Proceedings of the 4th International Symposium on the Analysis of Seismicity
and Seismic Risk
Bechyně Castle, Czechoslovakia, September 4 - 9, 1989

2-D VELOCITY STRUCTURE IN THE AXIAL PLANE OF THE BENIOFF ZONE IN THE SOUTH AEGEAN SEA

S.T.Tassos, J.S.Kalogeras
National Observatory of Athens, Seismological Institute,
P.O.Box 20048, GR 118 10 Athens, Greece

I.P.Kuzin
Seismological Laboratory, Institute of Oceanography,
Krasikova 23, Moscow 117218, U.S.S.R.

ABSTRACT

A method developed by Kuzin (1974) for the determination of body wave velocities with the use of travel-time residuals of events that lie on the axial plane of a Benioff zone has been applied.

Two polygons, one referring to the VLS station, at Cephallonia island, and another to ARG station, at Rhodes island, have been used. Events with focal depths $h > 30$ km and magnitudes $4.5 < M_S < 6.5$ have been considered. The two areas have been divided in blocks of one square degree and 50 km depth with 50% overlap. In each block the average travel-time residual, the average focal depth and the average epicentral distance, with reference to VLS and ARG stations, have been calculated. In each center of gravity of focal depth (h) and epicentral distance (Δ) the average travel-time residual has been assigned. Then the travel-time residual field has been constructed, the gradients of travel-time residuals along segments of ray-paths were determined and, finally, the observed values of velocities along the chosen segments of ray-paths were calculated.

INTRODUCTION

Various methods and techniques are being used for the determination of the velocity structure in the upper mantle. The main goal of all methods and techniques is to come up with a velocity model of the upper mantle and thus examine the connection that exists between deep mantle structure and large tectonic and seismically active zones, such as the Aegean Arc is.

The methods that require data from teleseismic events (Romanowicz 1980, Hovland et al. 1981, Hovland and Husebye 1981, Babuska et al. 1984, Pajdusak et al. 1986, Plomerova et al. 1988), and those that use teleseismic as well as regional local events (Spakman, 1986) are very sensitive to the spatial and azimuthal distribution of earthquakes.

Other more sophisticated inversion techniques (Granet and Cara 1988, Granet and Trampert 1988) require very high computer capacity, which nowdays cannot be offered by common computers.

The method being used in our work, was developed by Kuzin (1974) and gives a 2-D velocity structure in the axial plane of a Benioff zone, which nevertheless gives a pretty good 3-D picture, since it refers to aninclined plane. Its main advantage, which by far compensates its 2-D disadvantage, is its simplicity (no use of a computer is necessary), and the requirement of a relatively small number of regional events (300 - 500 events).

In this paper, we will discuss the method and we will attempt to apply it in the statically and dynamically complicated area of the South Aegean. It is the complexity of the region which requires a multivarious approach, using different methods and techniques relying on data and avoiding theoretical preoccupation and formalism in the interpretation of data.

THE METHOD

The method is based on travel-time residuals, of P and S waves, of events that lie on the axial plane of a Benioff zone. The idea is to find the gradient of travel-time residuals (dt/l) at different segments of a ray path, construct the travel-time residual field and from that the corresponding velocities of P and S waves at the central part of that particular segment of the ray path and finally the field of V_P and V_S. In our case only travel-time residuals of P waves have been considered.

For the South Aegean region we have two problems. The first, and most important, has to do with the curvature and the amphitheatrical shape of the Benioff zone (Papazachos and Comninakis 1969, 1971). The second, which is a result of the first, is its limited horizontal extension, since, in order to avoid the curvature, we are obliged to consider only a part of the Benioff zone.

Figure (1) shows the two regions, with respect to ARG station at Rhodes island, and to VLS station at Cephallonia island, that have been selected in our work. Events between 1970 and 1983, with focal depths h>30 km and magnitudes $4.5 \leq M_S \leq 6.5$ have been considered. The epicenters and the depths are taken from Comninakis and Papazachos' catalogue (1986), while the epicentral distances and travel-time residuals are taken from the ISC bulletins. In the cases that the travel-time residuals were found to be $5.0 < \Delta t_P < -5.0$ sec the events were rejected from calculations.

