Neotectonics and long-term seismicity in Europe and the Mediterranean region

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Abstract

We present a neotectonic model of ongoing lithosphere deformation and a corresponding estimate of long-term shallow seismicity across the Africa-Eurasia plate boundary, including the eastern Atlantic, Mediterranean region, and continental Europe. GPS and stress data are absent or inadequate for the part of the study area covered by water. Thus, we opt for a dynamic model based on the stress-equilibrium equation; this approach allows us to estimate the long-term behavior of the lithosphere (given certain assumptions about its structure and physics) for both land and sea areas. We first update the existing plate model by adding five quasi-rigid plates (the Ionian Sea, Adria, Northern Greece, Central Greece, and Marmara) to constrain the deformation pattern of the study area. We use the most recent datasets to estimate the lithospheric structure. The models are evaluated in comparison with updated datasets of geodetic velocities and the most compressive horizontal principal stress azimuths. We find that the side and basal strengths drive the present-day motion of the Adria and Aegean Sea plates, whereas lithostatic pressure plays a key role in driving Anatolia. These findings provide new insights into the neotectonics of the greater Mediterranean region. Finally, the preferred model is used to estimate long-term shallow seismicity, which we retrospectively test against historical seismicity. As an alternative to reliance on incomplete geologic data or historical seismic catalogs, these neotectonic models help to forecast long-term seismicity, although requiring additional tuning before seismicity rates are used for seismic hazard purposes.

1. Introduction

In this project, we present neotectonic models of ongoing lithosphere deformation for the eastern Atlantic, Mediterranean region, and continental Europe. Neotectonic models were computed in a quasi-static approximation, meaning that we neglected the effect of seismic cycles and attempted to determine the long-term averages of tectonic strain and motion over many earthquake cycles. In the past, such neotectonic modeling has been divided into two separate approaches: kinematic and dynamic modeling.

Kinematic (or inverse) modeling constrains the long-term neotectonic velocity field using all available geologic and geophysical information. Mature programs of this type almost always produce realistic output and have recently come to be used as deformation models [e.g., Field et al., 2013] in the estimation of long-term seismicity for earthquake hazard models. However, these models rarely explain the fundamental tectonic mechanisms and leave no independent datasets available for posterior testing (except future seismicity).

Dynamic (or forward) modeling solves the stress-equilibrium equation using estimated rock strengths and densities to determine the processes of the Earth as though our understanding of its structure and physics were complete. Dynamic models do not use geodetic velocities, geologic fault slip rates, stress directions, or seismicity patterns as input datasets. Instead, they attempt to predict these fields from first principles and use them only in posterior tests of model quality. Thus, dynamic models are excellent for testing basic hypotheses about geodynamics, but they are rarely realistic enough to contribute to seismic hazard estimates. In fact, despite the increasing number of articles on fully 3D models, little has been done to implement a rigorous approach (even for a simple 2-D model) as a tool for shallow seismicity forecasting. In addition, multilayered, fully 3-D formulations of several geodynamic settings have received considerable theoretical interest, but they are still severely hampered by the increased number (with respect to 2-D) of unconstrained natural properties in each model and by the huge computational costs. If independent constraints on important parameters (such as effective viscosities and temperatures) are difficult to acquire for a thin-sheet two-layered model, it is reasonable to assume that 3-D models will exhibit an even more severe underdetermination of several mechanical parameters, with dramatic consequences for experiment’s realism. Moreover, Bendick and Flesh [2013] showed that in geodynamics modeling the a priori vertical coherence of the lithosphere is more important than the fully 3-D formulation of the geodynamic problem. Consequently, the authors showed that newer 3-D mechanically heterogeneous numerical simulations are still strongly related to the 2-D bounding cases and do not add any new alternative to the a priori choice of vertical coherence, despite their 3-D nature. Lechmann et al. [2011], in their extensive study of 2-D versus 3-D formulations, found that thin-sheet models are still a good approximation for determining large-scale lithospheric deformation. The notable discrepancies between a thin-sheet model and a fully 3-D model are expected to be found only where buckling effects or lateral exchanges of vertical forces are large. In the Mediterranean area, these discrepancies occur only locally, as in foreland areas close to the subduction hinges and in basins with important sedimentation rates. Therefore, in this case, thin-sheet modeling is cost-effective and valid in the initial exploration of parameter space prior to further 3-D modeling [Garthwaite and Houseman, 2011].

All these considerations have prompted us to develop deformation models for the eastern Atlantic and the Mediterranean area using a thin-sheet "hybrid dynamic"-modeling code [Shells of Bird et al., 2008], which has been enhanced with a feature that substantially increases the realism of its simulations. Although plate velocities in each simulation are still determined from the balance among rock strengths, density anomalies, and boundary conditions, we iteratively adjust the (unknown) tractions on the base of each plate across a series of simulations until its rotation approximates that determined by geodesy. This procedure, typical of kinematic modeling, avoids the old problem of potentially finding no Earth-like models; in contrast, we would prefer a suite of Earth-like models to choose from or to express epistemic uncertainty. In addition, with respect to the previous global and regional models with the same theoretical background [Bird et al., 2008; Jimenez-Munt, 2003], this work makes several improvements to the lithosphere structure, given the availability of new data on the crustal thickness and updated datasets of fault traces. We also implemented and used a modified version of the code OrbData5 of Bird et al. [2008] to determine the total lithosphere thickness. More populated geodetic and stress orientation datasets allowed us to score the models presented in this work over a wider part of the study area with respect to previous models. We also propose a new plate model (MP2014) for the Mediterranean, which updates the PB2002 plate model of Bird [2003] and better reproduces the neotectonics of the Mediterranean. For each plate completely within the study area, we determined the role played by basal, side strength and lithostatic torques, providing new insights into the Mediterranean geodynamics.

In the final part of this work, we used the preferred dynamic model to derive the long-term seismicity rates and conduct a retrospective analysis comparing the predicted rates with the historical seismicity. This calculation assumes the hypotheses of the Seismic Hazard Inferred From Tectonics (SHIFT) model described by Bird and Liu [2007] and Bird et al. [2010]. Finally, we discuss the implications of the model predictions for the geodynamics and seismic hazard of the area.

2. Mediterranean neotectonics: the MP2014 plate model

Plate models approximate the kinematics of the lithosphere by neglecting a certain amount of distributed deformation, typically dismissed as very slow, in plate interiors. Bird [2003] described two conditions that his “plates” must satisfy: (1) any “plate” is deforming an order of magnitude slower than its surroundings and (2) residual velocities after fitting a plate model are “small” compared to measurement uncertainties. Bird [2003] also referred to complex regions with distributed deformation, such as the Mediterranean, as “orogens.” We will often refer to quasi-rigid plates for convenience in discussion, such as when we discuss the balance of driving and resistive torques on each plate. However, we will conduct our modeling with a code (Shells) that makes no sharp distinction between plates and orogens, allowing active faults to be placed into plates and permanent deformation to be released everywhere.

The first part of our experiment aims to model the lithosphere-mantle interactions and to discuss the related driving and resistive forces. A definition of a reference plate model is important for two main reasons: first, one of the important boundary tractions for each plate is basal shear traction applied below its lithosphere. Second, within Shells, the basal shear traction is assumed to be coherent and smooth across the base of each plate and is iteratively adjusted until the mean plate motion is approximately correct.

As a boundary zone between Africa and Europe, the Mediterranean region appears to include a number of microplates. The plate boundary model of Bird [2003] (PB2002) has only four plates in the Mediterranean region: Africa, Aegean Sea, Anatolia, and Eurasia. However, a substantial part of the Mediterranean falls within the Alpine-Himalayan orogen. The huge quantity of new GPS data and the related possibility to determine with greater detail the strain rate patterns paired with new marine geophysics, field geology data and theoretical reasoning on the rheology contrast between oceanic and continental lithosphere led us to increase the number of small plates in the Mediterranean area. Thus, we made a few substantial modifications to the PB2002 model and constructed a new plate model named MP2014 (Figure 1), available as Data Set S1 in Supporting Information.

2.1 Eastern Atlantic and western Mediterranean

The present-day tectonics of the eastern Atlantic and western Mediterranean result mainly from the convergence between the Africa (AF) and Eurasia (EU) plates, which has created a number of crustal-scale reverse and transpressive faults that are seismically active [Jiménez -Munt et al., 2001; 2003; Vernant et al., 2010; Vergés & Fernandez, 2012]. The activity of the Gibraltar subduction zone remains disputed. According to some authors, Pliocene sediments seal and overlay the outer thrust [Zitellini et al., 2009; Platt et al., 2013], indicating that the tectonic activity of the subduction has ceased. However, Gutscher et al. [2012], while acknowledging that the activity of the wedge has dramatically decreased, showed wedge deformation and evidence of recent fold and faulting. Furthermore, Duarte et al. [2013] proposed that a new subduction system may be forming to the west of the Gibraltar arc. These different hypotheses suggest that in the western Mediterranean, the plate boundary between the Africa and Eurasia plates should pass into a region of distributed deformation, possibly including the Betic Cordillera, Rif Cordillera, and Alboran Domain. In detail, the Rif Cordillera shows a southwestward motion of 2-3 mm/yr with respect to Africa [Koulali et al., 2011], whereas southern Spain shows a consistent west-to-southwest motion with respect to stable Eurasia. Koulali et al. [2011] assumed that southern Spain and the northernmost part of Morocco represent a microplate connecting the Rif Cordillera and the Betic Cordillera. However, according to Nocquet [2012], such a microplate is small, and the offshore boundaries are unclear. Similarly Martinez et al., [2012] described one of the longest faults of the Eastern Betics Shear Zone (Alhama de Murcia fault) as accommodating Africa and Eurasia plate convergence without mentioning other local microplates (Figure 1, Figure 2). Adopting this idea, we opted to leave PB2002 unchanged here and did not insert any additional plates in the western Mediterranean in the MP2014 model.

In the Tell Atlas, the active thrusts are accompanied by coeval strike-slip faulting [Yielding et al., 1989; Meghraoui & Doumaz, 1996]. The compressional kinematics of the Tell Atlas-Rif region began during the Miocene and continued through Pliocene and Quaternary times, when most of the south-dipping faults developed at the transition between continental and oceanic domains. We defined the Africa-Eurasia plate boundary along the Tell Atlas-Rif front, which joins the Sicilian basal thrust to the east. This interpretation is coherent with the geological evolution of the AF-EU convergence [Frizon de Lamotte et al., 2000].

2.2 Central Mediterranean

Several moderate-sized earthquakes in central-southern Sicily in the last 400 years have been ascribed to the activity of the Sicilian basal thrust [Lavecchia et al., 2007]. Across this thrust, GPS data indicate a N–S convergence rate of 1.1 ± 0.2 mm/a [Devoti et al., 2011] to 2.5 mm/a [Musumeci et al., 2014]. According to interpretations of geodetic and seismological data, the southern Tyrrhenian thrust, located offshore of northern Sicily, also accommodates active convergence [Goes et al., 2004; Pondrelli et al., 2004; Serpelloni et al., 2007] (Figure 2). These data suggest that in the central Mediterranean the plate boundary between AF and EU cannot be drawn as a line, but must be thought of as spread over a wide area, possibly encompassing Sicily and part of the Tyrrhenian basin (Mattia et al., 2012). We set the plate boundary between EU and AF along the Sicilian basal thrust because this system shares the same geological evolution as the Tell Atlas-Rif system (Figure 1). Consequently, the Hyblean foreland and the submerged Pelagian Block can still be considered as the northern edge of the Africa plate, though they are strongly perturbed by the rifting in the Sicily Channel [Corti et al., 2006; Musumeci et al. 2014].

To the east of Sicily, trench retreat in the Ionian subduction zone has been documented to have occurred well into the Pleistocene [Sartori, 2004; Malinverno, 2012], and the outer thrusts in the Ionian arc are still active. The activity of the Ionian subduction also causes the residuals of geodetic velocities in Calabria to be high, regardless of the reference frame (e.g., Eurasia, Adria, or Africa) [D'Agostino et al., 2011; Nocquet, 2012]. This observation is compatible with the ongoing subduction process at depth and a possible topographic collapse in the Calabrian crust. Currently, the normal-faulting mode dominates, as seen in the geologic data [Tortorici et al., 1995; DISS Working Group, 2010] and focal mechanisms of moderate-sized earthquakes [Chiarabba et al., 2005]. Calabria and the Ionian wedge are treated as part of Eurasia in the PB2002 model. However, high residual velocities (with respect to both EU and AF), historical earthquakes and active faulting led us to define an Ionian Sea plate (IO), which includes both the Ionian wedge and the Ionian Sea side of Calabria, in the MP2014 model. The boundaries of this new plate are the Ionian trench and the crest of Calabria, where a system of normal faults has developed [Tortorici et al., 1995]. We are aware that the IO plate is not rigid; it barely satisfies one of the two conditions of Bird [2003]. However, modeling the Ionian Sea plate as an independent plate and considering the associated mantle convection is more appropriate for the scope of this study than modeling the IO as part of stable Eurasia, as in the PB2002 model.

