The dynamics of the eastern Mediterranean and eastern Turkey

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SUMMARY
In this study we investigate the dynamics of the region that includes Greece, the Aegean Sea and Asia Minor. In a least-squares inversion we solve for a continuous strain rate field, and corresponding velocity field, that satisfies 872 GPS data. The estimate of the geodetic strain rate field provides constraints for our dynamic analysis. Next, we separately solve the depth integrated 3-D force balance equations for depth-integrated deviatoric stresses within the lithosphere, in which body force input comes from differences in vertically integrated vertical stress, or differences in gravitational potential energy per unit area (GPE). These GPE estimates calibrate the absolute magnitudes of deviatoric stresses that are acting within the lithosphere. Further, we investigate the sensitivity of our stress field solutions by using two different crustal structure models: one from compiled crustal structure estimates obtained primarily from relatively recent seismic observations and the other from the Crust 2.0 model. In an iterative least-squares inversion we then solve for stress field boundary conditions that, when added to the contribution of deviatoric stresses associated with GPE differences, provides a best fit to the directions of principal axes and relative magnitudes of the principal axes of the rates of strain obtained in the kinematic analysis. Robust features that arise from the boundary condition solution are NNE forcing along the southern boundary east of about 33° E (0.5–1.2 × 10¹² N m⁻¹), with a rapid anticlockwise rotation of forces to the west of this, along with an outward pulling force (~0.4 × 10¹² N m⁻¹) directed SSW along the entire Hellenic Arc segment. This force system along the Hellenic Arc can be interpreted as a result of slab rollback. The total depth integrated 3-D deviatoric stresses in the final dynamic solution provides an excellent match to the deformation indicators throughout the region, with vertically integrated stress magnitudes of order 0.5–2.5 × 10¹² N m⁻¹. We use constraints from derived stress magnitudes, together with GPS-defined scalar values of strain rate magnitude, to define bulk effective viscosities of the lithosphere. Depth-averaged effective viscosities for the entire lithosphere are high within the Black Sea, of order 0.7–3 × 10²³ Pa·s, relative to surrounding continental lithosphere. North Anatolian shear zone, northern Aegean Sea and Gulf of Corinth are characterized by low depth averaged viscosities of order 1–5 × 10²¹ Pa·s. Deviatoric stresses from GPE differences and boundary condition effects combine in surprising ways in some regions, resulting in near total stress cancellation in areas such as the southern Aegean Sea and portions of the central Anatolian block. GPE differences combine with boundary condition effects along the eastern segment of the North Anatolian Fault (NAF) in a way that is compatible with the hypothesis that motion on the NAF was facilitated by slab detachment beneath East Anatolia and dynamic uplift of East Anatolian Plateau. In general, GPE differences play a nearly equal role as boundary condition influences in their contribution to the total deviatoric stress field. The low depth integrated deviatoric stress magnitudes throughout the region suggest that zones of active deformation are facilitated by dramatic weakening mechanisms throughout the lithosphere.

Keywords: Space geodetic surveys; Fault zone rheology; Dynamics of lithosphere and mantle.
1 INTRODUCTION

The eastern Mediterranean (Fig. 1) is one of the most rapidly deforming areas on the planet (McKenzie 1972, 1978; Le Pichon & Angelier 1979; Şengör et al. 1985; Dewey et al. 1986; Jackson & McKenzie 1988; Taymaz et al. 1990, 1991a,b; Ambraseys & Jackson 1990; Goldsworthy et al. 2002; Nyst & Thatcher 2004; Reilinger et al. 2006; Aktug et al. 2009; Le Pichon & Kreemer 2010). A significant number of studies in the eastern Mediterranean have concentrated on the seismological evidence (Jackson & McKenzie 1988; Ambraseys & Jackson 1990; Taymaz et al. 1990, 1991a,b; Jackson et al. 1992; Hatzfeld et al. 2000; Kiratzi 2002; Koravos et al. 2003; Kotzev et al. 2006; Copley et al. 2009; Shaw & Jackson 2010) to obtain information on the styles and spatial scales of the deformation, as the region is characterized by widespread diffuse seismicity (Hollenstein et al. 2008, Fig. 2). Moreover, during the last two decades the kinematics of the region have been repeatedly refined using results from space-geodetic methods (Kahle et al. 1995, 1996; Clarke et al. 1997; Reilinger et al. 1997; Cocard et al. 1999; Wright et al. 1999, 2001; McClusky et al. 2000; Meade et al. 2002; Jenny et al. 2004; Kreemer & Chamot-Rooke 2004; Kreemer et al. 2004; Reilinger et al. 2006; Aktug et al. 2009). The main boundary conditions of the tectonics of the eastern Mediterranean are the northward motion of the Arabian Plate in the SE, the Hellenic subduction in the SW and stable Eurasia in the north (Fig. 1).

For the eastern Mediterranean (and the Aegean in particular) crustal motions have been described using rigidly rotating plates, or continental blocks (Le Pichon et al. 1995; Goldsworthy et al. 2002; Nyst & Thatcher 2004; Reilinger et al. 2006), confining deforming zones into narrow strips between rigid or elastically strained blocks. These models are simplistic and yet remarkably successful in describing the first-order kinematic patterns (Thatcher 2000). However, the velocity residuals with GPS that these block models produce are often non-negligible and they fail to explain the diffuse patterns of earthquakes that occur between the major fault zones. Moreover, these models are purely kinematic and do not address the dynamics, which involves the continuity of the stresses across blocks and major shear zones. Force balance on lithospheric blocks is not sufficient alone. Internal body forces associated with topography and crustal thickness variations must also be considered (England & McKenzie 1983; Ghosh et al. 2009). The elevation difference between the East Anatolian Plateau and the Aegean Sea led several workers to propose this lateral variation in gravitational potential energy as a major driving force for the lateral extrusion of the Anatolian block (see McKenzie 1972, 1978; Dewey & Şengör 1979; Le Pichon & Angelier 1981; Le Pichon 1982). Yet the relative roles of the gravitational potential energy differences and tectonic boundary conditions have never been quantified for the entire eastern Mediterranean and eastern Turkey regions.

A number of studies have addressed the dynamics of the Aegean region. Hatzfeld et al. (1997) and Martinod et al. (2000) have used analogue models to point out the importance of gradients in gravitational potential energy, slab rollback and Arabia indentation for explaining extension in the Aegean and strike-slip motion along the North Anatolian Fault (NAF) zone. Using a thin elastic shell parametrization Meijer & Wortel (1997) found the need for outward directed pull of uniform magnitude normal to the overriding margin of the Hellenic subduction zone. They interpreted this result as gravitational spreading of Aegean lithosphere accommodated in part by slab detachment and associated rollback. Lundgren et al. (1998) also used an elastic finite element approach to investigate the dynamics of the Aegean region. They also found the need for trench rollback, Arabian push, and gravitational collapse of east Anatolia to match rates of deformation inferred from GPS observations. Their model did not directly incorporate the influence of gravitational collapse resulting form internal body forces, but instead these forces were imposed in a NE–SW direction in western Anatolia using extensional truss bars. Cianetti et al. (2001) and Jiminez-Munt & Sabadini (2002) used thin shell methods to model the eastern Mediterranean region, and like others before them they found the importance of trench suction and Arabia push. Jiminez-Munt & Sabadini (2002) concluded that gravity potential energy was not responsible for the westward motion of Anatolia. Cianetti et al. (2001) obtained a reasonable long-wavelength fit to GPS, but suggested that future work needed to focus more on the role of density variations within the lithosphere, as well as a need for improved resolution and accuracy of modelling the tectonic domains of the eastern Mediterranean. Provost et al. (2003) have modelled the region using a 3-D model. They found best results if the NAF was weak. However, they also found a poor agreement with the GPS velocities in the Aegean and Marmara regions and they suggested that further work was needed.

