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Geometry of the Aegean Benioff zones

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Abstract
The morphology of the Aegean Benioff zone was reconstructed using 1366 hypocentres from the PDE catalogue 1973-1997. Two such zones are identified under the Aegean area, a large one (Main Aegean Subduction, MAS) spanning the whole Hellenic arc and a smaller and younger one under the Western Peloponnesus. The geometry of the MAS suggests that it blocks its own subduction and, therefore, causes the development of the smaller western zone as a result of a step back process in the Pliocene.

Key words *Aegean – subduction – Wadati Benioff zones*

1. Introduction

Probably the most obvious feature of modern subduction zones is the clustering of earthquakes at all depths in Wadati-Benioff zones (Benioff zones or WBZ for short). These are commonly accepted as expressions of deformation of the subducted brittle crust and, therefore, as a tracer for the subducted plate. Many interesting structures can be assessed directly from the spatial and vertical distribution of hypocentres. Examples are the very deep penetration of the Tonga slab, where seismicity reaches down to the 660 km mantle discontinuity (*e.g.*, van der Hilst, 1995), and the horizontal parts of the subducted Pacific plate beneath Peru and Argentina (Cahill and Isacks, 1992). Details of the structure as well as overall properties depend on the general geological setting of the subduction zone, like age of the subducted crust, convergence rate or the presence of continental lithosphere which may force the plate margins into a distinct shape.

Although the earthquakes that form a Benioff zone are likely to occur in the most brittle part of the subducted plate, they are not simply an indicator for the former surface, because source mechanisms change with depth. Seismicity begins outside the arc due to bending of the lithosphere (see, for example, Ruff, 1996). At depths between surface and Moho of the overriding plate, quakes are generated by the friction between the upper and lower plate and thrust faulting is the characteristic mechanism (Ruff and Tichelaar, 1996). In the deepest parts, the asthenosphere in contact with the plate top is too weak to concentrate stresses large enough for earthquake generation. Shocks occur as a result of internal deformation, phase transformations and compression. The maximum depth of WBZ seismicity coincides with the 700 C isotherm (Stein and Stein, 1996). Keary and Vine (1996) give typical figures for the different regimes.

It is often difficult to imagine the overall shape of a three dimensional structure like Be-

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nioff zones and slabs from cross sections, and it is even more difficult to compare different models of the same structure when the published cross sections do not coincide. An important question concerning the Aegean area is: how does the lithosphere fit into a subduction zone with a trench radius of curvature of only 200 km? What is the shape of the Aegean slab? The importance of this question arises from the fact that the Aegean arc is the strongest curvatured arc on earth (see table in Jarrard, 1986), while the subducted lithosphere can be expected to be thick. The tomographic results by Meulenkamp *et al.* (1988) show a 100 km thick slab (what may be an artifact of their coarse grid). In general, oceanic lithosphere near subduction zones can be expected to be about 100 km thick (Stein and Stein, 1996). A recent tomographic study by Papazachos and Nolet (1997) shows a high velocity structure similar to half a cup – but their model is not deep enough to show what happens with the lithosphere in its entirety.

In this paper, I try to find answers by good visualization of the problem. The WBZ, as a tracer of the downgoing plate is modeled along several profiles. These are shown as cross sections and 3D plots. By the curvature along these profiles (that is, the variation of slab dip), one can conclude that the slab hinders its own subduction. A by-product is the interpretation of the unusually diffuse hypocentre cloud beneath the Peloponnesus as a second slab.

2. The Aegean Wadati Benioff zone

The Aegean area is the seismically most active region of Europe, with many shallow earthquakes almost everywhere in the region and hypocentres down to depths of about 200 km in its southern and central part (fig. 1). The idea that especially the deep seismicity originates from the subduction of the African plate was developed in the late sixties. Papazachos and Delibasis (1969) demonstrated underthrusting by evaluation of fault plane solutions, while the distribution of deep hypocentres in Caputo *et al.* (1970) confirmed this. The position of the plate boundary is still under discussion. Some authors (*e.g.*, LePichon and Angelier, 1979; Giunchi *et al.*, 1996; Cianetti *et al.*, 1997) assume that the Hellenic Trench system is the expression of the plate boundary, but others (*e.g.*, LePichon *et al.*, 1995; Mascle and Chaumillon, 1997) assume that it is marked by the morphological deep between the Mediterranean Ridge and the North African continental slope.

The Aegean Benioff zone has been the subject of extensive investigations. Papazachos (1973) presents the distance of hypocentres from the volcanic arc *versus* hypocentral depth in a cross section that clearly defines a Benioff zone, although it is stacked along the whole arc and, therefore, only resolves global structures like the mean dip angle. Galanopoulos (1975) described a set of intermediate depth (> 90 km) earthquakes outside the trench which might reflect a young slab (pliocene) under the Mediterranean Ridge and suggested that several independent slab tongues are responsible for the complicated hypocentre distribution in the Aegean region. The only alternative (according to Galanopoulos, 1975) would be a 150 km thick African lithosphere. Richter and Strobach (1978) tailored a model of about 800 wooden pearls representing earthquake foci and divided the area into five profiles of different width, showing variations in extension and sharpness of the Benioff zone. From these profiles and from the complete model, they concluded that the slab may split up into two parts somewhere beneath Crete or the sea of Crete. Makropoulos and Burton (1984) divided the Aegean into 36 overlapping, vertical cross sections around the centre of the volcanic arc and rendered perspective views of the lower envelope of the hypocentres. These plots unfortunately do not distinguish between those parts of this envelope which connect hypocentres and those which artificially connect the slab with the earth's surface in the north through aseismic (seismically «empty») space. Kondopoulos *et al.* (1985) divided the Aegean into four sectors and found different WBZ angles for the western, southern and eastern parts. The western and eastern sectors of the WBZ are significantly steeper than the two sectors in between them. Spakman *et al.* (1988) and Meulenkamp *et al.* (1988) show tomographic cross sections through Crete in N-S direction, revealing a slab about 600 km in length and

