SUMMARY

5000 P-wave traveltime residuals from about 300 local and 400 teleseismic earthquakes recorded by the Athens seismological network were used to study the structure of the lithosphere in the Aegean region. Teleseismic data included readings from local stations which do not routinely report to the ISC. The large values of the residuals (up to ±3 s) and their variation with azimuth and epicentral distance indicate significant velocity heterogeneity in the upper mantle. P-wave traveltime residuals were inverted simultaneously for both local and teleseismic earthquakes. Tomographic results, which are best resolved at lithospheric depths, reveal a heterogeneous lithosphere down to a depth of 100 km or so. A high-velocity anomaly is found dipping at an average of 45° to the NE away from the Hellenic trench, and supports the hypothesis of subduction due to a cooler descending slab of oceanic composition. An area of low velocities is observed at depths less than 80 km in the Aegean back-arc, corresponding well with observed high heat flow values in the area. In contrast, an area of relatively higher velocity anomalies is resolved in the lithosphere underlying the Sea of Crete.

Key words: Aegean, P-wave traveltime residuals, seismic tomography.
chosen to be 30°–90°; the lower limit was set to avoid effects from the regional anomalies along the ray paths and the upper limit to avoid identification problems in the PKP range. The epicentral distance for the local events is 0°–7°, corresponding to a maximum penetration depth of about 80 km or so, for sub-horizontally propagating rays from the shallow earthquakes which dominate the seismicity. 

P-wave arrival times for teleseismic events were identified in a consistent manner, by a single observer (C.L.), from the original records and those for the local events were taken from the Athens Observatory archives. Teleseismic readings that are not routinely reported to the ISC were also included. Readings were taken from clear arrivals on short-period vertical components from local earthquakes greater than magnitude $m_b = 4$ and teleseisms greater than $m_b = 5$, and were weighted according to quality. The total reading error depends on the sharpness of the onset of the seismic pulse with respect to the background noise, and on the accuracy of the recording clock. Corrections for clock delays provided by the Athens Observatory network were also applied. The great majority of the readings used in the present work had uncertainties much less than ±0.3 s. The estimated traveltimes were also corrected for ellipticity effects (Dziewonski & Gilbert 1976), but no corrections were made for station elevation, the effect of which was estimated to be less than ±0.1 s.

P-wave traveltime residuals, defined as the difference between the observed traveltimes of the seismic wave and the computed traveltimes from a reference model, were calculated from the Jeffreys-Bullen (J–B) traveltime tables (Jeffreys & Bullen 1940), using standard ISC procedures. Residuals greater than ±4 s for teleseismic results were discarded in an attempt to avoid large errors in input data arising from unilaterally located events and observational errors. In most cases, the observed residuals were found to be much higher than the combined systematic and random reading errors (conservatively estimated to be ±0.3 s). The use of a single operator to read the teleseismic data reduces any random component of the error in reading residuals compared to that from the different stations that report to the ISC. Finally, more than 5000 station readings (2000 for local and 3000 for teleseismic events) were found to be of sufficient quality to be used in the tomographic inversion.

3 P-WAVE TRAVELTIME RESIDUALS

As the seismic wave travels from the source to the recording station through different parts of the Earth, many factors can affect its propagation.

Traveltime residuals can be ascribed to source effects, receiver effects, intermediate-path effects or any combination of these. However, our main objective was to examine the deviations from the J–B model due to heterogeneities of the lithosphere at the sub-receiver region. Teleseismic rays illuminate the deep mantle structure but the resolution in tomography diminishes with depth when teleseismic data are used. The interpretation of the residuals for the tomographic problem also includes the identification of all significant sources of systematic and random errors. In order to complement earlier work using the same tomographic technique for the whole of Europe (Spakman et al. 1988), a combination of teleseismic and local P-wave residuals from the Athens Observatory network was used. This guarantees good illumination of mid-lithospheric structure because of the near orthogonality of the teleseismic (vertical) and local (sub-horizontal) ray paths at these depths. The resolution of the tomographic results is critically dependent on the degree of intersection of crossing rays, and in this study the maximum illumination was specifically targeted at mid-lithospheric depths in the Aegean area. Residuals were also studied with respect to the variations of the mean values at each station and their dependence on azimuth.