Faccenna et al. (2006) used a 3-D analogue model to argue that slab detachment beneath the Bitlis-Hellenic subduction zone played a role in uplift of East Anatolia, increase in slab retreat rates at the Hellenic Trench, indentation of continent in the collision zone and initiation of NAF motion in late Miocene-Early Pliocene. Le Pichon & Kreemer (2010), further argued that the large-scale counter-clockwise GPS velocity pattern, from the Levant in the east to Anatolia and the Aegean in the west, ultimately has an asthenospheric driving source. Le Pichon & Kreemer (2010) hypothesized that this asthenospheric flow is associated with potential toroidal flow around the African slab edge and uplift of East Anatolia, which they argued was a response to slab delamination in the Late Miocene-Early Pliocene described by Şengör et al. (2003) and Faccenna et al. (2006). Le Pichon & Kreemer (2010) also suggest that this uplift of East Anatolia played a role in initiating motion on the NAF through the influence of the newly introduced gradient in gravity potential energy.

Based on previous work it is clear that higher resolution dynamic models of the Aegean and Anatolian regions are needed, along with a quantitative look at the relative role of gravitational potential energy differences and boundary force effects in driving the deformation (Cianetti et al. 2001). That is, there is a need to quantify the magnitudes of stresses, along with their spatial variability, associated with these different sources. Such an analysis will require models that incorporate topography and crustal thickness variations along with constraints provided by GPS observations.

The Aegean and Eastern Turkey regions have experienced a complex history of subduction (e.g. Şengör et al. 2003; Faccenna et al. 2006; Gogus & Pysklewyc 2008; Le Pichon & Kreemer 2010). Ongoing subduction and collision play a major role in affecting the boundary conditions acting on the deforming eastern Mediterranean lithosphere (Le Pichon 1982; Royden 1993; Meijer & Wortel 1997; Faccenna et al. 2006). Taking into account the dynamics of plate/slab interaction is clearly important for understanding mechanisms of stress transfer from slab to overriding plate, slab rollback effects, slab bending effects, among many other important factors (e.g. De Franco et al. 2007; Capitano et al. 2009, 2010). Therefore, a full 3-D treatment of the region, with full influence of the slab subduction history, is ultimately necessary.

We argue here, however, that an improved understanding of the dynamics of the lithosphere in this region can still be obtained...
through solutions to the depth-integrated 3-D force balance equations (e.g. Flesch et al. 2001; Ghosh et al. 2008, 2009). First, the aspect ratio of the width of deformation zone to lithosphere thickness within the region considered (Figs 1 and 2) is greater than 10. This aspect ratio justifies an approach where one solves for depth-integrated values of deviatoric stress. Second, the kinematics of the region has been elucidated over the past two decades due to increasingly better GPS coverage (see recent review by Le Pichon & Kreemer 2010). Our objective is to thus obtain self-consistent dynamic solutions that satisfy the components of the deformation field defined by GPS motions.

Although Kreemer & Chamot-Rooke (2004) and Jenny et al. (2004) have investigated the kinematics of the depth-averaged contraction within the accretionary complex within and to the south of the Hellenic Trench, we do not. As mentioned above, the dynamics of this zone are explicable only with full 3-D models of subduction, where depth-dependent displacement rates within the accretionary complex and bending of lithosphere near the trench are important (De Franco et al. 2007; Capitanio et al. 2009, 2010). Instead, our dynamic approach of solving the depth integrated 3-D force balance equations involves the determination of the influence of internal body forces and boundary conditions of stress (both described further below). The boundary conditions of stress act on the edges of the overriding lithosphere within the Aegean, and Central and Eastern Anatolia. The finite element grid where we solve the kinematics and dynamics of the lithosphere is shown in Fig. 2.

Figure 1. The major large-scale tectonic features in the region modelled, modified from a United States Geological Survey (USGS) map.

Figure 2. The grid used in both kinematic and dynamic calculations with earthquake focal mechanisms between 1976 and 2007 from the Global Centroid Moment Tensor Catalog (Dziewonski et al. 1983; Ekström 1994; Ekström & Nettles 1997). Focal mechanism size is plotted independent of magnitude.
The first step involves a kinematic analysis using GPS data. We adopt a solution routine that favours continuity of the strain rates. Regardless of the differing scales over which block-accommodated deformation can be subsumed into a more continuous process within the eastern Mediterranean, we use the strain rate tensor field inferred from the GPS only as a first-order proxy for the expected orientation of the continuous deviatoric stress–tensor field that is responsible for the longer term deformation. That is, our objective is not to argue for or against block-like models. Rather, our goal is to quantify the sources and magnitudes of deviatoric stresses acting within the lithosphere in the eastern Mediterranean region; the orientations of the geodetic strain rate field provide important constraints for this analysis.

Following the kinematic analysis the next step is the determination of the deviatoric stress field associated with gravitational potential energy differences per unit area (hereby referred to as GPE differences) through solutions to the depth integrated 3-D force balance equations. This solution corresponds to the non-homogeneous problem, without any forcing through the domain boundaries. This method has an advantage that the depth-integrated deviatoric stress magnitudes are defined by the effective body force magnitudes (GPE differences) and are nearly invariant to the absolute magnitude of effective viscosity of the lithosphere. This solution method thus offers a distinct advantage in that it (1) enables us to directly address the role of GPE differences in the deviatoric stress field within the Eastern Mediterranean and Turkey region and (2) these body forces calibrate the absolute magnitudes of deviatoric stresses acting within the lithosphere. Based on benchmarking results with global 3-D models (Klein et al. 2008) and with regional models (Klein et al. 2009), our solutions are expected to recover accurate depth-integrals of horizontal deviatoric stress in all regions within the overriding lithosphere of the eastern Mediterranean.

Finally, we use the separability of the problem to add a number of solutions to the unforced (homogeneous) problem to obtain a stress field boundary condition that, when added to the solution associated with GPE differences, matches the orientations and relative magnitudes of the principal axes of the geodetic strain rates. Yet another advantage of the methodology we employ is that the boundary conditions are free of any a priori bias as to possible sources of stress associated with assumed behaviour (kinematic or dynamic) at plate boundaries. Although the boundary condition solution satisfies force-balance constraints, it is defined only by the need to optimize the match with deformation indicators after it is combined with the GPE solution. Because we do not model the 3-D effects of subduction or continental collision, we cannot explicitly ascribe the boundary condition solutions to any input forcing that would arise through the 3-D interaction of overriding and subducting plates. Thus, although our determinations of the boundary forces are quantitative, and indeed are required to match observations, our description as to the ultimate origin or processes that give rise to these boundary forces must remain qualitative. Finally, our quantification of the relative role of GPE differences to boundary condition effects allows us to evaluate a hypothesis for the initiation of NAF motion (e.g. Faccena et al. 2006; Le Pichon & Kreemer 2010).

2 TECTONIC FEATURES

The major kinematic boundary conditions within the eastern Mediterranean tectonics are Arabian–Eurasia motion, the Hellenic subduction process and Africa–Eurasia convergence. A significant part of the motion of the Arabian plate is transmitted to the west by the NAF, on which the yearly slip rate is around 23–26 mm yr\(^{-1}\) (McClusky et al. 2000; Reilinger et al. 2006; Le Pichon & Kreemer 2010, Fig. 1). Part of the motion of the Arabian plate simply gives rise to the crustal shortening in Eastern Anatolia, and is also partially accommodated by NW–SE striking strike-slip faults with large compressional components in NE Turkey. The SW–NE striking Eastern Anatolian Fault accommodates the relative motion between the Arabian Plate and the Anatolian block; the slip on this left-lateral fault is 9 ± 1 mm yr\(^{-1}\) (McClusky et al. 2000).

