Geodynamics of the Aegean:
a microseismotectonic approach

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Abstract

From 1984 to 1989, we conducted 5 field experiments to study the seismotectonics of the Aegean area. We installed for periods of 6-8 weeks a temporary network of portable stations (from 25 to 85) and recorded the local seismicity. The intermediate depth seismicity confirms that the slab is dipping northward but is not symmetric, it is dipping more shallowly for 200 km beneath the Peloponnese than beneath the Cyclades, no intermediate depth earthquakes are observed beneath Epirus. The intermediate focal mechanisms shows T-axes plunging northward but P-axes trnding the same as the Hellenic trench suggesting that gravity is not the only force acting on the slab. Shallow seismicity is located mostly along the trench and does not define a small number of active faults suggesting that most of the brittle deformation is taking place on multiple preexisting small faults. The Sea of Crete is almost aseismic. Shallow fault plane solutions show a complex pattern. Reverse faulting, with P-axes trending ENE, is observed along the trench, and normal faulting is observed in a more internal position. The T-axes of the normal faulting are trending N-S in Epirus or northern Peloponnese, NW-SE in southern Peloponnese, and E-W in Crete. The non-uniform strain suggests that one mechanism of internal deformation in the Aegean is gravity spreading.

1. Introduction

Although the geodynamics of the Aegea is probably the most complex of the Mediterranean area, this region offers a laboratory for the study of several different tectonic processes ranging from localized subduction of oceanic lithosphere and continental collision to widespread crustal extension. Moreover, the close proximity of the various styles of deformation makes the Aegean region a particularly good laboratory for the study of the interaction of these processes.

The Aegean region is located between the European and the African plates, whose relative motion is convergent in a N-S direction with a velocity of about 1 cm/year. Yet the southern edge of Aegean moves southwestward relative to the African plate with a velocity of 7 to 10 cm/year (McKenzie, 1972, 1978; Le Pichon and Angelier, 1979), while the interior of the region undergoes rapid and large internal deformation.

This rapid motion of the margin of Aegea is responsible for the present-day lithospheric subduction of the African plate beneath Aegea, but the history of this subduction is not well known. The length of the slab, as well as the timing of the beginning of the subduction, are controversial. For instance the maximum depth of the earthquakes is 180 km (Papazachos and Comninakis, 1971), but tomographic studies suggest that the slab could be 600 km long (Spakman et al., 1988). The shape of the subducted slab is also disputed. For Papazachos (1973) it is symmetric with an amphitheatrical shape, and Richter and Strobach (1978) suggested that this conical shape implies stretching along the strike and therefore a possible tear within the slab. For Makropoulos and Burton (1984), however, although the geometry is complex, the slab deepens toward the East. Several kinematical models have been proposed to explain the geodynamics of this area. For McKenzie (1972,1978), on the basis of focal

mechanisms, the Aegean region moves south-westward, relative to Africa with a uniform velocity along the arc, but for the Le Pichon and Angelier (1979), on the basis of lateral variations in the length of the slab, and the trends of the slip vectors, the motion of the boundary between the plates is decribed by a rotation around a nearly pole, located near Keffalinia in western Greece. These 2 interpretations require quite different distributions and rates of deformation within the Aegean region. Paleomagnetic results (Kissel and Laj, 1988) suggest that the Peloponnese, in the western part of the arc, has rotated 25° clockwise since Miocene time, and that western Turkey, located on the east, has rotated 30° counterclockwise. Most of the information about internal deformation comes from tectonic observations gathered troughout Aegea. Mercier *et al.* (1979) and Angelier (1979) suggest that the strain is not uniformly distributed within the Aegean. They observe compression along the external part of the arc, trending mostly perpendicular to the trench, and extension in the internal part. This extension is oriented N-S in northern Aegea or around the Gulf of Corinth but E-W in the southern part of the arc, around Crete or the Dodecanese. Recently GPS measurements conducted around the Peloponnese, in the western part of Aegea (Billiris *et al.*, 1991), confirm this rapid internal deformation.

The seismicity, which can account only for 15% of the total deformation (Jackson and McKenzie, 1988), clearly defines the Hellenic arc from Corfu to Rhodos. Earthquakes deepen toward the center of the arc and are associated with the subducted slab. Shallow seismicity is spread over most of the Aegean. In the northern Aegean Sea it can be related to the North Aegean trough. Within continental Greece we observe clusters of earthquakes as in Epirus, in the Gulf of Corinth, or around Volos, but the seismicity is not related unambiguously to single active faults. Brittle deformation seems to be spread over wide areas.

Most of the reliable fault plane solutions have been determined using polarities recorded in stations from the World-Wide Standardized Seismograph Network, which requires the study of earthquakes of magnitude greater than 5.5 (*e.g.*, McKenzie, 1972, 1978). In a few cases the solu-tions were determined using body wave modelling (*e.g.*, Taymaz *et al.*, 1990, 1991).

Because the regional seismological network installed in Greece is neither very dense (there are only 4 stations on islands in the Aegean Sea) nor very sensitive (due to the microseismic noise of the sea), for earthquakes of magnitude smaller than 5.5 the control on focal depths is not very good, without which it is not possible to compute reliable focal mechanisms.

The installation of a dense local seismological network (with a spacing between stations of about the same order as the depth of the earthquakes) above a region allows one to study very accurately seismicity and fault plane solutions, but such networks are difficult to maintain. Therefore it is necessary to record only for short periods of time, during which, in general, earthquakes only of small magnitude occur. The installation of a dense network of several tens of stations during a short period of time (several weeks) over a specific area allows one to locate, generally with an accuracy better than 10 km, several hundreds of earthquakes of magnitude ranging between 2 and 5, and to compute several tens of focal mechanisms.

The dimension of the fault for an earthquake of magnitude 6 is of the order of 10 kilometers, but that for an earthquake of magnitude 3 is only a few hundred meters. Small earthquakes do not rupture the whole brittle part of the crust, and preexisting faults can be of significant importance in determining where and how such earthquakes occur.