There are many ways of addressing the problem of distinguishing between the different contributing parts in the...
overall residual values. Teleseismic P-wave residuals, irrespective of the direction of arrival of the rays, are often found to be related to the velocity structure in the underlying upper mantle rather than to deeper anomalies (Cleary & Hales 1966; Johnson 1967, 1969; Bott 1982). Moreover, the azimuthal dependence of residuals greater than several tenths of a second cannot be solely due to crustal heterogeneities (Spence 1974; Raikes 1980; Dzie-wonski & Anderson 1983; Babuska, Plomerova & Sileny 1984). Seismic rays from teleseismic events reach the stations with a very steep angle of incidence, and, therefore, a significant part of the overall residuals at the study area is acquired from the uppermost mantle underlying the station network. Nevertheless, source and intermediate-path effects cannot be ignored. The effect of structural changes with depth in the source area can be minimized by using events with a variety of focal depths (Raikes 1980). In tomography, deep mantle heterogeneities along intermediate paths are absorbed, up to a certain extent, by the estimation of event mislocation (Spakman & Nolet 1988). When local or regional events are used, P-wave residuals again are also often larger than that expected from variations in crustal structure alone, and also imply upper mantle heterogeneities (Nuttli & Bolt 1969; Kind 1972; Suyehiro & Sacks 1979).

A few more comments should be made on the effects of crustal structure. In this present work, crustal effects are considered to be small, since it is estimated that for every 1 km of crustal loss (excess) an average residual of (+)0.04 s is required for a typical crustal velocity of 6 km s⁻¹ and upper mantle velocity of 8 km s⁻¹. Therefore, large absolute values of the inferred residuals and, most important, azimuthal variations with epicentral distance indicate significant upper mantle heterogeneities. The ‘shallow’ structure illuminated by this study is therefore mainly on the scale of lithospheric rather than crustal heterogeneity. Any future studies, concerning specifically the crust, will require the use of even smaller aperture microseismic networks and much better understanding of the sedimentary cover in the whole of the Aegean area than exists at present.

Mean station values of P-wave residuals

The mean values of P-wave residuals at each station for both local and teleseismic events are shown in Table 1. Negative and positive values should be considered as an indication of relatively higher or lower velocities respectively, and mainly attributed to the anomalous structure at the sub-receiver region. A strictly quantitative approach in examining these values is not appropriate because of the problems introduced in estimating the residuals. For local events, the lowest residuals, that is early arrivals, (higher than normal velocities) are observed at PLG (Chalkidiki peninsula), PRK (north Aegean), and NPS (Crete). It should be noted that the absolute mean value for local residuals at each station may be biased by the azimuthal distribution of the local seismic rays, particularly for the stations located at the northern and eastern edges of the network (KZN, PLG, PRK, ARG). This occurs in part because of the lack of available local earthquakes to the north or east of these stations. For teleseismic events, negative residuals are observed at stations in northern Greece, part of western Greece and the central Aegean. In contrast to local mean station residuals, these values reflect local rather than intermediate-path or source anomalies because of the similarity of the ray paths reaching the stations. It is difficult however, to distinguish between very shallow and deeper effects at the sub-receiver region.

From the study of the mean station residuals two main observations can be made.

(a) The mean station residuals for most of the stations are not in general consistent for both local and teleseismic events, and, despite the ambiguity of the local mean values, indicate that they do not solely depend on near-station effects. This implies relative differences in the shallow and deep structure under the stations.

(b) There is a large contrast between the two stations located on Crete (VAM, NPS) for both teleseismic and local residuals, with VAM being the one with the more positive values (lower velocities). The difference is preserved down to at least 200 km (as it can be seen later on in the tomographic results) and also cannot be solely explained by station site anomalies at shallow depths.