With respect to the PB2002 model, we also included in MP2014 the Adria plate. Studies of the sedimentary facies, deformation history, and paleomagnetism indicate that Adria was an African promontory until the Meso-Cenozoic [Channell, 1996]. The focal mechanisms of large earthquakes occurring along the mountain belts surrounding Adria and the horizontal velocity differences seen in geodetic studies indicate extension in peninsular Italy and compression in the Dinarides-Albanides, which suggests a present-day rotation of Adria relative to Africa [Devoti et al., 2011]. Specifically, the Apennines act as a discontinuity in the velocity field; relative to EU, the velocities in the Tyrrhenian side trend toward the NNW and the velocities on the Adriatic side trend toward the ENE. The highest strain rate occurs along the belt, where extension has overprinted compression since 0.8 Ma [Bertotti et al., 1997; Piccardi et al., 1999; Villani and Pierdominici, 2010], and, in the present day, produces surface faulting and extensional earthquakes. Our choice was to establish a plate boundary (the western plate boundary of Adria) along the Apennine extensional belt and model the weakly active Apennine external thrust as a fault (Figure 2). The eastern edge of Adria lies along the most active thrusts of the External Dinarides [Kastelic & Carafa, 2012], whereas the northern boundary runs along the South-Alpine thrust. The southern edge of the Adria plate, however, is unclear. Along the Apulia escarpment, which represents a passive continental margin of Triassic-Jurassic age in the Ionian Sea, strike-slip faults and transcurrent tectonics have been documented [Catalano et al., 2001; Presti et al., 2013, and references therein]. We tentatively identified the southern margin of Adria as the Apulia escarpment. The Adria “neotectonic” boundaries set in this work and obtained from present-day deformation (strain rate and seismicity) are slightly different than the “geological” boundaries, especially on the westernmost side; these boundaries are typically placed along the most active thrusts of the Adria surrounding belts [Faccenna et al., 2014].

Based on GPS measurements, some authors have proposed a subdivision of Adria into two (or more) microplates [e.g., Calais et al., 2002; D'Agostino et al., 2008] and assumed the existence of a transcurrent plate boundary running from Italy to Croatia. However, other authors have rejected the idea of subdividing Adria into microplates based on GPS data alone; indeed, the high variability of GPS measurements indicates that none of the block subdivisions adequately represents the complexity of Adria [e.g., Devoti et al., 2008]. Because of the difficulty in determining the internal deformation and the paucity of earthquakes, we drop the hypothesis of a line acting as either a plate boundary or a shear zone in the middle of Adria microplate (Figure 1). Certainly, the seismicity described in Console et al. [1993], which was used by some authors as proof of a plate boundary between two rigid blocks, is well accounted for by the known active faults oriented in a NW-SE direction [Scisciani & Calamita, 2009; Kastelic & Carafa, 2012]. Consequently, we rule out the subdivision of Adria into two microplates because (1) such an alleged boundary is not particularly compatible with the stress regime and (2) the known active faults are nearly perpendicular to the supposed plate boundary. The only notable exception is the well-known dextral strike-slip Mattinata-Gondola shear zone [Di Bucci et al., 2009], which cuts the Adriatic foreland and whose easternmost termination is closer to the Southern Adriatic basin [Del Ben, 1994] than to the Croatian coastline. Given the above considerations, in the MP2014 model, we opted for the conservative choice of Adria as a single plate, although it exhibits some internal deformation [Kastelic & Carafa, 2012].

2.3 Eastern Mediterranean

The main geological feature in the eastern Mediterranean region is the Hellenic trench, an arcuate thrust fault system caused by the Aegean Sea plate overriding the Africa plate (Papadopoulos et al., 1986). The western portion of the Hellenic trench reaches the

Adriatic Sea, where the dextral strike-slip Kefalonia fault accommodates the transition from continental collision to oceanic subduction [Louvari et al., 1999]. The Aegean Sea plate is affected by widespread deformation. The northernmost region of the N-S-oriented crustal extension extends into the Balkans, as shown by GPS data [Pérouse et al., 2012]. Here, fault kinematics reflect the recent N-S to NNE-SSW extension during Late Pleistocene and Holocene times [Radulov, 2007; Vanneste et al., 2006]. N-S extension also occurs in northern Greece [Caputo et al., 2013], reaching the highest rates in the Corinth Rift [Armijo et al., 1996] and in the western part of the Anatolia plate. Radically different models seem to fit the GPS velocity field and the possible neotectonics in the Aegean Sea equally well: a mosaic of several rigid blocks [Nyst & Thatcher, 2004] or a continuous velocity field in a flowing fluid driven by gravitational forces [Floyd et al., 2010]. As a compromise, we added a small number of plates in the northern part of Aegean Sea plate to the PB2002 model, which distributed the N-S extension over a large area. Consequently, we ruled out the use of a single Aegean Sea plate because it favors the localization of deformation only on one fault. We used the kinematic block model of Reilinger et al. [2006], which determined block boundaries from the mapped faults, seismicity, and historic earthquakes. The region of diffuse deformation located in the northern Aegean Sea is divided into two plates: Central Greece (CG) and Northern Greece (NG). According to Reilinger et al. [2006], the northern plate boundary of NG is less defined with respect to the other plate boundaries, but we maintained it in our model. The southern boundary of the Aegean Sea plate is the Aegean trench, which is placed in our plate model according to the seismicity of the CMT catalog [Ekström et al., 2012].

The combination of the Hellenic subduction and the continental collision in eastern Turkey causes the Anatolia plate to move westward and rotate counterclockwise with respect to EU. The movement is facilitated by two weak strike-slip fault zones: the North Anatolia fault zone (NAf) and the East Anatolian fault zone (EAf). The westernmost part of the NAf is divided into two branches, which are further split into sub-branches in the Marmara region. Similar to the Ionian Sea microplate, we prefer to define a new plate that corresponds to the Marmara microplate based on Reilinger et al. [2006].

Several other active faults have been recognized within the Anatolia plate. Koçyiğit & Beyhan [1998] described a large, active, sinistral intracontinental transcurrent structure, known as the Central Anatolian fault zone (CAf), linking the NAf to the Cyprus Arc, which represents the eastern continuation of the Hellenic trench. Westaway [1999] disputed the activity of the CAf, but Koçyiğit & Beyhan [1999] (and recently Akyuz et al. [2013]) have contributed additional field evidence of low-to-moderate fault activity. At a smaller scale, second-order fault zones, such as the Inonu-Eskisehir and the Kutahya faults, divide the Anatolia plate into smaller blocks [Ocakoğlu et al., 2007; Özsayin & Dirik, 2009]. The Aegean Sea – Anatolia plate boundary lies in an area of E-W trending grabens with bounding active normal faults (Bozkurt [2000], and references therein). It is not obvious that these grabens are all connected by transform faults as predicted in plate theory. Thus, we have drawn approximate boundaries through the strain rate maxima. The plate boundaries of Anatolia can be drawn with fairly high confidence, with the exception of that between the Aegean Sea and Anatolia. The southern plate boundary of Anatolia is localized along the convergent boundary with the Africa plate along the Cyprus arc, where marine geophysical studies have established that recent tectonic evolution offshore is compatible with compression generated by the continental collision [Ten Veen et al., 2004; Carafa & Barba, 2013].

3. Torque balance for each plate

Under the usual modeling convention, in which we ignore the rotation and ellipticity of the Earth, the total forces on each plate are a zero vector, and the total torque on each plate (about the center of the Earth) is also a zero vector:

$$\iiint_V \vec{r} \times (\rho\vec{g})dV + \iint_S \vec{r} \times (\tilde{\sigma}\hat{n})dS = \vec{0} \qquad (1)$$

where V is the volume of the plate, S is the outer surface (with unit outward normal vector $\hat{n}$), ρ is density, $\vec{g}$ is gravity, $\tilde{\sigma}$ is stress, and $\vec{r}$ is the radius in a spherical coordinate system. Assuming that purely radial gravity does not contribute to torques yields the simplified form

$$\iint_S \vec{r} \times (\tilde{\sigma}\hat{n})dS = \vec{0} \qquad (2)$$

The outer surface S of a plate is composed of top, basal, and side surfaces. We neglect atmospheric pressure on the top and include the volume of the oceans in the underlying plate(s). Consequently, the top-surface part of the integral can be neglected. We divide the stress-related side and basal torques into three parts:

1. The lithostatic pressure torque ($\vec{Q}_{LP}$), which has been recognized in the previous qualitative literature as the effects of “ridge push” and “trench pull” as well as contributing to “continental collision” resistance;
2. Side strength torque ($\vec{Q}_{SS}$), which has been recognized in the literature as “transform resistance” and also contributes to “continental collision” resistance;
3. Basal strength torque ($\vec{Q}_{BS}$), which is due to strength at (and below) the base of the plate and has previously been recognized as either ‘‘basal drag’’ (distributed shear traction), "net slab pull" (the difference between "slab pull" and "subduction resistance"), or "slab suction."

All plate driving forces and resistances in quotation marks in the torque definitions are defined as in the glossary of Bird et al. [2008].

Equation (2) becomes the following:

$$\vec{Q}_{LP} + \vec{Q}_{SS} + \vec{Q}_{BS} = \vec{0} \qquad (3)$$

which is one of the fundamental equations of Shells [Kong & Bird, 1995; Bird, 1999], the program used to compute the deformation models in this project.

In Shells, the basal surface of any non-subducting plate separates the large conductive thermal gradient found within the plate from the lesser gradient, approximately adiabatic, located below the plate. This definition typically represents an isothermal surface. In subduction zones, we consider the base of the "plate" (for torque-balance purposes) to be a horizontal extension of the base of a flat-lying plate outside the subduction zone, and we do not include most of the subducting slab. Thus, strength-related forces exchanged between the slab and the surface plate across this horizontal boundary are counted as components of the basal-strength torque and not as part of the side-strength torque. This convention will affect the interpretation of the results below.

Shells employs a simple model of the 3-D structure of the lithosphere by assuming different crust and mantle layers (each featuring a homogeneous flow-law, except for lower friction in designated faults), laterally varying geotherms, and local isostasy. The program solves the stress-equilibrium equation (also known as the momentum-conservation equation) in the quasi-static approximation for the 2-D surface of a spherical planet after determining the strength of the lithosphere by vertical integration of stress anomalies. Nonlinearity of the frictional-strength rheology (at low temperatures) and rheology representing dislocation creep (at temperatures above the brittle/ductile transition) is managed by iterating the entire solution until it converges. The predictions of each model include long-term average horizontal velocities of nodes, long-term average (permanent) strain rates of continuum elements, long-term average slip rates of modeled faults, and a 3-D model of stress anomalies and deviatoric stresses, including their principal axes. The program is efficient enough to permit both high-resolution regional modeling [see also Jiménez -Munt et al., 2001; Jiménez -Munt & Sabadini, 2002; Negredo et al., 2002; Liu & Bird, 2002a,b; Jiménez -Munt & Negredo, 2003; Jiménez -Munt et al., 2003; Negredo et al., 2004; Bird et al., 2006; Barba et al., 2008] and low- to moderate-resolution global modeling [Bird, 1998; Bird et al., 2008]. Specifically, we use Shells version 2006.08.29, described in Bird et al. [2008], which includes two new features: (a) more flexible models of density and temperature in the lithosphere, which can incorporate seismically determined Moho and lithosphere/asthenosphere boundary depths, and (b) a semi-automated method for iteratively adjusting the basal shear traction on each modeled plate until its Eulerian rotation is approximately correct. This second innovation makes all models appear more realistic and increases the potential for Shells models to play a role in estimating future seismicity and seismic hazards.

4. Finite element model of the lithosphere

4.1. Grid geometry and faults

The first step was to deploy a grid of nearly identical (quasi-equilateral) spherical triangles with sides of approximately 55 km, obtained by repeatedly subdividing the faces of a global icosahedron, following Baumgardner [1983]. The next step was to manually move the selected nodes onto the traces of active faults and to insert these faults via special linear elements. Almost all of the plate boundaries described above were modeled as faults, as were the major faults described in Section 2 and shown in Figure 2, because these faults' activities play a key role in the tectonics of the Mediterranean region. The majority of active faults inserted into our model are also present in the European Database of Seismogenic Faults [Basili et al., 2013], and this database was used to determine fault dips. For the faults lacking dip information, we assumed a priori values depending on the fault kinematics of 90°, 55°, and 30° for strike-slip, normal, and thrust faults, respectively. For the Ionian and Hellenic subduction zones, we set dips of 18° (appropriate for the potentially seismogenic megathrusts). Not all the faults delimiting the Northern Greece plate boundaries have been fully mapped and parameterized; for this reason, unbroken – but deformable – lithosphere was left in place. We acknowledge that - at least locally -incorrect or incomplete maps of active faults could bias our models. However, we assume and are reasonably confident that the deformation deficit ascribable to the slip-rates of missing faults is released in our models as nearby off-fault deformation.