To improve the image of the brittle deformation given by earthquakes within the Aegean, in collaboration with the Aristotelian University of Thessaloniki, the University of Athens and the Observatory of Athens, we have conducted a systematic investigation of the seismicity of this region. From 1984 to 1989 we installed networks of portable seismological stations (from 25 to 85) for a period of 6-7 weeks in different parts of Greece.

2. Data

In 1984 and 1985 we investigated Chalkidiki, in 1986 the Peloponnese and its surrounding area,

in 1988 the Hellenic arc, and in 1989 Epirus (fig. 1). Most of the instruments that we installed were Sprengnether MEQ-800 smoked paper instruments connected to a 1 Hz vertical seismometer. The recorders were run at 60 mm/mn and visited every 2 days for maintainance and to check the clock against an external radio time signal. During each experiment we gathered between 400 and 1000 records and recorded from 350 to 1200 earthquakes. We will describe briefly the procedure that we use to check the accuracy of our locations.

2.1. The V_p/V_s ratio

First, we costruct Wadati plots ($S - P$ arrival time differences *versus* P arrival times) for earthquakes having more than 5 S-wave arrival times. The slope of such a plot gives a value of ($V_p/V_s - 1$), which is independent of the earthquake origin time. With five points on such a plot, however, we observe a large scatter in the values of the slope. We then calculate the mean of the individual values of V_p/V_s. Second, we plot the differences in P arrival times *versus* the dif-
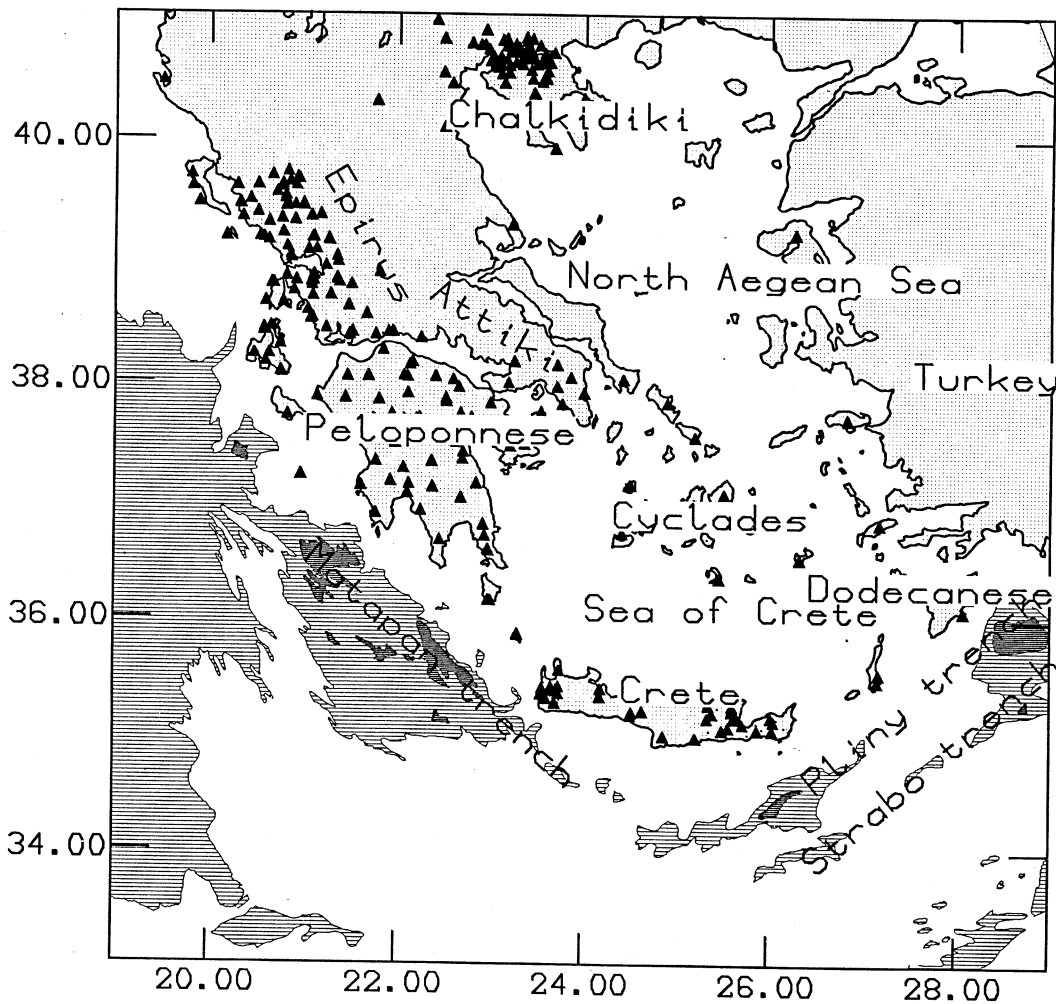

Fig. 1. Map of the temporary stations that were installed from 1984 to 1989 within the Aegean.

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ference in *S* arrival times for different pairs of stations for all the earthquakes. This method provides more data and therefore smoother results than studies of individual earthquakes. We observe spatial variations of the V_p/V_s ratio ranging from 1.74 ± 0.04 in the Mygdonian graben to 1.797 ± 0.003 for the Gulf of Corinth. This variation in the V_p/V_s ratio is not very large and therefore does not introduce important biases in locations when a mean value is used for the whole Aegean.

2.2. *The velocity structure*

We never know the real velocity structure in which the earthquakes have occurred. Although refraction profiles have been shot in some regions of the Aegean (Makris, 1978), only a few profiles are reversed, and thus they do not offer strong constraints. Travel times for P_n phases (Panagiotopoulos and Papazachos, 1985) show some variations in the crustal thickness between the Peloponnese and the Sea of Crete, both overlying an upper mantle with a velocity of about 7.7 km/s, but they provide no constraints on the velocity within the crust. Thus we have tried to find a more refined velocity structure, for each of the regions in which we have worked. Toward this end, and for every region, we select a few events, recorded by more than 8 to 15 stations. We first plot travel times *versus* the epicentral distance, but because there is a trade-off between the location and the travel time, these plots are useful only for deriving a rough mean velocity structure in the crust and upper mantle. Then, we locate the selected events using several hundred velocity structures beginning with a one-layered structure and increasing to 3 layers for the crust. By using reductions in the variance of arrival times as a guide, we select better structures and refine them. We consider the velocity structure that minimizes the residuals as the best (for our set of data).