The two regions have been divided in blocks of one square degree and 50 km depth with 50% overlap. In each block the average focal depth has been assigned, in order to approach the axial planes of the Benioff zone (figure 3). From the two regions only the one that corresponds to ARG station has been considered, because the VLS station, as it is shown in figure (2), does not lie at the top of the axial plane of the Benioff zone, which is a prerequisite for the application of the method in order to have some quantitative accuracy.

Then in each block of the ARG region the travel-time residual (Δt_P) - after the correction for the crust below the station was made, which for ARG is 0.5 sec according to Panagiotopoulos (1984) - and the average epicentral distance with respect to ARG station, have been calculated.

For the purpose of avoiding extreme values, due to reading and calculation errors of travel time, epicenter and focal depth, in each center of gravity of focal depth (h) and epicentral distance (Δ) the average travel-time residual

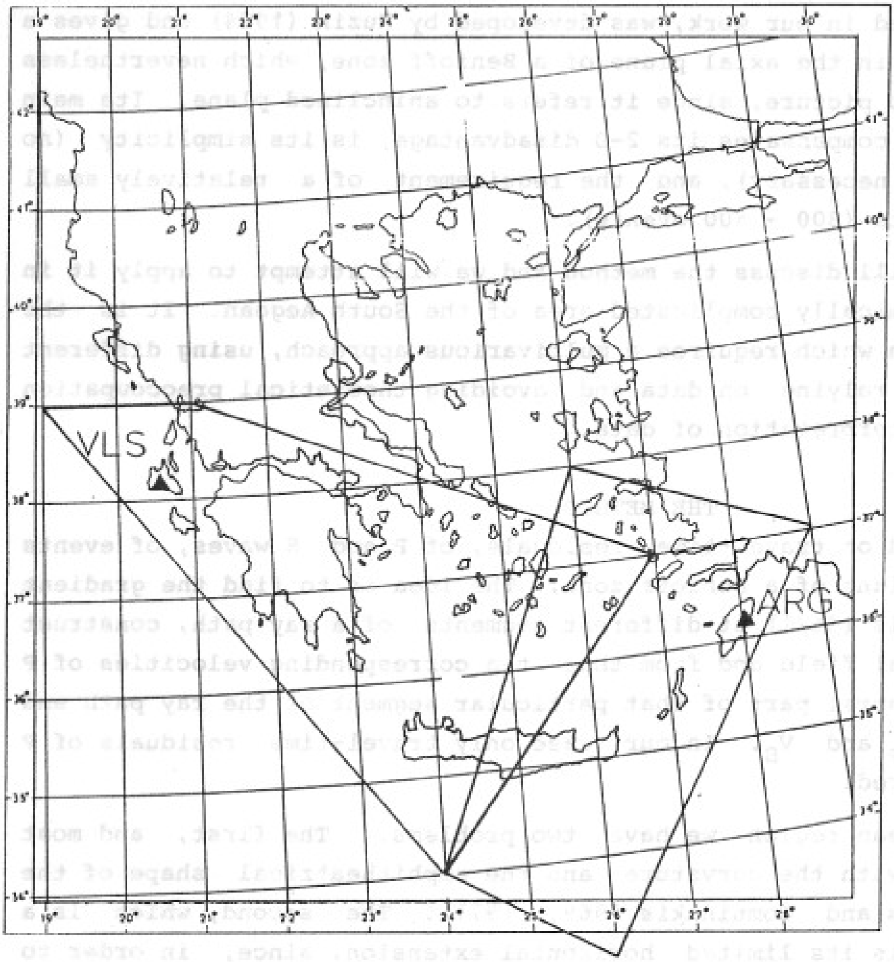

Fig. 1. Map of Greece and surrounding area showing the two polygons used in this study, one referring to the VLS and another to the ARG stations.