The 2-D grid incorporating these faults and certain local refinements consists of 6,049 nodes, 421 fault elements and 10,824 continuum elements (Figure 2 and available as Supporting Information as Data Set S2).

To describe the density and strength of the lithosphere, we define 6 scalar quantities for each node: elevation/bathymetry, heat flow, thickness of the crust, thickness of the mantle-lithosphere, a possible density anomaly due to petrologic variations, and a possible curvature of the geotherm due to transient heating or cooling. Co-located fault nodes receive matching data values; thus, all of these fields are continuous.

4.2. Elevation/bathymetry

Elevations (and ocean depths) were obtained from the 1 arc-minute model ETOPO1 by Amante and Eakins [2009]. ETOPO1 comes in both "Ice Surface" (top of Antarctic and Greenland ice sheets) and "Bedrock" (base of the ice sheets) versions, and we used the latter.

4.3. Thickness of the crust and its petrologic density variation

The thickness of the crust was obtained from the model of the European plate (EP-Crust) by Molinari and Morelli [2011]. EP-Crust was created by a review and critical merging of literature data, including active-source studies, receiver functions, surface waves, and a priori geologic information [Molinari et al., 2012]. Being more recent than the model CRUST2.0 [Bassin et al., 2000] and, therefore, integrating more information, EP-Crust returns a more realistic crustal structure, especially in Italy, the Alps, and Central Europe.

To avoid large departures from local 1-D isostasy, we allowed a vertically averaged density anomaly of petrologic origin in the lithosphere, represented by a nodal parameter determined within the codes OrbData5 and Shells [Bird et al., 2008]. To prevent the inference of fictitiously high-density anomalies along the subduction zone trenches (where actual 3-D regional isostasy involves the deep slabs), we limited these compositional anomalies to no more than ±50 kg/m^3.

4.4. Heat flow

The map of heat flow used in this project (Figure 3) was estimated using a combination of different approaches.

We started from the recently revised International Heat Flow Commission database consisting of 35,523 terrestrial data points and 23,013 marine data points (last revised January 12, 2011; available at http://www.heatflow.und.edu/index2.html). Surface processes (such as water circulation, sedimentation, and erosion) mask the deep conductive heat flow that is needed to study lithosphere processes. To reduce the effect of the surface processes, we detected the outliers in the global database as those heat flow values that fulfill one of the following three criteria: (1) heat flows greater than 0.14 W/m^2, which are usually from atypical volcanic areas, and either (2) values implying a Moho temperature that exceeds the melting point of gabbro at Moho pressures (which, according to petrologic barometry and thermometry, is approximately 950 °C = 1223 K) or (3) values under 0.037 W/m^2, which is the lowest stationary and conductive surface heat flow density predicted in Europe for the 2.8- to 1.8-Ga-old rocks in the central Fennoscandian shield [Kukkonen & Lahtinen, 2001]. The heat flow data points that remained after trimming the outliers were then interpolated into a 0.25° longitude/latitude grid using a kriging algorithm.

Where the oceanic crust age is known based on the seafloor age model (AGE3.6; Müller et al. [2008]), we determined the heat flow from the following equation:

$$Q(t) = \begin{cases} 0.3\, W/m^2 & t < 2.67 \text{ Ma} & \text{[Bird et al., 2008]} \\ C_Q t^{-1/2} & 2.67 \text{ Ma} \leq t \leq 80 \text{ Ma} & \text{[Jaupart et al., 2007]} \\ 0.048\, W/m^2 & t > 80 \text{ Ma} & \text{[Lister et al., 1990],} \end{cases} \quad (4)$$

where t is the seafloor age in Ma and $C_Q = 0.49 \pm 0.02\ W/m^2$. For $t < 2.67$ Ma, i.e., for spreading ridges, we imposed an upper limit of 0.3 W/m^2, which is approximately as high as the highest primary observations. This heat flow definition avoids zero strength and singular heat flow at spreading ridges, which is appropriate because conductive heat flow rarely exceeds approximately 0.3 W/m^2, the excess heat being removed by hydrothermal circulation. The resulting model of heat flow is available as Data Set S3 in Supporting Information.

Where detailed heat flow maps are available, as in the Central Mediterranean [Della Vedova et al., 2001], northern Africa, Alboran Sea and Dinarides [Milivojević, 1993; Pasquale et al., 1996; Lenkey et al., 2002], these maps were used in preference to our default estimation algorithms.

4.5. Mantle-lithosphere thickness

We followed the approach of Bird et al. [2008] by incorporating a lithosphere thickness that is constrained by teleseismic travel times. This estimate was first calibrated in the oceans and later extrapolated to the continents. We used recent global models of S-wave tomography (S40RTS by Ritsema et al. [2011]) and a compilation of seafloor ages (AGE3.6; Müller et al. [2008]). These models represent a substantial improvement with respect to the models of Ritsema and Van Heijst [2000] and Müller et al. [1997], which are used in Bird [2008]. Because even the best tomographic models lack the resolution needed to determine the thickness of the mantle lithosphere directly, we used a smoothed product of the tomographic model, specifically the travel time of vertically incident S waves above 400 km depth, after correcting for the crustal thickness (Figure 4). We then assumed that anomalies in S-wave travel time are proportional to mantle-lithosphere thickness, i.e., we approximated the anomalies at sub-lithospheric depths to be zero. Using the seafloor ages of model AGE3.6 of Müller et al. [2008], we found a relation between the vertical S-wave travel time anomaly (Δt_s) in S40RTS and the seafloor age (expressed in million years, A) by regression:

$$\Delta t_s = (1.5373\ \mathrm{s}) - (0.1446\ \mathrm{s})\ (A/1\ \mathrm{Ma})^{1/2} \qquad (5)$$

As A increases from 0 to 100 Ma, Δt_s decreases from 1.54 s to 0.09 s. For the linear regression, we used the program LINREG (http://www.pgccphy.net/Linreg/linreg_f90.txt).

Considering 85 km as the lithosphere thickness for 100 Ma oceanic lithosphere [Leeds et al, 1974] and 1.446 s (the difference between 1.54 s and 0.09 s) as the gain in S-wave arrival time with respect to 0 Ma lithosphere, we obtain the following scaling relationship (expressed in meters):

$$h_{ml} = (1.5373\ \mathrm{s} - \Delta t_s)\ (85000\ \mathrm{m}) / (1.446\ \mathrm{s}) \qquad (6)$$

We assume that this formula can also be used for continental thermal boundary layers. For example, the most negative S-wave travel-time anomalies in the S40RTS model (-2.7 s) imply a lithosphere thickness, h_{ml}, of approximately 300 km (Figure 5) in the East European Platform – Siberia Craton, compatible with the lithosphere thickness range suggested by Artemieva [2011] and references therein.

In ocean basins with known ages, we defined the plate thickness to be the geometric average of 95 km (in the preferred "plate model" of Stein and Stein [1992]) and the thickness of the thermal lithosphere, whose base is the 1673 K isotherm. This definition of lithosphere yields values ranging from 30 km at mid-ocean ridges to 95 km at the maximum seafloor age of 270 Ma.

4.6 Transient portion of the geotherm

To avoid implausibly high temperatures at the lithosphere/asthenosphere interface, Bird et al. [2008] inserted an additional value at each node to represent an extra curvature of the geotherm throughout the lithosphere associated with transient cooling (or heating). In their code, OrbData5, Bird et al. [2008] used a quadratic term to represent this transient portion of the geotherm, which is zero at the top and bottom of the lithosphere and reaches a maximum at mid-depth.

At some nodes, however, this extra curvature of the geotherm can generate an internal maximum in the lithosphere exceeding 1673 K, which represents the upper limit of the lithosphere-asthenosphere boundary. In such cases, Bird et al. [2008] used a two-step correction, first lowering the surface heat flow until the peak T equals the temperature of the asthenosphere and then reducing the thickness of the lithosphere to the point at which this temperature was reached. In our experiment, for points with unknown ages (those not represented by the AGE3.6 dataset), we considered our approach for estimating seismic lithosphere thickness more robust than the one typically used to define the thermal base of the lithosphere (the 1673 K isotherm). Thus, we used only the first step of Bird et al. [2008] and reduced the surface heat flow so that the thermal maximum in the lithosphere equals 1673 K. As a consequence of this condition, the greatest mid-lithosphere temperature anomaly above steady state (600 K) was obtained in the Pannonian Basin and Massif Central.

The resulting combined heat flow map is shown in Figure 3 and available in Data Set S2.

5. Lateral and basal boundary conditions

This regional model has side boundaries as well as a boundary at its base, where the subducting slabs, plumes, and general mantle convection control the plate motion. One important purpose of this project is to calculate the basal-strength and the side-strength torques acting on those parts of the lithosphere that are within our model.

Most of the lateral boundaries of our model involve fairly rigid lithosphere in the interiors of large plates that are only partially included (Figure 1): the Eurasia plate (EU) in the north and the Africa plate (AF) in the south. In this experiment, mainly focused on the Mediterranean region, total mantle torques cannot be determined for the greater AF and EU plates. Therefore, the unmodeled parts of these two major plates are represented by fixed-velocity boundary conditions on the AF and EU model edges with no basal traction, whereas the basal tractions of the smaller plates lying entirely within the model were determined by the iterative method described by Bird et al. [2008].

Computing basal-strength torques as efficiently as possible requires velocity regulation by means of distributed basal-shear tractions, as determined in the following 3-step procedure from Bird et al. [2008]:

1. Compute a non-physical model that lacks distributed basal-shear tractions but has velocity boundary conditions in the interior of each plate (Table 1). The velocities computed from an approximate plate model are applied at 2-4 internal nodes where the lithosphere is strong due to low heat flow. Applying such an internal condition causes local stress concentrations that are not plausible. Fortunately, the low spatial resolution of the F-E mesh keeps these unrealistic stress concentrations at reasonable levels, and they cause little deformation.

2. For each plate entirely within the model, compute the boundary-condition torque. Find a set of distributed basal tractions exerting the same torque. Then, model the region again with Shells, using these distributed basal-shear tractions but temporarily keeping the plate-specific internal velocity conditions for stability. In each iteration, the side-strength torques fluctuate slightly and the lithostatic pressure torques remain constant. Iterate step #2 until the torques imposed by the artificial velocity conditions become much smaller than the torques caused by distributed basal traction.

3. Compute a physically plausible model without internal velocity conditions, where each plate is moved through the correct Eulerian rotation by distributed basal-shear tractions (including side-strength and lithostatic pressure torques, which have varied only slightly and not at all, respectively, during this procedure).

6. Tuning the model

The aim of our experiment is twofold: determining a dynamic model with increased resolution (relative to previous global and regional models for the Mediterranean area) and forecasting long-term seismicity using the best-fitting (although preliminary) dynamic model.

Prior studies have shown that the value of the friction coefficient assigned to fault elements has large and systematic effects on the results and scores of any Shells model (see, e.g., Bird et al. [2008] and references therein). At a fixed rate of relative plate motion, fault friction determines one upper limit on the frictional strength in the shallow lithosphere [Bird & Kong, 1994]. This frictional upper limit usually controls the total strength-based resistance in seismogenic portions of plate boundaries, thus affecting the overall strength of plate boundaries. A parametric study of the fault friction shows its effect on the side-strength torque, which in turn affects the magnitude and direction of the modeled basal-strength torques. Therefore, to study the effects of this parameter on the Shells models, the numerical experiments are performed using 24 effective fault friction values that range from 0.025 to 0.6 (steps of 0.025).

In Shells, one distinctive fault type is the master interplate thrust of a subduction zone, which typically consumes large volumes of young wet sediment. This feature leads to compaction and elevated pore pressures, which may eventually escape by hydrofracturing. This distinctive set of physical processes may cause subduction zones to exhibit systematically different effective friction; accordingly, we tested different upper limits on the down-dip integral of shear traction (TAUMAX), varying it from 5×10^{11} Nm^{-1} to 1×10^{13} Nm^{-1}, with a step of 5×10^{11} Nm^{-1} (20 values). This range of

TAUMAX values corresponds to mean shear stresses from approximately 2.5 MPa to 50 MPa, respectively.

It can be argued that tuning a Shells model to single values of fault friction and TAUMAX across the whole model domain is too simplistic given the wide range of geological structures included in the study area. Indeed, it is true that we ignore much 3-D geologic variation that might influence and locally vary the effective fault friction. However, most geologists would not be able to guarantee the rock types shown on their cross-sections and the relative fault friction down to seismogenic depths of ~10-15 km. Thus, it is not an option to impose "geologically determined" variations in friction across faults. Conversely, constructing a hyper-inverse problem in which every fault has a free friction parameter to be determined leads to an underdetermined problem that typically has many local minima in the overall prediction error. Furthermore, in the absence of a considerable amount of data, any parameter optimization for each sub-region of our study area is challenging because all of the regions are interconnected in stress and kinematic balances. Separate and independent parametric studies for adjacent fault elements may result in severe instabilities connected to the overparameterization, which will not provide any robust answer regarding the lateral variation of FFRIC or TAUMAX.