2.3. *The accuracy of the locations*

The most difficult part of the work is determining how well located the earthquakes are.

There are 2 main causes of mislocation. The first due to uncertainties in the measured arrival times. These derive from errors in locations of the station, clock corrections, readings of arrival times, etc. This noise, assumed to be random, can be estimated from the covariance matrix in the location program, which gives us parameters as RMS (Root Mean Square), ERH and ERZ (respectively, horizontal and vertical uncertainties). In general, these quantities decrease with increasing numbers of arrival times, but occasionally, a small number of arrival times yields a small RMS. To guarantee against, including locations that seem reliable but may not be, we use only these earthquakes located by a minimum number of stations (usually more than 8, including 1 *S* wave) and with estimated uncertainties in location (ERH and ERZ) smaller than a given value (generally 10 or 20 km).

The second source of error, which is more difficult to estimate, arises from an inaccurate velocity structure or V_p/V_s ratio. Because we never know the exact location of an event, it is not possible to conduct realistic tests. Moreover, because of peculiarities in the geometry of the network, there can be large systematic biases arising from random noise. In order to reduce errors arising from an inappropriate velocity structure, we locate all the earthquakes that passed the statistical tests within a set of different but plausible velocity structures. Then, we keep only those whose epicenters and depths do not differ by more than a certain value (also usually 10 to 20 km). We consider these values to be the uncertainties in the locations.

The critical values that we adopt for ERH and ERZ, or for the value of the critical shift in location are functions of the problem that we want to study: for small-scale tectonics we need an accuracy of 5 km, whereas for large-scale geodynamics an accuracy of 20 km may be sufficient. Generally, following these rules, we keep about only half of the earthquakes that we can locate using more than 8 *P* and 1 *S* waves.

2.4. *The focal mechanisms*

Most of the reliable focal mechanisms for earthquakes in the Aegean region have been com-

puted with teleseismic data recorded by long period instruments. The main limitation of these techniques comes from the minimum magnitude needed ($m_b > 5.5$) to record clear signals at teleseismic distances (greater than 30°). Using short-period records to determine P-wave first motions at teleseismic distances is always risky because of the complexity of the signal for ray path travelling through the upper mantle. The same difficulty can arise with a local network if most ray paths are refracted. And in regions where the seismic acticity is moderate, as in Greece, there are not many reliable solutions. To overcome it we tried to use mostly direct waves. To accomplish this and to obtain reliable polarities homogeneously distributed on the focal sphere, it is necessary to have some stations at epicentral distances smaller than the depths of the earthquakes. For shallow seismicity the spacing between stations must be 10 to 15 km or less.

3. Seismicity

In this paper we will summarize results already presented in detail elsewhere (Hatzfeld *et al.*, 1987, 1989, 1990, 1992; Christodoulou, 1986; Pedotti, 1988; Besnard, 1991). We located a total number of 3416 earthquakes from the 5 experiments that we conducted for a total duration of 196 days. The smallest magnitude is –0.2, in the Mygdonian graben, and the largest is 5.2, close to the Matapan trench. This pattern of seismicity is not really representative of the amount of brittle deformation occurring in the whole area because of the inhomogenous distribution of sta-
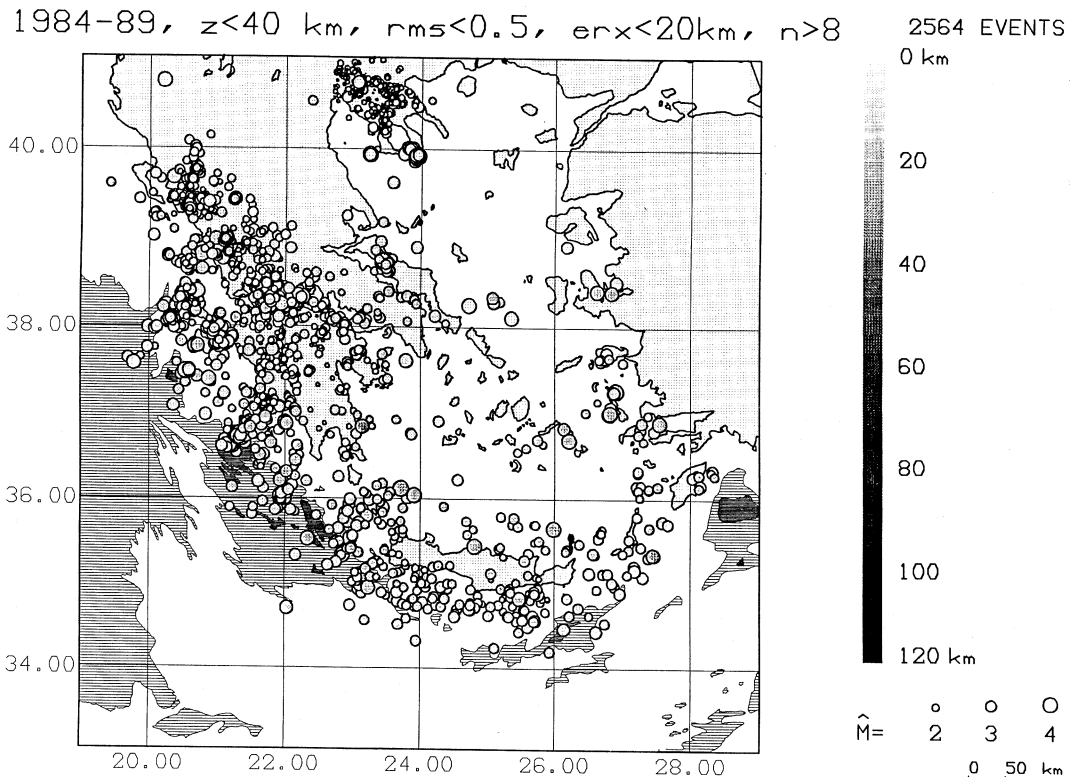

Fig. 2. Seismicity map of the 2564 earthquakes that pass the selection criteria and are thought to be located better than 20 km. Earthquakes do not define single active faults but are spread over most of the Aegean. Most of the earthquakes are located along the Hellenic trench and the Sea of Crete is almost aseismic.

tions that we used to locate the earthquakes during each of the field studies. Applying the criteria described above, we may draw a map of 2564 earthquakes located with uncertainties less than 20 km (fig. 2). We can use this map to discuss the places where we recorded earthquakes, but because the duration of each experiment was short, and because we did not cover the whole Aegean, we cannot draw conclusions concerning the lack of seismicity in some areas.

