Surface waveform tomography of the Turkish–Iranian plateau

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SUMMARY
The Middle East is a tectonically complex region consisting of terrains as diverse as Precambrian shields and young orogens. The dominant tectonic feature is the Turkish–Iranian plateau, a recently elevated, seismically active region along the Zagros–Bitlis suture, which results from the collision of the Arabian Plate with Eurasia. In this study, we use surface waveform tomography to elucidate the upper-mantle shear wave velocity structure beneath the Turkish–Iranian plateau and adjacent regions. The main large-scale feature in the tomographic model is a low shear wave velocity anomaly in the uppermost mantle beneath the plateau. This low-velocity feature correlates with a long-wavelength free-air gravity anomaly and with recent volcanism whose geochemistry has a lower lithospheric mantle signature. Seismology, gravity and volcanism all suggest the presence of a thin lithosphere and warm upper mantle beneath the Turkish–Iranian plateau. The upper-mantle low shear wave velocity zone, the high free-air gravity and the deep lithospheric source depth for the basaltic volcanism are all consistent with a partial delamination of the lithosphere as a result of earlier lithospheric thickening resulting from the continental collision.

Key words: low-velocity upper mantle, partitioned waveform inversion, surface wave tomography, Turkish–Iranian plateau.

1 INTRODUCTION
The Middle East is a tectonically complex region with a variety of crustal types and tectonic styles (Fig. 1). The most important tectonic feature is the high, young topography in the seismically active zone along the Zagros–Bitlis suture resulting from the collision of the Arabian Plate with Eurasia (e.g. Sengor & Kidd 1979). The final closure of the Neo-Tethys and its marginal basins, as a result of the northward motion of Arabia towards Eurasia, occurred over the last 15 Myr and involved the accretion of island arcs and other continental fragments as well as the subduction of intervening portions of oceanic lithosphere beneath Eurasia. For the past 5 Myr, the collision between Arabia and the accreted terrains that now form eastern Turkey and Iran has been purely continental in nature (Berberian & King 1981; Hempton 1987). Active subduction still occurs to the south and southwest of Turkey, beneath the Cyprian and Hellenic arcs, respectively, and to the southeast of Iran, beneath the Makran.

The tectonic framework of the Middle East is well understood. The northward motion of Arabia with respect to Eurasia is partially accommodated by the westward motion of Turkey towards the Aegean on the East and North Anatolian faults (e.g. McKenzie 1972, 1978), and is also related to the opening of the Red sea and strike-slip motion on the Dead Sea fault (e.g. Hempton 1987). The existence of subcrustal (> 50 km) earthquakes in the ISC and USGS catalogues has led some to postulate active subduction of the continental crust of the Arabian shield beneath the Zagros (Nouroozi 1971; Bird et al. 1975; Moores & Twiss 1995). However, neither the modelling of teleseismic waveforms for moderate earthquakes (Baker et al. 1993; Maggi et al. 2000) nor local seismic network earthquake recordings in the Zagros (Tatar et al. 2004) have yielded any evidence supporting the existence of earthquakes deeper than ∼20 km. The northward shortening is instead taken up by a combination of thrust and strike-slip faulting (Talebian & Jackson 2002) and crustal thickening (Hatzfeld et al. 2003) in the Zagros mountains and in the Caucasus (Jackson 1992). In eastern Iran, the northward convergence is taken up by major right-lateral strike-slip faulting (Walker et al. 2003), which turns into left-lateral E–W strike-slip faulting and thrusting further north in Iran. The south Caspian basin, possibly of oceanic origin, is being overthrust by the Alborz to the south and the Talesh mountains to the west (Priestley et al. 1994; Jackson et al. 2002).

The seismic structure of the Middle East is less well known. There are few studies of the crustal velocity structure of the Iranian plateau. A refraction profile consisting of sparse recordings along a line from central Iran to the Strait of Hormuz (Giese et al. 1984) shows arrivals of questionable quality identified as Moho reflections and indicating a crustal thickness of 40 km beneath central Iran. Using receiver function analysis Hatzfeld et al. (2003) find the crust in the central Zagros to be 44–48 km thick. Using ∼10 000 gravity measurements and the seismic results of Giese et al. (1984), Dehghani & Makris

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(1984) find that the crust beneath the Main Zagros thrust is 55 km thick, that it is less than 40 km thick beneath the Lut depression, 45–48 km thick beneath the ranges of eastern Iran and ~35 km thick beneath the Alborz mountains in northern Iran. Snyder & Barazangi (1986) use similar gravity and seismic data to Dehghani & Makris (1984) and find the Moho depth increases from ~40 km beneath the Persian gulf to ~65 km beneath the Main Zagros thrust. Analysis of surface waves dispersion in Iran (Asudeh 1982b) implies Moho depths of 43–46 km.