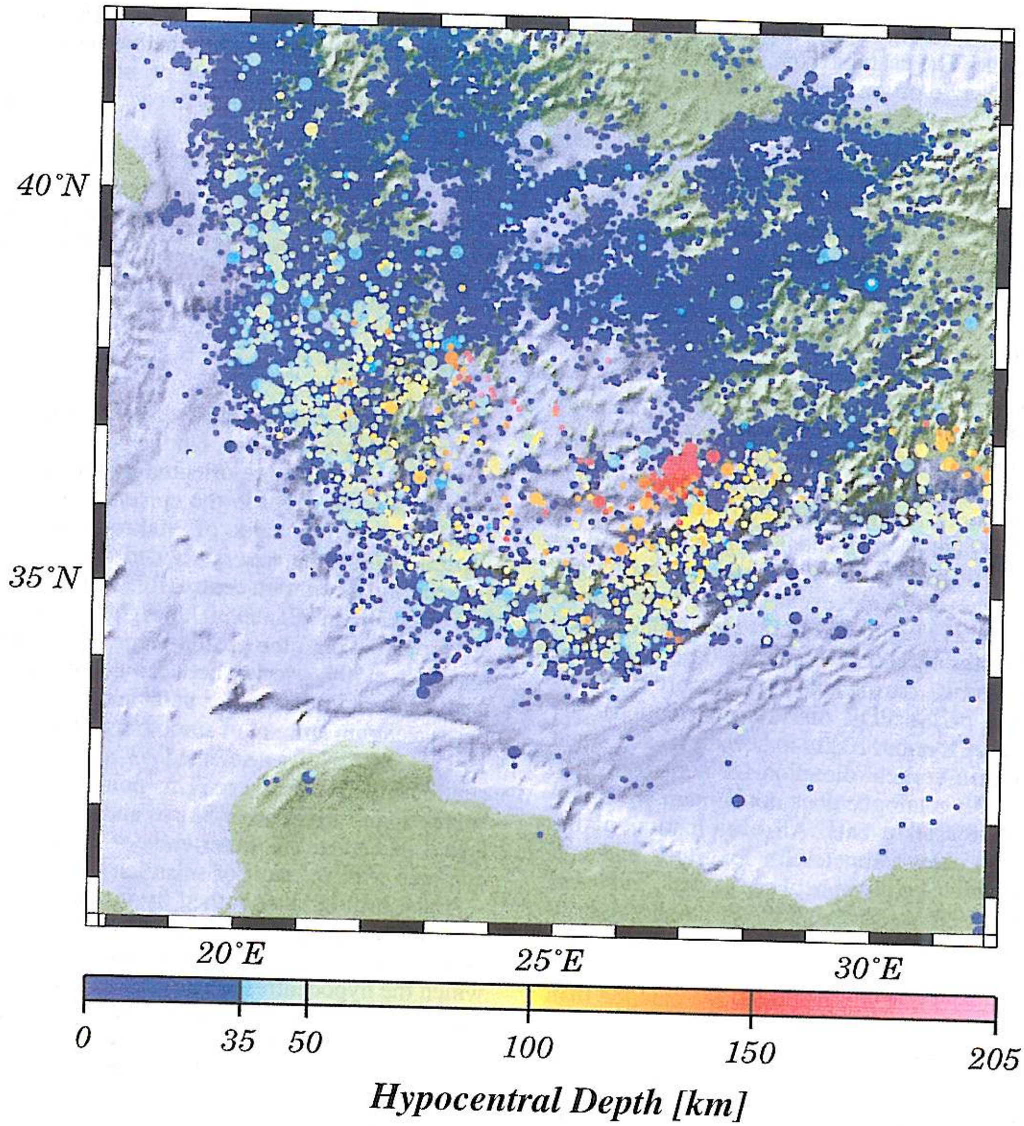

Fig. 1. Map of all 24942 PDE hypocentres between January 1, 1973 and August 25, 1997. Only those below 35 km are used in this study (colors other than dark blue).

100 km in thickness. Papadopoulos (1997) suggested that the high velocity anomalies visible in the tomographic results do not only represent the currently active slab but also remnants of a miocene paleosubduction which are no longer linked to the lithosphere of today's WBZ.

Kovachev *et al.* (1992) analyzed 105 microearthquakes between eastern Crete and Karpathos and identified two Benioff zones, one associated to the Pliny trench, the other to the Strabo trench. Christova and Nikolova (1993) showed a Benioff zone which consists of two flanks with different dip angles, separated by an aseismic zone. The eastern flank dips at 40 degrees, the dip angle of the western flank is 20-30 degrees. These different angles provide the two flanks from crossing. Muco (1996) presented evidence for an end of the subduction near the Greek-Albanian border by arguing that the deep foci sometimes reported in that area are probably artefacts of mislocation.