In summary, the average residuals cannot be ascribed solely to site effects or variations in crustal structure, and hence reflect significant heterogeneity in the upper mantle.

Variations of teleseismic residuals with azimuth and epicentral distance

In this section a more systematic study of residual variations with azimuth is described. Fig. 2 shows polar diagrams of
azimuth and epicentral distance at each station. Teleseismic P-wave residuals are plotted against azimuth and epicentral distance. Again, it should be noted that the marked variations with azimuth (up to 2s in certain regions), as well as the absolute magnitude of the residuals, confirm the existence of significant upper mantle heterogeneity. Moreover, if azimuthal variations were due solely to crustal effects, large variations of crustal thickness would be required to occur over very short distances, of the order of the station separation. This is most unlikely to happen according to the observed crustal thickness’ variations over the Aegean area (Makris 1973). The example of the two stations on Crete, with almost identical crustal thicknesses (Makris 1973) but completely different average residuals, is perhaps the most striking.

Azimuthal variations of the stations located on or near the sedimentary Hellenic arc (JAN, VLS, RLS, ITM, VAM, NPS, ARG) clearly suggest that all waves arriving from the inner part of the arc, to the north and east, are fast relative to those coming from the outer part, to the south and west. This systematic orientation is compatible with the existence of a high-velocity zone dipping to the NE in an 'amphitheatre shape' under the Hellenic trench (Agarwal, Jacoby & Berckhemer 1976; Gregersen 1977; Le Pichon & Angelier 1979, 1981; Spakman 1986). This effect may be due to the predominance of available teleseismic ray paths, from the north and east, that travel up the high-velocity subducting lithosphere. Because of their small angle of incidence, it is most unlikely that the ray paths for the stations at the sedimentary arc cross the upper mantle under the thinned crust of the Aegean, which might be expected to have low velocities.

4 TOMOGRAPHIC INVERSION OF P-WAVE TRAVELTIME RESIDUALS

Having established some systematic deviations from the J–B reference model in the upper mantle of the Aegean area, the seismic data were inverted by tomographic techniques in order to illuminate the heterogeneity further. The inversion algorithm that was used is known as the 'LSQR: An algorithm for sparse linear equations and sparse least squares' (Paige & Saunders 1982), which has been tested and confirmed for seismic data by Nolet (1985), Spakman (1988) and Spakman & Nolet (1988).

Initially, the study area was divided into 10 layers—one ‘crustal’ and nine ‘upper mantle’ layers (Table 2)—which were subdivided in 1° x 1° cells in plan (Fig. 3). The conventional Earth's coordinate system was rotated with a new origin at 38°N, 24°E, to produce horizontal cell dimensions of approximately 111 x 111 km. The depth range of the cells is not uniform. This particular pattern was chosen to enable the most significant heterogeneities of the uppermost mantle to be detected within the resolution of the method.

Some characteristics of the inversion method [for details of the method used see Spakman & Nolet (1988)] are as follows.

(a) Event and station corrections were used. This means that the residuals can be interpreted by factors other than the lateral heterogeneity of the Earth. Even though this approach has the risk of increasing the number of poorly constrained unknowns (such as these particular corrections are) in the tomographic equations, it can significantly improve the reliability of the final tomographic images. The event corrections not only reflect a strong physical requirement, but are also used to absorb large mislocation errors and the influence of mantle heterogeneities outside the cell model (Spakman 1988).

Table 2. Cell depth layering.