Therefore, by generating 480 candidate models, we are not claiming to verify uniform FFRIC or TAUMAX values as conclusions; rather, we claim only to consider them to be starting points for future work investigating local variations in both parameters when sufficient data become available.

Consequently, with the exception of TAUMAX and FFRIC, all physical parameter values input to Shells were identical to those in Table 1 of Bird et al. [2008]. Varying both TAUMAX and FFRIC in all combinations yields 480 candidate models. To select the preferred model, we computed the misfits of models with respect to two geophysical fields: interseismic geodetic velocities and horizontal most-compressive principal stress directions. This model was then used to determine the torques acting on the MP2014 plates of our study area and to forecast long-term seismicity.

6.1. Geodetic velocities

The modeled horizontal velocities were compared with the interseismic horizontal velocities of Nocquet [2012], who combined a number of published velocity results. The resulting velocity field includes 1495 sites, including 1276 sites located in the Mediterranean area, from west of the Alboran Sea to the Caucasus (Figure 6). We preferred to use the horizontal velocities from Nocquet [2012], instead of simply merging GPS solutions of different works, because the slightly different velocity reference frames that different research groups have defined as “stable Eurasia” may have a substantial impact on areas of slow deformation.

To determine the misfit, the scoring program applied two corrections (Bird et al. [2008]):

(a) To facilitate a comparison between model estimates and geodetic measurements, long-term velocities were corrected by assuming temporary locking of brittle parts of faults and by adding a mean coseismic velocity based on current model long-term slip. This step utilized solutions for displacement about a dislocation patch in an elastic half-space [Mansinha & Smylie, 1967; 1971].

(b) Different rotational reference frames of the data and the model were reconciled by adding a uniform Eulerian rotation.

In the final step (the summation of discrepancies after these two corrections), we attempted to compensate for uneven station density (due to, e.g., very extensive studies in Italy or Greece): we weighted each discrepancy by the surface area adjacent to that benchmark. Error measurements were also weighted by the inverse of the velocity variance in the data. Our summary geodetic error m_1 is the square root of the weighted mean discrepancy variance and is expressed in mm a^{-1}.

6.2. Stress azimuths

Within the perimeter of the model, the World Stress Map dataset [Heidbach et al., 2008] has 2582 stress-orientation indicators primarily derived from fault-plane solutions (1664), a number of borehole breakouts (659), and a few geologic fault-rake data (140). The data density is rather high in the Apennines, Alps, and Rhine graben, although the density is low for the ocean floors. Furthermore, the large scatter in the stress directions is rarely captured by the tabulated uncertainties, indicating that the error values may often be speculative. To compensate for the irregular sampling and inadequate description (for our purposes) of the uncertainties in the stress directions, we used the web tool SHINE, available at shine.rm.ingv.it [Carafa et al., 2015], which interpolates SHmax orientations (from any dataset) using the algorithm for "clustered data" developed by Bird and Li [1996] and updated by Carafa and Barba [2013]. We interpolated the World Stress Map dataset onto a grid with 0.25° point spacing using the following strategic parameters in SHINE: the search radius was set to 180 km (5th annulus), the minimum cluster number was set to 3, and the 90% confidence bounds were set to 45° (Figure 7). These interpolated azimuths were compared with the modeled principal axes of permanent strain rate to evaluate the model misfit for each numerical experiment. We tabulated the mean of the absolute value (L_1 norm) of the model-data discrepancy and refer to this as error measure m_2.

6.3. Combined misfit

Because misfits m_1 and m_2 are expressed in terms of different units, ranking the candidate models cannot be accomplished by adding or averaging them. We decided to summarize the two misfits of each model (the geodetic misfit m_1 and the stress discrepancy m_2) by tabulating the geometric mean $(m_1\ m_2)^{1/2}$. The rank order of these geometric means is independent of the system of units in use. The resulting geometric mean of the misfits is shown in Figure 8.

7. Preferred model

The two tuning parameters in our large set of models are the integrated shear-stress resistance to subduction (TAUMAX) and the effective friction coefficient (FFRIC) for all other faults. The preferred model features values of TAUMAX = 1×10^{12} N m and FFRIC = 0.075 and has a performance of m_1 = 3.18 mm a^{-1} and m_2 = 35.47°. The preferred model is available as Data Set S4 in Supporting Information.

Both preferred parameter values are in the lowest part of the tested intervals (TAUMAX: 5×1011 Nm-1 to 1×1013 Nm-1; FFRIC: 0.01 to 0.6), although not at the minimum values. Trench resistance, TAUMAX, strongly influences the amount of plate convergence that must be released as slip rate on fault elements or as compression on continuum elements. Greater TAUMAX values imply lower slip rates on the subduction interface and a relatively wide deformation area around the subduction zone. Lower TAUMAX values localize the convergence mainly as fault slip rates, leaving smaller compression to be accommodated in the continuum. Obviously, trench resistance TAUMAX exerts its greatest effect in and around the Aegean and Ionian subduction zones; high values of TAUMAX cause worse predictions of SHmax orientations and GPS velocities. These observations are not surprising because the complex pattern of the SHmax scoring dataset shown in Figure 7 suggests, for the Central and Eastern Mediterranean, a scenario where greater AF-EU convergence, locally expressed as AF-AS and AF-IO convergences, interacts with shorter-wavelength sources of stress (topography and/or lateral variations in crustal thickness) to re-orient SHmax to different angles than those expected in simple rigid-plate convergence. Similarly, GPS velocities describe a trench-ward, S-SE-oriented pattern in IO and AS that is anticorrelated with northward-oriented AF velocities. Both of these patterns in SHmax orientations and GPS measurements can be reproduced in Shells only with low values of TAUMAX, as in the case of our preferred model.

We found that the misfits with geodetic velocities also increase dramatically with increasing fault friction. As the modeled faults are mainly located along plate boundaries, increasing FFRIC and strength at plate boundaries implies higher rates of permanent strain in the intraplate lithosphere. If the opposite had been true, i.e., if the geodetic velocity misfits had decreased with increasing FFRIC, accommodating the deformation would have required a different fault dataset in the model. In other words, this result would have implied an incomplete plate model and fault dataset. The positive correlation between the misfit values and the fault friction thus indicates that our plate model and fault model are correct to the first order (Figure 8).

The low value of FFRIC of our preferred model can be ascribed to different phenomena distributed across the study area. Along the North Anatolia fault, dynamic weakening can lead to rupture propagation for a substantial distance across a statically unfavorable segment [Oglesby et al., 2008]. In the Northern Apennines and Carpathians, high fluid pressures and melt propagation may be weakening factors at the thrust interfaces [Faccenda et al., 2009]. Laboratory experiments suggest that elevated temperature, strain and the presence of pore fluids enhance frictional instability at seismogenic depths within carbonate–evaporite sequences, which are typical of the Mediterranean region [Scuderi et al., 2013]. Moreover, faults formed in weak clay-bearing rocks, which reduce friction, have been documented in different parts of the Mediterranean [Rutter et al., 1986; Solum and van der Pluijm, 2009; Tesei et al., 2012].

However, low FFRIC values can also have a modeling origin. The oversimplification of modeling closely spaced, parallel faults as a single fault requires a reduction in the friction of the modeled fault if the net fault slip rate is to be constant. We suspect that this issue is relevant in certain areas with complex fault patterns, such as the Betics or Dinarides, which are approximated in our work by a single fault. However, we believe that our resulting FFRIC and TAUMAX values, which are lower than laboratory results obtained from static experiments, are more than a model artifact. This point is noteworthy; it has strong implications for forward modeling as well as for seismic hazards because it influences the amount of deformation released seismically [Howe and Bird, 2010]. Because of these conflicting interpretations, we are not claiming uniformly low FFRIC to be a primary conclusion; rather, we consider this to be a starting point for future work investigating the local variations in both parameters, as performed by Kaneko et al. [2013].

The preferred model performs well in capturing plate velocities (Figure 9 and Figure S1, Figure S2, and Figure S3) and related fault heave rates (Figure 10 and Figure S4, Figure S5, and Figure S6). The velocity field of the preferred model, in the Eurasia reference frame, confirms the southward motion of the Aegean Sea, Central Greece, and Northern Greece plates. Across this area, our preferred model confirms the gradual increase in velocity instead of a sudden change localized along a single fault. The westward motion of the Anatolia plate is well reproduced in the preferred model, with the dextral North Anatolia fault slipping at approximately 20 mm/yr. The Adria plate is characterized by northeast-directed velocities respect to Eurasia. Extension along the Apennines mountain range is well resolved, even if in Northern Apennines it is spread over a bigger area. The Tyrrhenian coastline velocities show the typical north-to-northwestward direction described by Devoti et al. [2011] and Nocquet [2012]. The cause of this motion must be ascribed to the distant far-field effects of AF-EU convergence, which cannot be detected in Calabria because its velocity field is locally controlled by Ionian subduction. In northwestern Africa, the AF–EU convergence is partly accommodated in the Tell Atlas mountain range and partly in the south-dipping thrust offshore, where a new southward subduction zone appears to be in an early stage of development [Déverchère et al., 2005; Billi et al., 2011]. In the westernmost part of the Mediterranean, Iberia north of the Betics can be considered to be part of stable Eurasia, and in the Atlantic Ocean area, the velocity field appears to favor a scenario with deformation mainly localized along the AF-EU plate boundary.

Conversely, the most important local defects of the velocity field are as follows:

- The higher residuals in western Anatolia, supporting our notion of a more complex plate boundary and fault configuration between the Aegean Sea and Anatolia.
- The bias in velocity residuals in northern Greece, which are oriented between eastward and northeastward directions. This defect of the preferred model may be attributed to two causes: (1) the Southern Adriatic basin and the Albanides may be weaker and feature different material properties than assumed in our model and/or (2) the southern boundary of Adria may not be correctly mapped. These causes may interact with one another.
- The fictitious lithosphere-scale "landslides" of the Atlas Mountains toward the Mediterranean Sea, where we located the Eurasia-Africa plate boundary. Such lithospheric-scale gravity-tectonic artifacts have appeared in previous Shells models [e.g., Bird et al., 2008] and are well understood. They can be suppressed by local reductions in heat flow and/or increases in friction coefficients. However, such burdensome computations are not justified because the artifact is not clearly biasing the model scoring.
- The bias in velocity residuals in southern Iberia, which are westward directed. A different and more complex fault configuration, possibly with the insertion of an Alboran Sea microplate and a weakly active subduction front, as suggested by Gutscher et al. [2012], may reduce the misfit in this area.

The good representation among the plate velocities in our models was enforced by the first two modeling steps. In fact, we imposed velocity boundary conditions in the interior of each plate, computed the boundary-condition torque, and subsequently found a set of distributed basal tractions that exert the same torque. However, in the final step, we computed the side strength, lithostatic pressure, and basal shear tractions that lead to a physically plausible equilibrium. This convergence to equilibrium removes the initial apparent circularity, while the known information still contributes to determining the preferred model, which must be close to the truth and useful for seismicity forecasting.

Inevitably, the model’s local defects depend on our limited understanding of geodynamics and the variability of model parameters from one place to another.

The map of predicted permanent (non-elastic) intraplate strain rates confirms the widespread deformation in the Mediterranean region (Figure 11 and Figure S7, Figure S8, and Figure S9). The highest values are found in western Anatolia, the northern part of the Aegean Sea, Central Greece, and Northern Greece. In these areas, the intraplate deformation is very localized and kinematically consistent with respect to the active faults described in the European Database of Seismogenic Sources [Basili et al., 2013]. The strain rate found along the Tyrrhenian coastline of Italy is also a well-known aspect of the area, which deforms mainly aseismically due to a high heat flow. Large portions of Europe are rather stable and have very low strain rate values. The highest values for stable Eurasia are found where lateral variations in heat flow or topography are most pronounced (the Western Alps and Pannonian Basin).

Comparing stress azimuths with those in the preferred model, we found a very good fit for the whole Mediterranean area (Figure 12 and Figure S10, Figure S11, and Figure S12). First-order compressive stresses, due to Atlantic ridge push and Eurasia–Africa convergence, are also well reproduced. The SHmax orientations are uncorrelated mainly in the south of France and along the Atlantic passive margin. These local deviations from the predominant principal stress directions may reflect a regional variation not accounted for in the model material properties (rheological parameters) or an incorrect lithosphere thickness.

Concerning possible problems with our lithosphere structure model, we note the tendency of measured heat flow to increase at the coast (moving offshore) along the Atlantic margins. This tendency may be an artifact of different measurement methods, for example, deeper boreholes on land and shallow piston-cores in marine sediment. Shallow measurements are more strongly affected by the assumed histories of basal ocean temperatures during the Pleistocene ice ages and interglacials. Other, short-wavelength features (such as lateral density contrasts or diapirism), which may be missing from our structure model, would interfere with regional stresses and might cause SHmax reorientation depending on the stress magnitudes and angular differences. Typically, a regional-scale model similar to the one presented in this study is not able to capture small-scale variations in SHmax orientations. For this reason, our model does not reproduce the complex patterns of stress axis orientation, such as the pattern in the southeastern Carpathians [Müller et al., 2010].