3. Data

Since late 1997, PDE hypocentre data of NEIC are accessible in a condensed format via World Wide Web (Earthquake Search, 1997). The database covers the whole world since 1973. All hypocentres from January 1, 1973 to August 25, 1997 between 18 and 32 degrees east and 30 and 42 degrees north were extracted forming a catalogue of 24942 events. Most hypocentres (85% or 21279 out of 24942) are fixed to standard depths of 5 km, 10 km, or 33 km. According to the NEIC Earthquake Data Report 1997 (included in the PDE diskette distribution), the accuracy is some 10 km in horizontal and about 25 km in vertical direction for non-fixed foci. The PDE catalogue does not contain error estimates for all events. An examination of the 117686 error estimates for longitude and latitude which are given between January 1990 and March 1998 shows that 80% are smaller than 20 km. Depth errors are less frequently reported than epicentral errors. Again, 80% of 86296 estimates for the same time are smaller than 20 km.

Events above 35 km depth were not used in this study because they represent mainly crustal seismicity (for the Aegean crustal thickness see Makris, 1978, and Makris and Stobbe, 1984). But only the southernmost part of the crustal activity is due to the deformation of the slab. The topmost part of subduction, including the backstop, will therefore not be mapped here. All events with fixed depths were removed from the catalogue. This does not only affect depth values 5, 10, and 33 km. Depths which are integer multiples of 10 km were also considered «fixed-depths»: the depth distribution shows peaks at those values which are probably caused by human preference for round numbers. The problem with these depths is not whether the events are a few kilometers deeper or not, but that fixing one coordinate affects the accuracy of all others.

The remaining set of 1366 subcrustal events with magnitudes (mostly ML, measured by the Greek National Observatory) between 3.0 and 6.1 represent the Aegean Wadati Benioff zone(s) (fig. 2).

4. Method

The Benioff zone could be reconstructed from arbitrary oriented profiles, but the best choice is a set of lines which are oriented parallel to the dip which is apparent in the epicentre map.

Similar to the work of Makropoulos and Burton (1984), the area is divided into 23 profiles around a common centre (fig. 2). Hypocentres up to 30 km left and 30 km right of a profile are projected onto the profile plane. This width results in a 50% overlap between neighbouring profiles even in their outer parts and hence imposes a certain amount of smoothing. The point 25.12E 38.32N was selected as approximate centre of the Hellenic arc. This point is located southeast of the island of Skiros and west of the island of Chios. The outer ends of the profiles coincide with the start of seismicity.

Each profile is smoothed by visual and numerical interpolation. For easier handling, this is done in a rectangular coordinate frame, in which the hypocentres are projected. The line is then projected back into spherical coordinates. The interpolation sometimes causes sidelobes at the outer end of the profiles. These resemble a small southward dip of the Benioff zone. The fitting-by-hand allows single hypocentres which are obviously misplaced or which belong to another subduction to be ignored. Profiles 1-4 (fig. 3) show this behaviour: several hypocentres in the western part, at a depth of slightly more than 100 km, are interpreted as part of a second, smaller Benioff zone (see below), which is treated independently.

I decided to draw these lines as «mean» lines for two reasons. First, a mean line is easier to find than a top or bottom line, because the hu-

