide data on the seismicity of a region or urban areas and to define the seismic hazard on the basis of the existing seismological, tectonic, geophysical and other involved data. All above mentioned data are necessary for elaboration of various seismological maps, (epicentral maps, seismic zoning maps, seismic hazard maps, microzoning maps, etc.). These maps have been prepared, depending on whether the purpose of the maps is to present basic data, to present data augmented by judgment, or to specify criteria by design.

It is desirable that engineers know the potential hazard of future events. For this purpose, detail investigations and studies for definition of seismic hazard level of a given site, are necessary. Such studies can, in general only be justified for large, important projects such as nuclear power plants, major dams, etc. However, even for design of all other structures in seismic zones data about seismic hazard level are necessary, that is data on the level of design ground motion on seismic loading. Usually, these data are available from special seismic maps (iso-seismal, iso-acceleration maps, etc.) and graphs developed for certain geographical regions. These seismic maps and graphs are practical tools in earthquake resistant design because they provide useful guidance when it is not feasible to make thorough studies on the earthquake hazard at particular locations. In these maps and graphs, elaborated for engineering purposes, seismic hazard is usually presented in terms of probability of exceedence for given ground motion parameters with a given time period t.

Determination of the maximum earthquake intensity and its incorporation in structural design is one of the specific features of earthquake engineering. Considering the random nature of the earthquake occurrence, it is rather difficult to formulate its model and include it in the design process without certain simplifications and assumptions related to the earthquake occurrence mechanism and structural behaviour under seismic effects.

Therefore, due to uncertainties in the earthquake definition and some physical parameters defining structural behaviour, in aseismic design it should be always taken care not the exceed the calculated parameters of both earthquake and structures. Compromise solution of the problem is to adopt some seismic risk level in all the stages of creation of the structural: design, construction and use.

Due to the complex character of this problem which makes impossible the presentation of a more detailed review of all the aspects connected with the seismic hazard analysis, special emphasis will be put on elaboration of the various seismic hazard maps as well as for determination of the seismic design parameters and its application in the process of design.
GENERAL CONSIDERATIONS

Since scientists, engineers and planners are unable to predict and prevent earthquakes, they have tried to limit the damage caused by earthquakes. Defense against earthquakes relies, first, on evaluation of the seismic risk, which means making graphs and maps of seismic hazard zones and second, on their strict application in seismic design of structures.

It should be emphasized that the reliability for evaluation of the seismic hazard parameters influence mostly the reliability of the results obtained by the seismic risk analysis. They are directly associated with the decision making on the acceptable seismic risk level and the selection of the optimum structural concept for specified seismic conditions, i.e., the assessment of the possibility for these structures to exist in such conditions and the economical and social justification for their construction.

The seismic hazard analysis can be formulated by combination of parameters which define the seismicity as a seismic source model, recurrence and attenuation relationships and by application of certain forecasting models for earthquake generation.

Generally speaking, the parameters defining the seismic hazard, i.e., the probability of earthquake occurrence are the following: cumulative probability of maximum ground acceleration, return period of earthquakes with certain intensity of acceleration and probability level of maximum ground acceleration depending upon the return period.

These parameters are used for practical purposes: (1) elaboration of seismic hazard map and (2) elaboration of return period diagrams of maximum ground accelerations for certain sites.

The best way for presentation of the seismic hazard, i.e., the distribution of some ground motion parameters for a certain geographical area is the seismic hazard map (Fig.1) presenting the earthquake parameters in the most favorable form depending upon the purpose of the map. The necessary data for the elaboration of these maps are obtained by the function of the cumulative distribution of probability, namely, the return periods of maximum ground accelerations. The maximum acceleration for different return periods and probability level is considered to be the most appropriate parameter to be used in the analysis and design of structures. Therefore, it is most commonly used to percent the seismic hazard.

The practical importance of the return period diagrams of maximum ground accelerations (Fig. 2) is that they enable to obtain data on the seismicity of a site and compare the seismic activity and the seismic hazard level at certain sites.

In this way, the probability characteristics of the earthquake occurrence, i.e., the seismic parameters for different return periods and probability levels can be obtained.
SEISMIC HAZARD MAPS

All known methods and solutions of seismic hazard mapping techniques take into account the seismological and geological information to arrive at the probabilistic forecasting of future earthquakes. Description of the earthquake occurrence phenomenon as a basic problem is closely related to the availability of sufficient number of data. If they are available, the description can point to the expected future earthquake characteristics. Unfortunately, only limited number of data are available, and they mainly refer to the earthquake intensity. Least available are strong motion accelerogram records sufficient to define seismic loads in each zone, which are the most important from the engineering point of view.

This lack of data, especially instrumental data of recorded earthquakes, or their unreliability if older data are in question (before 1900) makes impossible complete definition of the seismic history and on this basis to draw conclusions on the future earthquakes. This refers even to regions for which many records are available since the period they are related with (the last 100 years) is relatively short compared to the periods of significant tectonic activity being from several tens of thousands to several millions of years.

Therefore, more tectonic parameters should be included in the analysis of earthquake phenomenon namely in modeling of the site seismicity. In such a way the complexity of probable and physically possible conditions for earthquake generation could be defined in more details.

All above mentioned data are necessary for elaboration of different seismological maps. A wide variety of seismicity maps (fault maps, seismotectonic maps, epicentral maps, etc.), seismic zoning maps, seismic hazard maps, etc. have been prepared, depending on whether the purpose of the maps is to present basic data, to present data augmented by judgment, or to specify criteria for design. Within the last category there can be major differences depending on the application for which the map was prepared as well as on the relative emphasis to the economy and safety. This map can be categorized by the extent to which they portray a mixture of data and judgment.

There is a real need to express the parameters of the zoning maps in probabilistic terms, that is in terms of hazard. Seismic hazard here is taken to mean probability of any physical phenomenon (e.g. ground shaking, ground failure) associated with an earthquake which may produce adverse effects on human activities at specific locations in a given interval of time. Without going in details, we should mention that there is a basic difference between seismic zoning map and seismic hazard map, primarily due to the results, the way they are presented and the adopted methodology for their elaboration.

Being basically mostly well known, the methods for seismic hazard analysis and elaboration of maps based on the seismic hazard study will not be explained and repeated in this paper. Many of the mentioned references comprise basic information on these methods.

The hazard maps are practical tools in designing structures because they provide useful information of the earthquake hazard at particular locations. On these maps, elaborated for engineering purposes the seismic hazard is presented in terms of: (1) peak value (or other more descriptive
parameters) of acceleration, velocity and displacement, (2) frequency content, and (3) duration.

There is a considerable work done in the area of probabilistic estimation of seismic loading. In particular, probabilistic maps forecasting in terms of seismic hazard maps like iso_seismal or iso_acceleration maps, have been developed by many researchers. Fig.1 is a typical example. The analytical development of such a map contains some restrictions and assumptions related to the parameters of earthquake occurrence.

None of the currently available procedures for hazard mapping provide the complete information listed above. This is caused by the reasons explained previously, i.e., due to the fact that it is impossible to provide sufficient and reliable data required for developing of these maps; some shortcomings and problems related to the seismic hazard analysis and elaboration of the maps based on these analyses will be discussed.

**SEISMIC DESIGN PARAMETERS**

The seismic hazard parameters, defined applying analytical and graphical relationships (Fig.2) between the maximum ground accelerations and the corresponding return periods, makes possible to assess the seismicity of a site and to compare it with the global seismicity of the region. However, applying only these relationships, i.e. their results, is not sufficient for aseismic design based on seismic risk methods. Some parameters of vital importance for rational design are missing, such as the serviceability period of the structure, its purpose and importance in the post earthquake conditions, as well as the level of the acceptable seismic risk, etc.

The next step in seismic risk assessing is including of these parameters in the seismic risk model. For various values of the life time of the structure and level of acceptable seismic risk corresponding return period values are obtained. The graphical presentation of these relationships represents the relationship between the risk level, the serviceability period of the structure and the return period (Fig.3). These diagrams are spatially independent and they are used for each location at any region, under the condition the return period diagrams of maximum ground accelerations for the considered locations have been previously determined.

On the basis of these diagrams design seismic parameters -maximum ground acceleration - corresponding to the life time of the structure and and the seismic risk level can be determined (Fig.4). In other words, the seismic loads are possible to be defined in function of the return period and the probability level for not exceeding the design parameters during the life time of the structure. In this way, the seismic risk level or the probability for exceedence of the design parameters can be defined.

Considering the fact that it is practically impossible and economically unjustified to provide through design and construction equal protection of all structures against damage and failure, it is necessary to classify the structures and categories according to their usage and importance and define for them the acceptable seismic risk level.
RELATED PROBLEMS AND CONCLUSIONS

A seismic map is a map predicting the geographical distribution of a certain quantity, either without time limitation or during a specified period of time. There is a principal question which everyone will have it clarified before starting the work, i.e., what information will the zoning map provide? The answer involves problems related to the existing building codes or to the requirements of regional planning.

There are several weak points of the procedure. First is the lack of seismological and other information on the processes governing the origin of earthquakes and propagation of strong motions in real media. At the beginning of the work faced is the problem of definition of potential earthquake source and of the recurrence of strong shocks. It is decided to consider all regions where earthquakes have originated so far as regions of at least the same activity in future. Other problems are delineate the potential regions which now display only low activity and to estimate the future frequency of occurrence of strong earthquakes. In this paper some aspects, ideas and experience on these problems are presented. On this basis the following conclusions are made:

- Seismicity, earthquake statistics, seismotectonics, geophysics and seismic hazard studies are the most important steps for the realization of seismic zoning map of a given region and define the seismic design parameters.

- A detailed analysis of the seismicity is possible only by introduction of more tectonic parameters. In this way all possible and physically achievable conditions for generation of earthquakes, i.e., a mathematical model of seismicity can be formulated and it is more suitable for application in the seismic risk analysis.

- From engineering viewpoint it has been considered that peak ground acceleration is the most adequate parameter for evaluation of the seismic hazard. Therefore, it is presented by: seismic hazard maps for different return periods, cumulative distribution function and return period diagrams of peak ground acceleration. The seismic hazard analysis and zoning mapping technique, considered in this paper, is one of the possible solutions of the problem. It involves many theoretical and practical issues which should be taken into account.

- The formulation of the mathematical model of the seismic design parameters based on the seismic risk for practical purposes is possible only if certain assumptions and simplifications are adopted, since a strict analytical solution of the problem requires an application of completed mathematical tools. For certain variable, random process, like the earthquake mechanism, performance of structures during earthquakes, etc., there are relatively small number of available data. herefore, the need for probabilistic solution of the problem is more evident since it is the only one leading to accurate and usable results for analysis.

- More detailed investigations are still needed in all fields of science (seismology, mathematics, earthquake engineering, economy, etc.) in order to define the actual theoretical scope for the purpose of definition of a general, complete seismic hazard model for practical purposes. Due to the necessity for introduction of certain assumptions and simplifications, the present solutions of seismic zoning should be treated as incomplete ones.
REFERENCES


Fig. 1. Seismic hazard map of the Republic of Macedonia. Distribution of maximum ground acceleration for a return period of 100 years.
Fig. 3. Relationship between return period, serviceability period and acceptable risk level.

Fig. 2. Return period diagram for maximum acceleration.

Fig. 4. Design seismic parameters. Relationship between peak acceleration risk level and depreciation period of structure.
Many different studies devoted to the knowledge of the Italian seismicity started after the 1976 Gemona earthquake. They were developed by the cooperation among geologists, geophysicists, and engineers in the frame of the "Progetto Finalizzato Geodinamica" of the "Consiglio Nazionale delle Ricerche" (CNR). One of the products was the hazard maps of Italy (Gruppo di Lavoro Scuotibilità, 1979) based on a probabilistic treatment of the seismological data. The CNR's proposal of seismic zoning was based mainly on the results of that study (Petrini, 1980), and was accepted by the Italian government and translated into a series of decrees by the Ministry of Public Works between 1980 and 1984.

The GNDT project

Ten years have passed from the first CNR's proposal of seismic zoning and the increase of seismotectonic knowledge (see Albini e Barbano, 1991 and GNDT, 1991 for the state-of-the-art at the end of 1991) suggests an updating of the classification. A project for the seismic hazard assessment of the Italian territory has been being undertaken since 1989 in the frame of the activities of the CNR's "Gruppo Nazionale per la Difesa dai Terremoti" (GNDT) and a detailed analysis of the computational approaches as well as of the input data set is planned. The scheme of the GNDT project is shown in the Figure; the five main topics, which are in progress at present and among which there is a continuous transfer of information, are: the analysis of the main probabilistic approaches to identify that/those most suitable for the Italian situation; the collection and revision of the seismological source data (earthquake catalogue and intensity maps); the analysis of the seismotectonic information for deriving a seismogenic zoning of whole Italy; the analysis of the available attenuation relations (for macroseismic intensity and horizontal peak ground acceleration, PGA), in order to define the most suitable for the Italian earthquakes; the modeling of the seismic hazard by deterministic approaches for checking the results obtained by probabilistic approaches.

The probabilistic approaches

It is planned to apply three increasing complexity level probabilistic approaches: 1) simple statistical analysis of the actual shakings at the site, referring to the extreme value theory; 2) memoryless approach for the seismicity recurrences, given homogeneous seismogenic zones (SZ's) following the method developed by Cornell (1968); 3) a memory dependent mixed procedure based on the renewal process (Grandori et al., 1991). The first can be considered a new application of a standard method and is possible because a long in time documentation on the felt shakings is available for many Italian cities. As it is clear that this data set is not sufficient for a robust statistical analysis all along the territory, it will be, therefore, integrated by data computed by using the epicentral parameters and an attenuation relation. The
Cornell (1968) approach is applied considering PGA as hazard parameter, while the others consider macroseismic intensity. The main difference between the Cornell (1968) and the mixed approaches is that the first operates with an exponential distribution for magnitude and a Poisson process for the occurrences and the second determines the distribution of intensity, as well as of the occurrences, directly from the data set.

**Seismological source data**

The complete revision of the major earthquakes (epicentral intensity larger than VII MCS) which affected Italy from 1000 has been planned by the retrieval and analysis of all the information already used and cited in the earthquake catalogues (see GNDT Macroseismic WG, this volume). The macroseismic parameters are derived from intensity point maps compiled after 1985 by several investigators or agencies, including GNDT. In the mean time the assessment of instrumental magnitude for the 20th century earthquakes, still missing in the Italian seismological files, has been performed (Margottini et al., 1992).

**Seismotectonic information**

Very recently a seismotectonic model for Italy has been proposed (Scandone et al., 1991). It consists of about 70 areas with different seismotectonic behaviour, which are used as SZ's. For each of them the characteristics of the seismicity (frequency versus magnitude and intensity relations) are computed.

**Attenuation relations**

Macroseismic intensity and PGA are considered as hazard parameters, and, therefore, attenuation relations are needed for both. For PGA the relation recently established by Sabetta and Pugliese (1987) for Italian earthquakes has been chosen. It is an unique circular relation which is considered valid for all the SZ's. For macroseismic intensity the Grandori et al. (1987) relation has been chosen and its parameters are calculated separately for each SZ considering the best intensity maps. The intensity attenuation, in the Grandori et al. (1987) formulation, is not necessary the same in all directions but eccentric elliptical relations can be accounted as well.