Central Anatolia internally deforms very little and rotates counter-clockwise with respect to stable Eurasia (Platzman et al. 1998; McClusky et al. 2000). Western Turkey is dominated by a combination of right-lateral shear and extensional deformation, with principal axes of extension in a NNE direction. The NAF bifurcates at the eastern end of the Sea of Marmara, and the main northern branch of the NAF cuts across the Sea of Marmara and enters the Aegean Sea on the west of Sea of Marmara. The Aegean Sea has a bathymetry dominated by SW–NE striking deep channels, which indicate the multibranch continuation of the right-lateral shear regime of the NAF from the east (Jackson et al. 1992). However, there are also bathymetric features, such as the North Aegean Trough, which are unlikely to have been created by a purely strike-slip regime; modern-day seismicity includes normal faulting events with N–S oriented extension there (Taymaz et al. 1990). The right-lateral shear regime continues across the Aegean Sea and connects with central Greece. Central Greece is mainly dominated by extensional regimes. Extensional deformation continues into Thrace (towards the Turkish border), but with lower seismicity rates. Goldsworthy et al. (2002) claim that the kinematics of central Greece has a strong component of block rotation. There, the faults between the blocks have been responding to the shear imposed from the North Aegean Sea by rotating clockwise with respect to Eurasia. The Late Quaternary faults in the block model of Goldsworthy et al. (2002) strike SW–NE in Northern Greece and roughly E–W in central Greece. The right-lateral Kefalonia fault zone connects the reverse faulting on the western coast of central Greece with the western tip of the Hellenic Arc. Whereas there seems to be little deformation going on in the Peloponnesos, the Corinth Strait has the highest extension rates in the area (Armijo et al. 1996; Clarke et al. 1997; Curtis et al. 1997; Davies et al. 1997; Briole et al. 2000; Avallone et al. 2004). Lithosphere just north of the Hellenic Arc forms the SE boundary of our computational domain. It has long been argued that the Hellenic Arc region is a zone of seismic deficiency (see, e.g. Jackson & McKenzie 1988) where the slab rollback effect has been commonly thought of as a major driving force of the tectonic processes over the Aegean (Le Pichon 1982; Royden 1993; Meijer & Wortel 1997; Wortel & Spakman 2000; Reilinger et al. 2006; Becker & Meier 2010). Recent seismological studies also established that to the south of the arc there have been earthquakes whose depth reach 42 km (Shaw & Jackson 2010). This might indicate that the relative motion of Africa causes shortening within the entire thickness of the lithosphere immediately to the south of Crete.

3 KINEMATIC METHOD

In our kinematic solution we use the GPS data from 872 velocity observations within the eastern Mediterranean and eastern Turkey regions (Cocard et al. 1999; McClusky et al. 2000; Clarke et al. 2002; Burchfiel et al. 2006; Kotzev et al. 2006; Reilinger et al. 2006; Aktug et al. 2009; Hollenstein et al. 2008) and four IGS stations (GPSVELv1.0; Davies & Blewitt 2000, Fig. 3a) and adopt the same approach as Holt et al. (2000) and Beavan & Haines (2001) and Kreemer et al. (2003) to infer a continuous velocity gradient.
Figure 3. (a) The continuous model velocity field in Eurasia frame for the modelled domain, obtained from the best-fit interpolation of 872 GPS velocities. Observed motions are the broad, grey vectors and the model predictions at the sites and on a regular 1° × 1° grid are the thin bold vectors. Confidence ellipses for observed and model are 95 per cent. (b) Velocity residuals between the observed GPS values and the model velocity field estimate at the GPS site locations. Confidence ellipses are 95 per cent. (c) Model strain rate tensor field estimate, obtained from our best-fit inversion of GPS, plotted on top of the second invariant of strain rates. Compressional principal strain rates are bold arrows, and extensional principal strain rates are the wide, open arrows. Each strain rate tensor value corresponds to the average horizontal strain rate tensor within six areas from the grid in Fig. 2. (d) Rotation rates in degrees per million years, with positive values counter-clockwise, associated with our kinematic solution. See text for description of the kinematic method.
tensor field estimate for the entire region (Fig. 2). The technique is based on finding a distribution of rotation vectors \( W(\mathbf{x}) \) defined on the knot points of the curvilinear grid (Fig. 2), and a bi-cubic spline interpolation (Beavan & Haines 2001) is performed between points to define a continuous velocity field.

\[
u(\mathbf{x}) = r(W(\mathbf{x}) \times \mathbf{v}).\tag{1}\]

Here \( r \) is the Earth’s radius and \( \mathbf{v} \) is the vector of direction cosines for the point under consideration. The continuous velocity field is determined in a least-squares inversion in which we seek to match the GPS velocities while minimizing the sum of squares of differences between observed and predicted strain rates. The objective function that is minimized is of the following form:

\[
\chi = \sum_{\text{cells}} \sum_{ij,kl} \left( \mathbf{v}_i - \mathbf{v}_j^{\text{obs}} \right)^T V_{ijkl}^{-1} \left( \mathbf{v}_k - \mathbf{v}_l^{\text{obs}} \right) + \sum_{\text{known } i,j} \left( v_i - v_j^{\text{obs}} \right)^T C_{ij}^{-1} \left( v_j - v_j^{\text{obs}} \right),\tag{2}
\]

where \( \mathbf{v}_i \) are the cell-area-averaged strain rates and \( v_i \) are the GPS velocities. Note that \( V_{ijkl} \) and \( C_{ij} \) are the variance-covariance operators corresponding to the strain rates and the velocities, respectively. The components of the strain rate tensor are tied to the rotation function \( W(\mathbf{x}) \) through partial differentiations executed on the sphere (Haines & Holt 1993). The variances that we use for the strain-rate part of the objective function have the following forms for each component of the tensor:

\[
\text{Var}(\epsilon_{\phi\phi}) = \frac{v}{S}, \quad \text{Var}(\epsilon_{\theta\theta}) = \frac{v}{S} \quad \text{and} \quad \text{Var}(\epsilon_{\phi\theta}) = \frac{v}{2S}.\tag{3}
\]

where \( v \) is a tolerance coefficient, which we have taken as uniform in our solution, \( S \) are the areas of the grid cells, \( \phi \) is longitude, positive east and \( \theta \) is latitude, positive north. The tolerance coefficient, \( v \), corresponds to the square of the \textit{a priori} expected change in velocity across one grid cell area. Our isotropic condition implies that this change in velocity can be in any direction and, moreover, the covariance terms have zero values. We assume zero \textit{a priori} strain rates for the cells. The quadratic term to be minimized therefore reduces to the following form:

\[
\begin{bmatrix}
\epsilon_{\phi\phi} \\
\epsilon_{\theta\theta} \\
\epsilon_{\phi\theta}
\end{bmatrix}^T \begin{bmatrix}
S & 0 & 0 \\
0 & S & 0 \\
0 & 0 & 2S
\end{bmatrix} \begin{bmatrix}
\epsilon_{\phi\phi} \\
\epsilon_{\theta\theta} \\
\epsilon_{\phi\theta}
\end{bmatrix} = \frac{1}{v} \left( \epsilon_{\phi\phi}^2 + \epsilon_{\theta\theta}^2 + 2 \epsilon_{\phi\theta}^2 \right) S.\tag{4}
\]

Thus, optimization of (2) provides a match to the GPS observations, while also minimizing the second invariant of the model strain rate tensor field, weighted by the areas and normalized by the tolerance coefficient. We chose the coefficient \( v \) in such a way that the final sum of the squares of the misfit between observed and predicted velocity, obtained from the inversion, divided by the number of degrees of freedom, is around 1.0.

The strain rate tensor field obtained from the GPS observations likely contains a significant component of elastic strain and is therefore more distributed spatially than the actual long-term strain rate field that is accommodated on a variety of discrete structures. Transient effects related to past earthquakes can also distort the magnitude of the strain rates away from the long-term magnitude of strain rate and also may not reflect the strain rates occurring at depth within the lower crust and upper mantle (Hetland & Hager 2003; Pollitz 2003). Nevertheless, since our goal is to model the larger scale dynamics, we assume that the orientations of principal axes and the relative magnitude of principal axes embedded in the geodetic strain rate tensor field that we estimate, when averaged over several grid regions in Fig. 2, reflect to first order the orientations and relative magnitude of principal axes of the strain rate tensor field that accommodates finite deformation over the much longer time scales. These timescales are where the loading from GPE differences and boundary condition effects would be reflected in the deformation field over many seismic cycles. We refer to the relative magnitude of principal axes as the strain tensor style (thrust, strike-slip, normal or a combination of thrust and strike-slip or normal and strike-slip). Note that Jenney \textit{et al.} (2004) have demonstrated a good agreement between geodetically determined strain rates and patterns of seismic moment tensor orientations within the eastern Mediterranean region.