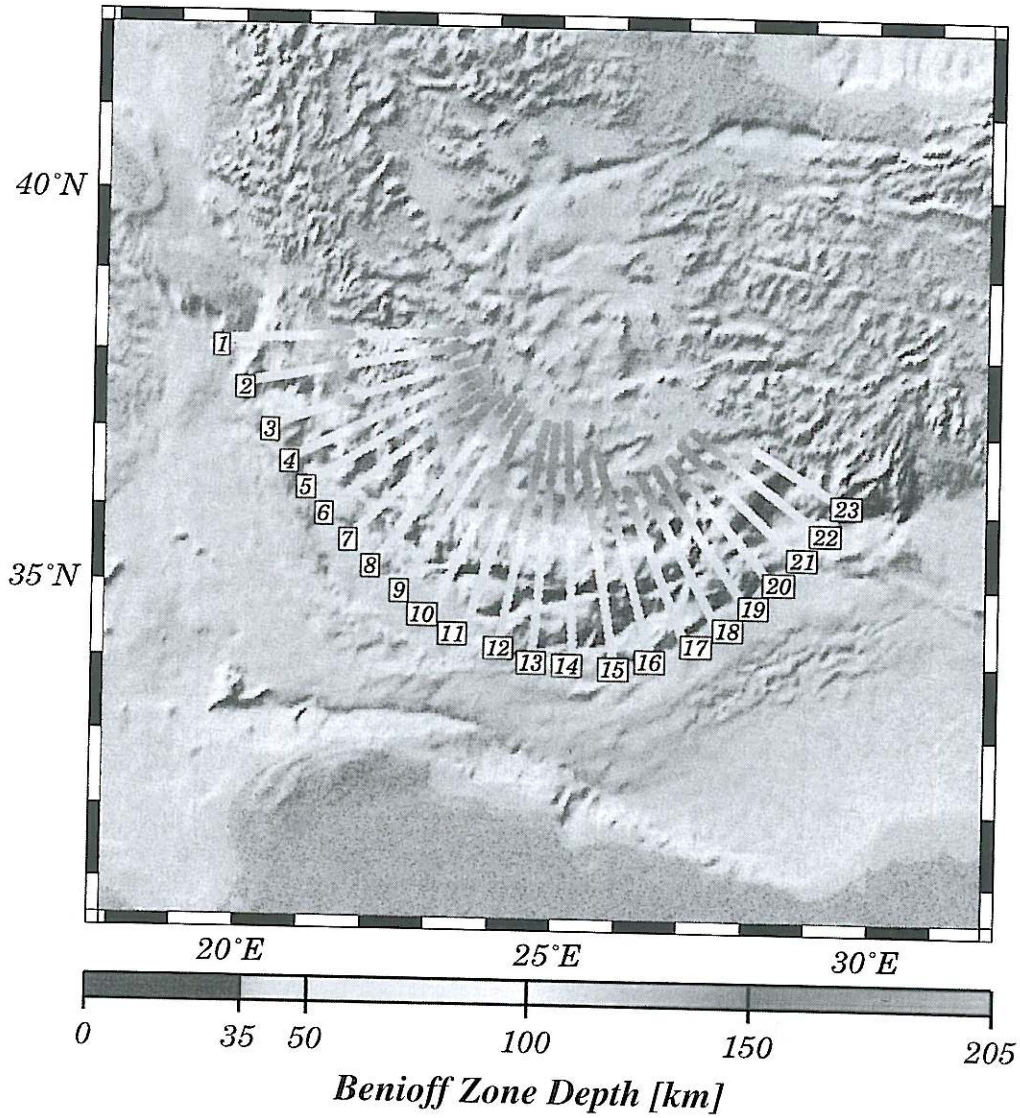

Fig. 2. Map of all profiles and BWZ depths on these profiles. Profiles 1 to 4 contain two dipping seismogenic zones.

man eye sees the overall structure first. Second, the geological meaning of a top or bottom line changes with depth: The largest events are interplate earthquakes which are generated at depths down to about 40 km directly at the interface between underthrusting and overriding plate. This interplate zone may extend up to the surface and seems to end at the depth of the Moho of the upper plate (Ruff and Tichelaar, 1996). Deeper events presumably occur within the downgoing plate (Ruff, 1996) and are «near the top surface of slabs» (Kirby *et al.*, 1996).

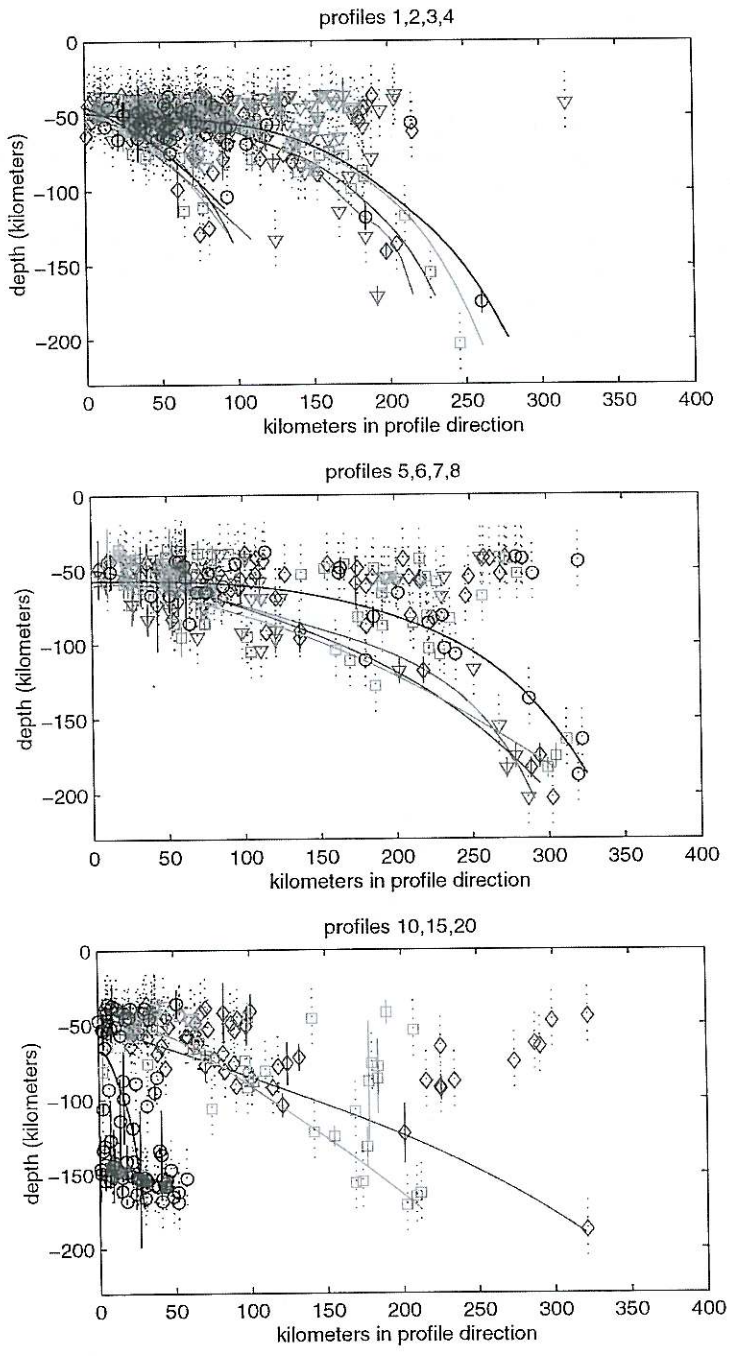

Fig. 3. Hypocentres and curve fits of several profiles. Error bars: solid = individual depths errors given by PDE, dotted = assumed 20 km global error for events without explicit error information.

An example of what is meant by «near» is the Alaskan Benioff zone, for which Abers (1996) has shown that WBZ thickness is no more than 10 km.

The central assumption of mapping slabs by their seismicity is that the relative focal depth with respect to the slab surface is constant in the whole slab. This may not be true at the lower edge of the interplate regime, where a certain «jump» of relative depth may occur due to the fading of seismicity in the upper plate. But given the small thickness of Benioff zones, the assumption probably holds for the deeper part considered in this study.