**Deterministic helps**

An automatic procedure for the seismic zoning of the Italian territory has been developed (Costa et al., 1992). The results consist of the deterministic computation of acceleration time series distributed on a regular grid over the territory. For the estimation of the accelerations, complete synthetic seismograms are computed by the modal summation technique. A first rough zoning can be accomplished by producing a map showing the distribution of PGA. The structural and source models necessary to compute the synthetic signals have been fixed after an extensive bibliographic research, and the seismogenic zone model has been used to limit the spatial distribution of sources. Information on historical and recent seismicity has been taken from the most updated Italian earthquake catalogues. The preliminary estimated PGA's have been found to be compatible with available data, both in terms of intensity (historical earthquakes) and accelerations (recent earthquakes).
SEISMIC HAZARD PROJECT
in ITALY
(by CNR - GNDT)

PROBABILISTIC APPROACH
- Gumbel statistics at sites
- Cornell zoning approach
- Renewal methodology

SOURCE DATA
- Revision of existing catalogue (FFC)
- Macroseismic magnitudes
- Integration of instrumental data
- Intensity maps of main historical events

SEISMOTECTONICS
- Seismotectonic zoning for Italy
- Seismic characteristics of areas and faults

ATTENUATION
- Macroseismic intensity
- Uniform hazard response spectra
- Indian relation (Selvini & Pugliese)
- Direct use of intensity maps

DETERMINISTIC HELPS
- Constant attenuation zoning
- Acceleration
- Max PGV from synthetic seismograms
FINAL CONSIDERATIONS

The forthcoming CNR's proposal for seismic zoning of the Italian territory will be based on the evaluation of the different hazard results which will be obtained, and on the expert evaluation of the regional seismotectonic characteristics. In such a way, the complete available information will be used, sometimes quantitatively and sometimes qualitatively, in the global hazard evaluation. In addition, an estimate on the expected damage (first damage and building collapse) will be performed and, therefore, the proposal will contain also some aspects of risk analysis. As future plans, a regional characterization in terms of uniform hazard response spectra will be provided.

REFERENCES

CHARACTERISTIC FEATURES OF GEONOMIC PROGNOSTIC FUNCTIONS FOR THE MAXIMUM POSSIBLE EARTHQUAKE

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ABSTRACT and CONCLUSION

The recent artificial intelligence techniques are commonly applied to solving problems in which statistical analyses of various quantities and their modelling prognostic functions can be used.

The prognosis of the maximum possible earthquake based on the knowledge of all available geonomic data closely related to geodynamic processes in the Earth's crust and the Upper Mantle, allow a relatively reasonable earthquake magnitude assessment to be obtained. The recent methods of the artificial intelligence techniques extend the possibility to analyze the existing data by multidimensional statistics, pattern recognitions, etc., and to create prognostic fields of certain parameters with respect to different input data and quantities.

Geonomic quantities are reviewed and categorized with respect to their influence upon the estimation of the maximum possible earthquake. It is clear that plausibility of the prognosis depends both on the quality of input data and on the general character (or trend) of prognostic functions of the individual geonomic quantities.

The paper attempts to summarize all applications of different geonomic quantities made as yet in the maximum possible earthquake prognosis. It is the first attempt to generalize the prognostic functions in this way and therefore, all the above described relations have to be understood as preliminary ones. Nevertheless, we believe that such a summary can be used as a guide in recent applications of the methods and algorithms of the artificial intelligence techniques in the maximum possible earthquake prognosis.

The paper will be submitted for publication on a special issue of journal NATURAL HAZARDS
MACROSEISMIC INVERSION AND VARIATION IN INTENSITY ATTENUATION IN
BRITAIN

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INTRODUCTION
In countries of low to moderate seismicity, where there are also
long established populations and documented histories, historical
seismicity is the natural foundation of any regional seismic hazard
study. Well researched historical and contemporary macroseismic
data are needed to calibrate the historical earthquake catalogue
for hazard analyses and also to determine intensity attenuation to
facilitate quantification of strong earthquake occurrences, damage
potential and seismic risk estimates. This note will highlight and
extract some results illustrating variability in intensity
attenuation and also estimate an optimal value of the attenuation
coefficient, α, representative of the best available historical
macroseismic data in Britain. The methodology used is fundamentally
Kovesligethy's, with modifications for optimisation, and
modernisation in general so that it is readily programmable on a
PC.

THE DATA
During the 1980s several studies were mounted in Britain with the
objective of reevaluating historical seismicity. These included
studies by the British Geological Survey, Principia Mechanica Ltd.
and Soil Mechanics Ltd. The data used here draws from the
isoseismal maps published in the BGS studies. These studies rely
whenever possible on reworking of primary historical data; they
document all pertinent primary data and use the MSK intensity scale
in a systematic way throughout. The macroseismic methodologies are
well documented by Burton et al. (1984) and data on specific
earthquakes supplements the earlier work, for example Musson et al.
(1990) have recently reworked the 1884 Colchester earthquake used
herein. The highest intensity isoseismal in the data set is 8 MSK
and no observed I₀ exceeds 8 MSK. The largest number of isoseismals
for any earthquake is seven (Colchester 1884) and frequently three
useable isoseismals have to suffice. More detailed selection of
data is described below.

METHOD AND SPECIMEN RESULTS
The physical basis of Kovesligethy's (1907) formulation of
intensity attenuation is firstly that macroseismic intensity, I,
relates to the peak ground acceleration, a, through a power law:

\[ I = \log_{10}a^3 + \text{constant} \]  

(1)

and secondly the energy in the seismic wave decays geometrically
according to an inverse square law with distance and is also damped
with distance described by an exponential function with damping
coefficient α km⁻¹. This leads to

\[ I_0 - I_i = 3\log_{10}(r_i/h) + 1.303\alpha(r_i-h) \]  

(2)

where \( I_0 - I_i \) is the difference between the epicentral intensity, \( I_0 \),
and intensity at hypocentral distance \( r_i \) km for a focus at depth \( h \) km.
Fig. 1 The upper diagram shows values of attenuation coefficients, $\alpha$, and focal depths obtained from optimisation of $[I_0, h, \alpha]$ from $(l_i, s_i)$ for each of a set of British earthquakes. The lower diagram shows results from joint optimisation of 13 earthquakes for optimal $I_0$ and focal depth for each earthquake but with an optimal average $\alpha$ for the set of earthquakes. Constraints as indicated.
There are several ways equation (2) can be manipulated to obtain solutions for both specific earthquakes and also sets of earthquakes. Firstly, for a given earthquake with observed $I_o$ and set ($I_i, s_i$), where $s_i$ are the isoseismal radii or isoseismal-epicentre distances, the values of a set of $h$ may be systematically tested or searched at a given $\alpha$, and the process iterated over different $\alpha$, to obtain optimum values of $\alpha$ and $h$ fitting equation (2). Alternatively, $I_o$ can also be treated as unknown and the root mean square difference between observed and theoretical intensities, $I_{oh,i}$ and $I_{t,i}$, for the $n$ isoseismals of one earthquake computed as a function of $I_o$, $h$ and $\alpha$ and the optimal set of $[I_o,h,\alpha]$ at minimum rms selected (Burton et al., 1985):

$$\text{rms} = \sqrt{\frac{\sum (I_{ob,i} - I_{t,i})^2}{n}}$$ (3)

The results of such an optimising process for over 30 earthquakes are shown in the upper half of Fig.1. Reasonable constraints on allowable values of the solution parameters are invoked. These are that the focal depth is less than 60 km, that the attenuation coefficient is positive but less than 1.5 km$^{-1}$ and the $I_o$ is no more than 2 MSK degrees greater than the value of the innermost isoseismal. Two typical types of result are seen. Attenuation coefficients for focal depths less than 40 km (an acceptable range for Britain) show scatter from very small values up to about 0.06. There is also a cluster of results around 60 km focal depth, a depth which would be considered quite unusual for British earthquakes during the instrumental era. These solutions probably correspond to either poorer data or isoseismals which do not entirely adhere to the Kovesligethy model of attenuation.

This suggests a refinement in data selection is appropriate. Choosing isoseismal data which are well behaved in terms of the Kovesligethy formalism, that is, a graph of $I_o-I_i$ against $s_i$ is concave with respect to distance at all points, isolates 13 earthquakes (these are: 1705 Northampton, 1816 Mansfield, 1832 Swansea, 1884 Colchester, 1903 Derby, 1904 Derby, 1906 Swansea, 1916 Stafford, 1925 Oban, 1940 Stirling, 1957 Feb 11 Derby, 1957 Feb 12 Derby, 1972 Todmorden). Optimising these 13 sets of isoseismals for individual optimal values of $I_o$ and $h$ and a joint value of $\alpha$ gives the results in Fig.1 (lower). The average optimal value of $\alpha$ is 0.00474. Optimal values of $I_o$ decrease systematically with focal depth. The greatest scatter in optimal $I_o$ is for mid-crustal earthquakes. Some shallow, high intensity (but relatively large felt area) earthquakes, for example 1884 Colchester, conform with the Kovesligethy formalism.

The rms surface defined by equation (3) allows examination of $I_o$ and $h$ at a given $\alpha$ as a set of contours of rms value for any earthquake. The Colchester earthquake of 1884 is examined in this way as an example in Fig.2. The value $\alpha=0.00293$ is obtained as optimal for 1884 Colchester alone, using seven isoseismals and the observed $I_o$ and iterating over $h$ and $\alpha$. The contours in Fig. 2 (right) result from optimisation of 13 earthquakes together, as described above, producing the average optimal value $\alpha=0.00474$, and subsequently these rms contours of $[I_o,h]$ for the Colchester earthquake. In Fig. 2 (left) the minimum point is at $I_{ob}=8$ MSK and $h=4.0$ and in Fig. 2 (right) at 10 MSK and 0.9 km respectively. In both cases the innermost contour (rms=0.3 and 0.4 respectively)
indicate both best fit and a measure of the zone of parameter uncertainty or resolution in the inversion. Both sets of contours of \( [I_o, h] \) in Fig. 2 are consistent with a focal depth around 2.5 km and epicentral intensity of 8 MSK but not reaching 9 MSK.

**Fig. 2** Contours of rms goodness of fit of 1884 Colchester isoseismal data to the Kovesligethy attenuation law. The rms contours for \( \alpha = 0.00293 \) (left) correspond to the best fit of the Colchester isoseismals and observed \( I_o \) to the law whereas rms contours (right) correspond to \( \alpha = 0.00474 \) obtained by joint optimisation of 13 "well behaved" earthquakes.

**CONCLUSIONS**

In summary, a set of programs operable on a PC are being developed to analyse macroseismic attenuation. Best results are obtained when the isoseismal data are selected to conform with the Kovesligethy law, suggesting that either some data are poor or Kovesligethy formalism requires further development or both. For British earthquakes an average value of \( \alpha \) optimised over 13 "well behaved" earthquakes is 0.00474. This well behaved data set shows optimal \( I_o \) to correlate well with focal depth and also shows the most scatter for mid crustal earthquakes, the depth at which many tectonic British earthquakes originate. Exceptional earthquakes such as 1884 Colchester (large felt area but very shallow focal depth and high \( I_o \)) also adhere to the Kovesligethy formalism. Taking lowest rms value contours of \( [I_o, h] \) as evidence of parameter uncertainty in the inversion suggests that the Colchester earthquake was about 2.5 km deep with an \( I_o \) of 8 MSK, not reaching 9 MSK.
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PROBABILISTIC TREATMENT OF THE SEISMIC ATTENUATION

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ABSTRACT

The attenuation laws with different analytic expressions and different techniques for estimating the parameters have been proposed for the seismic hazard evaluation. These attenuation laws are usually considered in deterministic way or sometimes associating a random error to the intensity decay. But the attenuation itself, considered as a physical process, depends on many random elements. This suggests we should deal with it in a probabilistic framework.

In fact this work compares two different approaches to the treatment of the attenuation law: the deterministic and the probabilistic one. Some tests have been performed varying the probability distributions according to different physical interpretation of their parameters. Data from some Italian regions have been analysed and the obtained results show the efficacy of the proposed probabilistic treatment; it seems stable with respect to the distance between site and source area and it is flexible to the degree of available knowledge on the seismic source.

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INTRODUCTION

During the last decade, increasing interest has been shown in the application of probabilistic procedures for the quantitative evaluation of the reliability of lifelines under earthquake loads. This interest stems from the demand of assuring adequate safety in urban communities which may be subjected to secondary disasters during and after destructive earthquakes. Transportation systems, pipelines, communication and electrical distribution systems are examples of lifelines. Such lifelines can be modeled as networks of interconnected links (edges) and vertices (nodes). The basic aim of this study is to develop a computationally efficient algorithm through which the reliability of a lifeline, modeled as a network, can be evaluated considering the seismic hazard. A computer program is coded to perform the numerical computations and a case study is presented to show the implementation of the proposed method.

SEISMIC HAZARD ANALYSIS MODEL

The probabilistic formulation developed in this study is based on the "classical" seismic hazard analysis model (Cornell, 1968). Lifeline networks generally extend continuously over large areas; therefore the seismic hazard analysis should be conducted to cover the whole region where the lifeline is located. For this reason, the probability distribution of the appropriate ground motion parameter is computed at a number of points along each spatially extending component of the system to identify the highest seismic demand for each component. As a result, the output of the seismic hazard analysis will be a set of probability distributions describing the seismic hazard at the most critical point of each component.

DETERMINATION OF THE CAPACITY OF THE ELEMENTS OF THE SYSTEM

Determination of the seismic capacity of the elements of a lifeline system, such as pipelines, aqueducts, tunnels, highways, bridges is a problem to be addressed by structural engineers and is beyond the scope of this study. Here, it is assumed that a set of probability distributions, describing the "capacity" in terms of the same ground motion parameter used for the demand, is available for each component.

If the seismic capacity and demand for an element are known, then it is possible to estimate the state of the element. In the most general case, there will be multistates described by a continuous or a discrete variable indexing the degree of damage. Here, we will consider the simplest two-state description: namely survive or fail. If C and D denote the random variables quantifying the seismic capacity and demand of an element, respectively, then the
failure probability \( P_f \), is:

\[
P_f = \Pr(C < D)
\]  

On the other hand, the probability of being in survival state, \( P_s \), called reliability, is:

\[
P_s = \Pr(C > D) = 1 - P_f
\]  

NETWORK MODELING AND RELIABILITY ANALYSIS

In assessing the overall reliability, a lifeline is generally idealized as an equivalent network. Then, the reliability of the network is computed as a function of the reliability of its components, which are the links and nodes. Such an approach seems to be simple conceptually, but in its implementation certain computational difficulties arise; namely the computer CPU (central processing unit) time and storage requirements may become restrictive as the number of components increases and as the connectivity of elements becomes more complex. Yoo and Deo (1988) have compared four different algorithms for the terminal-pair reliability problem and showed that the algorithm due to Dotson and Gobien (1979) is the fastest. They have improved this algorithm to reduce the execution time even further. This improved algorithm (Yoo and Deo, 1988) is used for the evaluation of the network reliability in our study.

APPLICATION

Panoussis (1974) and Taleb-Agha (1977) have analyzed the seismic reliability of the network composed of major highways surrounding and leading into the Boston area. This transportation lifeline system is modeled as a network consisting of 18 nodes and 23 links (see Fig.1). In the above-mentioned studies the objective of the network was to reach from node N1 to node N5. Here we will evaluate the seismic reliability of the resulting network based on the method proposed in this study and we will compare our results with those of Panoussis (1974) and Taleb-Agha (1977).

The potential seismic sources surrounding the Boston region were identified by Cornell and Merz (1975). The geographical distribution of the eight seismic source zones are shown in Fig.2. The values of the seismicity parameters that will be used in the seismic hazard analysis are taken directly from Cornell and Merz (1975). Computations are conducted by a computer program and for each link the probability distribution of \( D \), where \( D \) denotes the peak ground acceleration (demand), is obtained at a number of points searching for the point having the highest seismic hazard. Once this point is identified, the associated seismic hazard is taken as the seismic demand for that link. This is equivalent to reducing a spatially extending element to a "point" element. This approximation is justified, since the length of links are small compared to the overall size of the region containing the seismic sources (note the difference of scales in Figs.1 and 2). Also the assumption of a perfect correlation for the demand within each link supports this simplification from a statistical point of view.