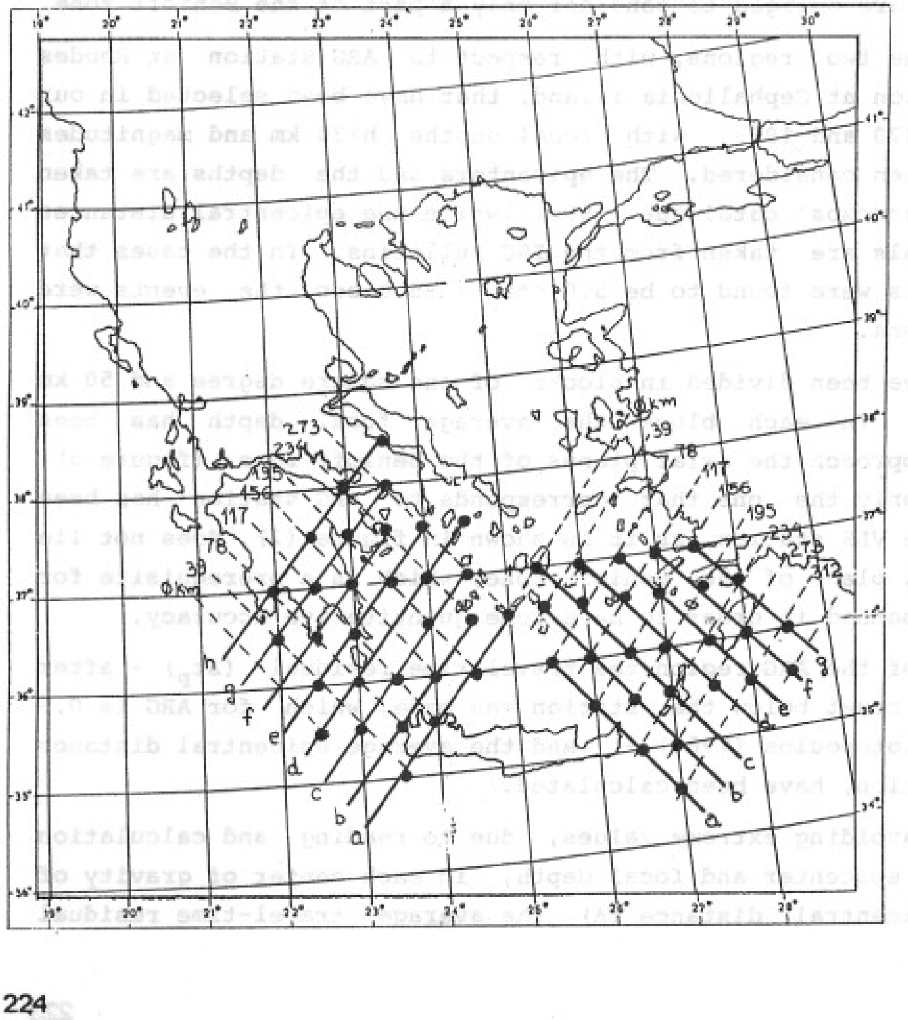

Fig. 2. Map showing the center of gravity of each block, to which the average focal depth, the average epicentral distance and the average travel-time residual have been assigned. It is also shown that VLS does not lie at the top of the axial plane of the Benioff zone.