In the grid that we use to model both the kinematics and dynamics (Fig. 2), we define the southern portion of the boundary between \(36^\circ\text{E} \) and \(44^\circ\text{E} \) as rigid and part of the Arabian plate. The entire northern boundary of the grid is also rigid and is defined as part of the Eurasian plate. All other points along the remaining boundaries and within the interior are free to strain. We impose Arabia–Eurasia velocity boundary conditions by moving all rigid points along the southern boundary between \(36^\circ\text{E} \) and \(44^\circ\text{E} \) using the angular velocity (\(28.4^\circ\text{N}, 18.4^\circ\text{E}, 0.428^\circ\text{Myr}^{-1}\)) defined by Reilinger \textit{et al.} (2006). We solve for a rigid body rotation for each geodetic data set such that when applied to the GPS velocities, places them into a best-fit model Eurasian reference frame (e.g. minimizes eq. 2). With a normalized tolerance coefficient of \( v = 0.05 \), corresponding to an
a priori expected value for the one standard deviation in change in velocity across any grid area in the model to be 1.4 mm yr$^{-1}$, the final reduced Chi-squared value for the fit to the GPS was 1.0 and the weighted root mean squared (WRMS) value is 1.82 mm yr$^{-1}$ for 872 GPS points (Figs 3a and b). Given that the average grid area size is of the order of 30 km dimension, the $\nu = 0.05$ value that enables optimal reduced Chi-squared value for the entire GPS data set appears small, because rapidly deforming regions involve larger velocity gradients over these length scales. However, much of the region modelled contains many slowly deforming, or nearly rigid regions (e.g. Anatolian block, Black Sea), and this explains the lower value for $\nu$.

In regions where the strain rates are essentially localized on smaller spatial scales than the grid cells, there will be the problem of loss of resolution through the spreading of the strain rates over more than one grid cell width. To account for this smearing problem, it is possible to use $\nu$ values that differ spatially, as Kreemer et al. (2004) have done. One could further use anisotropic variance-covariance matrices instead of isotropic ones to better delineate zones of concentrated shear (Özeren 2002). Another problem is that the GPS measurements are not homogeneous in space, leading to an unbalanced picture for the data density. This becomes a serious matter when significant portions of the major faults are under the sea (the Sea of Marmara, Aegean Sea and the Gulf of Corinth). In the absence of geological and geodetic observations, it is difficult to judge how localized the strain rates are in those areas. In this paper, however, we seek the simplest, most general solution that can be obtained from the GPS, as our goal is to isolate the styles and directions of principal axes of rates of strain within broader regions, rather than precisely define the geometry of specific blocks.

4 KINEMATIC RESULTS

The continuous model velocity field and velocity residuals (Figs 3a and b) show a good match to the GPS observations. Because the spacing of GPS stations generally exceeds the dimension of our grid areas, we compute and plot averages of strain rate from our continuous solution within areas that combine six grid areas (Fig. 3c). Within central Greece and the western portion of the NAF, however, the spacing of GPS is finer and the strain rate solution for these regions is of higher resolution.

The strain rate solution for the region of eastern Turkey (Fig. 3c) is dominated by N–S to NNW oriented compressional strain rates, and roughly E–W to ENE oriented extensional strain rates (dominant strike-slip style). On the Bitlis Suture, in the south, the dominating style of strain rate is N–S compression. Principal axes of extensional and compressional strain rate are approximately equal in the vicinity of the Karliova Junction, consistent with nearly pure strike-slip style of faulting (left-lateral on NE–SW striking planes and right-lateral on NW–SE striking planes). West and northwest of the Karliova junction, the strain rate principal axes are in accord with the right-lateral styles of faulting imposed by the NAF. The higher strain rate magnitudes associated with the NAF and East Anatolian Fault are visible in the solution (Fig. 3c).

There is very little internal deformation in central Anatolia where the velocity field solution shows a dominantly rigid body rotation (Figs 3a–c). Internal deformation within the nearly rigid portion of central Anatolia, where rotation rates relative to Eurasia are of order 1° Myr$^{-1}$ anticlockwise (Fig. 3d), amounts to less than 2 mm yr$^{-1}$ of total motion across central Anatolia. This is confirmed in our

strain rate solution with low strain rates there everywhere less than $1 \times 10^{-15}$ s$^{-1}$ south of 39° N (Fig. 3c).

South of 40° N, between the longitudes of 29° E and 31° E, strain rates show a dominantly pure extension, with principal axes of strain rates oriented both N–S and E–W, depending on location (Fig. 3c) and results are in general good agreement with Aktug et al. (2009). Rotation rates approach 2–4° Myr$^{-1}$ anticlockwise in this region (Fig. 3d), with peak values in excess of 5° Myr$^{-1}$ in southwestern Turkey coastal region. This anticlockwise pattern also is clearly evident in the geodetic velocity field (Fig. 3a). Further west of 29° E, although strain rates are predominantly extensional, they contain a strike-slip component of deformation (Fig. 3c). North of 40° N, the strain rates show the pure strike-slip style of strain rate with NE–SW-oriented extension and NW–SE compression, corresponding to the right-lateral shear on the E–W striking NAF as it enters the Sea of Marmara (Fig. 3c). Spatially concentrated clockwise rotation rates of 3–8° Myr$^{-1}$ are associated with this North Anatolian shear zone (Fig. 3d).

The main features of our kinematic result for the western part of the domain are in good agreement with the earlier results obtained from geodetic observations (Kahle et al. 1995, 1996, 2000; Briole et al. 2000; Davies et al. 1997; Kreemer et al. 2003, 2004; Jenny et al. 2004; Hollenstein et al. 2008, Figs 3a–c). Our continuous velocity field solution shows a clockwise rotation for parts of central Greece, with peak values in excess of 8° Myr$^{-1}$ clockwise (Figs 3a and d). Within central Greece, between 38° N and 40° N, which is between the northern coast of Penoplolessos and Thessaly, strain rates are dominated by N–S extension (Fig. 3c). Strain rates are highest within the Gulf of Corinth region (Fig. 3c).

Within the scope of this study we have no means of constructing a kinematic picture to the south of the trench using space-geodetic constraints. The kinematic solution (Figs 3a–c) is picking up primarily along-trench extension within the plate that overrides the subducting African plate (Fig. 3c), along with minor arc-perpendicular shortening there. Along-strike extension rates are highest in the western and eastern Hellenic Arc regions, with a minimum in Crete. Focal mechanisms to the south and west of the arc indicate predominantly thrust mechanisms. The kinematic solution, which satisfies existing GPS observations, does not record strong arc-perpendicular compression in locations near the trench. This compression must lie further to the south and southwest of the modelled domain (see Kreemer & Chamot-Rooke 2004). Our objective is to determine strain rates within the overriding lithosphere north of the Hellenic Trench and north of the acting collision between African and Arabian plates. We will argue below that this strain rate pattern north of the collision and subduction boundaries is sufficient to define the nature of the far-field component of acting stress field boundary conditions.