The nodes are assumed to be perfectly reliable. The seismic capacity of each link can be treated as a deterministic or a random
variate. Here we will consider the following three cases based on the resistance values given by Taleb-Agha (1977) and Panoussis (1974). Case 1: Deterministic seismic capacity of 75 cm/s² (Taleb-Agha, 1977); Case 2: Random (Gaussian) seismic capacity with \( \mu = 100 \) cm/s² and \( \sigma = 2 \) cm/s² (Panoussis, 1974); Case 3: Random (Gaussian) seismic capacity with \( \mu = 75 \) cm/s² and \( \sigma = 2 \) cm/s². The adjacency matrix and resulting survival probabilities for these three cases are shown in Figs.3 and 4.

A computer program is prepared in order to implement the algorithm proposed by Yoo and Deo (1988). The bounds on the annual reliability (probability of survival), \( R \), are found to be: Case 1: \( 0.999043 < R < 0.999045 \); Case 2: \( 0.999746 < R < 0.999748 \); Case 3: \( 0.996252 < R < 0.996254 \). The reliability bounds given in all of the three cases are very high due to the high component reliabilities and due to the existence of redundant paths. However, the assumption of random link capacities (Case 3) yields to a lower reliability compared to the deterministic case (Case 1), when the mean capacities are assumed to be the same. The difference increases directly proportional to the degree of uncertainty involved, which is quantified by the standard deviation.

CONCLUDING COMMENTS

The initial results of an ongoing research on the evaluation of reliability of lifelines under seismic threat are presented. The network evaluation algorithm is based on the one developed by Yoo and Deo (1988). In this study, the length of the links do not appear as a parameter, since each link is discretized at a point along its length where the seismic hazard is maximum. However, in the model which is currently under development, the length of the links and the spatial correlation among the failure events are taken into consideration.

The network of major highways surrounding the Boston area is analyzed based on the same data given by Panoussis (1974) and Taleb-Agha (1977). The results obtained from the methodology developed in this paper are consistent with those obtained in these two studies; furthermore our methodology eliminates the CPU time and computer storage problems that are encountered in large networks.

ACKNOWLEDGEMENT

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Figure 1 A network model for the major highways within the Boston area (from Taleb-Agha, 1977)

Figure 2 Seismic source zones used in assessing the seismic hazard (from Taleb-Agha, 1977)

Figure 3 The adjacency matrix and the computer output for the network example (Case 3)

Figure 4 Reliability vector for the links: (a) Case 1; (b) Case 2; (c) Case 3

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ATTENUATION OF INTENSITY WITH DISTANCE FOR VRANCEA INTERMEDIATE EARTHQUAKES.

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INTRODUCTION

In studying the hazard from earthquakes at any location it is necessary to know the relations of attenuation of intensity with distance from the epicenter. This problem was analysed in more papers and in some cases were established relations between intensity (I), epicentral distance (A) and depth of the focus (h), (Gutenberg, Richter, 1942; Neumann, 1954; Cornell, 1968; Kárník, 1969; Evernden et al., 1973; Howell, Schultz, 1975; Radu, Apopei, 1977; Estava, Chavez, 1982; Chavez, Castro, 1988).

Howell and Schultz (1975) analysing the attenuation of intensity with epicentral distance pointed out that for the weaker earthquakes the attenuation is greater than for the stronger earthquakes.

In this paper we studied the attenuation of macroseismic intensity for some Vrancea intermediate earthquakes versus epicentral distance in the four directions (E, W, N, S).

DATA

The available data for this study are the macroseismic maps for 22 intermediate earthquakes occurred in Vrancea zone during 1790-1966, proposed by Shebalin in the frame of UNESCO Program "UNDP-UNESCO survey of the seismicity of the Balkan region". We can see that these isoseismal maps show a quasielliptical shape with the major axes oriented NE-SW and this maps have similar shape for every magnitudes.

The principal parameters of the analysed earthquakes are shown in Table 1.

Table 1

<table>
<thead>
<tr>
<th>No</th>
<th>Date</th>
<th>h:m:s</th>
<th>φ°N</th>
<th>λ°E</th>
<th>h km</th>
<th>I o</th>
</tr>
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<tbody>
<tr>
<td>1</td>
<td>1790.IV.06</td>
<td>19:29</td>
<td>45.7</td>
<td>26.6</td>
<td>VIII</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>1821.XI.17</td>
<td>13:45</td>
<td>45.7</td>
<td>26.6</td>
<td>VII</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>1829.XI.26</td>
<td>01:40</td>
<td>45.8</td>
<td>26.6</td>
<td>VIII</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>1836.I.23</td>
<td>18:45</td>
<td>45.7</td>
<td>26.6</td>
<td>IX</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>1893.VIII.17</td>
<td>14:35</td>
<td>45.8</td>
<td>26.6</td>
<td>VIII</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>1893.IX.10</td>
<td>03:40</td>
<td>45.7</td>
<td>26.6</td>
<td>VII</td>
<td></td>
</tr>
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<td>7</td>
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<td>45.7</td>
<td>26.6</td>
<td>V</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>1904.II.06</td>
<td>02:49</td>
<td>45.7</td>
<td>26.6</td>
<td>75</td>
<td>VI</td>
</tr>
<tr>
<td>9</td>
<td>1912.V.25</td>
<td>18:01</td>
<td>45.7</td>
<td>27.2</td>
<td>90</td>
<td>VII</td>
</tr>
<tr>
<td>10</td>
<td>1914.X.26</td>
<td>02:52</td>
<td>45.7</td>
<td>26.6</td>
<td>120</td>
<td>V</td>
</tr>
<tr>
<td>11</td>
<td>1929.XI.01</td>
<td>06:57</td>
<td>45.9</td>
<td>26.5</td>
<td>160</td>
<td>VIII</td>
</tr>
<tr>
<td>12</td>
<td>1934.III.29</td>
<td>20:06</td>
<td>45.8</td>
<td>26.5</td>
<td>90</td>
<td>VIII</td>
</tr>
<tr>
<td>13</td>
<td>1938.VII.13</td>
<td>20:15</td>
<td>45.9</td>
<td>26.7</td>
<td>120</td>
<td>VI</td>
</tr>
<tr>
<td>14</td>
<td>1940.VI.24</td>
<td>09:57</td>
<td>45.9</td>
<td>26.6</td>
<td>115</td>
<td>V</td>
</tr>
<tr>
<td>15</td>
<td>1940.X.22</td>
<td>06:37</td>
<td>45.8</td>
<td>26.4</td>
<td>125</td>
<td>VII</td>
</tr>
<tr>
<td>16</td>
<td>1940.XI.10</td>
<td>01:39</td>
<td>45.8</td>
<td>26.7</td>
<td>130</td>
<td>IX</td>
</tr>
</tbody>
</table>
As we can see from this table the values of intensity belong to the range V-IX and the depths of foci are belonging to the range 75-160 km.

RESULTS

The figures la-e show the attenuation of macroseismic intensity versus epicentral distance. For every maximum intensity I (belonging to the range V-IX) we plotted four diagrams corresponding to the four directions (E,W,N,S).

Analysing these observational data we conclude that for every given point the gradient of intensity variation is greater, as the intensity is greater therefore.

\[ \frac{\Delta I}{\Delta r} \sim I \]  
(1)

Because I is decreasing whereas epicentral distance is increasing, follows that \( \Delta I < 0 \) for any \( \Delta r > 0 \), so that we can introduce a constant value \( K \) and the relation (1) becomes:

\[ \Delta I = - K I \Delta r \]  
(2)

or:

\[ \frac{\Delta I}{I} = - K \Delta r \]  
(2')

By integration,

\[ \ln I = - K r + C \]  
(3)

or:

\[ I = I_0 e^{-kr} \]  
(3')

It is obvious that the curves from figures la-e are well-modeled by relation (3').

This result are usefull for seismic risk studies.

REFERENCES


Fig. 1 a-b

(a)

(b)
Fig. 1 c-e
APPLICATION OF PROBABILISTIC MODEL TO VRANCEA INTERMEDIATE EARTHQUAKES.

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In this paper we try to apply the first Gumbel type distribution in modelling of the seismogenic process for intermediate earthquakes occurred in Vrancea zone.

In this order we used a new set of data which is more complete than any previous similar work due to adding the data obtained from the records performed by the modern Romanian telemetred seismic network. The whole set of data consists both from macroseismic data (1901-1933) and instrumental data (1934-1991).

By processing of these data we obtained the values of fundamental parameters $\alpha$ and $\beta$ from cumulative distribution function:

$$ G(M) = \exp (-\alpha \exp (-\beta M)) ; M \geq 0 $$

These fundamental parameters $(\alpha, \beta)$ allowed us to determine a series of statistical parameters as: annual expected number of earthquakes $(N_M)$, the mean return periods $(T_M)$, the mean magnitude of all earthquakes with magnitudes $M > 0$ $(M_0)$, the mean magnitude over the data range set $(M)$, the nodal annual maximum magnitude $(M)$, the modal magnitude in a $D$ years period $(M_0)$, the maximum annual magnitude which is exceeded with the probability $p$ $(M_p)$, the maximum magnitude which is exceeded with the probability $p$ in a $D$ years period $(M_p(D))$ and the seismic hazard $(H(D))$.

Some of these statistical parameters $(M_0, M_p(D), H(D))$ were calculated for diverse values of time $(D)$.

The accumulation of additional data (1983-1991) allowed us to improve the precision in the computed values of the statistical parameters.

The modal magnitude is 4.9 and the corresponding return period is one year.

Some of these statistical parameters are usefull in antiseismic design.

Remark: Full text will be published in Natural Hazard.
M.I.A. LAWS FOR THOSE CRUSTAL EARTHQUAKE SOURCE REGIONS, WHICH DETERMINE SEISMIC HAZARD FOR ROMANIAN TERRITORY

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The Vrancea Seismogenic Region is the most subcrustal earthquake province of the Europe and one of the most peculiar seismic zone of the World, it putting the heaviest signature on seismic hazard of Romania (as well as on neighbouring countries). But there are also other internal (crustal) earthquake source regions (Fagaras, Banat, Maramures etc.) and external (crustal) earthquake source regions as well as neighbouring countries (Bulgaria, Yugoslavia) which all have an important contribution on seismic hazard of Romania.

This study was done in general aim of the seismic hazard evaluation on the basis of the spatial distribution of Macrosismic Intensity Attenuation (M.I.A) with distance, considering the directions of maximum and minimum attenuation. Upon global seismological considerations, the Romanian territory was divided in five zones and three adjacent zones from neighbouring countries (Bulgaria and Yugoslavia) in order to study a regional approach of the problem.

The attenuation of observed MSK-64 intensity from eighteen earthquakes, with epicentral intensity larger than III, was studied by fitting the observational data to different attenuation equations published in the seismological literature. An attempt is made to determine the rate of absorption of the seismic energy released in each zones, as well as the influence of the seismotectonic characteristics (Dumitrescu and Sandulescu, 1974), on the propagation of the seismic hazard at different distances from the epicentre. A possible explanation for the differences in spatial intensity attenuation within Romanian Territory is that younger and old structures of the crust contribute substantially to the absorption of seismic waves.

The macroseismic intensities have been extensively studied in Romania (Radu, 1976). The UNESCO project for the seismicity of the Balkan countries has offered detailed information for the revision of intensity of past earthquakes (Shebalin et al., 1974). We use the MSK-1964 scale to carry out the macroseismic studies of the earthquakes.

Eighteen isoseismal maps for the area of the Romanian and adjacent zones have been analysed, and the attenuation of intensities with distance and azimuth has been studied. Intensity attenuation relationship have been derived for different seismogenetic zones.

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Remark: Full text will be published in Natural Hazard.
INTRODUCTION

The realized graphical interface is a tool of immediate use for the visualization of results, elaborated by Seismic Hazard computational programs, providing raw punctual results. This package enables the user to realize a friendly access to these numerical data otherwise not easily interpretable. It is, then, possible to obtain a first and immediate analysis of Seismic Hazard parameters through cartographical bases and coloured maps. To reach this aim it has been developed a user-friendly interface, in order to reduce the effort required by the user during the learning phase.

It has been carried out an application, running on Personal Computers and hardware device independent: the Graphical User Interface of MicroSoft Windows 3.0 has been chosen for its worldwide employment and diffusion just on P.C. systems.

INTERFACE STRUCTURE

The graphical interface GEP5 has been projected to deal with input-output data of programs for the study on seismic hazard and risk, paying particular attention to the visualization of computed results.

Analysing the system structure, schematized in figure 1, we can see that the main in-out data channels are clearly shown. The input data flow is directed to two software packages, the expert system CZAR and a data post-processing routine, which are connected to the user through the interface.

The output data, produced by different modules, are collected in three separate databases.

CZAR - Classificatore Zone A Rischio

The expert system CZAR enable us to estimate how the choice of subjective parameters, made during the phase of hazard calculation, affects the seismic classification of the Italian territory (Cella, 1992). The implemented prototype allows our analysis on different seismic classifications, obtained using the same procedure for seismic hazard calculation, but varying the seismic catalogue and consequently the threshold values of parameters. Besides, the expert system produces a relative index, which combines the achieved different classifications, basing on the theoretical structure of Bayes' theorem.

CZAR inferential module has been carried out using the expert system shell Nexpert Object (Neuron Data, 1990), through which the knowledge base, constituted by object structures and production rules, has been developed. A completely independent module, under Windows 3.0 graphical environment, has been then yielded, connected and activated by the interface through a proper procedure.

Moreover, it is also possible to call a run-time version of the expert system CZAR, working on municipality scale but only for the Toscana region, for which the digitized municipality boundaries are available.

Post-processing routines

Sometimes, our analysis requests the visualization of a great quantity of data, particularly when we are working on national scale. It is useful, in these cases, not to represent the parameters associated to each single municipality, but to consider areal values estimated on the territory.

The developed routines, which can be called by the interface, enable the generation of a regular grid basing on different parameters, singly computed for each Italian municipality.
The post-processing routines can work on any kind of input-output data relating to calculation modules, just described in this system. It is, therefore, possible to elaborate CZAR input values and obtained classification results. Adding to this, there is one more possibility: that one concerning the grids calculation of seismic hazard and risk parameters (i.e. exceeding probability values of fixed intensity or acceleration levels).

Selecting the parameter to be elaborated, the grid step and its origin point are settled. At this stage, we can decide how to proceed: we could choose the calculation of the medium, the maximum or the minimum values relating to those municipalities included in each grid mesh. After the routine execution, a file of data, visualized by GEOSP and containing a regular representation of the investigated territory, is generated.

**Data bases**

As shown in figure 1, the interface links three distinct data bases containing information in ASCII and DbaseIII+ format. The first one contains digitized data, useful as cartographical support, where the results, to be visualized and only referenced by couples of coordinates, are overlaid. To this end some maps of the Italian territory and its borderlands have been collected. Though these maps are in very different formats, there is however, the possibility to convert them by different projection systems. Anyhow, inside the application, the used system has been unified, adopting in the visualization process, the Lambert's projection system expressed in binary format. If the selected map is in a different format, the interface converts it automatically.

The second data base contains data processed by programs for the study of seismic hazard and risk. These data are stored in ASCII format and organized according to the scheme: latitude, longitude and some parameters values. As well as the various computed hazard parameters, this data base included a set of values, used in the seismic classification rules by the expert system CZAR: historical maximum intensity ($I_{\text{max}}$), expected intensity for a given return period ($I_{500}$) and finally a design seismic coefficient, named $C/C_{\text{ref}}$.

The third data base collects results generated by the expert system: a global measure of classification changing the classification strategies, and a relative measure for discrepancies for different seismic classification strategies (Cella, 1992). Inside these files there is also stored geographical and administrative informations like the number of inhabitants for each municipality, its province and region.
Hazard & Risk routines

This group of procedures is external to the system structure and through the interface it is only possible to visualize results yielded and contained in the hazard and risk data base.