No crustal seismic refraction profiles have been recorded in Turkey and the only measurements of the crustal structure come from analysis of earthquake data. Necioglu et al. (1981) find the crust in western Turkey is 25–32 km thick from travel time analysis of local and regional earthquake arrival times. Saunders et al. (1998) found from receiver function analysis that the crustal thickness beneath western Anatolia varies from 30 km in the region of western Turkey where active normal faulting reveals present-day extension, to 34 km farther east where extension is still evident but less pronounced, to 38 km beneath Ankara where there is no present-day extension. Mindevali & Mitchell (1989) analysed group velocity data from regional earthquakes recorded at Ankara and found the average crustal thickness and upper-mantle shear velocity to be ~40 km and ~4.2 km s⁻¹, respectively. Receiver function studies in eastern Turkey (Zor et al. 2003) find the average crustal thickness is 45 km. Mangino & Priestley (1998) and Priestley et al. (2001) find that the south Caspian basin is flanked by 40–50 km thick continental crust, but that beneath the basin the crust is ~33 km thick and consists of ~20 km of extremely low shear wave velocity sediments lying on an oceanic-like basement.

The crustal structure of the Arabian shield and platform are known from seismic refraction experiments (e.g. Prodehl 1985; Badri 1991), gravity and aeromagnetic observations (Gettings et al. 1986), receiver function analysis (Sandvol et al. 1998) and surface wave analysis (Rodgers et al. 1999; Mokhtar et al. 2001). The crustal thickness varies from approximately 35 km near the Red sea to 40 km near the NE edge of the shield, with only a thin sedimentary cover. The Arabian platform to the east of the shield is a large basin consisting of Palaeozoic and Mesozoic sediments that lie unconformably on the eastern edge of the Arabian shield and dip gently eastwards (Powers et al. 1966), reaching a thickness of 7 km or more beneath the Persian gulf (Ross et al. 1986).

$P_n$ velocities, primarily determined from ISC arrival time data (Chen et al. 1980; Kadinsky-Cade et al. 1981; Asudeh 1982a; Hearn & Ni 1994) are variable over the region. The $P_n$ velocity is well below the continental average of 8.1 km s⁻¹ under most of the Turkish–Iranian plateau, reaching values of less than 7.6 km s⁻¹ in places. The $P_n$ velocity beneath the Arabian shield is approximately 7.9 km s⁻¹. $S_n$ propagates weakly or not at all beneath the Turkish–Iranian plateau, but propagates efficiently beneath the south Caspian and Black Sea basins and beneath the Arabian Plate (Kadinsky-Cade et al. 1981; Rodgers et al. 1997; Sandvol et al. 2001). There have been several continental-scale group (Ritzwoller & Levshin 1998; Pasyanos et al. 2001) and phase (Curtis et al. 1998) velocity surface wave studies, which all find low velocities at 50–100 s periods along the Turkish–Iranian plateau.

In this study, we use surface waveform tomography to elucidate the upper-mantle shear wave velocity structure beneath the Turkish–Iranian plateau. The shear velocity structure from the tomography, along with gravity observations and measurements of the geochemistry and timing of the volcanism, provides insight into the rheology and dynamic processes of the upper mantle beneath the region.

2 PARTITIONED WAVEFORM INVERSION

The partitioned waveform inversion (PWI) technique (Nolet 1990; Van der Lee & Nolet 1997; Das & Nolet 1998) allows the imaging of laterally heterogeneous Earth structure on a regional scale. The procedure consists of two steps:

(i) matching mode-summation synthetic seismograms and observed regional surface waveforms from earthquakes with known locations, focal mechanisms and depths to produce an ensemble of
constraints on the average 1-D velocity model along the propagation paths between sources and receivers; and

(ii) combining the ensemble of 1-D constraints into a single linear system and then performing a cellular linear inversion to determine the 3-D velocity model for the region.