5. Results

Most of the 23 profiles dip with an angle of about 30 degrees over most of their length. Some become steeper at greater depths, and some dip angles are significantly smaller (fig. 4). Three groups of profiles are distinguishable: group A consists of flat and long profiles (profiles 1-6, visible slab lengths between 350 km and 450 km), group B are steep and long profiles (profiles 7-15, lengths about 350 km), and group C are steep and short profiles (profiles 16-23, lengths about 250 km). The flat profiles of group A reach the greatest depths.

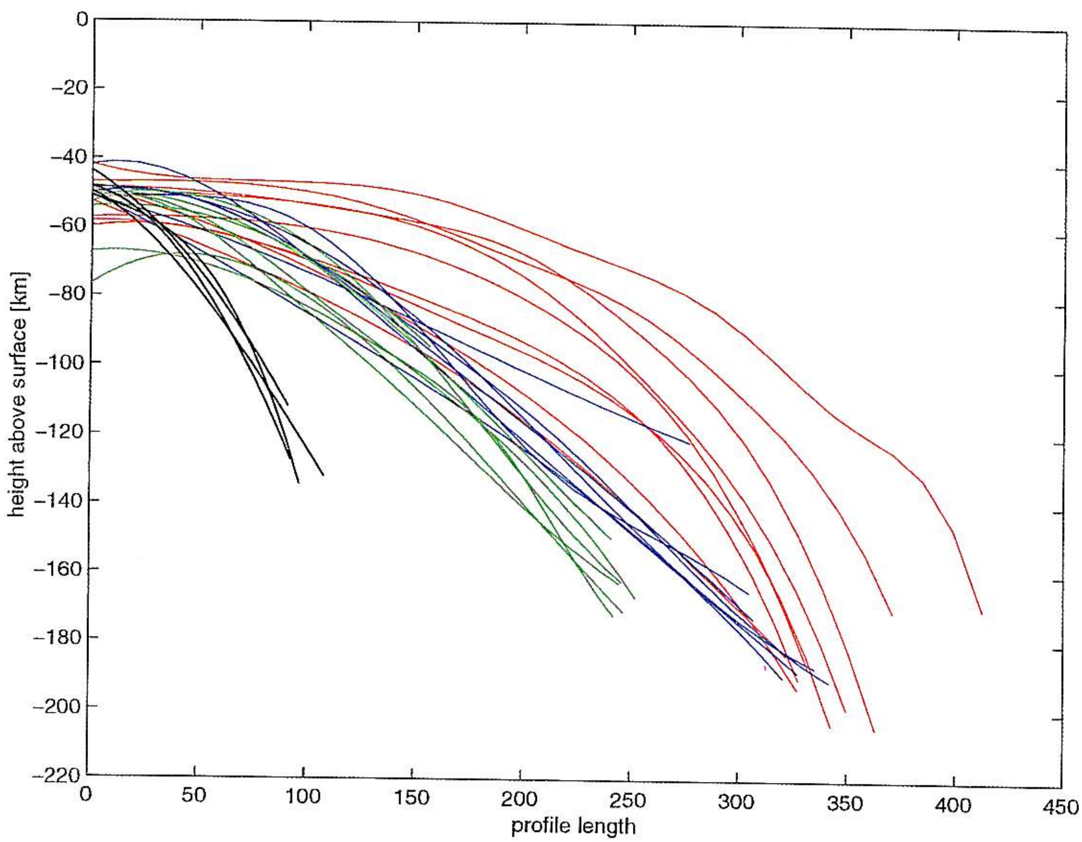

Fig. 4. Comparison of all profiles. Blue: main Aegean WBZ, red: Western Peloponnesus WBZ. The left end corresponds to the outward ends of profile lines in fig. 2, leftward dips are artefacts. Three groups of blue profiles are clearly distinguishable: flat and long profiles, long profiles dipping at 30 degrees, and short profiles dipping at 30 degrees. No vertical exaggeration.

Profiles 1-4 obviously contain two seismically active zones (red profiles in fig. 4 are from the smaller one. The events belonging to the smaller WBZ are listed in table I), but these should not be confused with double Benioff zones as known from Japan or Alaska (*e.g.*, Abers, 1996): double Benioff zones are separated by 20-40 km and run parallel over their entire length, whereas the two seismogenic zones beneath the Peloponnesus are separated by up to 100 km and are not parallel. The larger one is part of the Main Aegean Subduction zone (MAS) which wraps around the whole Aegean arc. The smaller, western part seems to be a subduction zone of its own, referred to as Western Peloponnesus Wadati Benioff zone. It is less active and therefore less well defined than the MAS, but the hypocentre distribution is clear. The dipping angle is steeper than in the rest of the Aegean, the maximum depth is about 120 km.

The flatness of the western profiles is in part an artifact due to the existence of the Western Peloponnesus WBZ: lines are drawn to the outermost part of the profiles for both WBZ (the backstop), therefore the first 100 km overlap because it is not clear which event is caused by which slab.

In fig. 2, the length of the «visible» slab for each profile of the Main Aegean Subduction can be seen. Remarkable profiles are 8 and 9, which are less steep than their neighbours 7, 10, and 12 in their outer parts, but seem to reach the same depths at their ends. The latter is an artefact due to overlapping: the deep foci defining the lower ends are the same for 7, 8, and 9. The volume mapped into profile 11 does not contain deep events, therefore it is shorter than its neighbours. These profiles correspond to the Kythira strait between the Southern Peloponnesus and Crete and may well reflect a real change in slab geometry. But it must be kept in mind that subduction may be aseismic: Okal and Talandier (1997), for example, have shown that an aseismic part of the slab beneath Bolivia does not represent a tear in the slab because it contains a continuous waveguide.

The surfaces constructed from the individual profiles shown in fig. 4 are given in perspective view in fig. 5.