The examination of fig. 2 shows that seismicity is spread over most of the Aegean area, but most earthquakes are located along the Hellenic trench from Corfu to Rhodos. Along the arc the earthquakes are located North of the trench and very few reliably located earthquakes are located south of it. It is therefore likely that the Hellenic trench (and not the Mediterranean ridge) is the locus of convergence between Africa and Aegea. Earthquakes are spread over wider areas in Epirus or in Peloponnese (located in continental Greece) than they are around the Hellenic trench. This diffuse seismicity could be associated with a difference in the type of convergence observed South and North of the Ionian islands. The convergence rate would be higher South of the Ionian islands where we observe active subduction, the convergence between Apulia and Aegea would be slower, perhaps because of the collision-type tectonics associated with the buoyancy of the continental Apulian lithosphere. The sea of Crete is almost free of earthquakes. This lack of seismic activity is surprising, because this area is supposed to be undergoing important extensional strain beginning in Pliocene time (Le Pichon and Angelier, 1979).

The seismicity reveals no clear evidence of individual faults, but we observe clusters of earthquakes from places to places, such as around the Ionian islands where strong earthquakes of magnitude greater than 6 have occurred in the past 20 years, and near the Gulf of Patras and the Gulf of Corinth where tectonic observations suggest active faulting. We observe also seismic activity around the Kythira strait, located between the Peloponnese and Crete, which has not experienced any strong earthquake since the last century and is thought to be a seismic gap (Ambraseys, 1981). Therefore it is likely that the brittle part of the deformation (as far as small

earthquakes are concerned) is not localized on single major active faults, but instead is distributed on many small faults along the Hellenic arc.

To determine the shape of the subduction zone beneath the Aegean we investigated the locations of subcrustal earthquakes. According to Makris (1978) the thickness of the crust is about 40 km in the Peloponnese and in the Hellenic arc, and considerably less in the Sea of Crete. Of the 101 earthquakes located at depths greater than 40 km, only 85 were assigned uncertainties less than 20 km (fig. 3). Amount these, 16 events occurred deeper than 80 km and therefore can be related without ambiguity to the subducted slab. All earthquakes deeper than 40 km are located in the inner part of the Hellenic arc, beneath the Peloponnese, the Cyclades, and the Dodecanese islands. We observe no subcrustal earthquakes beneath the Ionian islands or the Epirus, and hence we find no indication of rapid active subduction in this area. Very little activity deeper than 80 km is located in the middle of the arc beneath the Cyclades, and most of the seismicity is observed either beneath the Peloponnese or within a very dense cluster located beneath the Dodecanese islands. Moreover, the earthquakes located deeper than 80 km suggest that the slab does not dip at the same angle at both ends of the arc: beneath the Peloponnese we observe a shallow (about 15°) dipping zone toward the NE for 200 km, and farther NE a steeper (about 30°) slab. This wide zone with a very gentle dip is not common for subduction zones and could probably be explained by the important amount of deformation which is observed in the overriding lithosphere above the slab, around the Gulf of Corinth and Attiki. Beneath the Dodecanese, we observe a steeper slab, starting at 100 km only from the Pliny trench and dipping toward the NW. Thus the 100 km isodepth contour is not parallel to the Hellenic trench, and the nature of subduction at the ends of the arc are different from one another. Sections across Epirus, the Peloponnese and the Dodecanese islands (fig. 4) show clear difference in the seismicity.

Crustal seismicity is observed within the whole crust from the surface to the Moho. Although most of the earthquakes are located within the upper part of the crust, there is also activity in the lower crust. Therefore the brittle-ductile transition is not clearly defined, at least by earth-
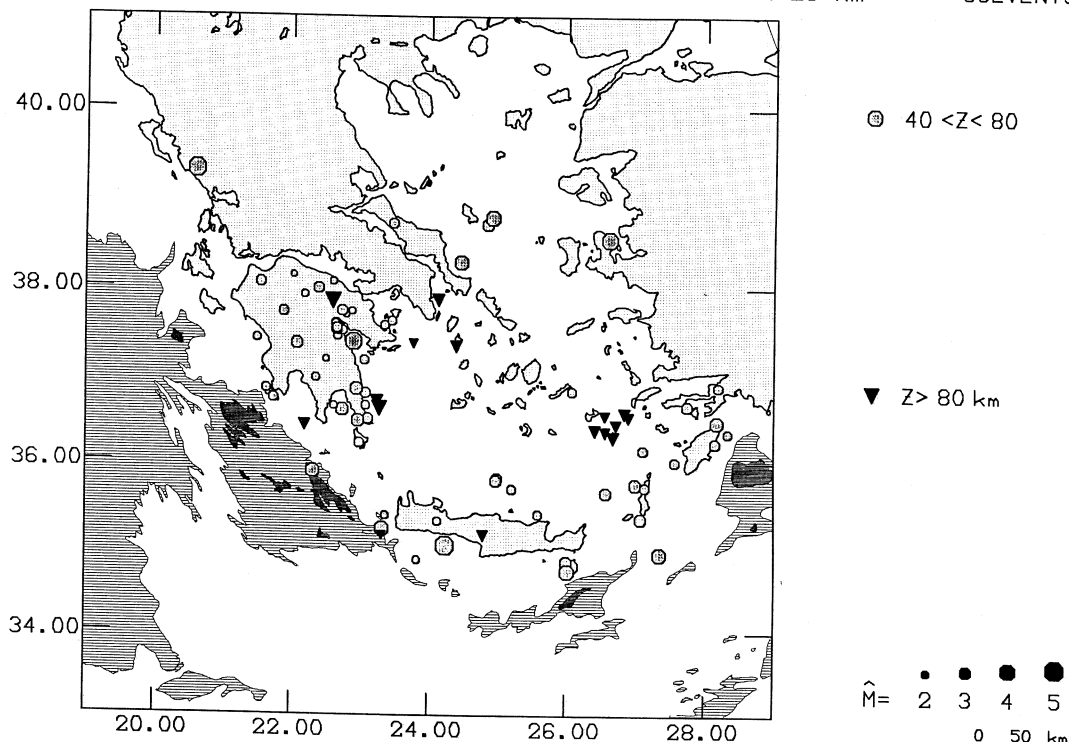
Fig. 3. Seismicity map of the earthquakes located deeper than 40 km. The isodepth contour of the earthquakes located deeper than 100 km is not parallel to the trench.