Rather than recalculate the modal expansion for each iteration in the waveform matching, PWI uses Frechet derivatives to relate model perturbations directly to wavenumber perturbations. This approximation is only accurate for small model perturbations (Nolet 1990); therefore, the background model for each path must be reasonably close to the actual 1-D average Earth structure along that path. The PWI formalism allows inversion for any combination of the isotropic Earth structure parameters \( V_p, V_s, \rho \) and additionally the depth of one major interface (Moho, 410 or 660 km discontinuities). We invert for \( V_s \) or \( \beta \) structure and Moho depth, as these are the parameters to which Rayleigh waves are most sensitive. Tests designed to measure the sensitivity of the waveform inversion to individual elements of the background model (Maggi 2002) show that the average crustal thickness must be correct to within \( \pm 2 \text{ km} \), the average crustal shear wave velocity must be correct to within \( \pm 130 \text{ m s}^{-1} \) and the average uppermost-mantle shear wave velocity must be correct to within \( \pm 330 \text{ m s}^{-1} \) for the inversion to converge correctly. Outside of these bounds, the inversion converges on an incorrect model, in some cases with a very small waveform misfit. To adhere to these limits, we create a different background model for each path using the isotropic version of PREM (Dziewonski & Anderson 1981) for the mantle structure and a path-specific crustal model made by averaging the velocities of the CRUST2 (Bassin et al. 2000) model along the path. Although there are discrepancies of up to 10 km in some places between the crustal thicknesses given by CRUST2 and those found locally by receiver function methods, we have found that these discrepancies are not systematic and that our paths are sufficiently long that the discrepancies do not significantly alter the average crustal thickness along the paths. We do not include the PREM upper-mantle low-velocity zone in the starting models, but allow the waveform inversion to incorporate a low-velocity zone if required by the data.

The PWI method assumes that the observed waveform can be matched by a multimode waveform in which each mode propagates independently and along the great-circle path. These assumptions are valid for a smoothly varying medium without strong lateral velocity gradients (Woodhouse 1974) and are also valid for mildly heterogeneous structure and lower order surface waves with frequencies below 20 mHz (Kennett & Nolet 1990; Kennett 1995). Sharp transitions such as continent–ocean boundaries induce coupling between more distant modes and can cause scattering effects such as shear-coupled Rayleigh waves (Meier & Malischewsky 2000). Marquering et al. (1996) find that neglecting mode coupling in PWI tomography can lead to artefacts in the deeper parts of the inversion model and a tendency for the polarity of heterogeneities in the inversion model to reverse with increasing depth.

Kennett (1995) examines the validity of the great-circle approximation for surface wave propagation at regional/continental scale and concludes that it should be suitable for periods between 30 and 100 s, and that it remains valid at longer periods (>50 s) even where surface waves cross major structural boundaries, such as an continent–ocean transition. Significant deviations from great-circle propagation have been observed for short period (<40 s) surface waves (Alsina & Snieder 1996; Cotte et al. 2000), but surface wave ray tracing in Earth models similar to ours confirms that off-great-circle propagation can reasonably be neglected for the fundamental and first few higher modes at periods greater than ~40 s and for paths shorter than ~10 000 km (Yoshizawa & Kennett 2002). Ritzwoller et al. (2002) examine the breakdown of the great-circle approximation using Born or Rytov approximations and find that the great-circle approximation is adequate for short (<5000 km) propagation paths. Very few of the waveforms included in this study are fit at periods shorter than 40 s; the average path length is 2200 km and none of the paths are longer than 5300 km. By limiting the period range and path length of our data in this manner, we have minimized errors due to off-great-circle propagation.

3 THE DATA

Our data are vertical component broad-band seismograms from the Cambridge University Caspian deployment (Mangino & Priestley 1998), from the French GEOSCOPE seismic network and from the Incorporated Research Institutions for Seismology (IRIS). Before proceeding with the PWI analysis, we deconvolve the instrument response from the broad-band records to create displacement seismograms and decimate the time series to one sample per second.

Waveform inversion of the kind performed by PWI requires an accurate knowledge of the earthquake source parameters because it assumes the misfit between synthetic and observed waveforms is only the result of differences between an Earth model and the real Earth structure (in the limit of the waveform sensitivity). Errors in the source parameters can therefore be mapped into the inversion model as erroneous Earth structure. Lateral errors in earthquake location in the Middle East can be in error by as much as 50 km (Lohman & Simons 2002). These errors map directly into the 1-D velocity models via the apparent group-velocity dispersion curves, without altering the waveform misfit (see Fig. 2). Earthquake focal depths are often poorly constrained in earthquake catalogues and can also be in error by up to 50 km in the Middle East (Maggi et al. 2000). Depth errors lead to incorrect assumptions about the the modal and frequency content of surface waves, and we have shown that they can change the output 1-D velocity models significantly without necessarily having a large effect on waveform misfit (Fig. 3). Errors in focal mechanisms, not unknown in the Harvard CMT catalogue for this region (Dziewonski et al. 1981; Baker et al. 1993; Maggi et al. 2000), also affect the reliability of the 1-D models as they lead to incorrect assumptions about the phase of surface waves.