6. Interpretation

The construction of the Main Aegean Subduction zone using profiles parallel to the dip direction extrapolates these profiles into depth from their last few points by linear regression (only the last two points were used to take the steepness of the western part into account. If more points were used, the following interpretation would not be affected significantly).

Table I. Events of the Western Peloponnesus WBZ: day, origin time, longitude (deg.), latitude (deg.), depth (km), depth error (km), magnitude (M_L), number of stations used (data: PDE).

Day	Origin time	Long.	Lat.	Depth	Depth error	Magnitude	Stations
22 March 1977	20:02:13.0	20.80	38.41	84.00	?	4.3	?
23 December 1981	09:22:22.5	21.75	37.60	88.00	3.2	4.1	12
07 March 1994	17:55:41.3	21.45	37.53	99.00	18.2	3.5	?
05 November 1985	10:40:38.7	22.08	36.93	104.00	13.8	3.9	12
20 March 1974	23:38:51.6	21.70	37.23	111.00	?	3.9	14
08 April 1986	10:52:51.1	21.58	37.17	113.00	13.6	3.7	15
25 November 1975	20:54:18.1	21.68	37.61	125.00	?	3.4	6
22 May 1974	07:40:15.2	21.55	37.62	129.00	?	3.9	12
31 December 1975	18:19:20.7	21.73	38.37	133.00	?	2.8	12

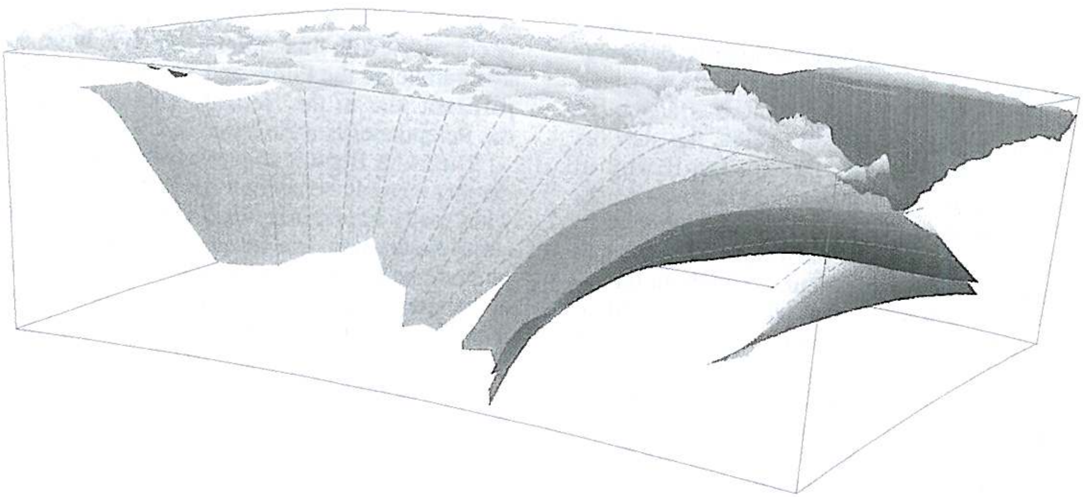

Fig. 5. Perspective view of the Benioff zones in spherical geometry, surface topography exaggerated by factor ten, WBZ not exaggerated. View from the north east.

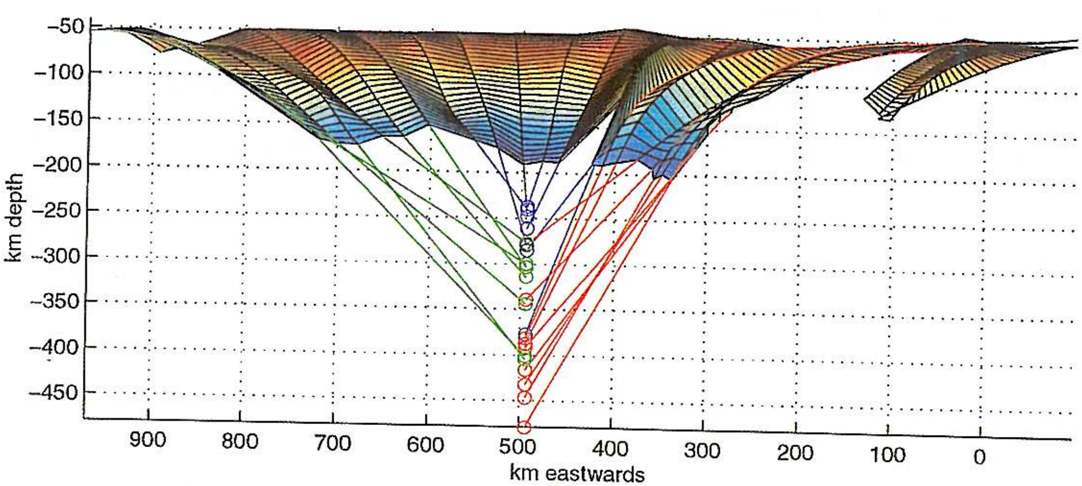

Fig. 6. Extrapolation of the main Aegean WBZ profiles to their common endpoint. Extrapolations are plotted as straight lines. Western profiles reach the greatest depth, followed by the eastern profiles, the middle part of the WBZ ends at smallest depths. Note that the Western Peloponnesus WBZ is significantly steeper than the MAS WBZ. The view is exactly from north, with no vertical exaggeration.