An important question to address is whether the insertion of five new plates increased the realism of the preferred model with respect to simpler models. To address this idea, we performed two additional experiments. In the first experiment, we used a plate model with only the Aegean Sea and the Anatolia plates; the plate boundaries were located as in the MP2014 model. In this model, we essentially tested whether the indication of spread deformation to the north of the Aegean Sea and the northeastward movement of Adria relative to EU could be dismissed as being marginal. This model (named BASIC), with a simplistic plate configuration, has a performance of m_1 = 6.26 mm a^{-1} (*cf.* preferred model m_1 = 3.18 mm a^{-1}) and m_2 = 38.42° (*cf.* preferred model m_2 = 35.47°). These misfits indicate that a basic plate model with only Aegean Sea and Anatolia plates, ignoring the existence of other microplates in the Mediterranean, is less successful in reproducing the present-day deformation. In the second experiment, we only inserted the three microplates (i.e., MM, NG and CG) located to the north of the Aegean Sea plate; the Ionian Sea and Adria microplates were not included in this experiment. In this case, our aim was to discriminate how much of the m_1 and m_2 gain for the preferred model can be ascribed to a different plate configuration in the Central or in the Eastern Mediterranean. This second model (named NO_ADRIA) has m_1 = 3.90 mm a^{-1} (*cf.* preferred model m_1 = 3.18 mm a^{-1}) and m_2 = 38.29° (*cf.* preferred model m_2 = 35.47°). Considering the reduction in m_1 in the BASIC (m_1 =6.26 mm a^{-1}) and NO_ADRIA (m_1 =3.90 mm a^{-1}) configurations compared to the preferred model (m_1 =3.18 mm a^{-1}), it is clear that the more complex plate configuration north of the Aegean Sea plate is implied by geodetic measurements. Conversely, the abrupt reduction in m_2

in the NO_ADRIA model (m_2 = 38.29°) compared to the preferred model (m_2 = 35.47°) suggests that the Adria and Ionian Sea microplates are fundamental to reproducing SHmax orientations.

Another issue in our modeling work is the completeness and correctness of the fault network used in the experiment. This point becomes even more important when we compute long-term seismicity from the best model [Section 9]. We admit that the assumed geometries and kinematics of faults affect the results and that incorrect/incomplete models of active faults will bias the model, at least locally. However, this problem is common to all deformation models (and active fault databases), especially in areas of slow deformation. We point out that the bias due to missing active faults is notably less in Shells (where faults and the continuum have the same type of rheology) than in some other models (including any fault database) where the continuum is modeled as being rigid and/or purely-elastic and faults are represented as frictionless structures that take up all of the deformation. Consequently, we assume that the related long-term slip rate is represented in our model by an equivalent amount of off-fault deformation. The rationale behind Shells allows our preferred model to recover the deformation of a missing fault somewhere in the continuum near where the fault is missing. Similarly, an eventual deficit in fault slip rate is recovered in the elements near the fault itself. Such automatic compensation does not occur in other models, which commonly assume no permanent deformation between faults.

8. Discussion of Mediterranean geodynamics

Given the complexity of its plate boundaries, the Mediterranean region can be considered a natural laboratory where geophysicists have refined the plate tectonic theory since its formulation [McKenzie, 1970; McKenzie, 1972; Le Pichon and Angelier, 1979]. Eurasia–Africa convergence frames the present-day tectonic pattern, which involves different subduction zones at different stages of evolution and subducting plates with shapes that change over time.

In this work, we defined the MP2014 plate model by updating the PB2002 of Bird [2003] for the Mediterranean region via the insertion of new plates in accordance with new data made available over the last decade. Furthermore, we divided and classified all torques on plates into three types – side strength, basal strength, and lithostatic pressure – to better understand the contribution of each to the plate dynamics.

In this section, we first describe the improvements in terms of horizontal velocities and stress orientations of our preferred model with respect to previous models used to simulate the entire Mediterranean via the program Shells. Then, we discuss the findings in terms of force balances, with respect to the results of other models for the Mediterranean region published in the literature, focusing on the three main plates within the Mediterranean region: Adria, Aegean Sea, and Anatolia. Finally, as an introduction to the second part of this experiment, i.e., seismicity forecasting, we compare our preferred model to the Global Strain Rate Model version 2.1 [Kreemer et al., 2014], which has been commissioned by the Global Earthquake Model (GEM) Foundation as an additional resource for faults and earthquakes in seismic hazard studies.

8.1 Improvements with respect to previous Shells models

Several experiments using Shells have been carried out to model parts of the Mediterranean area [Jimenez-Munt 2001; Jiménez -Munt & Sabadini, 2002; Negredo et al., 2002; Jiménez -Munt & Negredo, 2003; Jiménez -Munt et al., 2003; Negredo et al., 2004; Barba et al., 2008, Kastelic and Carafa, 2012; Cunha et al., 2013], but to our knowledge, only two have considered the entire area: the regional experiment described by Jimenez-Munt et al. [2003] and the global model of Bird et al. [2008].

Several upgrades to the Shells code (e.g., a different formulation for defining the lithospheric thickness) and wider scoring datasets, especially GPS measurements, distance our experiment from those of Jimenez-Munt [2003]. In our work, we also did not rate models by the correlation between their long-term strain rates and the seismic strain rates calculated from an earthquake catalog, as done by Jimenez-Munt et al. [2003]. We could do so, by inserting a further scoring dataset, but we considered predicting the long-term seismicity of the Mediterranean with the preferred model to be a priority of our work. To avoid any circularity and keep the selection of the preferred model independent, we dropped this “seismic strain rate” scoring. Another issue is that the method used by Jimenez-Munt et al. [2003] does not consider likely variations in coupling due to different kinematics when determining long-term seismic strain rates starting from any Shells model (see Bird and Kagan 2004, Table 5). This reasoning further convinced us to omit any scoring based on correlation with any earthquake catalog.

Another difference is that our work models the whole Mediterranean in a single model with all forces transmitted self-consistently according to the theory underlying Shells, whereas in Jimenez-Munt [2003], the Mediterranean was divided into two models, with the horizontal velocities being applied along the trench regions of each model to simulate the subduction effects. Despite these differences, it is notable that our preferred model features low fault friction values similar to the best models of Jimenez-Munt et al. [2003].

Our experiment shares almost all the theory behind the Bird et al. [2008] study but uses a totally distinct finite element grid. The main differences in the Mediterranean between the two finite element grids are in the lithosphere structure because our experiment took advantage of a new seafloor age model [Mueller et al., 2008], new S-wave travel time tomography [Ritsema et al., 2011], a more refined fault database [Basili et al., 2013] and more detailed heat flow data. We also defined and used a newer plate model (MP2014) than the PB2002 model used in Bird et al. [2008]. Finally, in our experiment, we did not score models against the fast polarization azimuth φ of SKS waves due to uncertainties regarding the depths at which polarization occurs and the petrologic mechanism(s).

The origin of anisotropy is the deformed crystallographic fabric (lattice preferred orientation, LPO) of anisotropic minerals, primarily olivine, in the upper mantle [Karato et al., 2008]. Precisely, it is thought that the fast [100] olivine axis (A-type LPO) tends to align with the elongation axis under simple shear conditions. Consequently, φ may reflect simple shear across pseudo-horizontal planes in the asthenosphere. If so, φ could be used to score azimuths of basal shear traction implied by each Shells model. Unfortunately, there are several reasons to suspect that SKS polarizations in the Mediterranean do not result from A-type LPO. In fact, the water content plays a crucial role, and Karato et al. [2008] suggest that other LPO types (E or C) are more likely to be present in the asthenosphere and deep upper mantle. At the same time, the geological evolution of the Mediterranean suggests that the huge volume of subducted lithospheric material could erase, or at least strongly modify, the φ related to the asthenospheric flow. In addition, Mediterranean subduction is characterized by multiple slices of crust scraped off and delaminated from the mantle. As a consequence, the deepest part of thrusts can produce a strong anisotropy due to trench-parallel serpentinite-filled fractures [Jolivet et al. 2009]. For these reasons, in a complex framework such as the Mediterranean, it is better to develop ad hoc 3D models to link seismic anisotropy to the geological evolution of subduction and/or mantle kinematics than to assume a simplistic scenario to score neotectonic models with SKS azimuths [Baccheschi et al., 2009; Faccenna & Becker, 2010; Baccheschi et al., 2011; Faccenna et al., 2014].

To understand whether our preferred model exhibits a substantial improvement in the Mediterranean region, we scored the best model of Bird et al. [2008], namely, the Earth5-049 model, with the same datasets used in our experiment. We obtained m_1 = 6.88 mm a^{-1} and m_2 = 31.89°. These results suggest that our preferred model dramatically lowers (m_1 = 3.18 mm a^{-1} instead of m_1 = 6.88 mm a^{-1}) the misfit relative to

GPS measurements and slightly worsens the misfit relative to the interpolated SHmax dataset (m_2 = 35.47° instead of m_2 = 31.89°). Within the Mediterranean area, i.e., approximately 8°W – 36°E and 30°N – 46°N, the misfits become virtually identical, with m_2 = 32.32° for our preferred model and m_2 = 33.09° for the Earth5-049 model. Notably, the area affected by the primary changes from plate model PB2002 to MP2014 (5°E – 30°E and 30°N – 46°N), features m_2 = 30.89° for our preferred model and m_2 = 33.96° for the Bird et al. [2008] model. These results confirm two main points: the MP2014 model better describes the neotectonics of the Mediterranean, and our preferred model represents a substantial improvement with respect to the Earth5-049 model, with the possible exception of the Atlantic passive margin (mainly due to its incorrect lithospheric thickness).

8.2 Comparison with previous geodynamic models

The basal-strength torques, side-strength torques, and lithostatic-pressure torques for the inner plates (i.e., for the plates located entirely within the model) are displayed in Figure 13. To visualize the torques of each plate, we used the graphical representation of Bird et al. [2008]: a torque (whose rotational axis is geocentric) can be replaced by a point force anywhere on the great circle defined by that axis. All torque vectors acting on one plate are coplanar because their sum is zero. Consequently, their great circles intersect at two antipodal points (Figure 13), one of which is chosen for the plotting of these 3 vectors. Naturally, statically equivalent point forces are more concentrated than the actual applied tractions, but this simplification is useful when economically representing the force balances for several plates in one map.

As discussed in Bird et al. [2008], the resulting estimates of basal-strength traction (whether expressed as local shear stress, total horizontal force, or total torque) are only as good as the assumption that the friction of plate-bounding faults is laterally homogeneous. There is always a possibility that unmodeled variations in fault friction might be mapped into incorrect basal-strength stresses, forces, and torques. For this reason, the torque models shown in Figure 13 have to be seen as preliminary results, whereas the relative sizes of the torques are likely to be stable, at least for the Adria, Aegean Sea and Anatolia plates.

The preferred-model direction of the NE-directed basal-strength forces acting on Adria must be considered "forward" or "active," rather than "resistive" or "passive" in either the AF or the EU velocity reference frame. In contrast, the side-strength torque is SW directed and resistive. This drag is associated with fault friction along the Dinarides (compression) and Apennines (extension). Additionally, the topography is quite low in this region; thus, there should not be any large errors in computing the density moments that are summarized by the lithostatic-pressure vector., which does not suffice in reproducing the deformation pattern of the area. Given the low FFRIC value of our best model, we suspect that any other choice in FFRIC (i.e., greater values) will only enhance the SW-directed side-strength torque. Consequently, the AD basal-strength torque must also increase.

Therefore, regardless of uncertainty in the torque-triplet, our preferred model suggests that side-strength torques oppose the Adria motion and basal strength enhances it. Several studies agree on the importance of basal-shear tractions on Adria. These tractions are ascribed to different causes, possibly acting together, such as lateral variations in temperature, induced convection in subduction zones, and eastward mantle flow [Doglioni, 1990; Barba et al., 2008; Jolivet et al., 2009; Faccenna & Becker, 2010; Petricca et al., 2013, Faccenna et al.; 2014]. Expressing a different interpretation, D'Agostino et al. [2008] speculate that Adria, although subdivided into two rigid blocks, is driven by the edges, and forces such as basal tractions or horizontal gradients of gravitational potential energy, in their opinion, are not required. Indeed, the internal inconsistency of the GPS velocity field shows that no rigid-plate rotation can describe Adria's motion exactly; however, none of the proposed block subdivisions of Adria seem adequate to illustrate its complexity [Devoti et al., 2008]. This problem suggests that experiments dealing with a deformable crust, such as the one performed in this paper, are better suited to studying the geodynamics of Adria than rigid block models.