The constraints from the 1-D inversion models are weighted according to the waveform misfit before being combined for the 3-D inversion, so it is particularly important to avoid 1-D inversion artefacts that do not affect the quality of fit. Many surface wave tomographic studies (e.g. Lebedev & Nolet 2003; Priestley & Debye 2003) use earthquake source parameters from the Harvard CMT catalogue and discard events with a poor fit. This method does not eliminate events with significant epicentral distance errors and may well retain events with significant errors in focal mechanism or depth. The authors of these studies argue that the residual errors average out in their large data sets.

In this study, we take a different approach and only consider data from events for which focal mechanisms and depths have been independently determined by body waveform modelling. This approach drastically reduces the size of our data set and it means we cannot assume that the effects of epicentral mislocation will average out (we take locations from the Engdahl et al. 1998, catalogue for events prior to 1999 January 1 and from the Harvard CMT catalogue for later events). However, artefacts in the 3-D velocity model caused by errors in focal depth and source mechanism should be minimized. In order to isolate cases of significant epicentre
Figure 2. Sensitivity of 1-D waveform inversions to a ±50-km epicentral mislocation for (a) an event with epicentral distance \(\sim 1700\) km and (b) an event with epicentral distance \(\sim 2500\) km. Inversion velocity models and dispersion curves for the correct epicentral location are shown as thick black lines. If the epicentre is closer to the event than the true epicentre, then the phase arrivals will be late and the waveform inversion will make the inversion velocity model slower to compensate; if the epicentre is further from the event than the true epicentre, the phase arrivals will be early and the resulting inversion velocity model will be faster. The effects are more pronounced for shorter epicentral distances, but the quality of fit is always unchanged because the mislocation produces a timing error that does not change the amplitude and relative phase of any part of the seismogram.

Figure 3. Sensitivity of 1-D waveform inversions to focal depths. The effects of varying focal depth for the two events in Fig. 2: (a) focal depth 2 km (b) focal depth 9 km. Fits of final synthetic (dashed) to observed (solid) seismograms for shifts in focal depths of 10–50 km are shown on the left and the corresponding inversion velocity models (thin lines) are shown on the right along with the velocity model for the correct depth (thick line). The misfit and maximum frequency achieved by the waveform inversion are denoted to the right of each waveform fit.
mislocation, we compare final inversion models and their dispersion curves for clusters of similar paths that should produce similar 1-D velocity models (Fig. 4). Comparison of inversion models within each cluster enables us to identify and remove inconsistent data but is still a majority vote method and does not guarantee that the source parameters used in determining the remaining velocity models are accurate. We therefore compare velocity dispersion curves calculated from the final 1-D Earth models with the group velocity dispersion measured by Ritzwoller & Levshin (1998) and Pasyanos et al. (2001; Fig. 4d) to isolate any residual erroneous 1-D models.

Of the 1100 seismograms originally chosen for analysis, the above data selection procedure accepted 550 good quality seismograms, many of which have very similar propagation paths. The resulting uneven geographical distribution biases the tomography results towards the structure of the regions with highest path density: multiple sampling along certain paths reinforces the structure along those paths compared with that of the crossing paths and leads to smearing artefacts in the 3-D model. We therefore thin the paths so as to render the path coverage as uniform as possible, selecting only the highest signal-to-noise ratio seismograms from each cluster, leaving us with 303 good quality and approximately evenly distributed paths (Fig. 5).

4 3-D TOMOGRAPHY

Waveform inversion of the final set of 303 seismograms provides a total of 5483 linear constraints on Earth structure in the Middle East. We perform a cellular inversion of these constraints using the method of Van der Lee & Nolet (1997) on a grid with 200 km horizontal and 100 km vertical grid spacing. The inversion assumes great-circle path propagation of surface waves and approximates their finite-width influence zone by a Gaussian that spans multiple cells. We use a Gaussian width of 400 km, as this is the approximate width of the influence zone for 100-s Rayleigh waves (e.g. Yoshizawa & Kennett 2002). Using smaller widths results in more pronounced structures that are more strongly affected by smearing. Before the 3-D inversion, we reformulate the constraints on 1-D Earth structure, which had been determined with respect to different background models for each path, as variations with respect to a common average background model (Nolet 1990; Van der Lee & Nolet 1997). Our
Figure 5. Data coverage. (a) Ray paths. Events are shown as white circles, GSN stations as upright triangles and stations from the Cambridge University Caspian deployment (Mangino & Priestley 1998) as inverted triangles. (b) Number of rays per $4 \times 4^\circ$ square. Note that the influence zone of a 100-s Rayleigh wave is approximately 400 km wide (Yoshizawa & Kennett 2002). (c) Standard deviation of the azimuths of the rays passing through each $4 \times 4^\circ$ square.