RESULTS VISUALIZATION

Data are visualized by the interface in three different ways: by punctual manner, by evaluation on a regular grid and on areal representations.

In figure 2 we have an example of punctual representation where each symbol (circle), whose size could be calibrated, locates a municipality and different chromatic shades show up the selected parameter value.

In the second type of representation (figure 3), each element in the grid reports a calculated value (medium, minimum, maximum) relating to all the municipalities inside of it.

Applying the parameter value on the entire municipality territory of the considered region, we get an area representation, as shown in figure 4.

GRAPHICAL TOOLS

It must be said that the user has, at his disposal, very efficient tools, developed thanks to the capacity of the adopted graphical environment. In figure 5, we find all the commands menus, controlling the graphical utilities.

![Figure 2: Punctual representation of the Italian seismic classification - Toscana region.](image-url)
Figure 3: Regular grid representation of the Italian seismic classification (medium value between the municipalities inside each mesh) - Toscana region.

Figure 4: Areal representation of the Italian seismic classification - Toscana region.
Through the menu Cartographical Bases we manage the facilities which better make "user friendly" this graphical interface. Boundary File selects a geographical map, where seismological data are overlaid; whereas the option Geographical Grid provides a checking on the visualization of the geographical grid (parallels and meridians), overlapped onto the map. It is possible to manually or automatically modify the grid step, when the investigated area is resizing. Legend, the third option, allows the setting and the sizing of the map legend in an interactive way: it is easy to do it, thanks to the immediate visualization of the final result inside a dialog window. Through the following option Range we can insert from the keyboard the geographical limits of the map to be displayed, choosing the minimum and maximum coordinates; by default the Italian peninsula borders are selected.

When the GEOP5 module is started, some operative options, like the legend visualization, the geographical grid step and the reference scale design, are then selected. These characteristics can be set through the option Default. From the sub-menu Parameters. In order to improve contrast and print neatness on the maps, the option Grey Scale can interactively select the different classes of grey shades and so get the best effect using any printer.

Pointing and selecting a zone to be displayed through the mouse, the Zoom procedure can increase a geographical map detail in an interactive way. Of course, it is possible to make further operations, as the interface keeps memory of any performed action and also allows to come back to the previous visualizations through the option Unzoom.

Finally, using the menu file, it is possible to start the output routines for print the maps on paper. Through this output routines we can activate the printer's software driver setting, the chosen colourshade test and the print of the work.

Besides, the interface provides a further possibility: an interactive query of the map elements. It is, then, possible to inquire the NexterP data base connected with the interface and extract those geographical-administrative informations, which cannot be graphically represented, but as to be displayed by characters. Selecting a main town point on the map by the right mouse button, we get the requested information, dynamically yielded, inside a window at the screen bottom.

![Figure 5: Menus structure of the interface GEOP5.](image)
The interface is entirely developed under MS-Windows 3.0 environment with Microsoft Software Development Kit (Microsoft, 1990). Such a toolkit allows to write applications on Personal Computers with all the features of a multtasking environment and provides a graphical library for typical windows, buttons and menus design of the used windows environment.

It must be, finally, remarked that the described system is not a Geographic Information System (G.I.S.), though it has some similar characteristics, like the capacity of the results visualization on cartographical basis, but it is only a working tool for seismic hazard and risk analysis.

REFERENCES


ATTENUATION OF VERTICAL STRONG-MOTION OF SHALLOW EARTHQUAKES IN GREECE

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INTRODUCTION
The empirical characterization of strong ground motion attenuation has mainly emphasized horizontal shaking. This is due to the principal use of strong-motion amplitude data in the earthquake resistant engineering of structures, where the transient vertical earthquake loads have been viewed as relatively unimportant perturbations of the loads imposed by the earth's gravitational field. In some building codes it is generally assumed that the peak vertical acceleration is simply a fraction of the peak horizontal acceleration. Usually, a value of 2/3 is used as the maximum effective ratio between vertical and horizontal accelerations (Newmark and Hall 1982). This value has been recently considered unconservative in the near field (Abrahamson and Litehiser 1989, Niazi and Bozorgnia 1992). In this paper, based on strong-motion data set from shallow earthquakes in Greece, the attenuation of vertical acceleration, velocity, displacement and spectral velocity is studied. Comparison with the attenuation of horizontal strong-motion of the same data set is made, as well as, with other relations of vertical strong-motion recently derived for other regions, is given.

DATA-METHOD-RESULTS
The data set for peak vertical acceleration, velocity, displacement and spectral velocity (fig.1), consists of 51 components from 34 shallow earthquakes in Greece with surface wave magnitudes 4.5≤Ms≤7.0 and focal depths h≤18 km (Theodulidis and Papazachos 1992). Recording sites were classified into two broad categories, rock and alluvium. The grouping was made on the basis of the stiffness of the material at the instrument location, together with a general knowledge of some of the individual sites, following a methodology proposed by Trifunac and Brady (1975).

The adopted model for the regression analysis was of the form,

\[ \ln Y = C_1 + C_2 \times M_s + C_3 \times \ln (R + R_0) + C_4 \times S + \sigma_{\ln Y} P \]  

\[ P = \begin{cases} 0 & \text{for 50-percentile level of non-exceedence} \\ 1 & \text{for 84-percentile level of non-exceedence} \end{cases} \]

where \( Y \) is the strong motion parameter to be predicted, \( M_s \) surface wave magnitude, \( R \) epicentral distance, \( S \) a binary variable which takes values 1 for rock and 0 for alluvium local soil conditions, \( P \) is 0 for 50-percentile and 1 for 84-percentile level of non-exceedence. Coefficient \( R_0 \), accounts for saturation in the near field and was taken equal to 15, 10, 5 for acceleration or spectral velocity, velocity and displacement, respectively. Scaling coefficients \( C_1, C_2, C_3, C_4 \) and the residuals root mean square, \( \sigma_{\ln Y} \), calculated from a four step regression analysis which is described in our previous paper (Theodulidis and Papazachos 1992). For spectral velocity, scaling coefficients and \( \sigma_{\ln Y} \) are a function of period (T-sec) and critical damping (D).

The resulting equations for peak ground vertical acceleration (cm/sec\(^2\)), velocity (cm/sec) and displacement (cm) are,

\[ \ln a_v = -5.23 + 0.87 \times M_s - 1.81 \times \ln (R + 15) + 0.04 \times S + 0.72 \times P \]  

\[ \ln v_v = -4.82 + 1.04 \times M_s - 1.79 \times \ln (R + 10) - 0.41 \times S + 0.75 \times P \]  

\[ \ln d_v = -5.14 + 1.55 \times M_s - 1.53 \times \ln (R + 5) - 1.58 \times S + 1.62 \times P \]

Similar relations, for certain periods and critical damping \( D=0.05 \), of vertical spectral velocity (cm/sec) may be derived by using scaling coefficients of the table I and equation (1).
Table I. Scaling coefficients of equation (1) for 5%-damped vertical spectral velocity, for shallow earthquakes in Greece.

<table>
<thead>
<tr>
<th>T (sec)</th>
<th>C1(T)</th>
<th>C2(T)</th>
<th>C3(T)</th>
<th>C4(T)</th>
<th>C5(Y)</th>
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<tr>
<td>0.05</td>
<td>1.677</td>
<td>0.998</td>
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<td>0.10</td>
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<td>-1.910</td>
<td>0.082</td>
<td>0.776</td>
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<tr>
<td>0.15</td>
<td>2.848</td>
<td>0.818</td>
<td>-1.853</td>
<td>-0.062</td>
<td>0.756</td>
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<td>0.30</td>
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<td>1.156</td>
<td>-1.788</td>
<td>-0.145</td>
<td>0.757</td>
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<tr>
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<td>1.230</td>
<td>-1.526</td>
<td>-0.528</td>
<td>0.800</td>
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<td>0.75</td>
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<td>-1.660</td>
<td>-0.980</td>
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</table>

DISCUSSION

The attenuation relations of the vertical strong-motion of shallow earthquakes in Greece, proposed in this study, can be considered valid for distances 1 ≤ R ≤ 130 km and magnitudes 4.5 ≤ Mₛ ≤ 7.0. However, the lack of data in the near field, R < 20 km, for large earthquakes, Mₛ > 6.2, must be taken into account when applying these relations. The ratio of vertical to horizontal peak ground acceleration is shown in fig.2 as a function of distance. It seems that the commonly used in engineering applications value aₐ/aᵥ = 2/3, is generally conservative in the far field, which is in quite good agreement with conclusions of relevant studies (Abrahamson and Litwhiser 1989, Ambraseys and Bommer 1991). Using a gross geology classification scheme of rock or alluvium, the site effect on vertical acceleration is negligible. However, the site effect on the aᵥ/aₐ ratio is quite significant with alluvium sites yielding larger ratios than rock sites. As can be deduced from relations (3),(4) for velocity and displacement, site effects become evident with alluvium sites yielding larger values. Comparing the attenuation of peak vertical acceleration in Greece with similar relations proposed for other regions (fig.3) we found them in quite good agreement up to about 50 km. For larger distances the vertical acceleration in Greece seems to attenuate faster than the others which, however, fall into 84-percentile of the vertical acceleration amplitudes proposed in this study. Comparing (fig.4) the predicted vertical spectral velocity response spectra, 5%-damped, with those proposed for horizontal ones (Theodulidis 1991), for 0 and 30 km distances, it is apparent that in the near field vertical strong-motion contributes significantly to the high frequency excitation. This is in agreement with recent results obtained from SMART-1 array in Taiwan (Niazi and Bozorgnia 1992). However, intermediate and low frequency amplitudes are drastically controlled by horizontal strong-motion.

ACKNOWLEDGEMENTS

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REFERENCES


Fig. 1. Magnitude-distance distribution of the data used.

Fig. 2. Vertical to horizontal peak ground acceleration ratio versus epicentral distance, for shallow earthquakes in Greece.
Fig. 3. Comparison of the peak ground vertical acceleration attenuation proposed in this study with other relevant relations.

Ms=5.0

Fig. 4. Predicted 50 and 84-percentiles vertical velocity spectra (dashed line) in comparison with horizontal spectra (Theodulidis 1991), for distances 0 km and 30 km.
ANALYSIS OF NONSTATIONARY PROPERTIES OF GROUND MOTION RECORDED DURING THE MONTENEGRO EARTHQUAKE OF APRIL 15, 1979

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The accelerogram of a strong ground motion represents the detailed characteristics of an earthquakes and contains a large number of data both for seismologists and earthquake engineers. Therefore, recently, accelerograms analysis has become a widely accepted method for fundamental investigations referring to understanding of the background of the seismic waves generation and propagation mechanism and investigation of the strong ground motion nature and the parameters defining the ground motion for the needs of seismic resistant design.

The method of physical spectrum was used as an approach to investigate the time variable characteristic of strong ground motions recorded during the Montenegro earthquake of April 15, 1979 with $M = 6.8$ (ISC). The application of the physical spectra for the design of a realistic synthesized accelerogram is also presented.

The presentation of the results obtained by the analysis in terms of two- and three-dimensional spectra points out clearly to the nonstationary character of the intensity and the frequency content of the earthquake ground motion at all the considered sites. This method enables a more sophisticated control of the energy release process from the earthquake origin and the considered case points out to the fact that the earthquake consists of several individual shocks originating from different parts of the fault and with different time of occurrence. This characteristic and knowledge that the dynamic properties of the structure can be considerably modified during the initial part of the earthquake and that the additional arrival of similar intensity energy from the earthquake origin can have a devastating effect on the structural stability should be on the mind for the reasons of occurrence of heavy damage and failure of structures over a large area during this earthquake.

Also, the design of artificial accelerograms by using physical spectra enables that the amplitude, the time duration and the frequency content be simultaneously specified.
BLIND PREDICTION OF THE SITE EFFECTS AT ASHIGARA VALLEY, JAPAN, AND ITS COMPARISON WITH REALITY

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ABSTRACT

Weak and strong ground motions were numerically predicted at three stations of the Ashigara Valley test site. The prediction was based on the records from a rock-outcrop station (KR1), one weak-motion record from a surface station at sediments (KS1), and the standard geotechnical model. The data were provided by The Japanese Working Group on the Effects of Surface Geology as a part of the international experiment. The finite-difference method for SH waves in a 2-D linear viscoelastic medium (a causal Q model) was employed.

Comparison with the real records shows that at two stations (KS2, KD2) the predictions were better than at the third one (KS1). Strangely, the two better predictions were for stations situated at larger distances from the reference rock station (one station was on the surface, KS2, the other in a borehole, KD2). The strong ground motion (the peak acceleration of about 200 cm/s/s) was not predicted qualitatively worse than the weak motion (8 cm/s/s). A less sophisticated second prediction (not submitted during the experiment), in which we did not attempt to fit the available weak-motion record at the sedimentary station, agrees with the reality better than the first one.

The example in Fig. 1 compares the observed and the predicted strong ground motions at the most problematic station KS1. Fig. 2 refers to the more successful prediction at station KS2.

Full text of this contribution has been submitted to Natural Hazards.
KS1, NS, STRONG
max. accel.: 199 cm/sec/sec
observed

Fig. 1

KS2, NS, STRONG
max. accel.: 220 cm/sec/sec
observed

Fig. 2

Normalized acceleration

Time (seconds)
COMPARATIVE STUDY OF DIFFERENT TECHNIQUES FOR THE GENERATION OF SYNTHETIC SEISMOGRAMS IN SIMPLE GEOLOGIC MEDIA.

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R. FREGONESE, O. BERTULETTI, F. PACOR
ISMES, v.le G. Cesare, 29, Bergamo, Italy.

INTRODUCTION
With the final aim about a correct prediction of the soil motion during strong earthquakes utilizing a parametric description of the source and considering the waves propagation and site effects, we acquired three different synthetic seismogram calculation codes which span the frequency range 0.1 - 20 Hz. These codes are: SISMA and SEIS83 for the calculation of low frequency synthetics up to 3 - 4 Hz; STOCH for the calculation of high frequency synthetics (3 - 20 Hz). The program SISMA calculates the superficial response of a horizontally layered visco-elastic halfspace from an embedded point source; a discrete wavenumber approach (Bouchon 1979, 1981) is used to calculate the complete wavefield. The program SEIS83 generates synthetic seismograms using ray tracing theory (Cerveny, 1977, 1984), for propagation in a 2D laterally inhomogeneous medium. Direct arrival P and S waveforms are synthesized, together with the principal reflected and converted waves. The program STOCH generates synthetic seismograms based on Boore's technique (Boore, 1983). White gaussian noise is modulated in the frequency domain with a target spectra. The target spectra is generated by specifying a point source parameters, (Mo, fo), the distance between the source and receiver, some attenuation parameters and the local site amplification transfer function. In a first phase of this study, the range of applicability of the three codes was determined based on a comparison of the resulting synthetic seismograms with theoretical results. The second phase, which is presented here, consisted in simulating data from a real earthquake. We choose an event in the Padana valley, (North Italy); where the medium is well studied and can be adequately represented by a horizontally layered halfspace. For this earthquake good quality data have been recorded, by two local seismometric arrays and by a regional one, all installed by ENEL (1983 1988). (Table n.1 show the velocity model used for the Padana valley, also tabulated are the principal focal mechanism parameters).

SIMULATION WITH SISMA
As an example, figure n.1 compares synthetic velocity traces with those actually recorded at station RVT. Both synthetics and real data were bandpassed filtered between 1.5 and 3.0 Hz. Synthetics and data show a good agreement both in arrival times and waveform shape.

SIMULATION WITH SEIS83
For the comparison of synthetics generated with real data, it was necessary to combine the NS and EW components of the record in order to coincide with the reference system defined by the 2D cross section between the source and receiver. The figure n.2 displays the simulation of records at station RVT. Both data and
Synthetics were bandpassed filtered between 0.6 - 5 Hz. There is a
poor agreement in waveform shape although the synthetics are able
to reproduce the arrival times. The synthetic and record are
similar in amplitude and in arrival times. Table n.2 compares
observed P, S, and S - P times with those calculated from the
synthetics generated by SEIS83 and SISMA programs at three
station.