quakes of this range of magnitude. This is particularly clear for the section across the Peloponnese, which represents the transition between the active subduction zone located further south and the collision- type tectonics located further north.

4. Focal mechanisms

We computed a total of 391 focal mechanisms of earthquakes recorded during the 5 experiments. All of them are based on a minimum of 8 polarities and a maximum azimuthal gap between stations with P-wave first motion of 180°.

4.1. *Depths greater than 70 km*

Figure 5 shows the 14 mechanisms which are

thought to be related to the subducted slab. The 7 solutions located beneath the Dodecanese islands are very similar to one another and are characterized by T-axes plunging steeply NNE and P- axes nearly horizontal and trending N-S to NNE-SSW.

The 5 solutions located beneath the northern Peloponnese and the sea of Crete are also similar to one another, and they show also T-axes dipping steeply, but the P-axes trend NNW. Therefore the T-axes plunge nearly down-dip in the subduction zone, as is common for intermediate depth earthquakes at subduction zones, but the P-axes trend roughly horizontal and parallel to the local strike of the Hellenic trench along the arc. Commonly, P-axes of intermediate depth earthquakes are oriented perpendicular to the downgoing slab, the P-axes parallel to the seismic zone imply that the downgoing slab
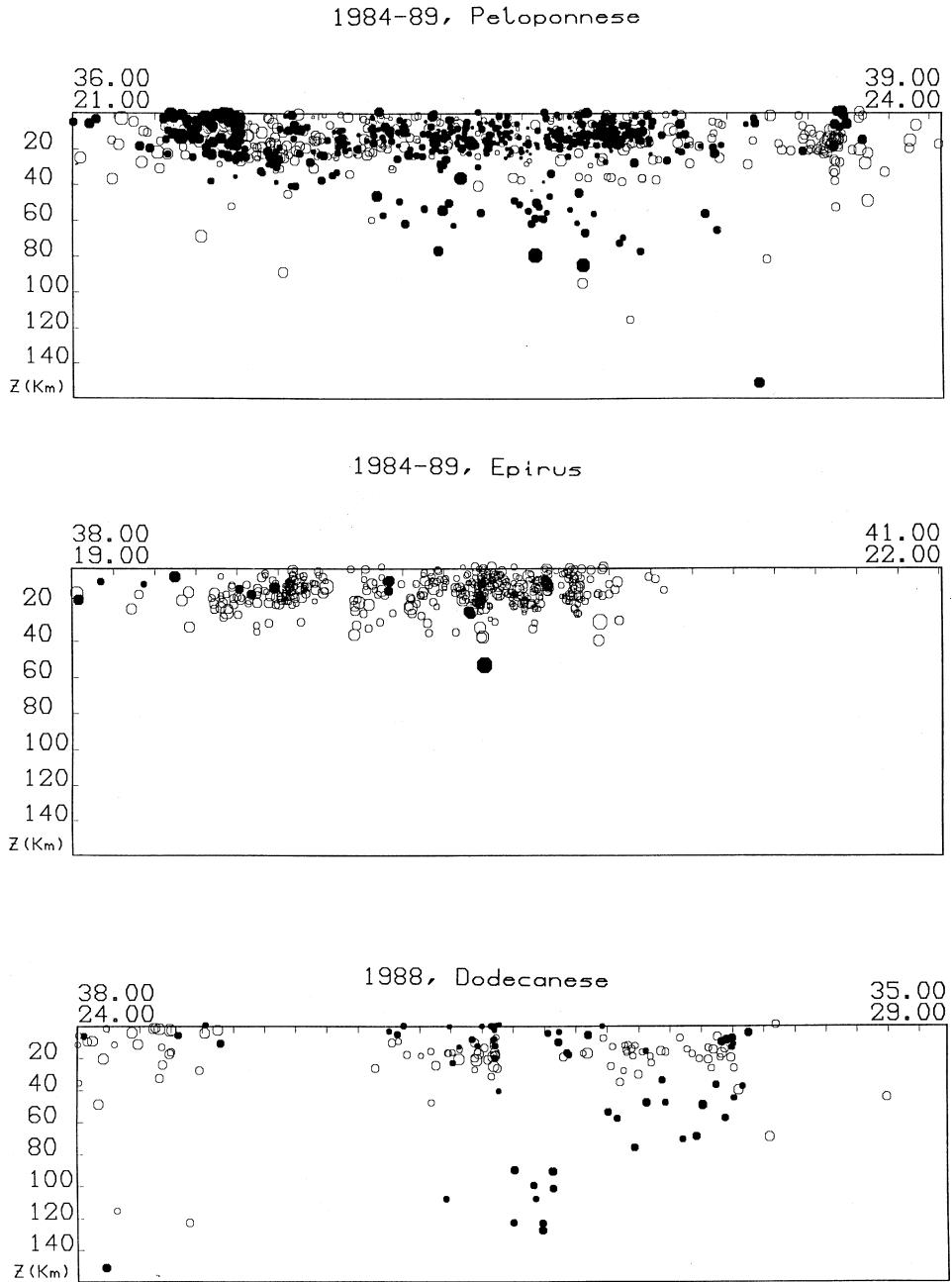
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1984-89, Peloponnese