<table>
<thead>
<tr>
<th>Depth layering</th>
<th>Mean Velocity</th>
</tr>
</thead>
<tbody>
<tr>
<td>km</td>
<td>km s⁻¹</td>
</tr>
<tr>
<td>&lt;33</td>
<td>6.064</td>
</tr>
<tr>
<td>33-66</td>
<td>7.799</td>
</tr>
<tr>
<td>66-110</td>
<td>7.975</td>
</tr>
<tr>
<td>110-170</td>
<td>8.091</td>
</tr>
<tr>
<td>170-245</td>
<td>8.282</td>
</tr>
<tr>
<td>245-320</td>
<td>8.526</td>
</tr>
<tr>
<td>320-400</td>
<td>8.786</td>
</tr>
<tr>
<td>400-500</td>
<td>9.277</td>
</tr>
<tr>
<td>500-600</td>
<td>9.958</td>
</tr>
<tr>
<td>&gt;600</td>
<td>10.404</td>
</tr>
</tbody>
</table>
The number of ray paths crossing each of the cells is important in determining how well the cell solutions are constrained, although the degree of crossing (independence) of the ray paths is much more significant for the spatial resolution in the cell model (Spakman 1988). In this present work, a reasonable sampling of the cell model with various ray paths, in terms both of number of rays and their intersection, was expected for lithospheric depths (<80 km) by the good geographical distribution of epicentres and the combined use of both local (sub-horizontal propagation) and teleseismic (sub-vertical) events.

The 'cell hitcount' for two cross-sections (N–S and SW–NE) can be seen in Fig. 4 (a and b respectively). The contouring indicates the grade of ray illumination of the cells solely in terms of the number of rays intersecting a cell. The true degree of illumination however, is also dependent on the angles of intersection of the crossing ray paths in each cell and the best achievable azimuthal range within which the rays are distributed (e.g. Walck & Clayton 1987). Hence, the best resolution is obtained at lithospheric depths, where the shallow zone in black in Fig. 4(a) and (b) corresponds to more than 225 rays crossing each cell, approximately half sub-vertically and half sub-horizontally, and with a good distribution in azimuths. Below this depth (100 km or so) the sampling of the cells with rays decreases significantly, except for a narrow zone in the central part of the cell model (under mainland Greece and central Aegean Sea). The predominance of available teleseismic rays from the north and east also affects the quality of the deeper parts of the tomographic images. For example, the elongation of the 'hitcount' contours to the north and east, below 120 km or so (Fig. 4a and b), is due to this non-uniform illumination of the cells. This preference in ray illumination raises the question of direction bias of the tomographic images at these depths. Artifacts, as we shall see later on, can be observed at the outer parts of most tomographic pictures, where essentially there are insufficient data to resolve the velocity structure. This lack of resolution at the edges is a common feature of most tomographic inversions.

Figure 4(c, d, e and f) shows horizontal views of the 'hitcount', with the best sampled layers corresponding to the black shaded areas. For mid-lithospheric depths (33–66 km) the coverage (in terms of number of events and the degree of intersection) is very good for mainland Greece, Crete, the sea of Crete, the volcanic arc and most of the Aegean basin. However, the 'hitcount' results for deeper layers and the crustal layer show evidence of station bias, with the number of rays heavily dependent on the presence of a station directly above (cf. figure beneath Table 1). This is particularly serious for the deeper layers (>110 km), where the area of high 'hitcount' underneath the stations implies near-vertically travelling waves through the cells, and hence a low orthogonality of ray paths. This severely diminishes the resolution at these depths, and hence any new results from the present work are strictly in terms of lithospheric structure at depths of 33–110 km, in mainland Greece and the southern and western part of the Aegean Sea.

A cell-spike test, as described in Spakman (1988), was also carried out to further assess the resolution of the final tomographic image. A cell-spike model is a model of zero slowness, except for certain cells which are attributed
the same slowness values. In this work, a 5 per cent velocity contrast cell-spike model consisting of spikes placed at the centre of specific cells, was used (Fig. 5). The maximum detail that can be resolved with the LSQR algorithm in this particular study can be tested by carrying out a model inversion. A model delay vector is obtained by using the ray-path matrix for the *a priori* known 'ideal' solution of the spike model slowness anomalies. The solutions of the inversion (cell-response functions), which uses the spike delay vector, are then compared to the initial 'model' solutions to obtain an estimation of the resolution of this particular spike model. For the data set used in this study, they were found to be quite sharply peaked around the cells, known to contain the anomalies, down to a depth of...
110 km. This implies that the algorithm is capable of resolving velocity anomalies of this magnitude (5 per cent) to a length scale within the dimensions of each cell at mid-lithospheric depths. The spike-test results, when a realistic amount of normally distributed noise is added to approximate the test to the real data case, show that only about 30 per cent of the input spike amplitude is recovered in the SE Aegean, but up to 75 per cent in mainland Greece can be resolved for depths between 33 and 110 km. Resolution is significantly poorer in the NE Aegean area within these depth limits. These results can be seen in Fig. 5 for the first four horizontal layers. The thick-lined circles indicate the position of the spikes. Note the low degree (10 per cent) of recovering of the spike in central Aegean for layer 1 and the ‘leaking’ (i.e. anomalies outside the pre-defined spikes, their position implying bias from the layers below and above) in the same area for depths greater than 110 km. The half-width of the cell response function is taken as a measure of spatial resolution and for the volume, defined by black on Fig. 4(d), this was found to be less than, but of the same order as, the cell size.