The preferred model confirms the importance of the northeastward basal tractions on Adria. We recall that the basal torques in Shells include the widely distributed basal strength and any anomalous tractions on a plate caused by an attached subducting slab (see Section 3). This “definition” depends on the fact that in Shells, the “base of the plate” is defined for each element as horizontal, even though this reference surface cuts through strong slabs of down-going oceanic lithosphere. Unfortunately, Shells cannot differentiate how much of Adria’s basal torque comes from attached slabs and how much comes from widely distributed shear tractions. However, in agreement with Faccenna et al. [2014], we definitively reject the concept of D’Agostino et al. [2008], in which side-strength torques and block interactions drive Adria, which is fragmented as two microplates.

Our preferred model also has strong implications for torque orientations in the Aegean Sea. As explained in Section 7, although the low and constant FFRIC and TAUMAX in the preferred model may be unrealistic, given the lithostatic pressure force directed toward the trench, any increase in FFRIC or TAUMAX consequently increases the side-strength force and the trench-ward basal-strength torque. In fact, increasing FFRIC along the northern boundaries of the Aegean Sea would require an increase in the basal strength to fit the long-term horizontal velocities and the equilibrium equation. Similarly, increasing TAUMAX would imply an increase in the basal strength opposing the African compression along the Hellenic subduction zone. Therefore, the preferred model suggests that the Aegean Sea extension partly originates from basal traction and partly from variations in lithostatic pressure, whereas the side-strength forces, mainly due to Africa’s motion, trend in the nearly opposite direction and are resistive. The smaller plates of Northern Greece and Central Greece also have substantial SE-directed basal-traction forces, whereas their lithostatic-pressure forces rapidly decrease to the north. Our preferred hypothesis is that the retreat of the slab induces a flow in the mantle that in turn creates gradients in the lithosphere thickness (mainly in Aegean Sea and Anatolia plates) and, thus, important pressure and shear forces directed toward the Hellenic trench. Our preferred scenario agrees with Kreemer et al. [2004] and Jolivet et al. [2009]; the SKS anisotropy vectors, aligned with the direction of extension in crustal stretching, indicate that the flow in the mantle drives the deformation in the Aegean Sea. Similarly, the conclusions of Le Pichon and Kreemer [2010], ascribing the deformation in the Aegean and the large-scale pattern in the geodetic velocities to a counter-clockwise toroidal flow in the asthenosphere, agree with the results of our preferred model. We also note that, to match deformation indicators, Özeren and Holt [2010] imposed a southwestward pulling force of 4×10^{11} Nm^{-1} located along the Hellenic Trench. Along with our result of TAUMAX = 1×10^{12} Nm^{-1}, we take the resulting trench-pull from Özeren and Holt [2010] as further evidence that the lithostatic-pressure forces must act together with basal tractions to balance the Africa-push side-strength force. Furthermore, the Aegean region shows a complex pattern of localized deformation (Figure 13); this pattern suggests that strong inherited crustal heterogeneities, both in thickness and rheology, play a crucial role by acting as weak zones during backarc extension [Jolivet et al., 2013].

For the Anatolia plate, the deviatoric stresses of the preferred model agree with those of Özeren and Holt [2010] in both style and magnitude. Our results clarify the roles of lithostatic-pressure forces (equivalent to differences in gravitational potential energy across a plate) and the side-strength forces, which are equivalent to the boundary forces of Özeren and Holt [2010]. We found that the lithostatic-pressure and the side-strength forces must have similar values to match the scoring datasets: the push of Arabia and Africa cancels out most of the lithostatic forces. Similarly, Özeren and Holt [2010] found an equivalent contribution from gravitational potential energy (GPE) differences and boundary conditions (analogous to density buoyancies and tractions located outside of their model domain) to the deviatoric stress field in Anatolia. However, these results contrast with those of Provost et al. [2003], who suggested that GPE differences have little influence within the region. Our results show that the equilibrium among torques is due to a small southwest-directed basal-traction force, which may originate from the sinking and/or retreat of the Hellenic slab and the consequent large-scale toroidal flow in the mantle [as in Faccenna & Becker, 2010].

8.3 Comparison with the Global Strain Rate Model

Recently, the purely geodetic Global Strain Rate Model (GSRM v2.1) [Kreemer et al, 2014] has been commissioned by the GEM Foundation as an objective addition to fault and earthquake data in seismic hazard studies. As such, GSRM v2.1 is a valid competitor in the Mediterranean to our preferred model, which is used for seismicity prediction in the second part of our experiment. Consequently, a comparison with GSRM v2.1 is necessary to understand how reliable (although preliminary) our preferred model may be with respect to the actual global-scale reference deformation model.

We downloaded the “Average Strain Rate values”, which are available at http://www.globalquakemodel.org/what/seismic-hazard/strain-rate-model/ (last access 31st January 2015). This dataset describes the average horizontal strain rate tensor for each cell with 0.25°/0.2° longitude/latitude spacing. Only cells within “deformation regions” are listed in this file and used for further analyses and comparisons between the GSRM v2.1 and our preferred model. Areas outside the “deforming regions” are assigned in GSRM v2.1 to rigid plates with no internal deformation.

In the first comparison, which is within the Mediterranean area, i.e., approximately 8°W – 36°E and 30°N – 46°N, we calculated the second invariant of the horizontal plane strain rate for each cell *i* in GSRM v2.1 as

$$\dot{\varepsilon}_{GSRM}(i) = \sqrt{(\dot{\varepsilon}_{1h})^2 + (\dot{\varepsilon}_{2h})^2} \tag{7}$$

We similarly calculated $\dot{\varepsilon}_{model}(i)$ using the strain rate tensor of each 0.25°-0.20° sized cell as the tensor of the Shells finite element in which the central point of each GSRM cell fell. The performance of the preferred model relative to GSRM v2.1 in the "deforming region" of the Mediterranean is given by the normalized strain rate correlation coefficient ($CORREL$) between the logarithm of GSRM v2.1 and preferred model second invariants:

$$CORREL = \frac{\sum_{i=1}^{N}\left[\log \dot{\varepsilon}_{GSRM}(i) - \overline{\log \dot{\varepsilon}_{GSRM}}\right] \cdot \left[\log \dot{\varepsilon}_{model}\,(i) - \overline{\log \dot{\varepsilon}_{model}}\right]}{\sqrt{\sum_{i=1}^{N}\left[\log \dot{\varepsilon}_{GSRM}(i) - \overline{\log \dot{\varepsilon}_{GRSM}}\right] \cdot \sum_{i=1}^{N}\left[\log \dot{\varepsilon}_{model}\,(i) - \overline{\log \dot{\varepsilon}_{model}}\right]}} \tag{8}$$

where the overbar designates the average value of the function over the model region, and $N = 5745$ is the number of cells inside the Mediterranean "deforming region". $CORREL$ can have negative values (up to -1) if the two models are uncorrelated, whereas positive values (up to 1) can be attained if they are correlated. For our preferred model with respect to GSRM v2.1, we obtained $CORREL = 0.36$, indicating modest agreement in the spatial distribution of the strain rate in the Mediterranean "deforming region". Perhaps many of the differences between these two models are related to the fact that strain-rates in GSRM v2.1 may be elastic or permanent (in unknown proportion), whereas strain-rates in Shells are always permanent by definition. We performed a further comparison between the GSRM v2.1 smallest eigenvalue azimuths with the SHmax orientations of our preferred model, obtaining a mean difference of 30.96°. To test the overall agreement of both models with respect to the measured SHmax data, we created a new SHmax interpolated dataset using SHINE [Carafa et al., 2015] and following the same procedure described in Subsection 6.2. This new SHmax dataset describes the SHmax azimuth at the center of each GSRM v2.1 cell in the Mediterranean "deforming region". The average misfit between the GSRM v2.1 smallest eigenvalue azimuth and the SHmax azimuth is 26.70°, whereas an average discrepancy of 29.41° was found between this new dataset and our preferred model. These two comparisons suggest that GRSM v2.1 and our preferred model are very close in terms of strain rate localization and stress orientations, with a slight preference for GRSM v2.1 in reproducing the interpolated SHmax in the Mediterranean "deforming region". However, GRSM v2.1 is a final product commissioned by the GEM

Foundation, whereas our preferred model is a preliminary one with several adjustments that are required and planned for the future.

Given the general agreement between these two models, we believe that a similar relevance can be assigned to our preferred model. Planned future improvements will only make it more relevant in terms of seismicity. Furthermore, in the broader Mediterranean area, i.e., outside the GSRM v2.1 "deforming region", GSRM v2.1 assumes a rigid plate behavior with no internal deformation; thus, the model is not relevant for understanding seismicity and its link with geodesy, whereas our preferred model remains relevant and useful because our model is capable of resolving continuum/intraplate off-fault deformation.

9. Seismic potential

9.1 Long-term seismicity

Since the introduction of the theory of plate tectonics, it has been widely expected to provide a basis for the quantitative prediction of long-term average seismicity and seismic hazard. In the last decade, knowledge of the structure and dynamics of the Earth have improved, and fully 3-D models of great theoretical interest have been developed. Unfortunately, this progress in the geosciences has rarely led to numerical deformation models that can compete with the older empirical-statistical approach, which basically projects past earthquakes into the future. Each numerical model adequately fitting several geophysical data (e.g., GPS measurements, SHmax orientations, and SKS azimuths) must have the aspiration of realizing the promise of plate tectonics theory and providing added value in seismic hazard assessments. For this reason, as a fundamental part of this project, fault slip rates and distributed permanent strain rates of the preferred model were converted to maps of expected seismicity in the area delimited by meridians 24°W – 35°E and parallels 30°N – 52°N (Figure 14). This calculation was performed using the program Long_Term_Seismicity_v3, which realizes the two hypotheses of the

Seismic Hazard Inferred From Tectonics (SHIFT) model [Bird & Liu, 2007; Bird et al., 2010]: (a) calculations of the seismic moment rate for any deforming volume should use the coupled seismogenic thickness of the most comparable type of plate boundary and (b) calculations of the earthquake rate should use the seismic moment rate in conjunction with the frequency-magnitude distribution describing the most comparable type of plate boundary. The seismicity coefficients (coupled seismogenic lithosphere thickness, corner magnitude, and asymptotic spectral slope of the tapered Gutenberg-Richter frequency/moment relationship) were determined by Bird and Kagan [2004] for classes of plate boundaries and updated by Bird et al. [2009] for subduction zones and continental convergent boundaries. An algorithm for assigning the "most comparable" plate boundary can be found in Bird and Liu [2007, Tables 1, 2]. However, the Hellenic and the Ionian subduction zones seem to be partially coupled, i.e., part of the deformation is released through some creeping mechanism [e.g., Papadopoulos, 1989]. Howe and Bird [2010] ascribe aseismic creep to the presence of Messinian evaporites, which could lubricate the subduction interfaces and reduce the seismic coupling factor to approximately 0.38, which describes the seismicity expected from plate-bending alone. The unusually large amount of evaporites makes the use of globally averaged earthquake productivity factors for subduction zones inappropriate. Thus, we used the globally averaged values from Bird and Kagan [2004] and Bird et al. [2009], with the exception of the coupling factor for the Hellenic and the Ionian subduction zones, where we adopted the value of 0.38 from Howe and Bird [2010]. Calculated earthquake rates for M_w>6 (Data Set S5), M_w>6.5 (Data Set S6), M_w>7 (Data Set S7), and M_w>8 (Data Set S8) are available in Supporting Information.

9.2 Earthquake rate forecast comparison

Our forecasts of long-term seismicity are independent of seismic catalogs because seismic catalogs were not used as inputs or to score the models. Primarily, our forecasts depend on the long-term rates of fault slip and the permanent deformation predicted based on bounding-plate motions, plate shapes, fault networks, nonlinear mantle and crustal rheology, topography, and heat flow. Because of this independence, it is not circular reasoning to test our forecast via a comparison with existing seismic catalogs, although the comparison is retrospective.

For this purpose, we refer to the European shallow earthquakes that occurred in Europe during years 1000–2006 CE (SHEEC), which were compiled by the EU project SHARE [Giardini et al., 2013]. The catalog consists of three portions. The first one was produced by Stucchi et al. [2013] for the years 1000-1899 CE. The spatial parameters are derived from either the macroseismic data points or the most updated regional catalogs, according to their reliability, and the magnitudes are weighted means. The relationships between the earthquake parameters and the macroseismic data points were regionally calibrated against a set of recent instrumental earthquakes. The catalog includes approximately 5000 events with magnitudes that range from M_w 4.0 to M_w 8.5. A second portion of the SHEEC catalog includes instrumental recordings of earthquakes that occurred during 1900–2006 CE, and this portion was produced by Grünthal et al. [2013]. This catalog included data records from Grünthal and Wahlström [2012] (with some inclusions from Italy [CPTI Working Group, 2008] and the North Atlantic), where the records related to Balkans have been replaced with those from Makropoulos et al. [2012]. The third catalog includes 12,331 earthquakes with magnitude values from M_w 2.5 to M_w 8.3 that occurred in central and eastern Turkey during 1000–2006 CE [Sesetyan et al., 2013]. The magnitude values were mainly taken from the Ambraseys and Jackson [1998] dataset and the ISC catalog.