1984-89, Epirus

1988, Dodecanese

Fig. 4. Cross-sections across the Epirus, the Peloponnese and the Dodecanese islands. There is no intermediate depth seismicity beneath Epirus. There is an important difference of the subduction geometry between the Peloponnese and the Dodecanese islands.
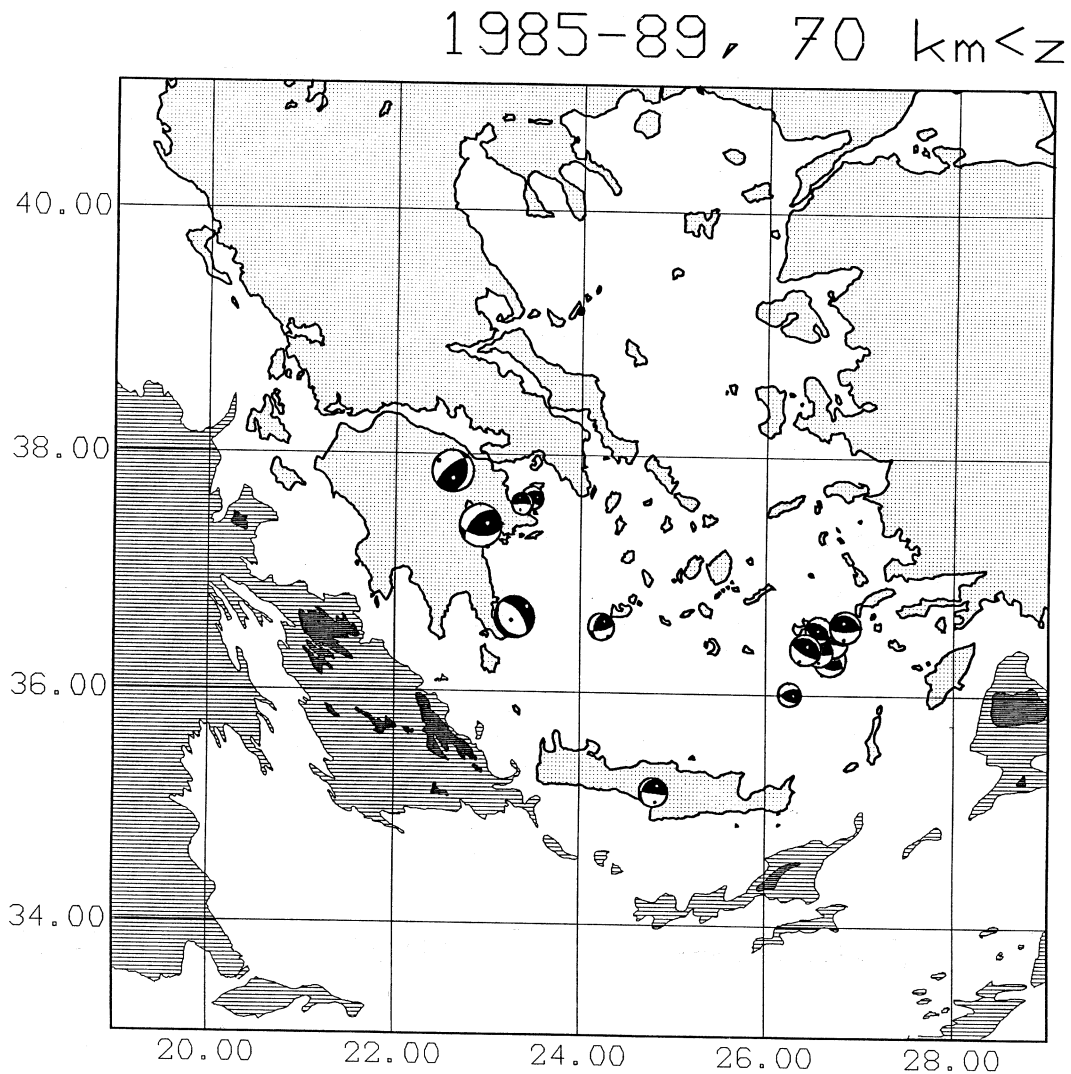
Fig. 5. Focal mechanisms of the earthquakes deeper than 80 km. The T-axes are plunging downdip the slab, but P-axes are trending the same as the Hellenic trench suggesting that the slab is put under stress by more than only gravitational forces.

is put under stress by more than just gravitational forces. The solutions for 2 earthquakes located south of Peloponnese and beneath Crete are more typical of those observed in subduction zones: the T-axes are aligned downdip and the P-axes are perpendicular to the slab. It is likely that these events are also located within the slab.