Figure 6. The common average 1-D background model for the 3-D tomographic inversion, shown alongside PREM for comparison.

background model is a smoothed average of all the best-fitting 1-D models from the waveform fitting stage and has a 44-km-thick crust overlaying an 100-km-thick low-velocity zone, which reaches a minimum shear wave velocity of $4.35 \text{ km s}^{-1}$ (Fig. 6).

4.1 Upper-mantle structure

Fig. 7(a) shows horizontal cross-sections through our 3-D tomographic model at 100, 150 and 250 km depth. The slices are colour shaded by absolute shear wave velocity perturbation with respect to the common background model described above; poorly constrained areas are masked in grey. Also shown for guidance are the ray density and azimuthal coverage from Fig. 5. The ray density and azimuthal coverage maps are essential for a correct interpretation of the tomographic images. For example, the 250-km-depth SE–NW trending slow anomaly between the Gulf of Oman and lake Balkhash in Kazakhstan passes at each end through zones of low path density and is almost entirely contained within a region with poor azimuthal coverage, strongly suggesting that the elongated nature of the anomaly is an artefact resulting from smearing.

We find that the Turkish–Iranian plateau is underlain by a strong low-velocity anomaly at least down to 150 km depth. There are low-velocity anomalies also under the Turkish peninsula and the Aegean sea, under the western Arabian peninsula, and under northern Iran, and high-velocity anomalies north of the Caucasus mountains, under the Caspian sea and in the northern portion of the Arabian peninsula. We will not discuss anomalies that lie east of $60^\circ\text{E}$ because of the low path density and azimuthal coverage in this region (see Fig. 5). The high-velocity anomalies fade out below 150 km depth, as do the low-velocity anomalies under northern Iran and the Aegean sea, while the other low-velocity anomalies fade out by 250 km depth.

Vertical cross-sections across and along the Turkish–Iranian plateau (Fig. 7b) clearly show that the low-velocity anomalies described above are confined within the upper 200 km of the model. The E–W profile AA' shows that the anomaly under the Turkish peninsula extends from 100 to 200 km depth and from the Aegean sea to eastern Turkey. The anomaly remains at constant depth, with a shallow anomaly appearing in central Turkey, until it strengthens and spreads to shallower depths in eastern Turkey, where it seems to dip strongly towards the west. The eastern end of the profile crosses the south Caspian sea, which is underlain by a high-velocity anomaly. The N–S profile BB' starts in the Turan shield, cuts across the Caucasus mountains and through the strong low-velocity anomaly in eastern Turkey before reaching the Arabian shield. The high-velocity signatures of the two shields are confined to the uppermost 100–150 km of the model. The strongest portion of the low-velocity anomaly is located under the easternmost
Figure 7. (a) Horizontal slices through the tomographic model at 100, 150 and 250 km depth. Also shown for reference are the geographic region, and the density and azimuthal coverage images from Fig. 5. Abbreviations on topographic map: BS – Black sea, C – Caucasus, CT – central Turkey, CS – Caspian sea, Ar – Aral sea, TS – Turan shield, TSh – Tien Shan, Z – Zagros, CI – Central Iran, EI – eastern Iran, RS – Red sea, AS – Arabian shield, PG – Persian gulf, M – Makran. (b) Vertical cross-sections both along and across the Turkish plateau and the Zagros mountains of southern Iran. Depths and distances along the profiles are given in km. Elevations, shown in black above the plots, are exaggerated by a factor of 10.

4.2 Resolution tests

Resolution tests are necessary to evaluate how well the 3-D tomographic images reflect the shapes and amplitudes of shear velocity heterogeneities. Resolution is traditionally assessed by checkerboard tests, in which an alternating input model is used to generate a synthetic data vector with the same path distribution as the real data and this synthetic data vector is then inverted in the same way as real data. Regions in which the synthetic pattern is faithfully recovered by the inversion are considered to be well resolved, at least for perturbations of similar or greater horizontal extent than those in the input model. Lévêque et al. (1993) have shown that this
intuitive interpretation of such tests is dangerous, as the resolution of fine features does not necessarily imply equally good resolution of coarser features: each test reveals only the resolution of perturbations with that particular input distribution. Hence, we first give a general picture of the resolution of the tomographic inversion using a traditional checkerboard test and then show the results of tests designed specifically to determine the resolution of particular features in the 3-D models.