In SHEEC, the assessment of completeness was based on a historical approach [Stucchi et al., 2011; 2013] that measures the capability of the earthquakes to leave traces in historical accounts. This approach does not assume that the seismogenic process is stationary; rather, the gaps in the earthquake history may indicate a lack of either earthquakes or historical sources. Based on this concept, Stucchi et al. [2013] first assumed that the earthquakes with $M_w \geq 6.8$ are likely to leave historical traces in

Europe and then considered the absence of such traces as incompleteness. Stucchi et al. [2013] found that the completeness interval for earthquakes of $M_w \geq 6.8$ starts as early as year 1200 CE in certain places, such as northern Italy, the Marmara region, and the Betic region, and as late as 1800 CE in western Turkey. Similarly, the start of the completeness interval for earthquakes of $M_w \geq 5.8$ ranges from 1300 CE (e.g., northern Italy) to later than 1900 CE (e.g., offshore Portugal and Aegean Sea).

9.3 Historical versus forecast earthquake rates

A serious question pertaining to any model is whether it predicts the appropriate rates of great ($M_w > 8$), major ($M_w > 7$), and strong ($M_w > 6$) earthquakes. Because the catalog record is short and includes only a modest number of such events and the model earthquake rates depend on the uncertainties in Shells and SHIFT parameters, we will not attempt to compare their map patterns in detail here and will limit our discussion to total rates in the region.

The comparison trapezoids represent supersets of the regions that have been used by Stucchi et al. [2013] for the completeness assessment of SHEEC. Within these trapezoids, we fix a time interval for earthquakes of $M_w > 6$ and $M_w > 7$ that is shorter and, therefore, more conservative than the completeness intervals of Stucchi et al. [2013], both for the threshold magnitudes (for which we choose higher values) and the start year of the completeness interval (which we set to be more recent). Within each region (Figure 14), we count the number of earthquakes that occurred during the observation period (after a certain year and during the completeness interval) and whose magnitudes are larger than or equal to the completeness threshold. We list in Table 2 the number of earthquakes within the observation period, expressed as earthquakes per century. In the same regions, we determine the number of earthquakes per century (above the threshold magnitude) predicted by the model using the SHIFT hypothesis.

For strong ($M_w > 6$) earthquakes, the number of earthquakes per century is more than 260. As a comparison, the model predicts approximately 280 earthquakes (region

ALLEU; Table 2). In the model area, the highest observed rate is approximately 70 earthquakes per century, corresponding to approximately two-thirds of the earthquake rate that is predicted by the model in the same area in the case of fully coupled subduction. For the Aegean plate, the presence of evaporites may lubricate the Aegean subduction and facilitate creeping mechanisms [Howe & Bird, 2010]. Here, assuming a coupling factor that is lower than the average produces a misfit that is reasonably small for earthquakes of magnitude $M_w > 7$. In Turkey and southern Spain, the model predicts a number of earthquakes per century that is between 50% and 100% larger than the number observed in SHEEC. In Italy, the observed and predicted rates nearly coincide.

Major ($M_w > 7$) earthquakes were recorded 26 times per century, or approximately every 4 years. This number is probably only a lower limit of the true rate because the completeness threshold of the catalog may be higher in unpopulated regions, such as the eastern Atlantic Ocean. The predicted rate is approximately 30, which is slightly higher than SHEEC. However, the possible creeping behavior of certain subduction zones would reduce the rate of $M_w > 7$ earthquakes. For most of the comparison zones, the discrepancy between observed and predicted is approximately 20%.

For great ($M_w > 8$) earthquakes, the observed rate is very low (approximately 0.12 events per century in the Hellenic subduction), and only two events of $M_w > 8$ are known: one in Crete in 365 A.D. [Shaw et al., 2008] and one east of Crete in 1303 A.D. [Guidoboni et al., 1994; Stucchi et al., 2007]. This observed rate is much less than our model predictions (0.42 events per century). However, the observed rate is more than double the predicted rate for the eastern Atlantic along the AF–EU plate boundary. These comparisons favor the hypothesis of creeping Hellenic subduction, but we recommend that this hypothesis be tested further with focused geodetic and seismic-array studies.

Considering the big picture, we can summarize the discrepancies between the observed and predicted number of earthquakes per century as being of order 50%. The distribution of earthquakes per century, corresponding to classes of magnitude from M_w

> 5 to M_w > 8 (in Figure 15), is dense around the bisecting line (predicted = observed), with a slight tendency of the model to overestimate the number of earthquakes. These 50% discrepancies in seismicity forecasts versus the catalog dataset are not surprising given the uncertainties involved in the computation, which include the magnitude calculations (see, e.g., Guidoboni et al. [1994]) and the completeness estimates. However, a strategy to lower the bias in overestimating the number of earthquakes must be implemented in the future.

The major differences between the predicted and observed rates may occur when large uncertainties affect both the crustal structure and the catalog data. Otherwise, the approach appears to be viable. The use of the seismic catalog as a calibration dataset requires us to embrace the seismic coupling factors from global analogs [Bird & Kagan, 2004; Bird et al., 2009; Howe & Bird, 2010]. However, the seismic coupling factor strongly affects the predicted rates. Within the SHIFT assumptions, the largest uncertainty in the forecasted seismic moment rate seems to result from the uncertainties in the coupled thickness (<c z> in Table 5 of Bird & Kagan [2004]), which in turn originates from the uncertainties in corner magnitudes. For these reasons, the comparison between the earthquake rates from the model output and catalog data represents a first-order attempt, whereas site-specific applications require a separate validation of the seismic coupling. Additionally, the choice of the tapered Gutenberg-Richter frequency/moment relation affects the predicted rates, especially for the great (M_w > 8) earthquakes. Here, the model systematically overestimates the number of M_w > 8 earthquakes with respect to the seismic catalog. However, this result may depend on the limited length of the historical catalog and on the slow deformation rates found in Europe. The neotectonic models presented here furnish an independent estimate of the probability of great (M_w > 8) earthquakes in Europe, especially over timescales that are greater or much greater than the seismic catalog length.

10. Conclusions

We modeled the neotectonics of the eastern Atlantic and Mediterranean areas using the latest release of Shells [Bird et al., 2008], a well-established code for producing robust and realistic long-term deformation models. This release was not previously tested on a regional scale, but it proved useful for modeling the complexity of the Mediterranean region.

We first defined a new plate model, revising and updating Bird's [2003] plate model by adding the quasi-rigid plates of the Ionian Sea, Adria, Northern Greece, Central Greece and Marmara to better detail the complex and diffuse deformation pattern of the area. We then integrated updated datasets (S-wave global tomography, heat flow measurements, crustal thicknesses, SHmax orientations and GPS measurements) to both generate the deformation models and evaluate their level of realism.

We conclude that the interaction of side and basal strengths drives the deformation in the Adria and Aegean Sea plates, whereas the lithostatic stress plays a critical role in Anatolia.

The preferred model shown in this work requires further improvements; its use for seismicity prediction represents a preliminary experiment that requires additional tuning of modeling parameters and additional testing on different plate boundaries and fault scenarios, e.g., Betics - Atlas. Moreover, our idea for improving the seismicity prediction of our deformation models is to follow a well-established procedure for quantifying the spatial uncertainty in predicted seismicity rates. At the present, the earthquake rates provided in this work should not be used in regions where inconsistencies with the scoring data sets have been identified. However, for majority of the study area shows consistency among data and model predictions and no apparent artifact in the preferred model. For these areas numerical models, e.g., the preferred model presented in this paper, can aspire to be a fresh and objective tool for probabilistic earthquake (and tsunami) hazard estimation. The inclusion of neotectonic models in the logic tree, even if they are assigned small weights, alleviates the need for long-term paleoseismic catalogs, especially in slowly deforming regions, and provides independent information where the seismic catalogs exhibit spatial gaps with respect to the tectonics.

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Calculations to obtain lithosphere thickness were performed using a slightly modified version of OrbData code, available at http://peterbird.name/oldFTP/neotec/. Finite element models were obtained using a slightly modified version of SHELLS code, available at http://peterbird.name/oldFTP/neotec/SHELLS/. The marginal modifications to the OrbData and SHELLS codes are explained in Section 4 and Table 1. Geodetic misfit m_1 and stress discrepancy m_2 were obtained using OrbScore code, available at http://peterbird.name/oldFTP/neotec/SHELLS/OrbScore/ . Earthquake rate forecasts were performed using the Long_Term_seismicity code, available at http://peterbird.name/oldFTP/Long_Term_Seismicity/ . Finite element grid and preferred model formats were modified to obtain seismicity forecast using Long_Term_Seismicity. The MP2014 plate model, the finite element grid (containing final values of topography, heat flow density, crustal thickness, upper mantle thickness, density anomaly due to petrologic variations, and curvature of the geotherm due to transient heating or cooling), the base map of heat flow density used where detailed maps were not available, the preferred model, the earthquake forecasts for Mw>6, Mw>6.5, Mw>7, and Mw>8 are available as in Supporting Information.

References

Aktug, B. et al. (2009), Deformation of western Turkey from a combination of permanent and campaign GPS data: Limits to block-like behavior, *J. Geophys. Res.*, *114*(B10), B10404, doi:10.1029/2008JB006000.

Akyuz, H. S., G. Ucarkus, E. Altunel, B. Dogan, and A. Dikbas (2013), Paleoseismological investigations on a slow-moving active fault in central Anatolia, Tecer Fault, Sivas, *Ann. Geophys.*, *55*(5), 847–857, doi:10.4401/ag-5444.

Amante, C., and B. W. Eakins (2009), ETOPO1 1 Arc-Minute Global Relief Model: Procedures, Data Sources and Analysis, *NOAA Tech. Memo. NESDIS NGDC*, *19*, 24, doi:10.1594/PANGAEA.769615.

Ambraseys, N. N., and J. A. Jackson (1998), Faulting associated with historical and recent earthquakes in the Eastern Mediterranean region, *Geophys. J. Int.*, *133*(2), 390–406, doi:10.1046/j.1365-246X.1998.00508.x.

Armijo, R., B. Meyer, G. C. P. King, A. Rigo, and D. Papanastassiou (1996), Quaternary evolution of the Corinth Rift and its implications for the Late Cenozoic evolution of the Aegean, *Geophys. J. Int.*, *126*(1), 11–53, doi:10.1111/j.1365-246X.1996.tb05264.x.

Artemieva, I. (2011), *The lithosphere: An Interdisciplinary Approach*, Cambridge University Press, New York.

Baccheschi, P., L. Margheriti, M. S. Steckler (2009), SKS splitting in Southern Italy: Anisotropy variations in a fragmented subduction zone, Tectonophysics, 462(1-4), doi: 10.1016/j.tecto.2007.10.014.

Baccheschi, P., L. Margheriti, M. Steckler, E. Boschi (2011), Anisotropy patterns in the subducting lithosphere and in the mantle wedge. A case study: the Southern Italy subduction system, J. Geophys. Res., 116(B8), 5,435 – 5,443, doi: 10.1029/2010JB007961.

Barba, S., M. M. C. Carafa, and E. Boschi (2008), Experimental evidence for mantle drag in the Mediterranean, *Geophys. Res. Lett.*, *35*(6), L06302, doi:10.1029/2008GL033281.

Basili, R. et al. (2013), The European Database of Seismogenic Faults (EDSF) compiled in the framework of the Project SHARE, , doi:10.6092/INGV.IT-SHARE-EDSF. Available from: http://diss.rm.ingv.it/share-edsf/

Bassin, C., G. Laske, and G. Masters (2000), The current limits of resolution for surface wave tomography in North America, *Eos Trans.*, *81*(48), Fall Meet. Suppl., Abstract S12A–03.

Baumgardner, J. R. (1983), A three-dimensional finite element model for mantle convection,Ph.D. dissertation, 271 pp., University of California Los Angeles.

Bendick, R. and L.M. Flesch (2013), A review of heterogeneous materials and their implications for relationships between kinematics and dynamics in continents, *Tectonics*, DOI: 10.1002/tect.20058.

Bertotti, G., R. Capozzi, and V. Picotti (1997), Extension controls Quaternary tectonics, geomorphology and sedimentation of the N-Appennies foothills and adjacent Po Plain (Italy), *Tectonophysics*, *282*(1–4), 291–301, doi:10.1016/S0040-1951(97)00229-1.

Billi, A., C. Faccenna, O. Bellier, L. Minelli, G. Neri, C. Piromallo, D. Presti, D. Scrocca, and E. Serpelloni (2011), Recent tectonic reorganization of the Nubia-Eurasia convergent boundary heading for the closure of the western Mediterranean, *Bull. la Soc. Geol. Fr.*, *182*(4), 279–303, doi:10.2113/gssgfbull.182.4.279.