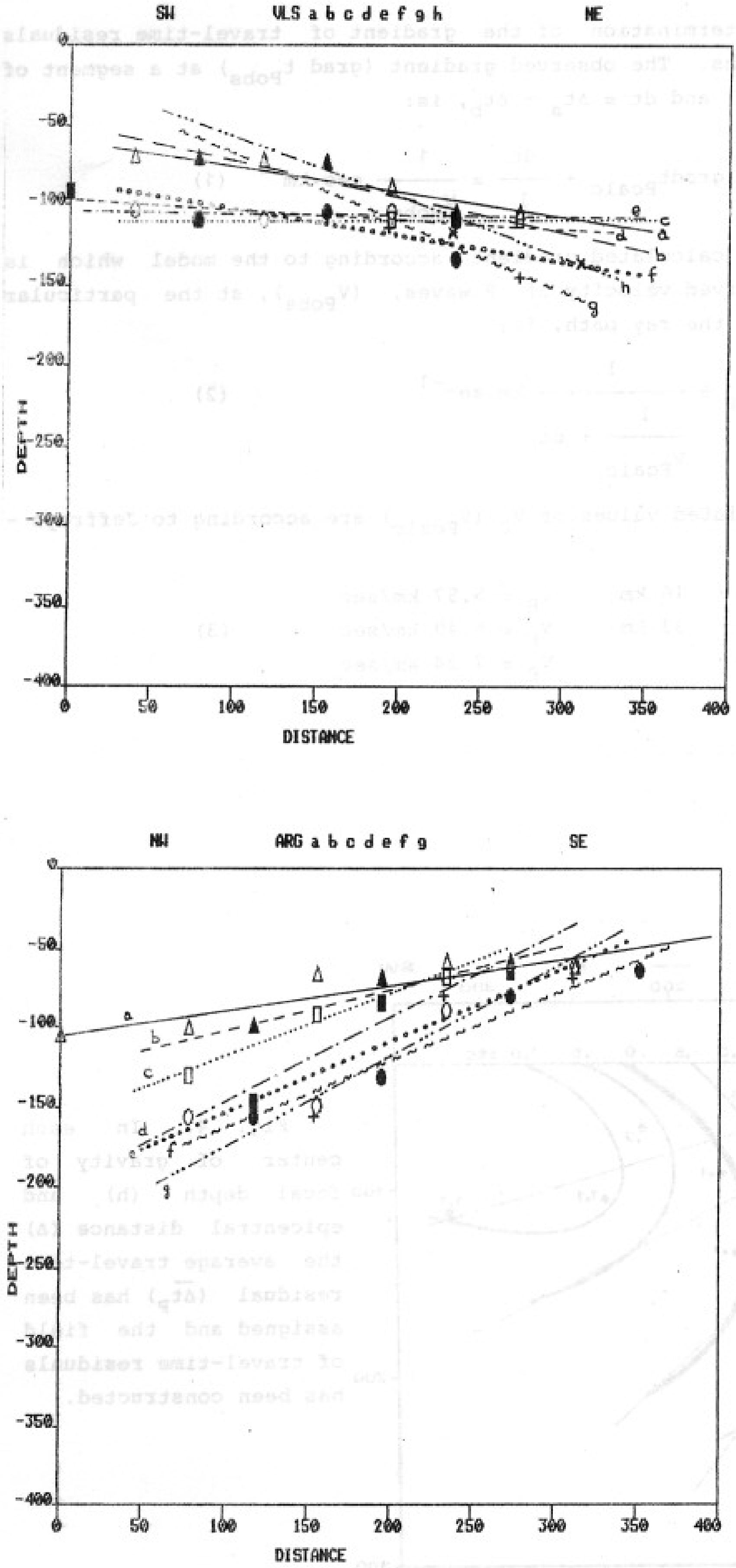

Fig. 3. Sections referring to the VLS and ARG stations, which are designed using the average focal depths. It is shown the SW-NE and the SE-NW inclination of the axial plane of the Benioff zone at SE and SW Aegean region respectively.

$(\overline{\Delta t}_P)$ has been assigned and the field of travel-time residuals has been constructed (Fig. 4).

Next step is the determination of the gradient of travel-time residuals along segments of ray-paths. The observed gradient (grad t_{Pobs}) at a segment of a ray path with length (l) and $dt = \Delta t_a - \Delta t_b$, is:

$$\mathrm{gradt}_{Pobs} = \mathrm{gradt}_{Pcalc} + \frac{dt}{l} = \frac{1}{V_{Pobs}}\ \mathrm{sec \cdot km^{-1}} \qquad (1)$$

where gradt_{Pcalc} is the calculated gradient according to the model which is being used, and the observed velocity of P waves, (V_{Pobs}), at the particular segment of length (l) of the ray path, is:

$$V_{Pobs} = \frac{l}{\dfrac{l}{V_{Pcalc}} + dt}\ \mathrm{km \cdot sec^{-1}} \qquad (2)$$

In our case the calculated values of V_P (V_{Pcalc}) are according to Jeffreys - Bullen model, which is

$$\begin{aligned} &0\ \mathrm{km} < h < 16\ \mathrm{km} && V_P = 5.57\ \mathrm{km/sec} \\ &16\ \mathrm{km} < h < 33\ \mathrm{km} && V_P = 6.49\ \mathrm{km/sec} \qquad (3) \\ &33\ \mathrm{km} < h && V_P = 7.24\ \mathrm{km/sec} \end{aligned}$$