5 DYNAMICS

In this work we vertically integrate the 3-D force-balance equations (see Flesch et al. 2001, 2007; Ghosh et al. 2008, 2009) and solve them for the vertically integrated deviatoric stress field. We first solve the force balance equations for the vertically integrated deviatoric stress field that balances the body force inputs associated with differences in GPE. We then seek a boundary condition solution that when added to the first contribution provides a best fit to the deformation indicators (e.g. Flesch et al. 2001, 2007). The boundary condition solution is equivalent to the integrated influence of density buoyancies and tractions outside of the
5.1 Crustal structure for defining GPE differences

We used crustal thicknesses inferred from the compilation of seismic and gravity studies to help constrain estimates of GPE. For most regions we used Moho depths obtained in seismological studies such as $P$ and $S$ receiver functions. In regions such as the Black Sea we also used results from gravity studies. In absence of detailed constraints, and for simplicity, we assumed fixed values for crustal and mantle densities of 2800 kg m$^{-3}$ for crust and 3300 kg m$^{-3}$ for mantle. In addition, we have computed stress calculations using the CRUST 2.0 model [Bassin et al. 2000; (Crust 2.0, G. Laske et al. Crust 2.0: A new global crustal model at 2 × 2 degrees, 2002, available at http://igppweb.ucsd.edu/~gabi/rem.html)]. Because there are some differences between these two crustal models, this comparison provides some calibration of potential uncertainties associated with crustal structure errors.

In this section we will give a short overview of our compiled crustal thickness map (Fig. 4a). The crustal thickness minimum in the Aegean Sea is located to the north of Crete where the Moho lies at a depth of around 25 km (Tirel et al. 2004; Karagianni et al. 2005; Sodoudi et al. 2006). Around Cyclades the Moho lies at around 30 km, but further north the thickness is confined between 25–27 km up to the North Aegean Trough where the crustal thicknesses increases northward, reaching about 30 km in the southern coast of western Thrace (Sodoudi et al. 2006). In Greek mainland the Moho depth increases towards the north where it reaches almost 40 km (Sachpazi et al. 2007). $P$- and $S$-wave receiver function analyses by Sodoudi et al. (2006) resulted in an average Moho depth of around 30 km under Corinth region. The multichannel seismic study of Zelt et al. (2004), on the other hand, indicates a significant E–W Moho depth gradient within the Corinth region, where crust is around 30 km thick in the east and 39 km in the west.

In Southern Adriatic and Albania the average crustal thickness is around 30 km (Papazachos et al. 1995). In Anatolia, overall, the crust thickens gradually from west to east. However, there are some variations, for example, around the Sea of Marmara. While the average thickness in the regions surrounding the Sea of Marmara is around 33 km, the value underneath the Sea of Marmara is as low as 27 km (Becel et al. 2009). In western Turkey the average crustal thickness is 33–35 km and it thins towards the south (Saunders et al. 1998; Sodoudi et al. 2006; Zhu et al. 2006). Crustal thicknesses are less well established in central Turkey where receiver function studies suggest an average thickness of around 36 km (including the central part of the Black Sea coast) and thicker crust towards Isparta region (Çakır & Erduran 2004; Erduran 2009). In places where we have no crustal thickness data (in Central Anatolia) we step-interpolated the values using the station data around the region. It is within west-central Anatolia where some of the greatest differences in crustal thickness exist between our compiled crustal model and the Crust 2.0 model (Figs 4a and b).

3-D gravity modelling and seismic studies indicate that the Moho depth contours achieve a minimum thickness on the west and east sides of the Black Sea of 20 km (Belousov & Volovsky 1992; Starostenko et al. 2004; Minsull et al. 2005). To the north, the crust under Crimean peninsula has a thickness of around 40 km (Starostenko et al. 2004). Also towards the west, under Bulgaria the crust is 37–38 km thick (Raykova & Nikolova 2006). Over the past decade there have been numerous studies on the crustal and upper-mantle structure of Eastern Turkey (e.g. Zor et al. 2003; Zor 2008). Thickness estimates show a dramatic variation from as low as 40 km in southern part of the high-plateau region to as thick as 50 km in the northern portion of East Anatolian Plateau, just south of the Black Sea coast. Thermal isostasy analysis pointed to the possibility that the mantle lithosphere under East Anatolian High Plateau could be very thin or entirely missing (Sengör et al. 2003). The geochemistry of the magmatism in eastern Turkey also shows that the lower-crustal material was subject to very high temperatures after the delamination (Keskin 2003; Sengör et al. 2008).

Given the crustal structure estimates and topography, the vertical stress at a common depth reference level is variable. This indicates the importance of some dynamic support [e.g. buoyant upper mantle in East Anatolia (Sengör et al. 2003) and also possible buoyant upper mantle within western Turkey]. Furthermore, modelling studies (Gogus & Pykslewyc 2008) show that the vertical motions that follow the delamination and slab break-off are currently continuing underneath Eastern Anatolian High Plateau. Ghosh et al. (2008) have shown that when GPE differences contain a contribution from dynamic topography that the deviatoric stresses associated with this dynamic contribution can be completely recovered as long as the density structure (crustal thickness and densities) is accurately known down to the common depth reference level. Accurate density estimates are problematic for our case, as we use a simple constant density upper-mantle model. However, in supplementary section we show that lateral density variations in the upper mantle of East Anatolia of order ±100 kg m$^{-3}$ result in relatively small perturbations to stress. The most important constraint on GPE comes from crustal thickness estimates where density contrasts are greatest, along with the need to incorporate present-day topography.

5.2 GPE contribution to deviatoric stresses

The vertically integrated deviatoric stresses we estimate from the GPE differences are defined as the 3-D deviatoric stress (Flesch et al. 2001, 2007; Ghosh et al. 2006, 2009). The solution for the vertically integrated deviatoric stresses associated with differences in GPE (Fig. 5a) are of the order of 0.2–1.5 × 10$^{12}$ N m$^{-1}$, which are in agreement with values obtained for the region, on a much coarser scale, from the global model of Ghosh et al. (2009) who used Crust 2.0 data to define GPE. As in the case for the strain rates in Fig. 3(c), the depth-integrated deviatoric stresses for each point plotted in Fig. 5(a) represent averages within regions containing six small areas of the finer grid in Fig. 2. The force-balance equations are, however, solved using GPE values defined within each of the smaller areas in Fig. 2. Note that regions with relatively high GPE yield deviatoric extension (eastern Turkey, western Turkey and central Anatolia), whereas the lower lying topography (Black Sea) possesses relatively lower GPE values; these regions of lower GPE yield deviatoric compression (Fig. 5a). Note that the highest GPE values are within southern portion of East Anatolia where upper-mantle asthenosphere is likely present beneath the relatively thin (40–45 km) crust there (Sengör et al. 2003; Zor et al. 2003). High values, accompanied by NNE–SSW extensional deviatoric stress are also observed in western Turkey. A paired belt of high-low GPE exists from the Penopollesos, through Northern Greece and into Albania. These regions are characterized by belt normal extension in the higher topography regions and belt-normal compression in low GPE areas offshore (Fig. 5a, see also Copley et al. 2009).

The GPE differences and associated deviatoric stress field from the Crust 2.0 model are shown in Fig. 5(b). The greatest difference...
between the two models lies within central Anatolian block region and to lesser extent within western Turkey. In western Turkey the compiled crustal model (thinner crust in western Turkey and western Anatolia) shows a greater amount of deviatoric extensional stress (due to thinner crust and higher GPE there, Fig. 5a) in comparison with the solution obtained from the Crust 2.0 model. Elsewhere the two solutions are for the most part similar. Both solutions show a GPE low in the southern Aegean Sea where force-balance solutions produce some contractional deviatoric stress there.