4.2. Depth shallower than 40 km

Figure 6 shows the complex pattern observed from the Albanian border to Turkey. In Macedonia, northern Aegea, Evvia, Attiki, and beneath the eastern Dodecanese, fault plane solutions are dominated by extension trending roughly N-S. These mechanisms are consistent with those
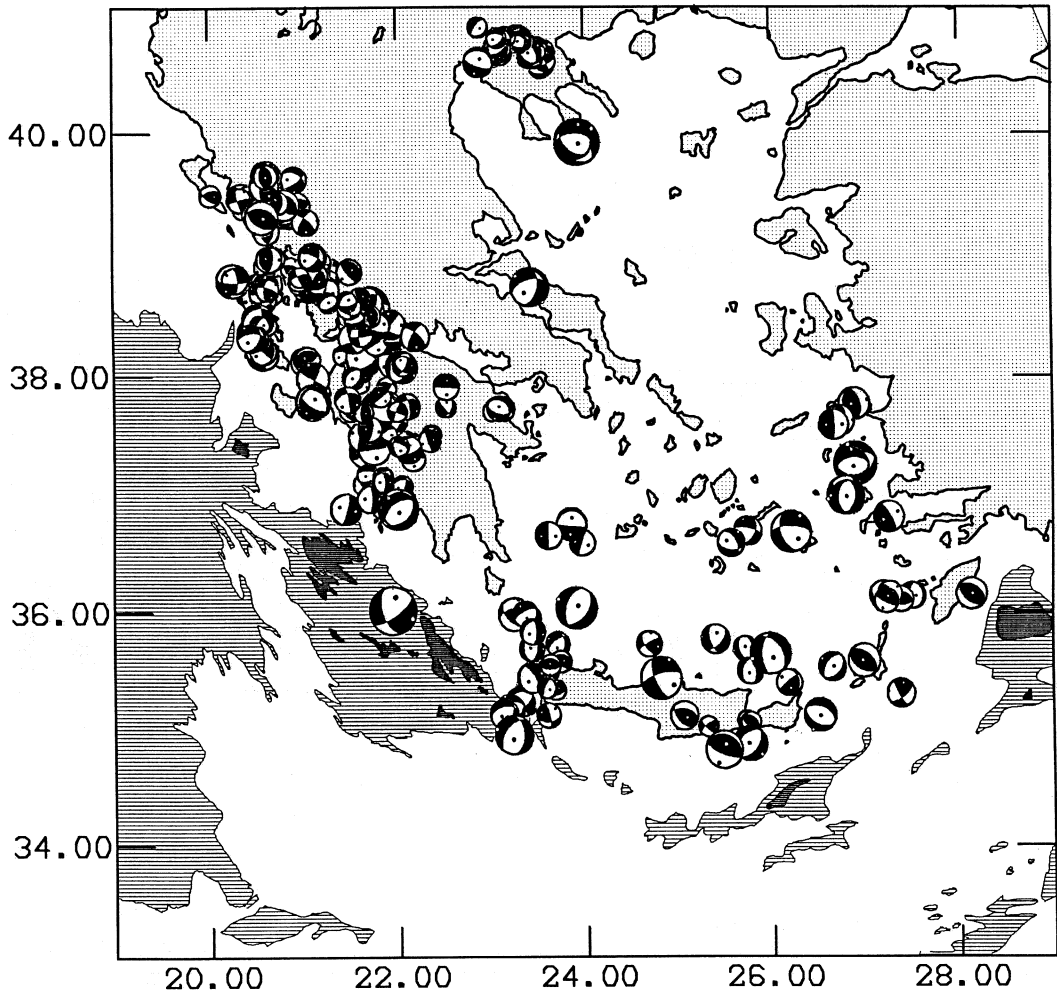
Fig. 6. Focal mechanisms of the crustal earthquakes. In an external position we observe reverse faulting, but in an internal position we observe mostly normal faulting.

computed for moderate earthquakes ($M{\sim}6$) in Macedonia (Soufleris and Stewart, 1981) and in northern Aegea (McKenzie, 1978; Lyberis and Deschamps, 1982; Taymaz *et al.*, 1991). In western continental Greece (Epirus, Ionian islands, western Peloponnese) we observe reverse faulting with *P*-axes trending NE-SW to E-W, as observed by McKenzie (1978) and Anderson and Jackson (1987). Around the Ionian islands, the pattern is more complex, and there are also some strike-slip mechanisms with *P*-axes trending NW-SE. This could be due to the presence of a strike-slip transfer zone west of the islands (Scordilis *et al.*, 1985; Papadimitriou, 1988).

In a more internal position, within the Aegean area, normal faulting is predominant. *T*-axes trend N-S from the Albanian border to the center of the Peloponnese, but *T*-axes vary from NW-SE

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in the southern Peloponnese and in the Kythira strait to E-W in Crete.

A plot of the *P*- and *T*-axes (fig. 7) shows clearly that *P*-axes trend roughly NE-SW along all the Hellenic trench, suggesting that the direction of convergence between Aegea and the African plate is almost constant all along the boundary, regardless of its orientation. In contrast the *T*-axes vary from N-S, in northern Epirus, to E-W in Crete, suggesting that the orientation of the strain pattern is not uniform within the Aegean.

5. Conclusion

In areas where the geodynamics is complex and where the seismic activity is only moderate, as it is for the Mediterranean, the information provided by large earthquakes recorded teleseismically is limited. It is therefore difficult and presumptuous to infer a pattern of tectonics from a sparse sample of earthquakes and fault plane solutions. The use of dense temporary networks of seismological stations can help overcome this difficulty by recording a significantly greater number of earthquakes, if nevertheless during a short period of time. Because of the denser distribution of ray paths in dip and azimuth, the earthquakes will be generally located with a greater accuracy than those recorded teleseimically. However, other types of difficulties arise. The duration of the observation is very short (several weeks) compared to the return period of big earthquakes, and therefore the image of the brittle deformation is very discontinuous. Consequently the lack of activity during such short periods is not significant. Moreover the energy released by the earthquakes cannot be used to quantify the rate or amount of deformation, because the earthquakes are of small magnitude. Moreover, because these earthquakes do not rupture the whole brittle part of the crust, the preexisting faulting, inherited from previous tectonic episodes, can be of fundamental importance. In order to make a significant interpretation of the results gathered during such small-scale experiments, it is necessary to take some precautions. One single earthquake, or one cluster of earthquakes (which generally follow a bigger shock) may not be related to regional strain, but might instead be unrepresentative of the regional strain. Thus it is necessary to smooth the observations in space, and to interpret only results over an area whose dimensions are similar to those of faults associated with large earthquakes (usually several km). The information provided by both big and small earthquakes should be consistent. With such precautions, small earthquakes can give valuable information related to geodynamical processes, and complementary to that provided by big earthquakes.

In the Aegean our observations give some constraints concerning the geodynamics of this area. The subduction zone is not symmetrical, and the slab at shallow depth dips gently and steepens further from the trench. The shape of the slab is contorted, and the intermediate depth fault plane solutions show that the subducted slab is deformed not only by gravity forces, but also by horizontal compression. Taymaz *et al.* (1990) suggested a similar observation, using a more sophisticated technique of body waves modelling, which slightly modified the focal depth and fault plane solutions of some earthquakes studied previously. The use of small earthquakes extends the observations over the whole Hellenic arc. The difference in dip of the subducted slab is probably related to the important amount of internal deformation observed above the slab. It is plausible that the 200 km length of the shallow dipping slab beneath the Peloponnese is mainly due to the convergence between Africa and Aegea (of about $(5 \div 7)$ cm/y) since only 5my and that this high rate is a consequence of the important extension observed within the Aegean for the last 10 my. The step portion of the slab might define the amount of lithosphere subducted into the asthenosphere and due to the convergence between Africa and Europa at a rate of only about 2 cm/y. This slow, but long-term convergence between Europe and Africa might account for both the 100 km of length of the seismic zone, and the 600 km of length of the high-velocity anomaly observed by tomography, which are not related to the same duration of time.