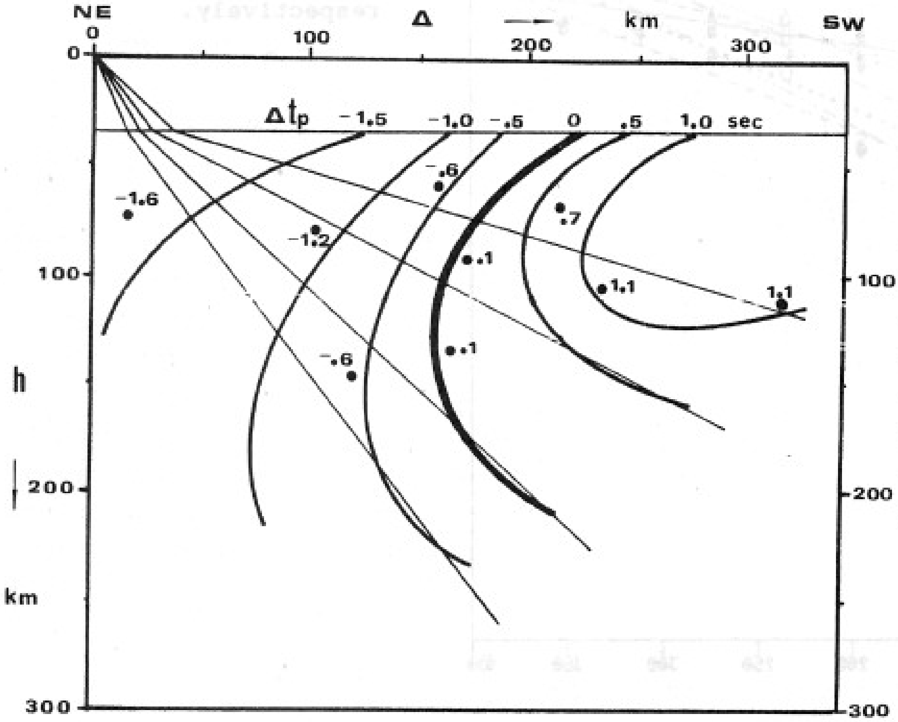

Fig. 4. In each center of gravity of focal depth (h) and epicentral distance (Δ) the average travel-time residual $(\overline{\Delta t}_P)$ has been assigned and the field of travel-time residuals has been constructed.

If the observed difference of travel-time residual (dt) between two points of a particular ray path is equal to zero, then V_{Pobs} is equal to the velocity of the model, which is being used.

RESULTS AND DISCUSSION

The 2-D velocity structure of P-waves in the axial plane of the Benioff zone in the SE part of the South Aegean region (Fig. 5), indicates the existance of two geodynamic regimes, as it has also been proposed by Tassos (1983). In the first one, at the northern and central part of the area, a high velocity zone of 8.5 - 9.0 $km{\cdot}sec^{-1}$ is surrounded by lower velocities down to 7.0 $km{\cdot}sec^{-1}$. The thickness of this high velocity zone is of the order of 20 km. The second one, at the southern part of the area, has an opposit structure; a low velocity zone (V_P = 6.0 - 6.5 $km{\cdot}sec^{-1}$) in the center is surrounded by a layer of $V_P \geq$ 7.0 $km{\cdot}sec^{-1}$.

The first regime has the characteristics of a lithothermal system of relatively low velocity $V_P \sim 7.5$ km/sec, ascending from the NE, which as a result of mechanical compression acquires a velocity of V_P = 8.5 - 9.0 km/sec at depths between 50 and 100 km. This high velocity zone coincides with the area of maximum Bouger gravity anomaly of +175 mgal (Makris, 1976) and high heat flow values in the order of more than 2 HFU (Jongsma, 1974).

In the south, we have a second regime, with the characteristics of crustal material (V_P = 6.0 - 7.0 km/sec), dipping to the NNW, thus exerting mechanical