**SIMULATION WITH STOCH**

The target spectra used in STOCH is defined by four fundamental
parameters: the attenuation parameters $Q_s$ (s-wave scattering), and
k (anelastic absorption), and the seismic moment and corner
frequency of the source. The values of $Q_s$ and $M_o$ (seismic moment)
are the same as those used in the previous simulation. It was
assumed that the dependence of $Q_s$ with frequency is linear ($Q_s =
Q_0 f$). The values of k and $f_0$ (corner frequency) were estimated
directly from the data: k with the technique of Anderson and Hough
(Anderson, J. G. and Hough, S. E., 1984); $f_0$ with the technique of
Andrews (Andrews, D. J., 1986); $M_o$ following the Brune's model
(Brune, J. N., 1970). Figure n.3 show the theoretical spectra
superimposed on the real acceleration spectra for the NS and EW
components of motion at station RVT and figure n.4 show the high
frequency spectral decay observed in the data and compared with
the attenuation given by k (kappa). Table n.3 compares peak and
average velocity and acceleration values obtained from the
 recordings considered in this study with the values estimated with
STOCH. The real values always lie within one standard deviation
from the theoretical values.

**CONCLUSIONS**

SISMA proved to be a reliable method for the calculation of low
frequency synthetic ground motion. One drawback of this code is
that only simple geological media can be accurately modeled
(horizontally layered media).

SEIS83 did not provide a good waveform match with the data. This
code, however, proved to be a good tool to analyze wave
propagation (calculation of arrival times of different waves,
identification of angles of incidence, identification of direct
and reflected waves).

STOCH provides a good statistical description of high frequency
ground motion given accurate estimates of a small number of
fundamental source parameters and attenuation parameters.

The use of any of
these three codes is
strongly related with
the problem that one
wishes to analyze. The
programs complement
one another and their
combined use affords
the possibility of a
global study of
earthquake ground
motion synthesis
through the techniques
of hybrid methods
(broad band signal).
REFERENCES


<table>
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<td>Ts (sec.)</td>
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Table n.2

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<th>R</th>
<th>ACC.P. VAR.</th>
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<td>-3.61</td>
<td>-3.63</td>
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</tr>
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</table>

Table n.3 Logarithmic values on PE simulation

Figure n.1 Simulation with SISMA. (Upper traces correspond at recorded data)
Figure n.2 Simulation with SEISS3

Figure n.3 Simulation with STOCH and spectral decay compared with the attenuation parameter K.
1. INTRODUCTION

Based on a high accuracy acceleration measurement technique in the space industry an integrated, miniaturised capacitive accelerometer has been developed for industrial purposes and made available for seismic strong motion networking technology.

The technology of capacitive sensors has been improved by introducing a fabrication procedure based partially on silicon wafers and partially on plates made of glass and silicon. The main aspects of this technology relate to:
- integrated process for circuit fabrication
- anisotropy etching of silicon by means of wet chemistry
- anodic bonding of glass to silicon

2. TECHNICAL PRINCIPLE

The key element of the acceleration sensor is a micro mechanical silicon chip comprising a movable plate suspended by flexure bars. This plate acts as an inertial mass and as a capacitor plate. This element is chemically etched to a rigid central mass suspended by thin, flexible membranes whose thicknesses are varied depending on the full scale acceleration range of the unit. It is deflected when an acceleration perpendicular to the surface is applied. This deflection is transformed into an electrical signal by measuring the variations of the capacitor formed by the movable plate and fixed electrodes placed on either side (See Fig. 1). The latter are made of aluminium thin films deposited on glass plates which are themselves fixed to outer silicon plates in order to provide good mechanical stability. The cavity containing the movable plate is sealed hermetically at reduced pressure to achieve controlled damping of the plate. The elimination of adhesive bonding and of moving components results in a design with extremely high reliability.

![Fig. 1: Schematic drawing of the accelerometer chip](image)
The micro sensors inertial mass deflects parallel to the top and bottom plates with minimal flexion or distortion. The cross-axis sensitivity is minimised due to the high membrane stiffness in the lateral direction and through the positioning of the inertial mass at the geometric centre of the element. As the acceleration level increases, the inertial mass deflection increases, eventually being restrained by overrange stops. This limits any further stress to the flexing membrane and provides excellent overrange capability.

The single crystal nature of the silicon, coupled with the elimination of any mechanical joints between the mass and the membrane, results in an extremely rugged sensor. The capacitive approach is inherently up to 20 times more sensitive than its piezo-resistive counterpart which has to be drastically amplified causing offset drift and instability.

3. PRODUCTION

The batch fabrication is based on CSEM's (Centre Suisse d'Electronique et de Micromecanique SA) proprietary technology which uses microelectronics photolithography techniques in conjunction with silicon etching and anodic bonding processes for wafer sealing (See Fig. 2). The final chip consists of a mechanical system enclosed in an hermetically sealed cavity.

![Fig. 2: The device after silicon etching](image)

4. MEASUREMENT ELECTRONICS

Measurement electronics bases on a self-balancing capacitance bridge. An AC signal of constant amplitude is applied to the fixed electrodes and the resulting current on the middle electrode is amplified and demodulated in such a way as to balance the capacitance bridge leading to an output signal proportional to the plate deflection (See Fig. 3).
This miniaturised capacitive accelerometer, which shows a high sensitivity in static and low frequency measurement ranges, is a highly competitive alternative for the monitoring of seismic motion. Based on this outstanding sensor technology, Syscom Instruments AG, Zürich, Switzerland has successfully developed a strong motion recording system. The instrument consists of a Sensor Unit, the MS2002, and a Data Acquisition Unit, the MR2002. Both are housed in rugged aluminium casings which are water and shock resistant (See Fig. 5). Also available is a version which incorporates the Sensor Unit within the Data Acquisition Unit.

The instrument detects vibration events and records them on an internal static RAM or on a digital memory card. The data may be downloaded onto a portable laptop in the field for immediate evaluation or the memory card can be removed and forwarded to a central location for analysis. Data may also be directly transferred via modem to any location over public or dedicated telephone lines.

**5. APPLICATION**

Fig. 3: Electrical diagram of the measurement Electronic

Fig. 4: The sensor as black rectangle unit on the board

Fig. 4 shows the hermetically sealed sensor as black rectangle unit on a board with the necessary electronics. The outer dimensions of this board are 39x39 mm. The three sensors for the three axes x, y and z are then mounted on a fixing bloc providing plane positions.

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6. SUMMARY

The high standard capacitive sensor shows an ideal behaviour for applications in the seismic frequency range. This is especially true for its ideal behaviour in amplitude and phase. No recalibration is needed and in-built on-line test-capabilities provide an easy use of the system.

Syscom Instruments AG has over the past few years successfully entered the seismic area with a new data acquisition system called the MR2002. The sensing element used by the MR2002, is composed of 3 monoaxis capacitive accelerometers, produced by the Swiss company Access Sensors SA.
ARTIFICIAL INTELLIGENCE IDENTIFICATION OF REGIONAL SEISMIC EVENTS

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D.Rouland, J.Bonnin, Institute of Physics of the Earth. Strasbourg University. France

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Quasi-real time interpretation of seismic records is of great interest for fast localization, depth identification, source history reconstruction and energy estimation of major and/or damaging earthquakes. Broad band records contain most of the information necessary to recover these source parameters, although it supposes years of training at the seismological observatory to recover partly this information and a lot of computer resources. Hence a promising interactive tool to solve the problem could be provided by artificial intelligence approach.

A source of knowledge is a collection of well identified seismograms, which can be considered as a set of examples in computer aided decision making. We explore Syntactic Pattern Recognition Scheme (SPARS) which is based on the cluster analysis of the nonlinearly aligned traces represented as sequences of parameterized (fixed length) fragments using dynamic programming technique.

To test the capabilities of the method we have analyzed only the vertical component of broad band records obtained in the GEOSCOPE station at Noumea (New Caledonia). The so called LGLP component recorded at a 1 Hz sampling rate has been selected for a group of 3 seismic regions: the Vanuatu – Loyalty region, the Fiji region, and the Tonga region, each of them including shallow, intermediate, and deep earthquakes. The Knowledge Base has been formed by a collection of 61 events (31 in Vanuatu, 15 in Fiji, and 15 in Tonga region, see Figure 1). Epicentral distances for the Vanuatu region vary from 447 to 1235 km, for the Fiji region from 1027 to 1462 km, and for the Tonga region from 1497 to 2310 km. The event magnitudes mb range from 4.3 to 6.4. In this first approach, using only one component, we have tried successfully to distinguish the regional characterization of each record.

To do this each seismogram (10 min duration) was divided into the set of non-overlapping fragments 16 sec duration. Each record has been filtered by a set of 10 Gaussian narrow band filters logarithmically centered between 6 and 25 sec and rescaled to have a dynamic range limited to 30 dB.

Using the one nearest neighbour decision rule we obtained only 3 errors among the 61 analyzed events: that means 95% correct classification rate.

The corresponding 3 events in fact are at the boundaries of the considered seismic regions. Two events are situated at the so called back arc of the Vanuatu subduction zone, and one event is at the northern part of the Vanuatu arc, close to the Solomon seismic region.
It is interesting to notice, that within each seismic region the deep events are well associated between themselves, what could be seen on the average linkage hierarchical clustering dendrogram.

For shallow events, which form the bulk of the data we used (42 events), 33 events are well associated by SPARS as well as by epicentral location (pairwise distances less than 150 km), and 22 among them are perfectly associated, it means that epicenters of these SPARS neighbours are closer than 50 km.

Only for 5 events the closest neighbours in SPARS are located far from each other. Furthermore, for 3 another records the closest SPARS neighbours do not correlate from a seismological point of view; they show an apparently good fit in surface waves, but they does not fit at all in body waves.

The further development of the algorithms can be seen in following ways.

First, the ambiguity in body wave interpretation seems to be easy to overcome by using polarization figures of three components records.

Second, it is possible to include into the algorithms the geophysical knowledge about travel time of the seismic waves together with the statistical estimate of the likelihood of the classification using such models like Hidden Markov Chains.

Third, the analysis of the seismic waves envelops similar to SPARS can be used for estimation of the energy released by the earthquakes giving information about magnitude and/or seismic moment.

![Map of the locations of the events.](image-url)
ACTIVE DATABASE FOR THE STRONG MOTION RECORDS

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INTRODUCTION

The strong motion accelerograph network throughout the Yugoslav territory was first installed at the beginning of 1970 year, covering all areas which are seismically active.

During the past period, more then 1000 records have been collected and analyzed in the Institute. Selected sets of this data have been published in various Institute publications.

Accordingly, the main purpose for developing a program package for base management of databases for strong motion data, was efficient manipulation and appropriate processing and storing of the data.

In this text we describe the program package and appropriate databases. A rule management procedure is added to the database management system (DBMS) for maintaining event-condition-action rules, concept used in HIPAC, an active, object-oriented database management system.

However it is planned to develop supporting software for statistical analysis of the data, for the purpose of the seismic hazard analysis. Such active and dynamic updating of the database enables permanent and continuous improvement of the data and knowledge regarding the problems associated with strong motion seismology and related topics.

Also, an automatic processing of the accelerograms including automatic digitization, processing and storing of data on an appropriate medium (hard or floppy disc) is planned for realization. Automatic digitization of accelerograms can be done using a scanner and appropriate application program package which will be developed.

ARCHITECTURE OF A DBMS FOR MANIPULATION OF A STRONG MOTION DATA

Active database management system for efficient manipulation of a strong motion data can be considered as it is shown in Figure 1.

Figure 1 Architecture of an Activated DBMS for manipulation of a strong motion data
Event appearing is signaled by the applicative detector of events. Rule management procedure determines which rules are fired by the event and calls the condition evaluation procedure.

During the condition evaluation process, queries on the databases can be executed for providing appropriated data. List of satisfied rules is passed to the rule management procedure. Depending on this list, actions are executed which include database operations or some other operations.

The database management system is activated with the use of event-condition-action rules. When a new record is processed, an event is signalized, a condition is evaluated, and if it is satisfied, or with other words, if the answer on the query is not empty, an action is performed which can include operations with the database records. Presently, the action is restricted on updating the database with the new record.

**STRUCTURE OF THE DATABASES**

Databases are organized so that the requirements for simplicity and minimal redundancy are satisfied.

In the projecting of the relational databases the standard methods are used: Every entity is presented as a relation (table) and the primary key is specified. Then, every association among entities is presented as a table, and the outer keys are used for specifying the associations. After that, relational databases are normalized, so that the redundancy is eliminated.

The result of that process are five databases which are described in the following text:

1. Database which consists of the data about the earthquakes - name of the earthquake, which is usually the name of the region where the earthquake takes place, date, local time, geographical coordinates of the earthquake, magnitude, intensity in MM, MCS or MSK64 scale, number of accelerograms for that earthquake

2. Database for locations of the instruments SMA-1 and SMA-2 - coordinates of the place where accelerographs are located and local side condition data

3. Database with data related to the accelerograms - maximum acceleration for each of the three components (longitudinal, transversal and vertical), epicentral distance, length of raw records, name of the publication, name of the file of digitized accelerogram

4. Database for the data of instruments - data for sensitivity, damping and period for each of the three components of the instrument, and the date of putting and taking out the microfilm of the instrument

   In accordance with the variation of the characteristics of the instruments, for each accelerogram there is appropriate record for the instrument.

5. Database with data from processed accelerograms - band-pass frequencies of the filter for the three components of the record, peak acceleration, peak velocity and peak displacements for each of
the three components of the accelerogram, length of processed accelerogram, method of processing, name of the file from which we can take plots of the three components of the accelerogram.

CURRENT CONDITION OF THE PROGRAM PACKAGE

Presently, program package consists of three main parts:

- Maintaining data for earthquakes, locations of the instruments and data from earthquake records

- Reports

- Maintaining data for the users of the databases.

Using the first option, with especially created masks, which significantly simplifies the communication with the user, can be added or updated (view, browse, delete) records from the databases.

Two types of reports can be created: summary and selective. Summary report is a complete view of the database records in which database records from different databases are connected using the cross-index fields.

The database records which satisfy some requirements regarding the magnitude and coordinates of the earthquake, location coordinates of the instrument, epicentral distance, peak acceleration, etc, can be selected from the database to obtain a selective report. (See Appendix 1)

Apart from printed reports, graph plots of the three components of the accelerogram can be obtained. Such a plot is shown in Figure 2.

![Graph plot of processed accelerogram](image)

Figure 2: A graph plot of processed accelerogram

Figure 3 shows the mask for selecting the range for magnitude, coordinates of the earthquake, location coordinates of the instrument, epicentral distance, peak acceleration, etc. This mask is used in creating a selective report.
CONCLUSION

Presently, classical databases of the strong motion data exist in the Institute. There is much still to be done before the program package for database management can be considered to be complete for connecting with the software for automatic processing of the accelerograms, or software for statistical analysis of the data for the purpose of the seismic hazard analysis. But in given form, it enables permanent and continuous improvements and is a useful tool for manipulation of the strong motion data.

REFERENCES

Dayal U. (1988). Active Database Management Systems, 3rd Int. Conf. on Data & Knowledge Based Systems

APPENDIX 1

<table>
<thead>
<tr>
<th>R.BR. IME NA ZEMJOTRES</th>
<th>DATUM</th>
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<td>27 BANJA LUKA LOKALEN</td>
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SEISMOGEOLICAL AND HYDROLOGICAL CRITERIA
FOR THE EUROPEAN MACROSEISMIC SCALE (MSK-92)

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One of the trend followed by the ESC "Macroseismic Scales" Working Group for updating the MSK intensity scale is to consider the response of three main categories of "sensors": humans, objects, buildings. These sensors are generally present in a number of units sufficient to be treated in a statistical way. On the contrary, seismogeological effects are by their nature such that: a) they generally don't appear in several instances, so that they cannot be treated statistically; b) most of them appear outside, or even far away from the localities where intensity is assessed, and therefore cannot be merged with the other observations.