Fig. 8 shows horizontal and vertical cross-sections through a 12° checkerboard pattern with input anomalies extending from 100 to 200 km depth. The input model is shown after application of the horizontal and vertical smoothing operators used in the cellular inversion, so that any differences with the output model are the result solely of the geographical distribution of paths and of the linear constraints. The horizontal slice at 150 km depth shows acceptable pattern recovery throughout the Turkish–Iranian plateau region and prominent horizontal smearing of anomalies east of 60° E. The northernmost vertical profile (AA′) shows relatively poor amplitude resolution but reasonable recovery of the alternating pattern between 30 and 65° E. The central E–W profile (BB′) shows good pattern and amplitude recovery along most of its length, except at the easternmost end, under the Tien Shan mountains. The southernmost profile (CC′) shows good recovery of the input anomalies especially in the central portion that crosses the northern Arabian peninsula and central Iran. The westernmost N–S profile (DD′) shows good anomaly recovery for the Black sea and the Turkish peninsula. The central N–S profile (EE) shows good pattern and amplitude recovery along most of its length, except for the central Arabian peninsula. The easternmost profile (FF′) shows good sensitivity in the Caspian sea region, as well as through central Iran. On all the profiles, the centre of the output anomalies has migrated vertically upwards compared with the input anomalies by approximately 25 km and the anomalies have smeared downwards by approximately 50 km.

The low-velocity anomalies that we have described as underlying the Turkish–Iranian plateau and the surrounding region between 100 and 200 km depth are elongated and trend E–W under the Turkish peninsula, N–S under the western margin of the Arabian Plate and NW–SE under the Zagros mountains of Iran (Fig. 7). Although most of this region has an elevated ray density and a good azimuthal coverage, the elongated nature of the anomalies makes them naturally suspect as possible artefacts of horizontal smearing. The checkerboard test shows good recovery of input anomalies for the region between the Black sea, the Caspian sea and the Arabian peninsula, with no discernable horizontal smearing (Fig. 8). However, as the anomalies in the tomographic images are both smaller and less regularly distributed than those in the checkerboard test, we cannot base a discussion of the reliability of these anomalies solely on the results of that test. We have therefore designed a synthetic test specifically to investigate the reliability of the elongated low-velocity anomalies.

The input model for our synthetic test is made up of a set of low-velocity vertical cylinders placed at the locations of the strongest low-velocity signals in our tomographic results. As our data set includes waveforms recorded at the GEOSCOPE station ATD in Djibouti that lies on top of the Afro-Arabian dome, we have included a low-velocity anomaly corresponding to the Afar plume in our synthetic input model. The amplitude of the input anomalies before horizontal and vertical smoothing is ±800 m s⁻¹ and was chosen so that the amplitude of the synthetic output anomalies would be comparable to that of the tomographic results. The input model is shown alongside the tomographic results and the synthetic tomographic output in Fig. 9. We have tested three elongated low-velocity anomalies in particular: the E–W trending anomaly under the Turkish peninsula, traversed by profile AA′ in Fig. 9(a); the NW–SE trending anomaly under the Zagros mountains, traversed by profile BB′; and the N–S trending anomaly under the western Arabian peninsula, traversed by profile CC′.
Figure 9. Synthetic tests for determining the reliability of the elongated low-velocity anomalies. The left column contains the tomographic results, the centre column the synthetic input model and the right column the output of the synthetic test for four slices through the model space: (a) horizontal slice at 100 km depth; (b, c, d) vertical cross-sections along profiles AA', BB', CC', respectively.

Profile AA', identical to profile AA' in Fig. 7, crosses the Turkish peninsula from west to east traversing three strong low-velocity anomalies: a mantle anomaly under the Aegean sea, a possible crustal anomaly under western Turkey and an extended anomaly that dips westwards under eastern Turkey and spreads out at crustal depths towards the Caspian sea. The three anomalies have been approximated by vertical cylinders of equal amplitude but varying radius, height and depth. The output of the synthetic test shows very little smearing between western and eastern Turkey, but some horizontal smearing of the easternmost input anomaly towards the south Caspian sea at depths between 100 and 200 km, indicating that the amplitude of the high-velocity anomaly under the south Caspian sea may be underestimated in the tomographic images.