5.3 Stress boundary condition solution

We next estimate a stress field boundary condition solution that when added to the contribution from GPE differences in Figs 5(a) and (b) yields a best fit with the deformation indicators (the geometrically determined strain rate tensor orientations and styles in Fig. 3c). The force-balance equations are linear in stress, and therefore any number of boundary condition solutions can be added to determine a final stress field solution (Flesch et al. 2001, 2007).
Using the method of Flesch et al. (2007) we define 141 stress field basis functions defined by forcing along 47 evenly spaced points along the boundary. Note that the boundary condition solutions involve no differences in GPE. The complete stress field boundary condition is a linear combination of stress field basis functions, calculated around the boundary of our grid. The coefficients to these stress field basis functions are determined in an iterative least-squares inversion such that when the final stress field boundary condition is added to our deviatoric stress field associated with GPE differences, an optimal match to deformation indicators is obtained (Flesch et al. 2007). The objective function we minimize is

$$\sum_{areas} (ET - e \cdot \tau) \Delta S,$$

where

$$E = \sqrt{\epsilon_{xx}^2 + \epsilon_{yy}^2 + \epsilon_{zz}^2 + \epsilon_{xy}^2 + \epsilon_{yx}^2},$$

$$T = \sqrt{\tau_{xx}^2 + \tau_{yy}^2 + \tau_{zz}^2 + \tau_{xy}^2 + \tau_{yx}^2},$$

$$e \cdot \tau = \dot{\epsilon}_{xx} \tau_{xx} + \dot{\epsilon}_{yy} \tau_{yy} + \dot{\epsilon}_{zz} \tau_{zz} + \dot{\epsilon}_{xy} \tau_{xy} + \dot{\epsilon}_{yx} \tau_{yx},$$

$$= 2\dot{\epsilon}_{xx} \tau_{xx} + \dot{\epsilon}_{yy} \tau_{yy} + \dot{\epsilon}_{yz} \tau_{yz} + 2\dot{\epsilon}_{xy} \tau_{xy} + 2\dot{\epsilon}_{yx} \tau_{yx}.$$
\( \Delta S \) is the grid area, \( \dot{\varepsilon}_{ij} \) is the strain rate from the kinematic modelling (Fig. 3c) and \( \tau_{ij} \) is the total 3-D depth-integrated deviatoric stress tensor. This objective function is minimized when the directions of principal axes of deviatoric stress are aligned with the directions of principal axes of strain rate and when the stress tensor styles are in agreement with the strain rate tensor styles. That is, we assume an isotropic relationship between deviatoric stress and strain rate. The minimization of (5) provides a unique global minimum (Flesch et al. 2007). This method of solving for stress field boundary conditions is objective in that we make no explicit assumptions about the nature of the tectonic boundary conditions (e.g. whether the slab is retreating, or whether plates are converging). Klein et al. (2009) have performed benchmarking exercises and have shown that this method of solving for stress field boundary conditions enables a recovery of the total deviatoric stress field (both magnitude and style) when (1) both GPE and deformation styles are accurately known and (2) when deviatoric stresses associated with GPE differences are of the same order of magnitude as deviatoric stresses associated with boundary condition effects. Problems of resolving the stress magnitudes (but not style) arise when the deviatoric stresses associated with boundary conditions are an order of magnitude or more larger than the deviatoric stresses associated with GPE differences.

The boundary condition solution in Fig. 6(a) that yields a minimum misfit to the stress indicators (when added to the contribution from GPE differences from our compiled crustal model in Fig. 5a) is dominated by mostly N–S oriented compressional deviatoric stresses south of Turkey, of order 1–2.0 \( \times 10^{12} \) N m\(^{-1}\). West of southwestern Turkey (west of 29\( ^\circ \)E) there is a dramatic shift in boundary condition such that the entire Hellenic Arc region involves extensional deviatoric stresses (0.1–1.0 \( \times 10^{12} \) N m\(^{-1}\)) that are trench parallel in the western portion of the arc and mixed in the eastern Hellenic Arc between trench perpendicular and trench parallel deviatoric extension. The forcing on the boundaries corresponding to the best-fit boundary condition solution shows a dramatic shift from NNE directed forces of 1.5 \( \times 10^{12} \) N m\(^{-1}\) in the eastern collision zone to southwest directed pulling force of 0.4 \( \times 10^{12} \) N m\(^{-1}\) along the Hellenic Trench (Fig. 7a).

We next performed a second inversion for optimal boundary condition solution (Fig. 6b) such that when it was added to the GPE stress field from the Crust 2.0 model it provided a best-fit to the...
deformation indicators (style and orientation of principal axes in Fig. 3c). This second stress field boundary condition in Fig. 6(b) is remarkably similar to the boundary condition solution in Fig. 6(a), with only minor differences in western Turkey. The forcing along the southern boundary is also nearly identical to the result obtained from the compiled crustal model (Fig. 7b).

5.4 Total deviatoric stress field

The total deviatoric stress field solution is the sum of the boundary condition solution and the deviatoric stress solution associated with GPE differences (Fig. 8a). Total deviatoric stresses are dominant compressional (thrust style) along the southern collision front, north of Arabia as well as north of Cyprus along the southern coast of Turkey. Deviatoric stresses are nearly pure strike-slip style in the high plateau of East Anatolia ($\sim 1$–$1.5 \times 10^{12}$ N m$^{-1}$), with N–S compressional principal axes (Fig. 8a). In central Anatolia the vertically integrated deviatoric stresses are low (less than $0.2 \times 10^{12}$ N m$^{-1}$) where strain rates are also low (compare Figs 8a and 3c). Within western Turkey deviatoric stresses increase dramatically in magnitude, oriented NNE–SSW, with depth integrated values of deviatoric extension of $1.0$–$2.0 \times 10^{12}$ N m$^{-1}$. In the western portion of the Hellenic arc region the depth integrated deviatoric stresses show arc-perpendicular contraction and arc-parallel extension. However, south and east of Crete the deviatoric stresses are primarily arc-parallel extension.

The distinctly different stress styles on the eastern and western portions of the Hellenic Trench might be due to the different ways in which these portions interact with the upper plate. Recent modelling work by De Franco et al. (2007) revealed that the interactions with the upper plate might play an important role in determining the stress styles along trenches. In that context, in a related study, De Franco
et al. (2008) categorize Aegean subduction in the same class as Makran and Barbados where the whole plate contact is represented with a weak subduction channel and subduction with significant backarc extension taking place with relatively small earthquakes. The absence of contraction on the eastern end of the Hellenic arc appears to be caused by the pulling boundary conditions there (Fig. 7), as the GPE differences produce arc-normal contraction everywhere except south of Crete (Fig. 5a). Another interesting feature is that the boundary condition solution produces extensional deviatoric stresses in the southern Aegean Sea that cancel with the contractional stresses in the GPE low there (Figs 5a and b and 7a and b). The total deviatoric stresses in the south Aegean Sea are thus small, involving roughly E–W extension (Figs 8a and b). The thermal and stretching history of this region has been discussed by Sonder & England (1989).

The total solution associated with the Crust 2.0 model (Fig. 8b) (sum of Figs 5b and 6b) is similar in most regions to the solution from our compiled crustal structure model (compare Figs 8a and b), with depth integrated deviatoric stresses around 0.5–2.0 \( \times 10^{12} \) N m\(^{-1}\). The greatest difference is within central Anatolia, where the solution from Crust 2.0 predicts NNW directed contraction. Moreover, within western Turkey the solution in Fig. 8(b) predicts lower magnitudes, by about 30 per cent, of NNE–SSW oriented deviatoric extension within western Turkey.

The stress magnitudes from our dynamic solution are smaller by about a factor of 3–8 than those obtained by Jiménez-Munt & Sabadini (2002). This is most likely due their choice of strength profiles for their lithosphere model. On the other hand, our stress magnitudes are calibrated by the body forces (GPE differences) and stresses from these are nearly invariant to lithosphere strength variations or strength magnitudes. Thus, if much higher stresses were needed to explain deformation indicators, then boundary conditions would dominate over contributions from GPE differences. Instead, we find that roughly equal contributions from both sources are needed to match deformation indicators. The pulling forces along the Hellenic trench obtained from our inversion of best-fit boundary conditions are smaller, by about a factor of 2, than the best-fit trench suction force (TSF) obtained by Cianetti et al. (2001). Meijer & Wortel (1997) found a best-fit distribution of outward pulling forces that were arc-normal along the Hellenic Arc, whereas our best-fit solution gives a fairly uniform SW direction along the arc (Figs 7a and b).