To give an adequate account of the subject within the scope of the scale proved not to be easy, and arduous discussions took place before an agreement was reached. A long textual essay would not be appropriate or easily applied, and a simple table would be in danger of misleading the user. The solution adopted was a graphical one, in which each phenomenon (or diagnostic) is rated for the overall range of intensities at which it is possible to observe it, the range of intensities which is most typical, and also the range at which the diagnostic is most useful in discriminating between different intensity values (for detailed explanation of the table, see Grünthal, ed., 1993). The use of arrowheads in the figure implies the possibility of the ranges being extended even further under extreme circumstances.

REFERENCES

The "SIRENE" macroseismic data base has been utilized to enable isoseismal maps to be drawn for 136 of the best-documented, historical as well as contemporary, French earthquakes; these have epicentral intensities of at least V and cover all parts of the country. A study of focal depths obtained from the decay laws of all available individual local intensities with distance (Sponheuer relationship) shows that a majority of the events processed have focal depths no greater than 10 km and that their distribution correlates positively with the broad characteristics of the ambient dynamic geology. A relationship is then computed between magnitude, intensity, and focal distance. It is based on 68 earthquakes that were recorded instrumentally (M_L between 3.3 and 6.3) and on 235 of their average radii (between 3 and 220 km) for isoseismals of VIII to III (MSK).

A comparison of this empirical relationship with others established from various collections of data made it apparent that the observed discrepancies could in part be explained by the influence of: 1) differing seismic wave attenuations from separate regions; 2) heterogeneities in the evaluation of seismic characteristics (magnitude and intensity).

The effort made in France to gather macroseismic observations, to evaluate the intensities in a homogeneous manner by the same team within a uniform scale of European practical experience (MSK) in order to compile a national reference file should be extended to the elaboration of a uniform scale of magnitudes. In our opinion, the reliability and homogeneity of the data have a stronger influence on the validity of the results than the sophistication of calculation models, which can, in majority of cases, be simplified in view of the scattered nature of macroseismic data inherent to their empirical basis.

This relationship is used to determine the magnitude of historical earthquakes, contained in the SIRENE file, which are known only by their intensities. Then, the results obtained in this study is the basis necessary to the assessment of seismic hazard on the French territory, following a deterministic as well a probabilistic approach. An atlas, presently at press, will include the 136 macroseismic maps. Each of these will be accompanied by a list of localities affected by the event, with the locally observed intensities, bibliographical references relating the effects and some characteristics of the source resulting from this study (epicentral intensity, average radii of the pleistoseismal areas, magnitude, focal depth) and from an instrumental determination in some instances.
MACROSEISMIC PRACTICE IN SLOVENIA

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ABSTRACT

In 1985 a project was started concerning the improvement of macroseismic data management in Slovenia. We have decided to establish a network of collaborators all over Slovenia (20251 km², approx. 2 million inhabitants), dense enough to allow the use of the MSK scale. Since the uniform spatial distribution with desired density of collaborators is hard to achieve we use several methods to obtain new collaborators. At present (September 1992) we have approximately 4100 collaborators in our computer supported data base.

When an earthquake is felt in Slovenia, the seismologist decides in what area the questionnaires are to be sent. The persons can be chosen by two criteria: density mark and quality mark. Density mark allows us to choose the number of observers in the same town or village, based mostly on the quality of answers and demographic conditions. Quality mark is computed, combined with the previous marks and updated every time a questionnaire is received from a person.

The average number of shocks for which we send questionnaires is about 25 per year during the normal seismic activity. The average number of questionnaires is approximately 300 per event. The percentage of the returned questionnaires is above 70%. We have developed an user friendly, interactive program: to manage data base with the information about collaborators, to select an area, density and quality threshold of inquiring, and to print addresses onto questionnaires.

(To be published in "Natural Hazards")
INTRODUCTION

Macroseismic intensity map is the most comprehensive and informative representation, for seismologists and urban planners, of an earthquake which occurred in the past, when no instrumental records were available. In some cases, this is true even during this century because the lack of seismographs has not enabled a good monitoring of the source parameters, or a detailed picture of the distribution of ground motion in the epicentral area. Frequently, the pure experimental data, i.e. a list of localities each with an associated evaluation of intensity, are not presented without an interpretation, or not presented at all: what is in fact an isoseismal map, or a so-called macroseismic epicenter, if not an interpretation?

The problem arises by the fact that the interpretations are almost never declared, and never follow well defined standards of work, giving different results from an author to another. The macroseismic epicenter, for example, could be defined theoretically by one of the "boxes" in Fig. 1, but rarely the sentence is strictly applied, and a generic "expert judgement" is invoked. Furthermore, Macroseismic Epicenter (ME) is a crucial quantity, because it is the common way of characterizing an event in an earthquake catalogue. At this point the same "object" is freely used both by seismotectonic studies, that look at the ME as a direct representation of the source location, and by seismic hazard analyses, for which ME is a point of application of an attenuation model, too.

The capability in reproducing the damage distribution of the past is particularly felt in seismic hazard assessment, but attenuation models of macroseismic intensity versus distance are often applied to an epicenter that has nothing to do with the "true" point of application of the chosen model, giving computed values of intensity that could completely differ from the observed ones.

![Diagram of Commonly used definitions of Macroseismic Epicenter (ME).](image-url)
A significant example is given in Fig. 2. The well documented Soncino earthquake of May 12, 1802 (Postpischl, 1985) is presented in Fig. 2a; the isoseismal lines of the highest intensities are quite well controlled and paint a field not far to be circular: the star shows the ME location proposed by the authors. It seems an earthquake easy to be reproduced by using a model. On the opposite, in generating a synthetic field from the ME and an attenuation relation we often obtain a very bad reproduction of the intensity map. Here Blake (1941) model has been chosen; its parameters have been calculated on Western Alps historical seismicity. Nevertheless, 8 localities over 12 of the maximum intensity VIII MCS, 2 over 2 of VII-VIII, 4 over 4 of VII, 1 over 2 of VI-VII, and 1 over 4 of VI are not predicted by the model (underlined intensity values in Fig. 2b); so only the 30% of the localities that experienced damage (with intensity greater than VI MCS) fall inside the right isoseismal lines of the model.

Figure 2: Soncino earthquake of May 12, 1802: a) intensity map with ME (star) and isoseismal lines (from Postpischl, 1985); b) isoseismal lines with Blake model (parameters \( \gamma = 2.54, h = 1.89 \)), underlined values of intensity are not correctly predicted by the model.
in this way, all the probabilistic seismic hazard methods that try to reproduce
the seismic process at a site starting from an epicenter and an attenuation
relation are strongly biased by the traditional approach to ME location.
The aim of this work is to describe the methodology followed to develop a
procedure that, locating the point of application of a given model, enables to
better reproduce a macroseismic intensity map. Its utilization is of great
importance in compiling an earthquake catalogue devoted to seismic hazard
analyses.

MACROSEISMIC EPICENTER FOR SEISMIC HAZARD PURPOSES
Attenuation models of macroseismic intensity have been proposed by many
authors; most of them were developed in the aim of obtaining some hypocentral
parameters, like for example in Blake (1941) where h is considered to be the
depth of the focus. Under these hypotheses, the discrete character of intensity is
neglected, uncertain attributions of intensity constitute a new class of values,
and the computation is performed using real numbers for intensity. Some
computer location programs have been developed by the practice in
instrumental quake location (see e.g. Console et al., 1990), forcing the original
destination of the attenuation relationships; a macroseismic attenuation law is
"a priori" accepted as right, the model (values of the coefficients, γ and h in the
previous cited Blake form) is adjusted together with the epicentral parameters
(coordinates and epicentral intensity) to fit the data. Often, the computation of
distances between the localities is extremely simplified, and the final solution (let
be x₀ = 13.20, y₀ = 45.50, h = 13.7, I₀ = 9.35, y = 2.7) is maked-up to be similar to
a common earthquake record (I₀ = 9.35 => 9.5 => IX-X). This habit completely
vanishes the possibility to reproduce the intensity map.

Our approach starts from the research of some arbitrary parameters, not
representative of the physical source of the earthquake. Like in instrumental
quake location, Macroseismic Epicenter with respect to Seismic Hazard
Assessment (MESHA) is obtained assuming a model to be true (attenuation of
intensity instead of crustal velocities) and minimizing the residuals:

\[ R_I = \sum_{i=1}^{N} |I_{(obs)} - I_{(cal)}| \]  

(1)

Here the main news introduced in the computer code:
1) only integer values are accepted for intensity: uncertain attributions (e.g. VIII-
IX MCS) are not treated like a separate class of values (8.5) but are put by the
user into the two contiguous classes of intensity with proper weights;
2) computed values are truncated, not rounded to the nearest integer; the
residuals in (1) are evaluated after truncation: so the common way of tracing
isoseismal lines is respected;
3) the location could be performed using only some intensities, for example the
highest ones that are the most important for hazard assessment;
4) a comparison between several models (in term of different attenuation
relation, and of parameters) is done; this practice adjusts the model without
performing inversion;
5) a good computation of distance of points on a sphere is performed.

In particular three attenuation relationships of macroseismic intensity are
considered. The first one follows the Blake form, where h is not a depth but only
a parameter; in addition the possibility of a finite dimension of the epicenter has
been introduced. The second one is the Grandori et al. (1987) relation, where a
finite epicenter is modelled by the parameter D₀. Last, a simple exponential
An example of the results is given. The previous described Soncino earthquake is located using different models and the best solution, obtained with the Grandori relation, is given in Fig. 3. The differences in coordinates between ME (Postpischl, 1985) and MESHA is quite small (see Tab. I), while the epicentral intensity has a significant change. Here only 5 localities of intensity greater than, or equal to, VI MCS are wrongly predicted in the northernmost area. So, the possibility of reproducing the intensity map is obtained, but both the model, and its parameters have to be reported in the earthquake catalogue.

Tab I: Epicentral parameters of Soncino earthquake.

<table>
<thead>
<tr>
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<tr>
<td>x₀</td>
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<td>9.810</td>
</tr>
<tr>
<td>y₀</td>
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<td>45.388</td>
</tr>
<tr>
<td>I₀</td>
<td>VIII</td>
<td>VII</td>
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Figure 3: Soncino earthquake of May 12, 1802: isoseismal lines with Grandori model (parameters in Tab. I), underlined values of intensity are not correctly predicted by the model.

REFERENCES
The correct way to obtain a reliable earthquake catalogue is surely to start from "primary information" as coeval descriptions, for the macroseismic data, and waveforms for the instrumental ones. After a phase of "selection and evaluation" of the dataset obtained it is possible to produce a "unified dataset" from which a new set of focal parameter "catalogue" can be derived. This process, that has as final goal the "hazard assessment", is very expensive in terms of time and human resources.

There are two main reasons to obtain a seismic catalogue by merging "existing catalogues":

- one is the need of evaluating the seismic hazard in short time;
- the other is to produce a preliminary assessment of hazard parameters using the dataset existing at the beginning of the research and repeat the assessment using the final dataset, for checking the importance of the revision.

Following the last reason, a sort of "methodological exploration" in merging catalogues is here described, made because, in Italy there are several catalogues that it is important to merge together, as better as possible, to perform a preliminary assessment of seismic hazard.

The area under study is northeastern Italy with a part of Slovenia and minor parts of Switzerland, Austria, and Croatia. It has been chosen as a "test" zone for the complex interaction occurred among some of the most important earthquake catalogues developed in the Italian area.

One of the main reasons for which this study has been performed is the very different evolution of the two main earthquake catalogues available for this area: the italian earthquake catalogue (namely PFG; Postpischl, 1985); and the Easten Alps earthquake catalogue (namely ALPOR; OGS, 1987) developed by Osservatorio Geofisico Sperimentale (OGS).

For a better understanding of the (reciprocal) connection between the two cited catalogues a tentative flow diagram of temporal evolution of them has been drawn. On the left side the evolution of PFG catalogue is given starting from its source catalogues (the rounded frame indicates the catalogues of the entire Italy area). On the right side the evolution of ALPOR catalogue with an indication of contribution received by various border nation catalogues (Switzerland, Austria, Slovenia, and Croatia) is reported. In particular the latest version of ALPOR includes a part of a recent version of the Slovenia and Croatia catalogues, and has also an area extension (from 15°E to 16°E) towards east.

The interaction of the two families of catalogues is pointed out by dashed lines. It can be seen that the GEOTECNECO catalogue is a "father" of both PFG and ALPOR catalogues, and that ALPOR version "0" contributed to the development of PFG catalogue, which has been partly used for the preparation of the final ALPOR version "7".
The mutual interaction of the two catalogues and the better updating of the ALPOR catalogue (mainly due to the recent merging of Slovenian and Croatian catalogues) lead to the need of an accurate merging and checking of ALPOR and PFG catalogues. For this reason a sub-catalogue, having the same geographical extension of ALPOR, has been extracted from the bigger PFG catalogue. This sub-catalogue has 4902 records and has been named PFG1045 (10°E and 45°N are the same left coordinates of the ALPOR catalogue area).

The sensible difference in term of record number between the two catalogues (4902 in the PFG1045, 7311 in the ALPOR) can be quite completely attributed to the greater area extension (towards east) of the ALPOR catalogue.

Analysing the bibliography distribution of the two catalogues it is possible to see that there are about 900 events in ALPOR that came from PFG catalogue and about 1700 events in PFG1045 that came from previous versions of ALPOR. For these reasons it is evident that, from a seismological point of view, a well reasoned merge of the two catalogues is a necessary step.

There are some considerations to do, before describing the merging catalogues process, about the history of an earthquake catalogue record.

The originary sources are, in the historical period coeval documents that have been interpreted and collected in intermediate documents before being evaluated by seismologists and compressed in a computer file. The work, that results often very difficult to do,
is, to go back to the original sources, starting from contemporary earthquake catalogues.

An analysis of the complex interaction of the bibliographical items of the two catalogues and the need to preserve a link to the original sources led to the conclusion that it was quite impossible to design an efficient computer code for the automatic catalogue merging.

The previous experience in merging the Yugoslav catalogues to ALPOR (Rebez, 1987) has indicated that it is possible to obtain a satisfactory automatic or semi-automatic merging of catalogues only if the hierarchy of the two catalogues is well known. In practice a computer code for merging catalogues can be designed when it is exactly known that one of them is surely more revised, or has a different area extension than the other.

In the present case a doubt existed that both PFG1045 and ALPOR catalogues had a great quantity of good information (but also of mistakes) in them.

For all these reasons, and also to develop a kind of "methodological exploration" it has been decided to perform a "manual" analysis and cleaning of data. The ALPOR catalogue has been translated into the PFG format and the records of ALPOR has been numbered in a different manner to make them easily recognisable.

The two catalogues have been joined and sorted out according to their temporal order. In the merged catalogue a new field (in column 1) has been inserted for marking earthquake records:
- symbol "=" if the event has been evaluated good or preferable;
- symbol "d" if record is doubtful, uncertain, difficult to interpret, or needs a historical research.

In case of doubleness (record ALPOR and record PFG) the event with a "blank" in column 1 is classified as discardable.

This methodology, based on the multiplicity of records, has the advantage to leave always a track of the operator choice (with the possibility of easily correcting mistakes) and also to extract the sub-set of reliable data (for seismological maps or computer elaboration).

The point of view of this kind of analysis is of producing a dynamic "working file" (not fixed, on which it is possible to think over and go back on decisions) that let one to restore possible wrong choices.