Tomographic results were also tested for a slightly shifted cell model and did not indicate any differences from the model used initially. This confirmed that the results are quite independent of the positioning of the cells and the spatial resolution at lithospheric depths is less than, or of the order of the cell size.

In conclusion, the three tests, outlined above, indicate sufficient resolution (within ±1° in plan) and good confidence in the results for mid-lithospheric depths, except for the far NE Aegean. The systematic use of local stations, which do not routinely report to the ISC, and a single operator to read the data imply that the final inversion of these data for these particular depths is a significant improvement on the earlier work of Spakman et al. (1988), whose study was based solely on ISC data. Spakman et al. (1988) used 500,000 ray paths, mainly of regional and teleseismic events, for the whole of Europe and the present study was based on 5000 ray paths of local and teleseismic events, concentrated on Greece and the Aegean Sea. However, the present study was specifically focused on the volume outlined in black on Fig. 4(d), and a careful optimization of the data quality was obtained in terms of systematic reading of arrival times and a maximum achievable orthogonality of ray paths for this volume.

**Tomographic results**

Tomographic results, using the LSQR algorithm, are presented in Figs 6 (layers 1–10) and 7 (a and b) for horizontal and vertical sections respectively.

**Horizontal layers** (Fig. 6) are presented for 10 different interfaces of depth corresponding to those of the cell model. The uppermost layer (‘crustal’ layer) is not particularly well
Figure 6. The inversion results for horizontal layers obtained with the LSQR algorithm. The shading denotes velocity anomalies in percentages relative to the Jeffreys–Bullen ambient mantle velocity. Low- (high-) velocity regions are (cross) hatched.
resolved mainly because of:

(i) the station bias and and the non-uniform illumination of the cells at the very top of the model, and
(ii) the near-vertical incidence of all rays close to the surface.

At this shallow layer relatively low velocities are found in Peloponnesos and the south Aegean, in contrast to higher velocities in the north Aegean and the rest of mainland Greece. This distribution is likely to be the result of local station effects rather than a consistent crustal pattern.

The tomographic pattern is best resolved in the mantle lithosphere (layers 2 and 3), but the resolution decays rapidly thereafter. The elongation of 'hitcount' contours to the north and east with depth (Fig. 4a and b), below 110 km, could contribute partly to the existence and the shape of the anomalies dipping to the north and east below this depth. The outer parts of all the tomographic pictures reflect instabilities of the solution or contouring inadequacies and are not real anomalies. These parts correspond to the areas that are very poorly sampled, primarily outside mainland Greece and the Aegean Sea, and at depths below 320 km.

A marked feature of these results is the high-velocity anomaly, under the Ionian Sea, south Peloponnesos and the south Aegean, moving NE with depth, although the penetration depth is unknown. This high-velocity zone is consistent with the results of Spakman et al. (1988), but is not as well resolved in the present study for reasons discussed previously. Above this high-velocity anomaly, a pronounced well-resolved area of low-velocity anomalies is observed in the mantle lithosphere under the central Aegean Sea and northern Greece. Relatively higher velocities than those in the volcanic arc and back-arc region are observed in the Sea of Crete for depths between 33 and 66 km.