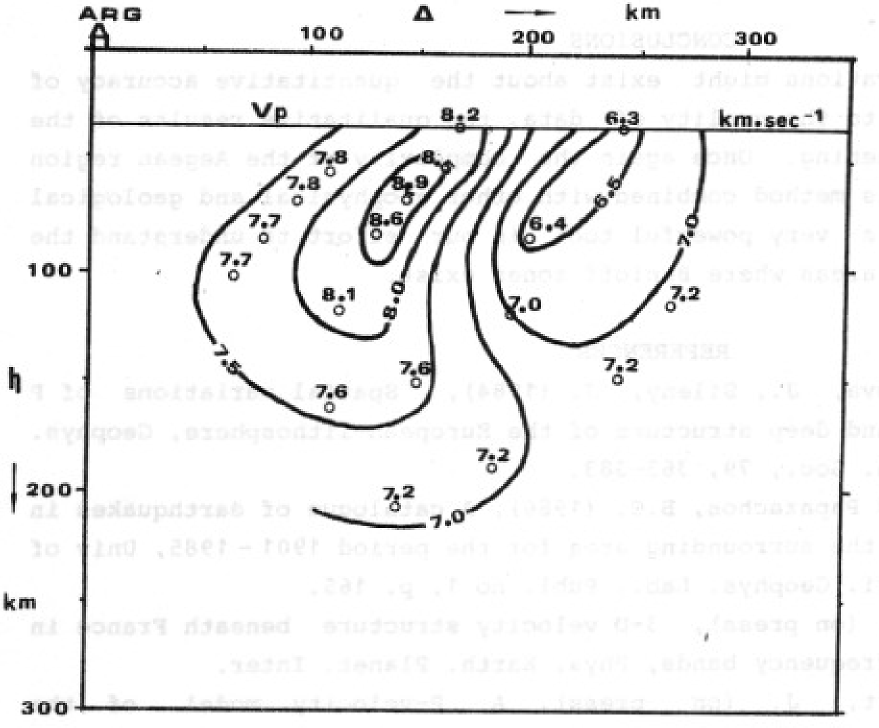

Fig. 5. After the construction of the travel-time residual field and the determination of the gradient of travel-time residuals along segments of ray path, the field of V_P is made. This indicates the existance of two geodynamic regimes in the SE part of the South Aegean region.

compression on the lithothermal regime to the north. It is at the boundary c the two regimes, where most of the events with focal depths less than 100 k occur. Below 100 km and down to 120 km we have an aseismic zone, as it was als noted by Soloviev et al. (on press). This aseismic zone seems to be relate with the area where the bending of the ascending lithothermal system take place, and it is a result of mechanical decompression and reduced rigidity, du to high temperatures, at that particular depth. A cluster of events appears a an average depth of 150 km and apparently is related with the ascending regime

McKenzie (1970) has found directions of relative movements betweeen th African and Aegean plates, which are similar to those of the ascendin lithothermal system from NE, for the Aegean plate, and to the descendin lithosphere from SSE down to a depth of about 100 km, for the African plat (Fig. 6).

As a result of that we have a compressional and a strike-slip effect at th boundary of the two regimes, and the formation of differently dipping Beniof zones, which has also been found by Gregersen (1977) for that particular area.

Figure (7), which is a cross-section along line d is a schemati representation of the proposed interpretation.

Line A, represents the average dip of cross sections a through g of Figur (3) and corresponds to the axial plane of the Benioff zone, on which we wer based for our computations. Line B, represents the average dip of lines . through d and corresponds to the descending African lithosphere, while line C represents the average slope of cross sections d through g and corresponds t the ascending lithothermal system. Line D, with the greatest dip, represent the site where compression and strike-slip motion take place and high velocitie of V_P = 8.5 - 9.0 km/sec are calculated. This is also the site, where due t the contact of the ascending hot mantle with the descending lithosphere we hav the generation of andesitic volcanism along the volcanic arc of SE Aegean.

CONCLUSIONS

Although some reservations might exist about the quantitative accuracy o the results, mainly due to the quality of data, the qualitative results of th method are quite enlightening. Once again the complexity of the Aegean regio has been proved, and this method combined with other geophysical and geologica data and methods can be a very powerful tool in our effort to understand th geodynamic properties of areas where Benioff zones exist.

REFERENCES

Babuska, V., Plomerova, J., Sileny, J. (1984), Spatial variations of P residuals and deep structure of the European lithosphere, Geophys. J.R. Astron. Soc., 79, 363-383.

Comninakis, P.E. and Papazachos, B.C. (1986), A catalogue of earthquakes i Greece and the surrounding area for the period 1901 - 1985, Univ of Thessaloniki, Geophys. Lab., Publ. no 1, p. 165.

Granet, M., Cara, M. (on press), 3-D velocity structure beneath France in different frequency bands, Phys. Earth. Planet. Inter.