This "reasoned cleaning of ALP-PFG catalogue has been executed for the whole "macroseismic" period (from 238 A.D. to 1899, 3933 records) which is the period more clearly influenced by subjective and often wrong interpretation of historical documents. The studied time interval comprises several different historical periods, with various and sometimes typical historical-seismological problems.

After this experience some general rules can be indicated for the operation of "manual cleaning" of catalogue:
- It is necessary that the operator had a good knowledge of the evolution and the bibliographical reliability of at least one of the catalogues.
- In case of double records it is preferable to maintain a link as clear as possible to the original historical sources (original bibliography).
- It is important to carefully evaluate an evident territorial competence of a catalogue.
• Records that came from recent studies or from other revised catalogues are preferable.
• If exist doubt that the record indicates simple "effects" of far earthquakes (in this area it is a typical situation for the town of Venice) it is better to mark with a "d" the record and maintain the record until it will be possible to study the earthquake in detail.

The first part of catalogue ALP-PFG (238 A.D.-1899) included 3933 events of which 1972 deriving from ALPOR and 1961 from PFG. At the end of the merging phase a catalogue of 2430 records (2070 "=" and 360 "d") has been obtained. The major effect of catalogue merging is the consistent reduction of defectives characterising both catalogues.

In the Fig. 2 the distribution of bibliographical items of ALPOR (characters) and of PFG (numbers) before (in black) and after the merging is indicated. It is noteworthy the decrease of the item GE (record from PFG in ALPOR) from 600 to about 20. In this case the records have been substituted by the original records with native bibliography and more informative format.

On the right side it is possible to see a clear decrease of 504 and 505 items (old versions of ALPOR catalogue): about 500 events. It is important to say that many of these records refer to earthquakes coming from an old version of the Slovenia catalogue, that have been substituted with new records recently updated by Slovene authors. Another remarkable decrease refers to the 501 item (Iaccarino and Molin, 1978., catalogue) that has been replaced by various bibliographical items of ALPOR.

REFERENCES
A DIFFERENT INTENSITY RECORDING FOR NOT LOSING INFORMATION: AN APPLICATION TO THE COMPLETENESS ANALYSIS OF THE CATALOGUE

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Considering the ordinal and qualitative nature of the seismic intensity, we deal with the problem of recording its value, keeping the information obtained from the experts as much as possible. It is well known that the intensity degree synthesizes a great deal of information of different nature (building damages, ground splits, natural phenomena, human reactions, etc.); hence the classical way of recording by only one integer value implies that much information collected by seismologists is lost. It is necessary to find a way of formalizing the uncertainty in the intensity assessment; a way of treating: a) the cases in which we are in doubt between contiguous or even not contiguous classes or b) situations, to be referred to the same degree, which we think different so that we could even say which is the most severe.

For doing that, we propose to record the intensity not by only one value but by a vector whose components indicate the belonging degree to the various classes of the scale. The complexity of the seismic phenomena and the large number of physical quantities involved lead one to study them from a probabilistic point of view. We consider the intensity as a random vector $X$ whose distribution is a mixture of multinomial distributions; according to this approach we study the probability of exceeding a certain threshold of intensity $i_c$ in a time interval and we give Bayesian estimates of the parameters.

Moreover, as application of this new approach, we deal with a remarkable problem: the completeness of a catalogue and we present a statistical test in order to evaluate the completeness degree in objective way. Since a catalogue with intensity records expressed by vectors does not exist yet, we simulate it by transforming the catalogue of an Italian zone, Sannio-Matese, according to different criteria; that is, with each event whose intensity is recorded as $i = i$ (or $i\pm 0.5$) in the original catalogue, we associate a vector $(0,\ldots,0,x_{i-1},x_i,x_{i+1},0,\ldots,0)$, where $(x_{i-1},x_i,x_{i+1})$ is a realization of a trinomial distribution.

The results obtained show (Figure 1) how, going from a "non-uncertain" version of the catalogue to a "more uncertain" one, the completeness tends to increase; this fact strengthens the opinion that failing to take into account the element of uncertainty in the available data
could lead to less reliable results and that the information lost, concentrating the intensity assessment in one value, is not to be neglected. The new way of recording is more flexible and it allows one to utilize more fully the information supplied by the experts. Of course, in order to evaluate accurately its validity, further one of the proposed probabilistic model, we need a revision of a considerable number of past occurrences so as to be able to express their intensity in diffuse way.

REFERENCES


Southeastern Sicily is one of the areas of the Italian territory with a higher seismic hazard (Purcaru and Berckemer, 1982; Mulargia et al., 1985). Furthermore, the present high density of both inhabited areas and petrochemical industries allows a very high level for the seismic risk, too.

A long series of ruinous earthquakes occurred in the region, as those of 1169 (intensity Io=XI MCS), 1542 (Io=IX MCS) and 1693 (Io=XI MCS). This one totally destroyed the cities of Catania and Siracusa, killing more than 60,000 people. Nevertheless, since 1850 all significant seismic activity ceased, suggesting a large quiescence pattern (Mulargia et al., 1985). From the analysis of both seismic and precision levelling data it was suggested (Mulargia et al., 1985; 1991) the near occurrence of a moderate-to-large (M=6.0) earthquake.

In fact, on December 1990, at 00.24 (U.T.), the Eastern Sicily was shocked by a 5.5 magnitude (Mb, NEIC) earthquake. The location of the event was few kilometers offshore Augusta at a depth of around 10 km. Very few aftershocks occurred, the strongest one (Mb=3.8) on December 16, at 13.50 (U.T.). The focal solution of the main shock shows a strike-slip mechanism with fault planes North-South and East-West trending; (for further information on the event, see Boschi and Basili, 1991).

In despite of the relative low magnitude of the event, damages were quite high. Seventeen people were killed and some thousand ones were homeless, on a large area (around 2500 square kilometers) including the towns of Catania and Siracusa. The three towns of Augusta, Lentini and Carlentini were the most seriously damaged, and the VIII degree of the MCS scale was here found. Damages induced in the small town of Carlentini have been studied, focusing the attention on the problems linked to site effects and response of soil. Maps of both building typology and induced damages were made, and compared with the geological one.

Carlentini is west located around 25 km from the epicenter of the December 13, 1990 earthquake. It is built in the upper smoothed part of a hill (200 m a.s.l.) having slopes ranging 10-60%.

A detailed (scale 1:2,000) field geology investigation was made allowing to generally find a calcarenite sub-stratum (5-70 meters thick), emplaced on Tortonian volcanoclastic layers, with thickness from 80 to 150 meters. A Quaternary NE-SW trending fault system
affects the Southern part of the village. Data from 25 down-holes, 30 meters deep each, supported the results of the geological mapping. Furthermore, using 15 geophones into every hole, data on velocities of both P and S waves, for each 2 meters layer, were available.

Fig.1: Map of Carlentini showing the distribution of both "class A" (blank areas) and "class B" (dashed areas) building types. See text for details.
A map of building typology was made, based on the following classification. In "class A" were 2-3 floors old houses, builded with blocks of stone, only partially with reinforced concrete, and generally without (or with very shallow) foundations. In "class B" the more recent buildings in reinforced concrete, with 2-3 floors and consolidated foundations.

Fig. 2: Map of Carlentini showing the distribution of the three classes of damages; "class 1" (blank areas), "class 2" (dashed areas) and "class 3" (dotted areas). See text for details.
The areal distribution of buildings consisted in a great part of "class B" edifices concentrated in the central part of the village; whereas peripheral areas were mostly characterized by "class A" edifices (Fig.1).

A map of damages induced by the shock was finally made. They have been divided into three classes. The first one ("class 1") included the slight damages (as small fissures in plasters); the "class 2" more serious damages (as many fissures in the walls and falls of rubble). "Class 3" of damages was referred to collapses of walls or to the total collapses of the building.

Both the Northern and the Southern peripheral parts of Carlentini showed damages of "class 2", while very few damages occurred in the down-town (Fig.2).

CONCLUSIONS

The distribution of damages induced in the village of Carlentini by the December 13, 1990 earthquake has been critically analyzed, in terms of both building typology and geological (and morphostructural) features of the area.

No evidence of site effects, neither due to the local geological condition, or to the soil-foundations coupling were found.

Results are that the highest damages occurred in both Northern and Southern peripheral zones of the village, where poor quality edifices ("class A") were builded. Furthermore, the high slopes (greater than 50%) characterizing those areas could have contribute to a local amplification of the ground shake.

REFERENCES


SOIL LIQUEFACTION DURING THE ANDALUSIAN EARTHQUAKE (25/12/1884)

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ABSTRACT

On December 25, 1884, an earthquake of epicentral intensity of 7 caused great damage in a large area in the provinces of Granada and Málaga, in the south of Spain. The reports of the Spanish, Italian, and French Commissions that studied the earthquake described ground phenomena in seven different sites, which can be identified as soil liquefaction.

By means of dynamic penetration tests carried out in the above sites, the corresponding soil profiles based on SPT data and water table depth were established, and the occurrence of liquefaction was proved in five out of seven of these sites. Also, the intensities at such locations and the magnitude of the earthquake were estimated.

From the geotechnical data and the cyclic stress ratio induced by the earthquake, liquefaction conditions were confirmed in all five sites which presumably liquefied. Then, possible values of the minimum ground surface accelerations necessary for the onset of liquefaction at each location were derived. At this stage, the results obtained were completed with similar data reported in six liquefaction case studies from Japan and U.S.A., from which design charts relating soil acceleration with normalized SPT values for different intensity levels were drawn.

Finally, by using standard attenuation curves, the above data were translated into epicentral distances, and a good fitness with the known epicentral area was obtained. Therefore, the method seems to be reliable considering both the historical and earthquake engineering approach.

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INTRODUCTION

Recent big earthquakes in Southeastern Europe and in the world demonstrated that there are two factors which can have a significant effect on the nature of recorded strong ground motions:

- Characteristics of strong ground motions for single and multi-shock events, in near and far field conditions, and their modification due to local site conditions.

It was demonstrated by recorded strong ground motions of the multi-shock of April 15, 1979 (Ms = 6.9, Io = IX), with the epicenter between Montenegro and Albania, in near and far field conditions, and of the single shock of January 9, 1988 (Ms = 5.1, Io = VII), with the epicenter in Tirana, in near field conditions. Based on ten years experience on seismic microzoning studies carried out in Albania, it was shown that the mapping of strong ground motions parameters and their effects on free surface are of primary importance, for seismic microzonation.

1. CHARACTERISTICS OF RECORDED STRONG GROUND MOTIONS

Our strong motion network started to work from 1986. Up to now only the earthquake of January 9, 1988 (Ms = 5.1, Io = VI-VII) was recorded as a single shock event, on baserocks of Tirana hills, just above the active fault.

It was a typical near field record (TIR) (Kociaj & Pitarka, 1990).

Mainshock of April 15, 1979 earthquake (Ms = 6.9, Io = IX), recorded on baserocks of Ulcinj (ALB), as a multi-shock event (Petrovski & Paskalov, 1980) was very similar to well-known record of El Centro (ELC). Both of them (ALB, ELC) were considered as far field records. These strong motion records: TIR (as a single shock) and ALB, ELC (as multi-shock), were used as input functions on baserock conditions, for further data processing.

2. THE MODIFICATION OF STRONG GROUND MOTIONS BY LOCAL SITE CONDITIONS

2.1. Recorded strong ground motions:

Based on publications: on the earthquake of April 15, 1979 (Petrovski & Paskalov, 1980) it was observed that the parameters of recorded strong ground motions were influenced a lot by the local site conditions.
2.1. Computed strong ground motions

For the computation of dynamic response of soil cross sections (1D case), wave propagation method (FREELEY program) (Mlutilinovic, 1982 & Skateriqi, 1990) and Thomson-Haskell method (NOLI5K program) (Boucheva & Kostov, 1987 and Pitarka, 1987) were used. For soil profiles (2D case), finite element method was used (Kocija & Pitarka, 1987). The modification of strong ground motions by local site conditions became very clear for 1D and 2D cases. It was observed that using input motions Alt or ELC, Sa spectra of analytical accelerograms at free surface can be classified in four groups, for 1D case (Fig. 2a), and in three groups for 2D case (Fig. 2b), depending mainly on the geometry and properties of soil cross sections or soil profiles. Using near field record (TIR) as input function, it was observed very small influence of soil cross sections or profiles on Sa spectra of computed accelerograms at free surface (Fig. 3). As many inhabited centers in Albania are situated on seismic active faults it is very important to take that into account during seismic microzoning. On the other hand, influenced by distant earthquake feel as well, which means that far field input motions should be considered as well.

3. THE PARAMETERS TO BE USED FOR SEISMIC MICROZONING

As there are very few data, on strong ground motions, recorded in different soil conditions, computed accelerograms at surface level and their spectra were used as a supplementary tool. That means we have to use other approaches for this purpose. Three approaches have been used for seismic microzoning, which means that we have to deal with different strong motion parameters based on:

a. The consequences of past earthquakes.
Based on isoseismal maps of historical earthquakes and on the earthquake model with a layer above the half space, with uniform absorption, it was shown that corrected epicentral intensities values can be considered as intensities of strong ground shakings at surface (Iok) and baserock levels (Ibr) (Kocija, 1989). Based on Iok and Ibr values, it was concluded that only they may be converted to physical values.
b. The instrumental measurements of elastic properties of soils:
- Spectral periods $T_s$
- Seismic intensities increments ($dI_m$).
- Soil categories according to $A_{max} = f(T_s)$ nomograms.

Based on seismic shear velocities measurements ($V_s$) and underground water level, were determined:
- Seismic intensities increments ($dI_r$).

- The analytical data of computed accelerograms at free surface:
  - Maximal amplitudes for following parameters were used:
    - Acceleration ($a$),
    - Response acceleration ($S_a$), velocity ($S_v$) spectra, for 5% damping,
    - Spectrum Intensity values ($S_I$),
    - Fourier spectra ($F_S$),
    - Transfer functions ($T_F$).

For 1D case, all $S_a$ spectra can be classified into three groups for majority of soils of our inhabited centers (Kociąj, 1986). Comparison of 1D and 2D cases outputs for $S_a$ spectra leads to better understanding of soil categories, using analytical data. Based on these maximal amplitudes the seismic intensities increments can be determined as ratios of these parameters in moving ($v$) and reference point ($o$), chosen usually on bedrocks at free surface as follows:

$$
\frac{dT}{T^{(v)}}; \frac{dI}{I^{(o)}}; \frac{S_v^{(v)}}{S_a} \quad \text{and} \quad \frac{S_v^{(o)}}{S_a} (1)
$$

The conversion of these ratios into seismic intensity degrees was made, taking into account that increase of these ratios two times, corresponds to the increase of seismic intensity one degree.

4. CORRELATION BETWEEN STRONG GROUND MOTION PARAMETERS

So called "complex intensity" ($I_k$) was determined as:

$$
I_k = I_{ok} + dI_k = I_{ok} + (dI_m + dI_r + dI_g)/3 (2)
$$

where: $dI_m$, $dI_r$, are intensities increments according to instrumental measurements and engineering geology data ($dI_g$). Comparing $dI_k$ values with those of $dI_T$ and $dI_S$ (Fig. 4a), it can be observed a linear correlation between them, which shows that last parameters can be used for strong ground motions evaluation through seismic intensity $dI_T$, $dI_S$, $dI_a$, $dI_v$ and $I_k$ scale. Concerning $dI_a$ and $dI_s$ values, a decrease of these values can be observed with the increase of $dI_k$ for 1D case (Fig. 4b). For 2D case the picture is different: with the increase of $I_k$, $S_a$ and $S_v$ peak.
values (i.e., $l_a$ and $l_s$) are increasing (fig. 5), which is in the favour of 2D modelling.

CONCLUSIONS

Strong ground motion parameters used for seismic microzonation purposes, is recommended to be based on three approaches:
- On seismic intensities as the intensities of strong ground shaking based on isoseismic / on maps of past earthquakes,
- On parameters based on instrumental measurements for elastic behaviour of soils,
- On recorded or analytical (1D or 2D modelling) strong motion data parameters. For analytical parameters 2D modelling is preferred.