The N–S and SW–NE cross-sections (Fig. 7) present a different view of the tomographic results. In the N–S cross-section a well-constrained heterogenous lithospheric structure is revealed at depths above 100 km or so beneath the Aegean after contouring. A high-velocity zone is observed with a dip angle of about 30° for the upper 170 km, agreeing with that of the observed Benioff zone (Papazachos & Comminakis 1971; Richter & Strobach 1978). A change in slope to a steeper dip of 45° is apparent from the diagram, but again the degree of confidence for such an observation is low. In contrast, a well-resolved low-velocity anomaly beneath the actively stretching Aegean basin is also observed in Fig. 7(a). At shallow depths, high velocities relative to the J–B model can be seen under northern Greece and low velocities under western Greece.

The tomographic results described above undoubtedly indicate a heterogeneous lithosphere in agreement with the azimuthal variations of P-wave residuals presented in Section 3.

The results show a high-velocity dipping structure, compatible with the existence of a subduction zone, dipping to the NE away from the Hellenic trench. They also indicate that the slab is curved in plan, in a similar sense to the Hellenic arc. The use of teleseismic ray paths allows the
the existence of a cooler and hence more dense, descending given elsewhere (Ligdas with other thermal and geophysical fields in the Aegean is consistent with an area of lower heat flow values (Spakman et al. 1988; Granet & Trampert 1989), or tomographic results of a larger cell model scale (Howland & Husebye 1981).

The following observations can be made with respect to the thermal conditions.

(i) The dipping high-velocity anomaly is consistent with the existence of a cooler and hence more dense, descending lithosphere (Davies & McKenzie 1969; Sleep 1973), which sinks under its own weight into the surrounding viscous mantle material.

(ii) The low-velocity anomaly in the Aegean back-arc correlates very well with an area of high heat flow (measurements indicate values from 1.64 to 2.73 HFU) occurring predominantly behind the volcanic arc (Jongsma 1974; Erickson, Simmons & Ryan 1977; Fyticas & Kolios 1979).

(iii) Similarly, higher velocities in the sea of Crete are consistent with an area of lower heat flow values (1.42–1.61 HFU, according to the same studies mentioned above).

The tomographic velocity reconstruction and the heat flow data are also compatible with a 3-D attenuation structure presented by Hashida, Stavrakakis & Shimazaki (1988), if areas of high (low) Q are correlated with high (low) velocities and low (high) temperatures. According to this study, low Q, down to a depth of 40 km, is dominant in the Aegean Sea, and a high-Q zone in the upper mantle, dipping to the NE is dominant along the Hellenic arc. Another interesting observation in this work is the area of relatively high Q values in the sea of Crete at shallow depths. Note that this latter area defines a zone of quiescent seismicity this century (Makropoulos & Burton 1981) and has the thinnest crust (Makris 1973).

A more detailed correlation of the tomographic results with other thermal and geophysical fields in the Aegean is given elsewhere (Ligdas & Main 1990, in preparation).

CONCLUSIONS

Seismic results, from the study of P-wave traveltime residuals in the Aegean region, confirm a heterogeneous upper mantle structure. A high-velocity anomaly dipping to the NE (with an average dip angle of about 45°) is interpreted, as in previous works, as a cold, dense subduction slab of oceanic composition. Relatively low velocities, consistent with high heat flow, are found in the mantle lithosphere, behind the volcanic back-arc region, and lower velocities are seen in the fore-arc region.

The tomographic inversion adequately resolves the signs and general trends in the velocity structure at lithospheric depths, but the geometry of the inferred anomalies should be examined with reference to the variable resolution of the method in different parts of the tomographic images. In particular, the resolution of the present work is inadequate to resolve crustal heterogeneity or to resolve any but the broadest features below the mantle part of the lithosphere. However, the upper mantle lithosphere in the area of mainland Greece and the Aegean Sea is well resolved in the present work, and the tomographic velocity structure in the whole of the Aegean Sea area is in very good agreement with the observed heat flow values.

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