Granet, M., Trampert, J. (on press), A P-velocity model of the Euro-Mediterranean Area from a simultaneous iterative

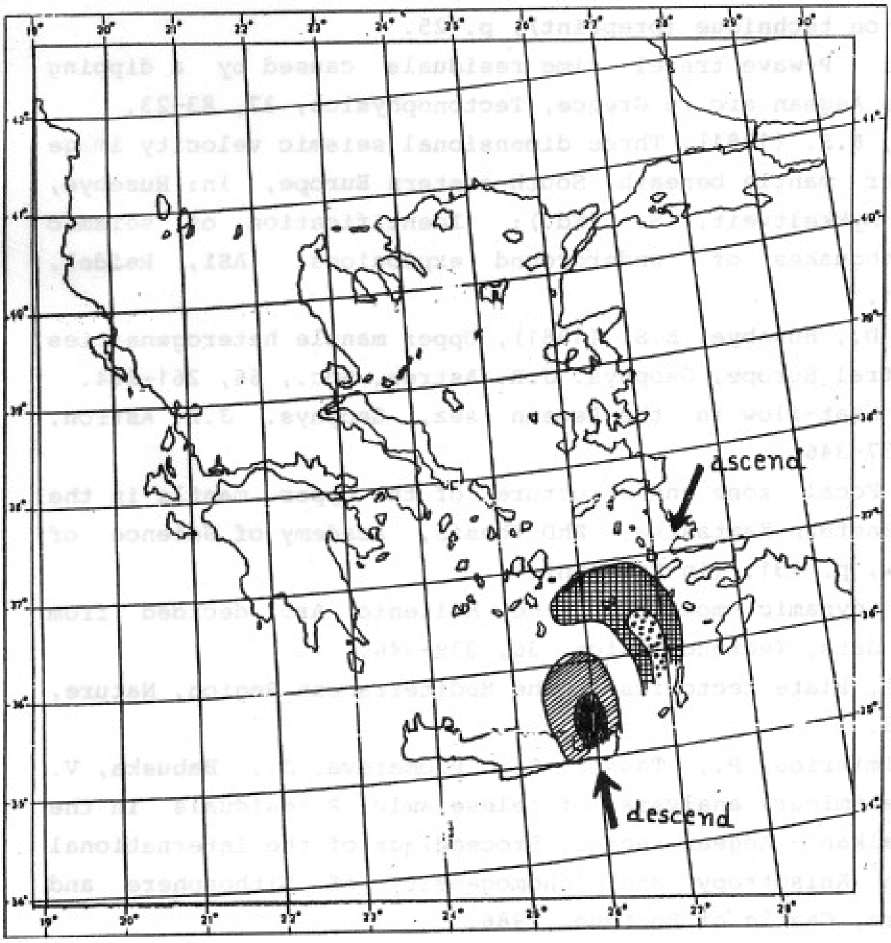

Fig. 6. High and low velocity zones on the axial plane of the Benioff zone in the SE Aegean region.