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**Figure 8.** (a) The final deviatoric stress field that is a linear sum of the contribution from GPE differences (Fig. 5a) and the stress field boundary condition solution (Fig. 6a). (b) Same as (a), except that linear sum involves solution in Figs 5(b) (Crust 2.0) and 6(b).
6 INVESTIGATION OF THE RELATIVE ROLE OF GPE SOURCES OF STRESS VERSUS BOUNDARY CONDITION SOURCES

Some insight into the relative contributions from GPE differences versus boundary effects can be obtained by examining the ratio of the stress magnitudes from these two different sources. We plot the ratio of second invariant of deviatoric stress associated with GPE differences to the second invariant of stress associated with our boundary condition solution, $T_{\text{GPE}} / T_{\text{Boundary}}$ (Figs 9a and b). These ratio plots show that stress magnitudes from these two sources vary laterally relative to one another in a considerable way. A ratio of 1.0 implies an equal contribution from both sources. A small

Figure 9. (a) Ratio of second invariant of deviatoric stresses from GPE differences in Fig. 5(a) (compiled crustal model) and second invariant of deviatoric stress from boundary condition solution (Fig. 6a). Contour interval = 0.2. (b) Same as (a), except that ratio of second invariant of deviatoric stresses are from solutions in Figs 5(b) and 6(b) (Crust 2.0) model. Contour interval = 0.2.
ratio implies boundary condition sources dominate in magnitude over those associated with GPE differences. A larger ratio than 1.0 implies that GPE stresses dominate over boundary condition sources. Both solutions suggest that the southern boundary of the collision with Arabia is dominated by boundary condition stresses, whereas they both indicate ratios close to 1.0 in most regions of East Anatolia. Western Turkey stresses are dominated by sources from GPE differences in both solutions. Most of Greek mainland, Penopellosos, Gulf of Corinth and regions to north up into Albania are also dominated by GPE sources. Both solutions indicate that the high stress magnitudes in the Black Sea region, particularly the western region, result primarily from GPE differences as opposed to boundary condition sources. Results for the two models differ within the Sea of Marmara and they differ in the region of central Anatolian shear zone where crustal structure differences are greatest.

7 VERTICALLY AVERAGED EFFECTIVE VISCOSITIES OF THE LITHOSPHERE

We argue here that our methodology enables us to quantify the absolute magnitudes of the depth integrated deviatoric stresses acting within the lithosphere. This is possible because we can quantify the stress levels (Fig. 5) associated with the input body force distributions (differences in GPE), which are known with reasonable accuracy. Secondly, we assume that the long-term deviatoric stress field that is responsible for producing finite deformation is coaxial with the present-day geodetic strain rate tensor field (Fig. 3). The linear sum of the boundary condition solution with the influence of GPE differences yields a deviatoric stress field that provides optimal agreement with the geodetic deformation field. Thus the GPE differences, along with the style and orientation of deformation, constrain the absolute magnitudes of deviatoric stress within the lithosphere (see benchmarking in Klein et al. 2009).

Because the absolute magnitudes of the stresses are known, one can use the geodetically defined strain rates to define volume-averaged effective viscosities of the lithosphere using

\[ B = \frac{T}{\dot{E}^{1/n}}, \]  

where \( T \) is the second invariant of deviatoric stress, \( \dot{E} \) is the scalar quantity of the second invariant of strain rate and \( n \) is the power-law exponent. These effective viscosities are subject to considerable uncertainty because the geodetic strain rate magnitudes may not represent the actual long-term strain rates within crustal and mantle shear zones (Hetland & Hager 2004).

To first-order, there is a significant and large effective viscosity contrast between the Black Sea lithosphere (0.7–3 × 10^{23} Pa-s) and the continental lithosphere to the south (1–20 × 10^{23} Pa-s), for the case of \( n = 1 \) (Fig. 10a). High effective viscosities are also found for southern Turkey coastline region (0.5–5 × 10^{23} Pa-s). Relatively low effective viscosity magnitudes exist within the NAF shear zone, the northern Aegean Sea and Gulf of Corinth (1–3 × 10^{21} Pa-s). The non-linear effective viscosity (Fig. 10b), assuming power-law exponent of three, yields a similar pattern as the linear case except that effective viscosities are surprisingly high in parts of western Turkey and within eastern Anatolia (within regions of high GPE) relative to parts of central Anatolia. The effective viscosities obtained from the Crust 2.0 model solutions are similar and can be viewed in the supplementary section.

Based on the results in Fig. 10, the low strain rates within parts of central Anatolia and southern Aegean Sea appear to be due to the low deviatoric stress magnitudes there, rather than these regions being anomalously strong. Within the southern Aegean Sea in particular, the GPE associated contraction is cancelled, in part, by the boundary condition pull-effect to the southwest (Figs 6 and 7). On the other hand, the lithosphere in eastern Turkey does not appear to be anomalously weak, but the moderately high strain rates there appear to result from the much higher depth-integrated deviatoric stress magnitudes. Likewise, high deviatoric stresses in western Turkey appear to be responsible for the higher deformation rates there. This also appears true for the southernmost Caucasus. The Gulf of Corinth, the northern Aegean Sea and the North Anatolian shear zone appear to possess significant weakness relative to surrounding lithosphere. Our results are in contrast to findings of Fischer (2006) who found the need for low effective viscosities within central Anatolia but high effective viscosities within the southern Aegean Sea region.

7 DISCUSSION AND CONCLUSIONS

Our modelling constrains the relative role of GPE differences and plate interaction in the Mediterranean region. GPE differences and boundary conditions contribute approximately equally to the total deviatoric stress field within much of the region. This is in contrast to the findings of Jiménez-Munt & Sabadini (2002) and Provost et al. (2003), where their models suggest that GPE differences play little to no role within the region. By contrast, we show that GPE differences constitute a major component of the total stress field within the entire region. Moreover, we argue that our stress magnitudes determined from the GPE differences (see Ghosh et al. 2008, 2009) are reliable as long as GPE values and GPE differences are reliable.

Examination of the GPE solution indicates that the region of western Turkey, central Anatolia and eastern Turkey have a propensity to extend in a roughly N–S direction in response to the GPE differences (Figs 5a and b). This influence has often been referred to as gravitational collapse (e.g. Liu et al. 2000) and its effect in the absence of boundary conditions have been calculated for central Asia (Flesch et al. 2001; Ghosh et al. 2006) and western North America (Flesch et al. 2000, 2007; Humphreys & Coblenz 2007). Both solutions yield a GPE low in central Anatolia, with highs flanking this relative low in western Turkey and East Anatolia. The influence of the collision of the Arabian plate, and African plate south of Turkey, as evidenced from the boundary condition solutions and force plots (Figs 6 and 7), cancels out much of the N–S extensional deviatoric stress associated with GPE differences. Within eastern Anatolia this cancellation is nearly complete (Fig. 8a). The transition along the southern boundary from compressional stress boundary conditions to ubiquitous extension to the west of southwestern Turkey (Figs 6 and 7) provides support to the hypothesis that Hellenic slab rollback is an important contributor to the dominant N–S extensional stress system in the Aegean Sea (Le Pichon 1982; Royden 1993; Meijer & Wortel 1997; Lundgren et al. 1998; Cianetti et al. 2001; Kreemer et al. 2004; Heidbach 2005; Reilinger et al. 2006; Le Pichon & Kreemer 2010).

A new result from our study is the large compressional stresses along the southern Turkey coastline (Figs 8a and b). Marine geophysical studies have established the recent tectonic evolution offshore south of Turkey and south of Cyprus (Ten Veen et al. 2004). Kempler (1998) documented incipient continental collision by looking at the marine drilling data around Eratosthenes and found that the Cyprus-Eratosthenes collision triggered the extreme uplift of
Figure 10. (a) Depth average of the effective viscosity of the lithosphere, inferred from the magnitude of deviatoric stresses in Fig. 8(a) and the magnitude of geodetic strain rates in Fig. 3(c), for power law exponent, $n = 1$ (eq. 6 in text). Scale is for $\log_{10}$ (effective viscosity in units of Pa$\cdot$s). (b) Depth average of the effective viscosity of the lithosphere, inferred from the magnitude of deviatoric stresses in Fig. 8(a) and the magnitude of geodetic strain rates in Fig. 3(c), for power-law exponent, $n = 3$ (eq. 6 in text). Scale is for $\log_{10}$ (effective viscosity) in units of Pa$^{-1/3}$.