The shallow seismicity is mostly located along the trench, where the relative convergence between the 2 plates is localized. Because of the sea, we could not install stations above the trench and determine precisely the depth of these earth-
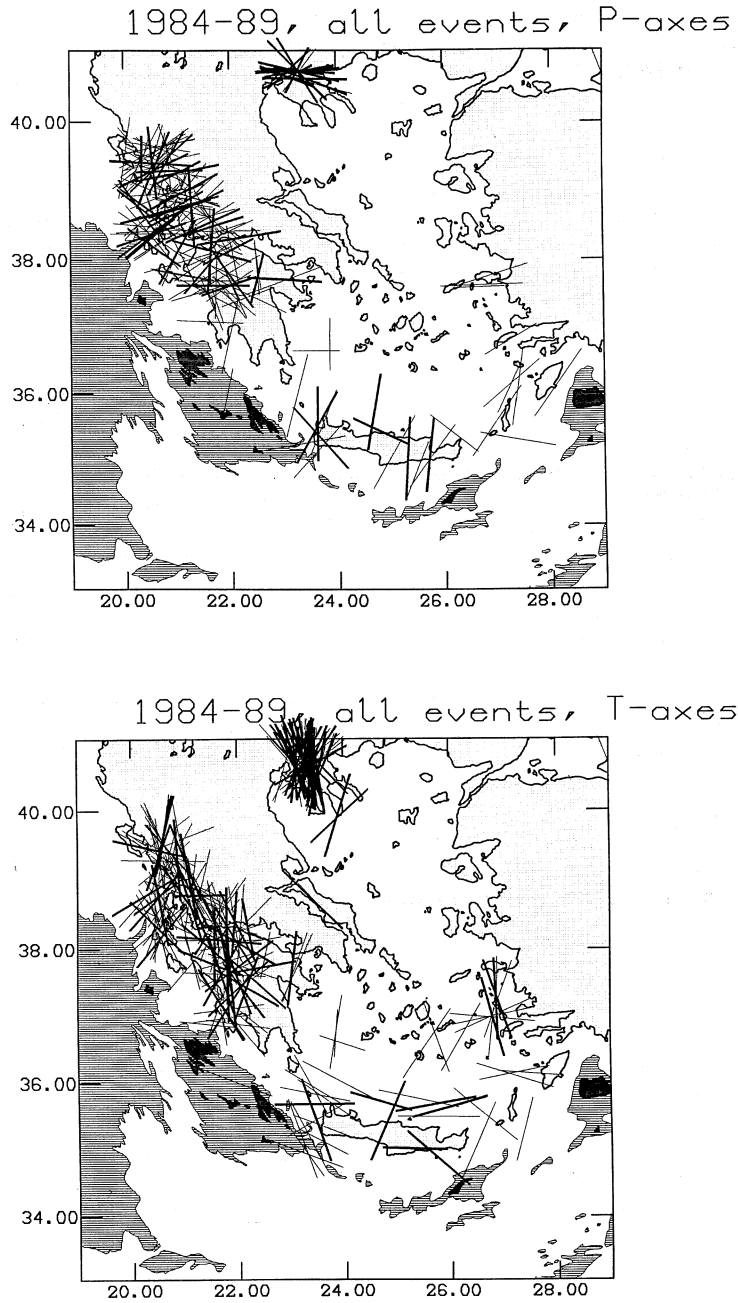
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Fig. 7. Map of the horizontal projection of the *P*- and *T*-axes, with plunge less than 45°. *P*-axes are trending ENE-WSW to NE-SW along the arc, but *T*-axes trend N-S in the northern Aegean, Epirus and northern Peloponnese, NW-SE in southern Peloponnese and E-W in Crete.

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quakes. Shallow seismicity is also observed throughout continental Greece, mostly in Epirus and Peloponnese. These earthquakes do not define a small number of active faults, suggesting that continental convergence can be accommodated by a relatively important amount of deformation on multiple preexisting small faults. Moreover earthquakes are observed from the surface to the Moho and are not confined only to the upper «brittle» part of it. This could be due to high strain rates near the boundary between the 2 plates. Crete and the Sea of Crete are almost aseismic, however, which is surprising considering the active tectonics observed at the surface (Angelier, 1979) and the important amount of total deformation. This deformation apparently is mostly horizontal stretching and therefore may modify the geotherm and the brittle-ductile transition sufficiently to render earthquakes rare.

Fault plane solutions of shallow earthquakes show a complex pattern. Reverse faulting is associated with the trench, and the directions of *P*-axes are roughly trending ENE to NE along the whole arc, as they are for moderate earthquakes. This confirms the direction of relative motion between the Aegean arc and Africa. We were unable to see any variation in the trend of the slip vectors, from West to East, to test the suggestion of south-westward translation of McKenzie (1978), or of rotation of Le Pichon and Angelier (1979). Behind the trench, in an internal position, we observe horizontal extension. *T*-axes trend roughly N-S in the Aegean sea, as in the Gulf of Patras, the Gulf of Corinth and the Dodecanese islands, which is consistent with the tectonic extension observed at the surface in the Aegean. Along the arc, in an internal position (Epirus, Peloponnese, Crete), extension is also predominant, but *T*-axes trend NNW-SSE in Epirus, NW-SE in southern Peloponnese, and E-W in Crete. The strain, therefore, is not uniform, and the extension is roughly perpendicular to the trend of the arc. This rotation in the trend of *T*-axes, consistent with neotectonic observations (Mercier *et al.*, 1987; Angelier, 1979) or paleomagnetic rotations (Kissel and Lay, 1988), suggests that one mechanism of internal deformation observed in the Aegean is gravity spreading, which could be associated with the rollback of the lithospheric slab. Because of the conservation of volume, the principal strain would be longitudinal extension near the boundary, and radial extension far from the boundary (Merle, 1989).

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