The extrapolated profiles meet, of course, beneath their common endpoint, but the sequence at which they reach this point is interesting (fig. 6): the western profiles reach the greatest depths, above them lie the eastern profiles, and the uppermost ones originate in the mid part beneath Crete. This sequence differs from that seen in the seismically active part (figs. 2 and 4), where the western profiles are the deepest, above them the mid part and the eastern profiles are on top.

This re-arrangement and separation into three parts implies at least intensive internal deformation between the eastern and middle part, and possibly a tear in the slab. The re-arrangement also answers the question of how the slab fits into the curvature: it doesn't. Even if we had three separate slabs, they would meet (or crash) at 400 km depth or above, depending on their thickness, and prevent each other from further downgoing. This behaviour is visible in the results of Papazachos and Nolet (1997). Meulenkamp *et al.* (1988) show a very long slab from the south, suggesting that the western and eastern parts are blocked, although plate movement is mainly in a NE-SW direction.

If further downgoing of a slab is not possible, a continuous (roll back) or discontinuous (step back) retreat of the subduction as a whole will occur. LePichon and Angelier (1979) suggest a roll back as a consequence of the asymmetric collision of the rigid African plate with the Anatolian plate in the east and with the European plate in the northwest. The two Benioff zones of Kovachev *et al.* (1992), which are not resolved in the PDE data, may indicate a step back from the Pliny to the Strabo trench. The Western Peloponnesus WBZ shown in fig. 6 is obviously due to a step back, which most probably occurs in the direction of strongest motion.

The length of the Western Peloponnesus WBZ (about 200 km, fig. 4), combined with the present day convergence rate (43 mm/a velocity of the Aegean with respect to Africa at Chrisokellaria, after LePichon *et al.*, 1995) and the assumption that there is no aseismic part of the slab implies that the step back started to occur about 4.5 Ma ago in the Lower Pliocene.

7. Discussion

LePichon *et al.* (1995) computed the present day plate velocities from recent laser ranging and GPS measurements, resulting in a movement of the Aegean area which is dominated by a rotation pole about 200 km north of the Nile delta. Cianetti *et al.* (1997) FE-modeled this velocity field, which causes a mainly southwestward plate motion. The decrease of WBZ length from the western to the eastern part roughly coincides with this motion, which is more or less parallel to the Pliny- and Strabo trench, but perpendicular to the Hellenic Trench between the Peloponnesus and Crete.

The MAS Benioff zone geometry suggests a distinct tectonic behaviour. An important key to what really happens beneath the Aegean is the stress field which can be assessed by the computation and evaluation of fault plane solutions or moment tensors. Unfortunately, not many focal mechanisms for subcrustal foci have been published. Hatzfeld *et al.* (1993) presented 26 newly determined mechanisms and 13 others quoted from earlier works by several authors. Most of these mechanisms coincide roughly with directions expected by the arcuate structure of the Aegean, but also reveal components different from those expectations, suggesting the existence of stress other than due to gravitational sinking of the slab. Almost all «really» deep foci (those with $z > 80$ km) are located in the Kos cluster, a dense group of events around 120 km depth beneath the Dodecanese islands (see fig. 2). The *P* axes described by Hatzfeld *et al.* (1993) show the existence of a NNE-SSW compression as it would probably occur when the eastern and the middle part of the slab crash into each other as suggested above.

8. Conclusions

Constructing a 2D surface as a representation of a Wadati Benioff zone has easily been done since the availability of high quality hypocentres provided by the USGS NEIC. A relatively small data set is selected, which is both easy to handle and detailed enough to serve as a basis for further considerations. It must be kept in mind that this method maps only the seismically active part of a slab and that the largest part of the Aegean slab seems to be completely aseismic (regarding the 600 km long high velocity zone in Meulenkamp *et al.*, 1988). The current seismicity is linked to the present-day tectonics.

The earthquake distribution as shown here confirms existing hypotheses about roll back or step back occurrence and suggests an explanation for the partly abnormal stress distribution mapped by focal mechanisms.

The existence of the Western Peloponnesus WBZ, although not as well defined as the Main Aegean BWZ, demonstrates that the Aegean is still subject to substantial structural evolution.

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REFERENCES

ABERS, G.A. (1996): Plate structure and the origin of double seismic zones, in *Subduction: Top to Bottom*, edited by G.E. BEBOUT *et al.*, *Geophysical Monograph 96*, American Geophysical Union, Washington, DC.

CAHILL, T. and B.L. ISACKS (1992): Seismicity and shape of the subducted Nazca plate, *J. Geophys. Res.*, **97**, 17503-17529.

CAPUTO, M.G., F. PANZA and D. POSTPISCHL (1970): Deep structure of the Mediterranean basin, *J. Geophys. Res.*, **75**, 4919-4923.

CHRISTOVA, C. and S.B. NIKOLOVA (1993): The Aegean region – deep structures and seismological properties, *Geophys. J. Int.*, **115**, 635-653.

CIANETTI, S., P. GASPERINI, M. BOCCALETTI and C. GIUNCHI (1997): Reproducing the velocity and stress fields in the Aegean region, *Geophys. Res. Lett.*, **24** (16), 2087-2090.

EARTHQUAKE SEARCH (1997): in «News and Notes», *Seismol. Res. Lett.*, **68** (4); see also http://www.neic.cr.usgs.gov/neis/epic/epic.html.

GALANOPOULOS, A.G. (1975): A new model accounting for the intermediate earthquakes at the Convex side of the Hellenic arc; *Annales Geologiques des Pays Helleniques*, 355-371.

GIUNCHI, C., A. KIRATZI, R. SABADINI and E. LOUVARI (1996): A numerical model of the Hellenic subduction zone: active stress field and sea-level changes, *Geophys. Res. Lett.*, **23** (18), 2485-2488.