Southern Cyprus at about 1.5–2 Ma. The scenario of the compression generated by the collision propagating smoothly northwards agrees with our results that show significant compression along the Turkish coast facing Cyprus. By analysing seismic reflections Mart & Robertson (1998) revealed the existence of an important northward-dipping thrust fault separating Cyprus from the Eratosthenes. Further to the west, Hall et al. (2009) also established a broadly N–S compressional regime in the Rhodes Basin with some left-lateral strike-slip component across NE–SW structures during the Pliocene-Quaternary.

Note that we have not included basal traction terms in our dynamic calculations. Ghosh et al. (2008) have calculated the global deviatoric stress field associated with GPE differences and with basal tractions associated with mantle circulation. The Aegean region is one of the few regions where basal tractions directly below the region of deformation play little or no role in affecting the dynamics. Furthermore, Ghosh et al. (2008, 2009) show that the global GPE differences provide a remarkably good fit to the deformation indicators within the Aegean region, including strike-slip deformation along the NAF zone. The difference between that solution and this solution is that Ghosh et al. (2008, 2009) had the effects of GPE differences far outside of the region we are considering, which play a role in providing the collision forces that we are calling for. We are thus confident that ignoring basal tractions does not pose a serious problem in this study. The advantage of this study is that we have addressed a much higher level of resolution in the Mediterranean.
and eastern Turkey regions than were addressed in the global study of Ghosh et al. 2008.

Hetland & Hager (2004) have pointed out problems with the approach we take to estimate effective viscosities. They argue that the strain rates inferred from GPS can be severely affected by transients within the earthquake cycle, so that the measured strain rates at the surface may not reflect secular strain rates within the deeper crust and upper mantle. Strain rates from the GPS determined solution plotted in Fig. 3(c) are averaged over areas that are of order 100 km in dimension. Due to the GPS station spacing, strain rates within central and eastern Anatolia can only be reliably resolved on spatial scales of order 100–150 km. Such averaging may smooth over transients to some degree. Within the western NAF and parts of central Greece, however, the GPS station spacing is sufficient to resolve strain rates within areas that are of the dimension of 1–2 grid areas in Fig. 2 (30–60 km). Problems in resolving the true long-term width of shear zones, and hence magnitude of strain rates, are also likely associated with the elastic component embedded in the field. This elastic component will be characterized by a smoothed pattern of strain accumulation, depending on locking depth and slip rate of active faults (Savage & Burford 1973). If strain rates within the lower crust and upper mantle are concentrated within narrow shear zones, then the spatial average estimates in Fig. 10 may be too large, by orders of magnitude, to represent the effective viscosities of such shear zones. For example, a more concentrated zone of shear within the NAF than what we determine may explain why Hearn et al. (2009) obtain viscosity values an order of magnitude lower than ours. Note that the width of the shear zones does not influence our estimates of depth-integrated deviatoric stress. Uncertainties in effective viscosity arise, therefore, primarily from uncertainties in accuracy of the volume averaged strain rates.

Estimates of effective viscosity, assuming power-law exponents of 1 and 3, show a strong lithosphere in the Black Sea, a strong lithosphere along the southern coast of Turkey and weaker continental lithosphere sandwiched between. Regions with lower effective viscosity lie within areas with active seismicity patterns (Fig. 2), such as along the NAF zone, northern Aegean Sea and Gulf of Corinth. The southern Aegean Sea and central Anatolia appear to not have anomalously high effective viscosity, but are instead characterized by very low deviatoric stress magnitudes. In both regions it happens that sources of stress associated with GPE differences are nearly cancelled by boundary condition sources. Parts of western Turkey and eastern Anatolia, where deformation rates are reasonably high, have surprisingly higher relative viscosity than sections of central Anatolia, where deformation rates are much lower (Fig. 10b). It is also interesting to note that the relatively higher effective viscosities within parts of western Turkey and eastern Anatolia correspond to regions of high GPE (Fig. 5a), where some elevation support may result from buoyant upper mantle (Şengör et al. 2003). Volcanic basaltic volcanism, and associated lower crustal underplating of mafic composition magmas, in east Anatolia in particular (Keskin 2003; Şengör et al. 2008), may in the end be responsible for the slightly higher effective viscosities there. That is, a dry diabase composition in the lower crust is expected to yield a higher depth-integrated strength in comparison with a quartz-dominated rheology (Kohlstedt et al. 1995).

A first-order strength contrast between the Black Sea and the Anatolian continental crust may explain why strain rates are focused along the NAF zone. Indeed the contrast of elastic properties across faults, not analysed directly here, can be important for properties of dynamic ruptures (e.g. Ben-Zion 2001; Ampuero & Ben-Zion 2008) and development of rock damage near the faults (e.g. Dor et al. 2006, 2008). The low depth-integrated deviatoric stresses of order 1–2.5 × 10^{12} N m^{-2} or lower suggest that there are substantial weakening mechanisms present within the actively straining regions of both the brittle frictional and plastically flowing regimes of the continental lithosphere (Ben-Zion 2008). That is, the depth-integrated deviatoric stresses present within the Mediterranean region are a factor of 5 – 10 smaller than what would be obtained from typical strength profiles for continental crust (Jiménez-Munt & Sabadini 2002).

Despite differences between the two GPE solutions (Figs 5a and b), the boundary condition solutions (Figs 6 and 7) are similar. The forcing along the southern boundary, in particular, appears to be a robust result. Faccenna et al. (2006) and Le Pichon & Kreemer (2010) outline a hypothesis for the forces responsible for initiation of motion along the NAF in late Miocene-Early Pliocene (see Şengör et al. 2004 for a description of NAF evolution). Both describe the importance of slab detachment, accelerated slab rollback, along with uplift of East Anatolia. Le Pichon & Kreemer (2010) suggest that the GPE differences that arose from the dynamic uplift of East Anatolia played a substantial role in forcing westward motion of Anatolia. Our results elucidate more precisely how this hypothesis may be generally correct. First, our boundary condition solution and associated forcing along the southern boundary (Figs 6 and 7) show a general N–S convergent deviatoric stress, along with a dominant SW-directed pull along the Hellenic trench. This represents the forcing for the present day. However, the convergent boundary condition, east of about 36° E, likely also applied during the period before slab break-off (Lower Middle Miocene) (Faccenna et al. 2006). Such a stress field is, however, not optimally aligned for strike-slip motion on the NAF (Fig. 6), but is instead more in accord with NNW contraction. Note that present-day GPE differences (Figs 5a and b) do not yield stresses in accord with strike-slip motion either; but instead provide wide-spread N–S oriented extensional deviatoric stress across most of Turkey, with some E–W contraction within central Anatolian block region. Nevertheless, the combined field yields a deviatoric stress field that undergoes a deflection of deviatoric compressional stress direction just south of the Dead Sea coastline (between longitudes 34° E and 40° E). This deflection is from N–S orientations north of the Dead Sea coast to NW–SE contraction just south of the Dead Sea coastline (Figs 5a and b). The combined results suggest that such a deflection of the deviatoric stress field would arise following the development of the GPE differences that are observed today, with a GPE high in East Anatolia. The dynamic uplift in east Anatolia, triggered by slab delamination (Şengör et al. 2003; Faccenna et al. 2006; Le Pichon & Kreemer 2010) would provide such a mechanism for the development of the present-day GPE pattern there. The missing component is the need for a weakening mechanism near where the deflection in direction of deviatoric stress occurs. Şengör et al. (2004) indicate that the NAF delimits the Intra-Pontide suture zone to the north at least as far west as the 28.5° E meridian. The presence of this suture would probably provide the necessary zone of weakness to accommodate strain initiation and strain localization.

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**SUPPORTING INFORMATION**

Additional Supporting Information may be found in the online ver-

**Supplement.** Procedure for smoothing the crustal thickness values.

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