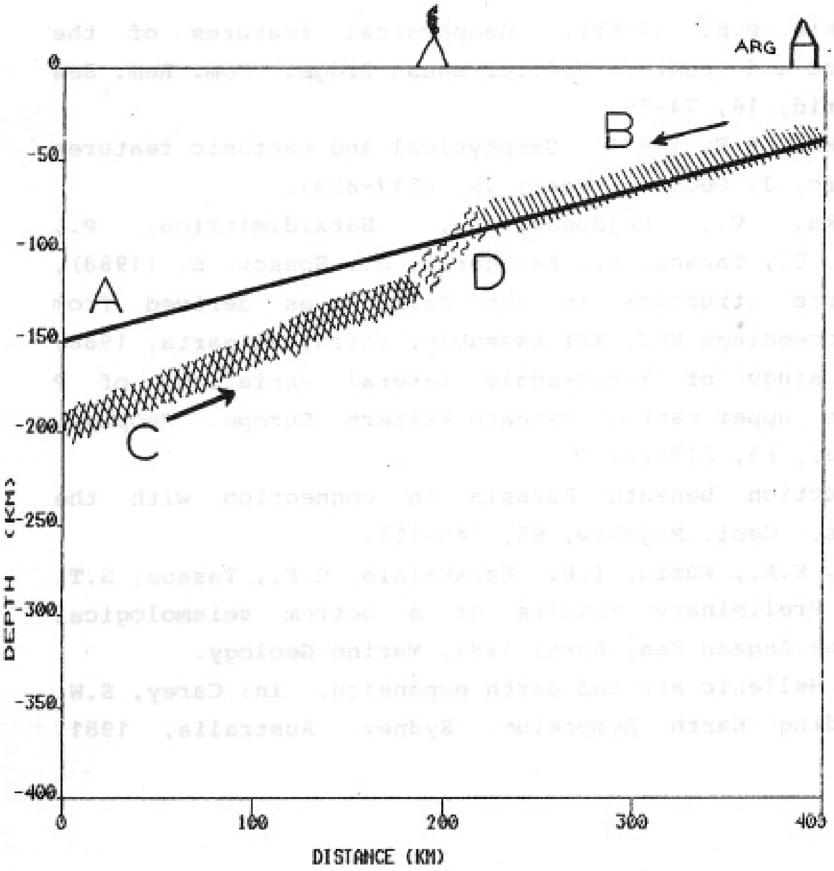

Fig. 7. Schematic representation of the geodynamic regimes in the SE Aegean region.

reconstruction technique (preprint), p. 25.

Gregersen, S. (1977), P-wave travel time residuals caused by a dipping plate in the Aegean arc in Greece, Tectonophysics, 37, 83-23.

Hovland, J., Husebye, E.S. (1981), Three dimensional seismic velocity image of the upper mantle beneath South-eastern Europe, In: Husebye, E.S. and Mykkeltveit, S. (Ed.): Identification of seismic sources-earthquakes of underground explosions, ASI, Reidel, London, 1981.

Hovland, J., Gubbins, D., Husebye, E.S. (1981), Upper mantle heterogeneities beneath Central Europe, Geophys. J.R. Astron. Soc., 66, 261-284.

Jongsma, D. (1974), Heat-flow in the Aegean sea, Geophys. J.R. Astron. Soc., 37, 337-346.

Kuzin, I.P. (1974), Focal zone and structure of the upper mantle in the region of Eastern Kamtsatka, PhD Thesis, Academy of Science of USSR, Moscow, p. 131, (in Russian).

Makris, J. (1976), A dynamic model of the Hellenic Arc decided from geophysical data, Tectonophysics, 36, 339-346.

McKenzie, D.P. (1970), Plate tectonics of the Mediterranean Region, Nature, 226, 239-248.

Pajdusak, P., Hatzidimitriou, P., Tassos, S., Plomerova, J., Babuska, V. (1986), Preliminary analysis of teleseismic P residuals in the Southern Balkan - Aegean region, Proceedings of the International Workshop on Anisotropy and Inhomogeneity of Lithosphere and Asthenosphere, Castle of Bechyne, 1986.

Panagiotopoulos, D.G., (1984), Travel time curves and crustal structure in the southern Balkan region, PhD. Thesis, Univ. of Thessaloniki, Thessaloniki, p. 173.

Papazachos, B.C., Comninakis, P.E. (1969), Geophysical features of the Greek island arc and eastern Mediterranean Ridge. Com. Rem. Sea Conf. Reun. Madrid, 16, 74-75.

Papazachos, B.C., Comninakis, P.E. (1971), Geophysical and tectonic features of the Aegean arc, J. Geophys. Res., 76, 8517-8533.

Plomerova, J., Babuska, V., Pajdusak, P., Hatzidimitriou, P., Panagiotopoulos, D., Tassos, S., Kalogeras, J., Spasov. E. (1988), Deep lithosphere structure in the Balkans as derived from P-residuals, Proceedings ESC, XXI Assembly, Sofia, Bulgaria, 1988.

Romanovicz, B. (1980), A study of large-scale lateral variations of P velocity in the upper mantle beneath Western Europe, Geophys. J.R. Astron. Soc., 63, 217-232.

Spakman, W. (1986), Subduction beneath Eurasia in connection with the Mesozoic Thethys, Geol. Mijnbow, 65, 145-153.

Soloviev, S.L., Kovachev, S.A., Kuzin, I.P. Karakaisis, G.F., Tassos, S.T. (on press), Preliminary results of a bottom seismological experiment in the Aegean Sea, April 1987, Marine Geology.

Tassos, S.T. (1983), The Hellenic arc and earth expansion, In: Carey, S.W. (ed).: Expanding Earth Symposium, Sydney, Australia, 1981, 161-164.