HATZFELD, D., M. BESNARD, K. MAKROPOULOS, N. VOULGARIS, V. KOUSKOUNA, P. HATZDIMITRIOU, D. PANAGIOTOPOULOS, G. KARAKAISIS, A. DESCHAMPS and H. LYON-CAEN (1993): Subcrustal micro-earthquake seismicity and fault plane solutions beneath the Hellenic arc, *J. Geophys. Res.*, **98** (B6), 9861-9870.

JARRARD, R.D. (1986): Relations among subduction parameters, *Rev. Geophys.*, **24** (2), 217-284.

KEAREY, P. and F.J. VINE (1996): *Global Tectonics* (Blackwell Science Ltd.), second edition, 146-149.

KIRBY, S., E.R. ENGDAHL and R. DENLINGER (1996): Intermediate-depth intraslab earthquakes and arc volcanism as physical expressions of crustal and uppermost mantle metamorphism in subducting slabs, in *Subduction: Top to Bottom*, edited by G.E. BEBOUT *et al.*, *Geophysical Monograph 96*, American Geophysical Union, Washington, DC.

KONDOPOULOU, D.P., G.A. PAPADOPOULOS and S.B. PAULIDES (1985): A study of the deep seismotectonics in the Hellenic arc, *Boll. Geofis. Teor. Appl.*, **27** (107), 197-207.

KOVACHEV, S.A., I.P. KUZIN and S.L. SOLOVIEV (1992): Microseismicity of the frontal Hellenic arc according to OBS observations, *Tectonophysics*, **201**, 317-327.

LEPICHON, X. and J. ANGELIER (1979): The Hellenic Arc and trench system: a key to the neotectonic evolution of the Eastern Mediterranean area, *Tectonophysics*, **60**, 1-42.

LEPICHON, X., N. CHAMOT-ROOKE and S. LALLEMANT (1995): Geodetic determination of the kinematics of Central Greece with respect to Europe: implications for Eastern Mediterranean tectonics, *J. Geophys. Res.*, **100** (B7), 12675-12690.

MAKRIS, J. (1978): The crust and upper mantle of the Aegean region from deep seismic soundings, *Tectonophysics*, **46**, 269-284.

MAKRIS, J. and C. STOBBE (1984): Physical properties and state of the crust and upper mantle of the Eastern Mediterranean Sea deduced from geophysical data, *Mar. Geol.*, **55**, 347-363.

MAKROPOULOS, K.C. and P.W. BURTON (1984): Greek tectonics and seismicity, *Tectonophysics*, **106**, 275-304.

MASCLE, J. and E. CHAUMILLON (1997): Pre-collisional geodynamics of the Mediterranean Sea: the Mediterranean Ridge and the Tyrrhenian Sea, *Ann. Geofis.*, **40** (3), 569-586.

MEULENKAMP, J.E., M.J.R. WORTEL, W.A. VAN WAMEL, W. SPAKMAN and S.E. HOOGERDUYN (1988): On the Hellenic subduction zone of Crete since the late middle Miocene, *Tectonophysics*, **146**, 203-215.

MUCO, B. (1996): The reliability of reported depths of some earthquake sequences in Northwestern Greece (Epirus and Corfu islands) and the geodynamic situation of the area, *Ann. Geophys*, **39** (6), 1125-1132.

PAPADOPOULOS, G.A. (1997): On the interpretation of large-scale seismic tomography images in the Aegean Sea area, *Ann. Geofis.*, **40** (1), 37-42.

PAPAZACHOS, B.C. (1973): Distribution of seismic foci in the Mediterranean and surrounding area and its tectonic implication, *Geophys. J. R. Astron. Soc.*, **33**, 421-430.

PAPAZACHOS, B.C. and N. DELIBASIS (1969): Tectonic stress field and seismic faulting in the area of Greece, *Tectonophysics*, **7**, 231-255.

PAPAZACHOS, C.B. and G. NOLET (1997): *P* and *S* deep velocity structure of the Hellenic area obtained by robust nonlinear inversion of travel times, *J. Geophys. Res.*, **102** (B4), 8349-8367.

RICHTER, I. and K. STROBACH (1978): Benioff zones of the Aegean arc, in *Alps, Apennines, Hellenides*, edited by H. CLOSS, D. ROEDER and K. SCHMIDT, Inter-Union Commission on Geodynamics, *Scientific Report No. 38*, 410-414.

RUFF, L.J. (1996): Large earthquakes in subduction zones: segment interaction and recurrence times, in *Subduction: Top to Bottom,* edited by G.E. BEBOUT *et al.*, *Geophysical Monograph 96*, American Geophysical Union, Washington, DC.

RUFF, L.J. and B.W. TICHELAAR (1996): What controls the seismogenic plate interface in subduction zones?, in *Subduction: Top to Bottom*, edited by G.E. BEBOUT *et al.*, *Geophysical Monograph 96*, American Geophysical Union, Washington, DC.

SPAKMAN, W., M.J.R. WORTEL and N.J. VLAAR (1988): The Hellenic subduction zone: a tomographic image and its geodynamic implication, *Geophys. Res. Lett.*, **15** (1), 60-63.

STEIN, S. and C.A. STEIN (1996): Thermo-mechanical evolution of oceanic lithosphere: implications for the subduction process and deep earthquakes, in *Subduction: Top to Bottom*, edited by G.E. BEBOUT *et al.*, *Geophysical Monograph 96*, American Geophysical Union, Washington, DC.

VAN DER HILST, R. (1995): Complex morphology of subducted lithosphere in the mantle beneath the Tonga trench, *Nature*, **374**, 154-157.

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