Profile BB' corresponds to profile CC' in Fig. 7 and runs along the Zagros mountains from NW to SE, starting at the Black sea and ending at the Gulf of Oman, and crossing two low-velocity anomalies: the eastern Turkey anomaly also crossed by profile AA' that dips to the NW on this cross-section and a broad anomaly under the central Zagros that thins progressively towards the SE. We have approximated the Zagros anomaly by a cylindrical low-velocity anomaly centred under the central Zagros, to test whether the low velocity under the southern Zagros is the result of smearing of the central Zagros signature towards the southeast. The results show very little horizontal smoothing along this profile, only a general broadening of the input anomalies, and specifically no smearing from the central to the southern Zagros.

Profile CC' runs from the Black sea, across central Turkey, down the western Arabian peninsula to the Red sea. The tomography images show a strong horizontal anomaly along the whole profile, with a peak in intensity at the latitude of the Gulf of Aqaba, which we have included in our input model. Our path coverage extends south to Djibouti, thought to be near the centre of a strong broad (~1000-km diameter) mantle upwelling referred to as the Afar plume (e.g. Debayle et al. 2001). Given the shape of our path distribution in this region and the preferentially N–S trend of the paths ending in
Djibouti, smearing of the Afar anomaly northwards is extremely likely. We have therefore approximated the Afar plume by a 1000-km-diameter cylindrical anomaly in our synthetic input model. The output of the synthetic test shows very little smearing north of the Gulf of Aqaba, but severe smearing between the Gulf of Aqaba and Djibouti, indicating that the southern portion of the elongated western Arabian anomaly may be affected by the Afar anomaly. We therefore conclude that two out of the three elongated anomalies, the one under Turkey and the one under the Zagros, are most probably reliable images of the mantle, while the N–S trending anomaly under western Arabia is unreliable given the poor ray coverage in this region.

5 DISCUSSION

The mantle shear wave velocity structure deduced from the surface wave tomography provides insight into the rheology and dynamic processes of the upper mantle beneath the region. The most significant upper-mantle feature of our shear wave velocity model is the low-velocity zone extending beneath the Turkish–Iranian plateau. A similar image of this structure exists in the continental scale surface wave group (Ritzwoller et al. 1998) and phase (Curtis et al. 1998) velocity maps for Asia. Variation in shear wave velocity is caused by changes in temperature and composition as well as by the presence of volatiles and partial melt. The low shear wave velocities observed beneath the Turkish–Iranian plateau and the recent volcanism suggest that the upper mantle in this region is above the solidus temperature.

Regional seismic wave propagation studies (Kadinsky-Cade et al. 1981; Rodgers et al. 1997; Sandvol et al. 2001) have found that the seismic phase $S_n$ is strongly attenuated for propagation paths across this same region. Modelling of $S_n$ propagation (Stephens & Isacks 1977) shows that efficient $S_n$ propagation requires a positive upper-mantle shear wave velocity gradient and that $S_n$ propagation is weak or absent in regions with a negative upper-mantle shear wave velocity gradient. Efficient $S_n$ propagation is cited as evidence of a relatively thick, cool seismic lithosphere and poor $S_n$ propagation as evidence of a relatively thin and warm seismic lithosphere (Molnar & Oliver 1969). The low upper-mantle velocities and negative uppermost velocity gradient suggest a warm (Jackson 2000) low-density upper mantle but do not necessarily require the presence of partial melt (Priestley & McKenzie 2002).

Fig. 10 compares the pattern of low shear wave velocity observed at ~100 km depth in our model with other geophysical and geological observations, suggesting a warm, low-density upper mantle beneath the Turkish–Iranian plateau. Fig. 10(c) shows long wavelength (800–3500 km) free-air gravity anomalies from the EGM96 data set (Lemoine et al. 1996). There is a striking correlation between the gravity high running under the Turkish peninsula and the Zagros mountains, and the low-velocity anomaly beneath the same regions (Fig. 10a). Long wavelength free-air gravity anomalies reflect density differences in the mantle: less dense mantle is buoyant and will tend to rise, creating an upward deflection of the surface. This deflection produces a larger positive gravity anomaly than the negative anomaly caused by the density deficit itself, thereby producing an overall positive anomaly and a correlation between the

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**Figure 10.** Comparative images of the Middle East. (a) Tomographic slice at 100 km depth; (b) regional topography low-pass filtered at 400 km; (c) free-air gravity anomalies (EGM96, Lemoine et al. 1996) low-pass filtered at 800 km; (d) regional seismicity (black circles) 1964–1998 from Engdahl et al. (1998) and Neogene–Quaternary volcanic outcrops (pink circles; Choubert & Faure-Muret 1976; Haghipour & Aghanabati 1989; Alavi 1991).
long-wavelength free-air gravity anomalies and long-wavelength toponography. The density differences in the mantle are most likely caused by temperature differences.