REFERENCE:

Abstract/Conclusions

Various stochastic models of strong ground accelerations have been proposed. However only very few analyses have been devoted to the rotational motion of the ground. Since only translation components are being measured, an algorithm for the rotational motion should be derived.

Consider a system of principal axes located on the surface of the site, with horizontal axis \( x \) directed toward epicenter and vertical axis \( z \) directed downward. In addition to three translation components \( u,v,w \), corresponding to axes \( x,y,z \) respectively, two rotational components (rocking \( \psi \) about \( y \) axis and torsion \( \phi \) about \( z \) axis) can be important for engineers. The rocking component can, for example, significantly contribute in the response of slender towers (e.g. chimneys, TV towers etc.).

Two methods of calculating the rotational component can be mentioned:

1) direct differentiation of the surface field of seismic motion,

\[ \psi = \frac{\partial}{\partial x} w(x,y), \]

2) decomposition of total motion onto respective wave contributions, differentiation and the summation of the motion back again.

This paper presents a brief review of both methods. As a result in the first method one obtains the rocking spectrum (spectral density) as a function of vertical acceleration point spectrum, the coherency of the random field \( w(x,y) \) and apparent wave velocity. In the second case the rocking spectrum is formulated in terms of time derivative of translation components, incidence angles of body P and S waves and the velocities of body and Rayleigh waves.
MODELLING OF EARTHQUAKES BY EXPLOSIONS OR ARTIFICIAL RECORDS

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INTRODUCTION

The number of obtained earthquake records is still not sufficient for the purposes of the expanding theory and practice of earthquake engineering. Because of specific local geological and tectonic conditions, it is not satisfactory to apply records obtained in one region automatically into another one. There exist many regions where the obtained data about the type and character of seismic motions are rather poor. Therefore, besides time records of natural earthquakes, the seismic records of explosions or artificial seismic records can be used as loading functions for calculation process or for laboratory and field experiments. The paper describes the properties of seismic records from explosions of large intensity in comparison with those of natural or synthetic seismic accelerograms.

NATURAL SEISMIC RECORDS

In a non-homogeneous medium the surface and body waves have different propagation velocities. Besides this, in a layered medium the surface waves propagate in a dispersive way, their velocities depending on the material properties of the medium, the frequency of the wave motion, and the geometric configuration of the layers. These phenomena together with the magnitude of earthquake and the distance from origin create the properties of the seismic vibration at the place of interest.

Few natural seismic accelerograms were chosen to show their seismic response spectra distribution in comparison with general standard response spectra as they are recommended in different codes. Uniform Building Code of USA [6], Eurocode 8 [2] and CS Standard 73 0036 [3] are included into comparison. Uniform Building Code and Eurocode 8 in the base approach propose nearly the same spectra with the accent on the low frequency range up to 20 Hz (Fig. 1). The CS Standard shifts the frequency band of interest up to 30 Hz (Fig. 2). The ratios of extreme response acceleration to the input seismic acceleration $S_a/a_0$ for our chosen accelerograms are in Table 1.

Table 1. Extreme values of accelerations and response ratios for damping $\zeta_d = 0.06$

<table>
<thead>
<tr>
<th>Accelerogram</th>
<th>$\ddot{x}_{max}$</th>
<th>$\ddot{y}_{max}$</th>
<th>$\ddot{z}_{max}$</th>
<th>$S_x/\dddot{x}$</th>
<th>$f_a$</th>
<th>$S_y/\dddot{y}$</th>
<th>$f_y$</th>
<th>$S_z/\dddot{z}$</th>
<th>$f_z$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Niš 1977</td>
<td>4.26</td>
<td>3.87</td>
<td>2.04</td>
<td>3.69</td>
<td>5.0</td>
<td>3.62</td>
<td>5.4</td>
<td>2.89</td>
<td>10.8</td>
</tr>
<tr>
<td>Dayhook 1978</td>
<td>3.86</td>
<td>3.76</td>
<td>1.87</td>
<td>3.60</td>
<td>6.2</td>
<td>2.95</td>
<td>2.6</td>
<td>3.58</td>
<td>13.2</td>
</tr>
<tr>
<td>Tabas 1978</td>
<td>8.66</td>
<td>9.63</td>
<td>7.13</td>
<td>2.75</td>
<td>4.8</td>
<td>3.74</td>
<td>5.0</td>
<td>3.07</td>
<td>11.4</td>
</tr>
</tbody>
</table>

The respective values of answering accelerations are used in Figs. 1, 2. Taking only $a_{max}$ for comparison means more realistic
relations of real spectra to standard spectra. Higher frequency components are remarkable mostly in vertical seismic acceleration records.

Fig. 1. Seismic acceleration response spectra in comparison with UBC [6] and EC8 [2]

Fig. 2. Seismic acceleration response spectra in comparison with Revised ČSN 730036 [1]

SYNTHETIC SEISMIC ACCELEROGRAMS

When simulating synthetic seismic accelerograms we must take into consideration all available informations about the seismic source and the peculiarities of the followed place. Group velocities which are representatives of the chosen place, may be evaluated either from the data of previous records using observational techniques of seismology or by theoretical calculations [3]. The obtained synthetic seismic accelerograms can have different spectral properties, as we can see in Fig. 3 a,b. This is influenced by input data when creating synthetic accelerograms, but essential features of them are in limitation of frequency content to some narrow frequency band until the high frequency Fourier spectrum is prescribed for the simulation. We must admit that both natural and synthetic seismic accelero-grams give very narrow displacement seismic response spectra with limitation to low frequencies.
SEISMIC EFFECT OF UNDERGROUND EXPLOSIONS

Another approach how to obtain the seismic loading functions and to follow the seismic response is the utilization of the soil vibration records from large explosions. Two case studies from our experimental research are presented here. In the first case the vibration of a massive rock surface has been followed in Central Slovakia near Mochovce. The large explosion charge $m = 22173 \text{ kg}$ was divided into 18 particular charges with time phase shift of $0.024 \text{ s}$. The phase shift of the front wave answers the wave velocity $3600 \text{ m/s}$. Prevailing frequency about $8 \text{ Hz}$ is remarkable from the obtained records and their spectra. The nearest point D was placed in a distance of $160 \text{ m}$ from the place of explosion. In Fig. 4 a,b we can see the seismic response spectra of acceleration $S_a$ and displacement $S_d$. The used time step $\Delta t = 0.001 \text{ s}$.

The second case concerns measurements which were realized under field conditions in Central Asia in Uzbekistan [7]. The concentrated charge varied from 2520 kg up to 6540 kg. The distance of measured points from the place of explosion was changing from 100 m to 360 m. The subsoil consists mostly from silty clay sediments. Typical frequency distribution in seismic response spectra of records on soil surface is in Fig. 6 a,b for acceleration $S_a$, displacement $S_d$ and pseudo-displacement $S_{pd}$. The time step in sampling was $\Delta t = 0.001 \text{ s}$.

Either the soil conditions in two presented cases are different, we can observe the larger effect of higher frequencies in acceleration response spectra. It appears consequently in
placement response spectra where the range of extreme response is wider than those obtained from the response spectra analysis of natural and synthetic seismic accelerograms. The large explosion measurements allow to follow simultaneously the seismic response of structures and models at large seismic loading when the effect reach the non-linear region. When following e.g. the seismic response of buried underground structure we can estimate the soil-structure interaction parameters and natural characteristics of underground structure [5].

CONCLUSION

The presented study shows that synthetic seismic accelerograms and the records of vibration from large explosions can be used as helping loading functions for seismic response analysis of structures similarly as the natural earthquake records. However, the user must be acquainted with the local geological and seismological conditions of the building site and their effects on loading functions properties. The shallow shape of standard seismic spectra can be accepted only with the account of expected non-linear behaviour in the seismic response of structure.

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SEISMOGEOLOGICAL ACCEPTANCE CRITERIA FOR RADIOACTIVE WASTE DISPOSALS

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ABSTRACT and CONCLUSION

An integrated safety analysis of the final underground disposal for highly radioactive wastes has been realized in the Czech Republic in accordance with the published international recommendations and national law aspects. Qualitative geological acceptance criteria and quantitative seismological acceptance criteria for radioactive waste disposals were developed. The background material for the initiation of site selection and for its earthquake hazard assessment was discussed. The recent movements of the Earth surface as well as the other mechanical properties of geological media and hydrological conditions had to be taken into account.

The formulated qualitative geological and quantitative seismological acceptance criteria were applied to the territory of the Bohemian Massif. All geological bodies situated in the territory under study were classified with respect to the above mentioned criteria so that the unsuitable, prospective and conditionally suitable regions could be defined. Attention was given to regions in which a lower movement activity of the Earth's surface, an unintensive circulation of underground (subsurface and juvenile) waters, and other specific mechanical properties of geological media are observed.

This first complex study was finished towards the end of 1991. Because of the geological conditions of the Bohemian Massif, the most prospective media for radioactive waste disposals seem to be granitic bedrocks, catazonal metamorphites or any clay formations. It is expected that the detailed analysis of each prospective locality for building underground final radioactive waste disposals will be accomplished within the next step of the methodological process of their construction. It is evident that an action like this has to be underpinned by the relevant legislation, in which individual steps of the procedure are justified and in a sense codified.

The paper will be submitted for publication on a special issue of journal NATURAL HAZARDS
INTRODUCTION

It is the rule that there are always found some persons or a group of persons who are waiting for the first stronger blast-firing to blame damages of houses caused by deformations in their foundation surface or by constructive defects, in the suitable way, for the organization which performs explosive works.

Also, it is truth that many blasters rely on their good existing experience and during the time when seismic effects are not be controlled they increase charges on time stage to such extend that damages of buildings occur round the site. In many cases, defects develop because of an abrupt change of geotechnical properties of the blasted rock and unfavourable reflected and refracted waves are developed. Their effects can be interfered and could overcome the critical vibration velocity at which first marks of damages come into being.

CAUSES OF THE FORMATION OF DAMAGES CAUSED BY BLAST-FIRING

It was possible to create the rough classification of damages the basis of experience in the vibration propagation laws and changes of geotechnical properties during dynamic stress (see Fig. 1).

Geotechnical conditions 1:
Superposed sedimentary rocks or made-up grounds or sand/loamy sedimentary rocks which are sensitive to the dynamic liquefaction, the bedrock is formed by rocks of the group R according ČSN (CS Standard) 731001

Geotechnical conditions 2:
Superposed layer is clay or loamy sedimentary rocks and the bedrock is formed by solid rocks

THE DESCRIBING OF CAUSES AND THE CHARACTERIZATION OF FAILURES

The following classification is presented on Fig. 1.

a) DEFORMATION IN THE FOUNDATION SURFACE
a-a Geotechnical conditions 1:
- foundation of the construction on the slope with different geotechnical properties (Object A, fissures 1,2)
- construction partly with cellars under part of building, the part without cellars is founded on compressible sedimentary rocks (Object B, fissures 3)
- additional veranda annex, founded in the depth of the frost penetration (Object C, fissures 5)
- decantation of fine particles by groundwater or due to defect of the gutter (Object A, fissures 1,2)
a-b Geotechnical conditions 2:
- additional changes of soil properties - congelation, drying due to the evapotranspiration by trees or by the boiler room (Object C, fissure 11)
- irregular surplus loading by the next additional building (Object C, fissure 11)

b) CONSTRUCTIONAL DEFECTS
- horizontal strengths created by the accelerating structure of the roof truss (Object D, fissure 6)
- constructional fissure on the interface of materials with different thermal expansivity
- The concrete ceiling is moving due to thermal changes in the flat roof. Concrete or steel piers transport these movements (Object E), the large fissures occur round window lintels (fissures 18). If the roof is not protected against heat absorption from sun rays (black paint) differences of temperature are rising to 20 centigrade degrees and thermal movements cause sound effects in the morning and in the evening.
- the concrete layer scale off (foundations or the whole structure) corrosion of the reinforcement (Object E, fissure 10)
- the hard plaster scale off which is coated over the soft corn (Object F, fissure 12)
- sinking of external walls due to an old age or due to loading by the ceiling (Object I, fissure 15)
- sinking of interior partition walls due to the deflection of the girder (Object J, fissure 4)
- replacement of small windows for larger ones when building is founded bad (Object B, fissure 4)
- disconnection of window frames or door frames due to drying of wood or door strokes (Object J, fissure 21)
- constructional fissures in the heat strained masonry of the chimney (Object J, fissure 23) or in its interface with the partition wall (Object J, fissure 22)
- structure carrier of sand-cement bricks (which do not shrink during 2 years) cause failures in higher floors (Object J, fissure 24)
- plaster notched on the fresh concrete of the ceiling is tearing due to the shrinking of the ceiling (Object J, fissure and the plaster spalling fissure 25)
- placing of the central heating radiator by the old wall under the window causes the unregular expansion and shrinking and fissures under the window occur B,4

c) DYNAMIC EFFECTS

c-a Interior
- horizontal strengths from dynamic effects caused tearing of cavettos (Object a, fissure 18, Object B, fissure 18, Object J, fissure 18) but they can be MISTAKEN with CONSTRUCTIONAL DEFECTS (Object C, E, fissure 18, Object B, fissure 18, Object F, Fissure 18)
- the separation interior partition from walls-carriers (Object J, fissure 19)
- slanting fissure over weakened places of the partition (Object J, fissure 20)
- fracturing of the cavetto (fissure 18) can occur even from acoustic pressure of surface charges blastings

c-b External
- disconnecting of the gable wall from the front (Object J, fissure 17, Object A, fissure 17)
- breaking centers of the vaulting or windows lintels (Object I, fissure 10, Object A, fissure 10)

c-c Due to the dynamic liquefaction of cohesive soil (geotechnical group 1 presented on Fig. 5)
   In the case of the soil with tendency to the liquefaction, the fissures can occur if the following connections come into being:
   If vibration velocity of given object is higher than the critical one and in the same time porosity of the soil with tendency to the liquefaction increasing of fissures developed in the foundation surface can occur during the dynamic strain. It concerns fissures 1, 2, 3 and 5 at objects A, B, C. They must be accompanied by fissures 16, 17 and 18. The enlargement of fissures also occurs due to changes of the pore pressure.

c-d Due to dynamic volume changes (Geotechnical group 2, see Fig. 1)
   Due to long-term affection of vibrations which destroy the flocculated structure (card houses) of overconsolidated soils, the enlargement of fissures can occurs at the object F. Blast-firing can not cause such fissures because of its short-term affecting.
   From this review is clear that many fissures are very similar (14, 26) (1, 3, 11), but their development can be different. Same fissures in cavettos do not have to be always caused by dynamic effects but always by horizontal static or dynamic strengths.
   But, a plaster scale off is typical for dynamic defects. Fissures coming into being due to deformations in the foundation surface are opening upwards, they have the diagonal shape and occur in the places of the stress concentration (door, window, room's corners).

REFERENCES
Alkut Aytun (1986). Outlines of soil dynamics nad soil structure interaction workshop on design of earthquake resistant building, Jemen.
GEOTECHNICAL CONDITION 1

CONSTRUCTIONAL DEFECTS - fissure 16, 17, 18 if vibration velocity higher than the critical one and in the same time velocity fissures developed

DEFORMATION IN THE FOUNDATION

CLAY cohesive soil

DYNAMIC EFFECTS - fissures 16, 17, 18 if vibration velocity higher than the critical one and in the same time velocity fissures developed

GEOTECHNICAL CONDITION 2

NONCOHESIVE SOIL

CONSTRUCTIONAL DEFECTS - fissure 21-26, 15

DEFORMATION IN THE FOUNDATION

SAND noncohesive soil

DYNAMIC EFFECTS - fissure 16-20

CLAY cohesive soil
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