The distribution of volcanism across the Turkish–Iranian plateau also suggests a warm upper mantle as the source for the low shear wave velocities. Fig. 10(d) shows the correlation between the locations of the low shear wave velocity zone and recent volcanism. Volcanic activity persisted throughout the Palaeogene and most of the Neogene over central Iran (Berberian & King 1981), but the spatial distribution and the petrology of the volcanics have changed with time (Yilmaz et al. 1998). Best studied are the suite of late Cenozoic volcanic rocks in eastern Anatolia extending across a broad SW–NE trending belt of the Arabian–Eurasian collision zone from the Arabian foreland basin in the southwest to the lesser Caucasus in the northeast. Keskin et al. (1998) and Pearce et al. (1990) found that the foreland volcanism is dominated by basaltic shield flows and fissure eruptions of transitional tholeiitic–alkaline composition whereas volcanism on the thickened crust north of the Bitlis suture zone varies from mildly alkaline in the southwest to calc-alkaline in the northeast. Isotope and trace element analysis suggests that the lavas from the foreland were derived from the mantle lithosphere of the Arabian continent, which had previously been enriched by small volumes of asthenospheric melts, and lavas from the alkaline volcanics north of the Bitlis suture zone were derived from a lithospheric source of a similar composition. The transitional and calc-alkaline lavas from the northernmost volcanics appear to be derived from lithosphere having a subduction signature inherited from pre-collision subduction events. The oldest dated volcanism is ~11 Ma, within ~2 Myr of the start of the rapid uplift of the region (Keskin et al. 1998). The chondrite-normalized rare earth element patterns for these basaltic show that the magmas came from a depth of approximately 70 km.

Seismology, gravity and volcanism all suggest the presence of a thin lithosphere and a warm (above the solidus temperature) upper mantle beneath the Turkish–Iranian plateau. Continental collision, a process that is thought to lead to cool, thick lithosphere, not warm, thin lithosphere, began along the Bitlis–Zagros suture zone approximately 12 Ma (Dewey et al. 1986) and has continued to the present. The upper-mantle low shear wave velocity zone, the high free-air gravity and the deep lithospheric source depth for the basaltic volcanism are consistent with a partial delamination of the zone approximately 12 Ma (Dewey lision, a process that is thought to lead to cool, thick lithosphere, of approximately 70 km. The transitional and fissure eruptions of transitional tholeiitic–alkaline composition whereas volcanism on the thickened crust north of the Bitlis suture zone varies from mildly alkaline in the southwest to calc-alkaline in the northeast. Isotope and trace element analysis suggests that the lavas from the foreland were derived from the mantle lithosphere of the Arabian continent, which had previously been enriched by small volumes of asthenospheric melts, and lavas from the alkaline volcanics north of the Bitlis suture zone were derived from a lithospheric source of a similar composition. The transitional and calc-alkaline lavas from the northernmost volcanics appear to be derived from lithosphere having a subduction signature inherited from pre-collision subduction events. The oldest dated volcanism is ~11 Ma, within ~2 Myr of the start of the rapid uplift of the region (Keskin et al. 1998). The chondrite-normalized rare earth element patterns for these basaltic show that the magmas came from a depth of approximately 70 km.

Seismology, gravity and volcanism all suggest the presence of a thin lithosphere and a warm (above the solidus temperature) upper mantle beneath the Turkish–Iranian plateau. Continental collision, a process that is thought to lead to cool, thick lithosphere, not warm, thin lithosphere, began along the Bitlis–Zagros suture zone approximately 12 Ma (Dewey et al. 1986) and has continued to the present. The upper-mantle low shear wave velocity zone, the high free-air gravity and the deep lithospheric source depth for the basaltic volcanism are consistent with a partial delamination of the lower lithosphere (Pearce et al. 1990; Keskin et al. 1998), caused by an instability resulting from earlier thickening of the lithosphere during the continental collision of Arabia and Eurasia. The low shear wave velocity, low density and recent volcanism suggest the upper mantle is above the solidus temperature either because of an increase in temperature, or an increase in volatiles, or a combination of both. When delamination occurs, the sinking cold material is replaced by rising, warmer material bringing the material in the remaining lithosphere closer to the melting point. However, prior to approximately 12 Ma, the region was underlain by a subducting, wet, oceanic slab. Rising fluids from an oceanic slab will reduce the melting point of the material above. Seismology does not allow us to distinguish between the relative importance of these two possible causes of the low-velocity, low-density upper mantle beneath the Turkish–Iranian plateau. Further analyses of the gravity data and geochemistry of the volcanics are required to investigate this problem.

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