Bird, P. (1998), Testing hypotheses on plate-driving mechanisms with global lithosphere models including topography, thermal structure, and faults, *J. Geophys. Res.*, *103*(B5), 10115–10129, doi:10.1029/98JB00198.

Bird, P. (2003), An updated digital model of plate boundaries, *Geochemistry, Geophys. Geosystems*, *4*(3), 1–52, doi:10.1029/2001GC000252.

Bird, P., and X. Kong (1994), Computer simulations of California tectonics confirm very low strength of major faults, *Geol. Soc. Am. Bull.*, *106*(2), 159–174.

Bird, P., and Y. Kagan (2004), Plate-tectonic analysis of shallow seismicity: Apparent boundary width, beta, corner magnitude, coupled lithosphere thickness, and coupling in seven tectonic settings, *Bull. Seismol. Soc. Am.*, *94*(6), 2380–2399, doi:10.1785/0120030107.

Bird, P. & Li, Y., 1996. Interpolation of principal stress directions by nonparametric statistics: Global maps with confidence limits, J. Geophys. Res., 101(B3), 5,435 – 5,443.

Bird, P., and Z. Liu (2007), Seismic hazard inferred from tectonics: California, *Seismol. Res. Lett.*, *78*(1), 37–48, doi:10.1785/gssrl.78.1.37.

Bird, P., C. Kreemer, and W. Holt (2010), A long-term forecast of shallow seismicity based on the Global Strain Rate Map, *Seismol. Res. Lett.*, *81*(2), 184–194, doi:10.1785/gssrl.81.2.184.

Bird, P., Y. Y. Kagan, D. D. Jackson, F. P. Schoenberg, and M. J. Werner (2009), Linear and Nonlinear Relations between Relative Plate Velocity and Seismicity, *Bull. Seismol. Soc. Am.*, *99*(6), 3097–3113, doi:10.1785/0120090082.

Bird, P., Z. Ben-Avraham, G. Schubert, M. Andreoli, and G. Viola (2006), Patterns of stress and strain rate in southern Africa, *J. Geophys. Res.*, *111*(B08402), 1–14, doi:10.1029/2005JB003882.

Bird, P., Z. Liu, and W. Rucker (2008), Stresses that drive the plates from below: Definitions, computational path, model optimization, and error analysis, *J. Geophys. Res.*, *113*(B11406), 1–32, doi:10.1029/2007JB005460.

Bozkurt, E. (2000), Timing of extension on the Büyük Menderes Graben, western Turkey, and its tectonic implications, *Geol. Soc. London, Spec. Publ.*, *173*, 385–403, doi:10.1144/GSL.SP.2000.173.01.18.

Calais, E., J. Nocquet, F. Jouanne, and M. Tardy (2002), Current strain regime in the Western Alps from continuous Global Positioning System measurements, 1996–2001, *Geology*, *3*(7), 651–654, doi:10.1130/0091-7613(2002)030<0651:CSRITW>2.0.CO;2.

Caputo, R., A. Chatzipetros, S. Pavlides, and S. Sboras (2013), The Greek Database of Seismogenic Sources (GreDaSS): state-of-the-art for northern Greece, *Ann. Geophys.*, *55*(5), doi:10.4401/ag-5168.

Carafa, M. M. C., and S. Barba (2013), The stress field in Europe: optimal orientations with confidence limits, *Geophys. J. Int.*, *193*(2), 531–548, doi:10.1093/gji/ggt024.

Carafa, M. M. C., G. Tarabusi and V. Kastelic (2015), SHINE: Web Application for Determining the Horizontal Stress Orientation, *Computers & Geosciences*, doi: 10.1016/j.cageo.2014.10.001.

Catalano, R., Dogloni, C., and S. Merlini (2001), On the Mesozoic Ionian basin, *Geophysical Journal International*, 144, 49–64.

Channell, J. E. T. (1996), Palaeomagnetism and palaeogeography of Adria, *Geol. Soc. London, Spec. Publ.*, *105*(1), 119–132, doi:10.1144/GSL.SP.1996.105.01.11.

Chiarabba, C., L. Jovane, and R. Di Stefano (2005), A new view of Italian seismicity using 20 years of instrumental recordings, *Tectonophysics*, *395*(3), 251–268, doi:10.1016/j.tecto.2004.09.013.

Console, R., R. Di Giovambattista, P. Favali, B. W. Presgrave, and G. Smriglio (1993), Seismicity of the Adriatic microplate, *Tectonophysics*, *218*(4), 343–354, doi:10.1016/0040-1951(93)90323-C.

Corti, G., M. Cuffaro, C. Doglioni, F. Inoocenti, and P. Manetti (2006), Coexisting geodynamic processes in the Sicily Channel, *Geol. Soc. Am. Spec. Pap.*, *409*, 83–96, doi:10.1130/2006.2409(05).

Cunha, T. A., L. M. Matias, P. Terrinha, A. M. Negredo, F. Rosas, R. M. S. Fernandes, and L. M. Pinheiro (2012), Neotectonics of the SW Iberia margin, Gulf of Cadiz and Alboran Sea: a reassessment including recent structural, seismic and geodetic data, *Geophys. J. Int.*, *188*(3), 850–872, doi:10.1111/j.1365-246X.2011.05328.x.

D'Agostino, N., A. Avallone, D. Cheloni, E. D'Anastasio, and S. Mantenuto (2008), Active tectonics of the Adriatic region from GPS and earthquake slip vectors, *J. Geophys. Res.*, *113*(B12), doi:10.1029/2008JB005860.

D'Agostino, N., E. D'Anastasio, A. Gervasi, I. Guerra, M. R. Nedimović, L. Seeber, and M. Steckler (2011), Forearc extension and slow rollback of the Calabrian Arc from GPS measurements, *Geophys. Res. Lett.*, *38*(17), L17304, doi:10.1029/2011GL048270.

Del Ben, A., I. Finetti, F. Mongelli, and G. Zito (1994), Seismic and heat flow study of the southern Adriatic Basin, *Boll. di Geofis. Teor. e Appl.*, *36*(141-144), 29–44.

Della Vedova, B., S. Bellani, G. Pellis, and P. Squarci (2001), Deep temperatures and surface heat flow distribution, in *Anatomy of an orogen, The Apennines and adjacent Mediterranean basins*, edited by G. B. Vai and I. P. Martini, Kluwer Academy Publishers, , The Netherlands.

Déverchère, J. et al. (2005), Active thrust faulting offshore Boumerdes, Algeria, and its relations to the 2003 Mw 6.9 earthquake, *Geophys. Res. Lett.*, *32*(4), L04311, doi:10.1029/2004GL021646.

Devoti, R., A. Esposito, and G. Pietrantonio (2011), Evidence of large scale deformation patterns from GPS data in the Italian subduction boundary, *Earth Planet. Sci. Lett.*, *311*(3-4), 230–241, doi:10.1016/j.epsl.2011.09.034.

Devoti, R., F. Riguzzi, M. Cuffaro, and C. Doglioni (2008), New GPS constraints on the kinematics of the Apennines subduction, *Earth Planet. Sci. Lett.*, *273*(1-2), 163–174, doi:10.1016/j.epsl.2008.06.031.

Di Bucci, D., D. Ridente, U. Fracassi, F. Trincardi, and G. Valensise (2009), Marine palaeoseismology from very high resolution seismic imaging: the Gondola Fault Zone (Adriatic foreland), *Terra Nov.*, *21*(5), 393–400, doi:10.1111/j.1365-3121.2009.00895.x.

DISS Working Group (2010), Database of Individual Seismogenic Sources (DISS), Version 3.1.1: A compilation of potential sources for earthquakes larger than M 5.5 in Italy and surrounding areas, , doi:10.6092/INGV.IT-DISS3.1.1.

Doglioni, C. (1990), The global tectonic pattern, *J. Geodyn.*, *12*(1), 21–38, doi:10.1016/0264-3707(90)90022-M.

Duarte, J.C., Rosas, F.M., Terrinha, P., Schellart, W.P., Boutelier, D., Gutscher, M.A., Ribeiro, A. (2013), Are subduction zones invading the Atlantic? Evidence from the SW Iberia margin. *Geology*, doi:10.1130/G34100.1.

Ekström, G., M. Nettles, and A. M. Dziewoński (2012), The global CMT project 2004–2010: Centroid-moment tensors for 13,017 earthquakes, *Phys. Earth Planet. Inter.*, *200-201*, 1–9, doi:10.1016/j.pepi.2012.04.002.

Faccenda, M., G. Minelli and T.V. Gerya (2009), Coupled and decoupled regimes of continental collision: numerical modeling, *Earth and Planet. Sci. Lett.*, 278 (2009), 337–349.

Faccenna, C., and T. Becker (2010), Shaping mobile belts by small-scale convection, *Nature*, *465*(7298), 602–605, doi:10.1038/nature09064.

Faccenna, C. et al. (2014), Mantle dynamics in the Mediterranean, *Rev. Geophys.*, 52(3), 2013RG000444, doi:10.1002/2013RG000444.

Field, E. H. et al. (2013), Unified California Earthquake Rupture Forecast, version 3 (UCERF3)-The time-independent model, *U.S. Geol. Surv. Open-File Rep., 2013-1165(Cal. Geol. Surv. Spec. Rep. 228, South. Calif. Earthq. Cent. Pub. 1792),* 97 pages. Available from: http://pubs.usgs.gov/of/2013/1165/.

Floyd, M. A. et al. (2010), A new velocity field for Greece: Implications for the kinematics and dynamics of the Aegean, *J. Geophys. Res.*, *115*(B10), B10403, doi:10.1029/2009JB007040.

Frizon de Lamotte, D., Saint Bezar, B., Bracene, R. and Mercier, E., (2000)- The two main steps of the Atlas building and Geodynamics of West Mediterranean, *Tectonics*, 19, 740-761.

Garthwaite, M.C. and G.A. Houseman (2011), Validity of the thin viscous sheet approximation in models of continental collision, J. Geophys. Res., 116(B2): doi: 10.1029/2010JB007770.

Giardini, D., J. Wössner and L. Danciu (2014), Mapping Europe's Seismic Hazard, *Eos Trans. AGU*, 95(29), 261.

Giardini, D., J. Woessner, L. Danciu, G. Valensise, G. Grünthal, F. Cotton, S. Akkar, R. Basili, M. Stucchi, A. Rovida, D. Stromeyer, R. Arvidsson, F. Meletti, R. Musson, R., K. Sesetyan, M. B. Demircioglu, H. Crowley, R. Pinho, K. Pitilakis, J. Douglas, J. Fonseca, M. Erdik, A. Campos-Costa, B. Glavatovic, K. Makropoulos, C. Lindholm, T. Cameelbeeck, Seismic Hazard Harmonization in Europe (SHARE): Online Data Resource, doi: 10.12686/SED-00000001-SHARE, 2013.Goes, S., D. Giardini, S. Jenny, C. Hollenstein, H. Kahle, and A. Geiger (2004), A recent tectonic reorganization in the south-central Mediterranean, *Earth Planet. Sci. Lett.*, *226*(3-4), 335–345, doi:10.1016/j.epsl.2004.07.038.

Grünthal, G., and R. Wahlström (2012), The European-Mediterranean Earthquake Catalogue (EMEC) for the last millennium, *J. Seismol.*, *16*(3), 535–570, doi:10.1007/s10950.

Grünthal, G., R. Wahlström, and D. Stromeyer (2013), The SHARE European Earthquake Catalogue (SHEEC) for the time period 1900–2006 and its comparison to the European-Mediterranean Earthquake Catalogue (EMEC), *J. Seismol.*, *17*(4), 1339–1344, doi:10.1007/s10950-013-9379-y.

Guidoboni, E., Comastri, A., Traina, G. (1994), Catalogue of ancient earthquakes in the Mediterranean area up to the 10th Century. ING-SGA, Rome-Bologna, Italy, 504.

Duarte, J.C., Rosas, F.M., Terrinha, P., Schellart, W.P., Boutelier, D., Gutscher, M.A., Ribeiro, A. (2013), Are subduction zones invading the Atlantic? Evidence from the SW Iberia margin. *Geology*, doi:10.1130/G34100.1.

Heidbach, O., M. Tingay, A. Barth, J. Reinecker, D. Kurfeß, and B. Müller (2008), The World Stress Map database release 2008, , doi:10.1594/GFZ.WSM.Rel2008.

Howe, T. M., and P. Bird (2010), Exploratory models of long-term crustal flow and resulting seismicity across the Alpine-Aegean orogen, *Tectonics*, *29*(4), TC4023, doi:10.1029/2009TC002565.

Jaupart, C., S. Labrosse, and J. C. Mareschal (2007), Temperatures, Heat and Energy in the Mantle of the Earth, in *Treatise on Geophysics*, vol. 7, edited by G. Schubert, pp. 253–303, Elsevier, Amsterdam.

Jiménez-Munt, I., and A. M. Negredo (2003), Neotectonic modelling of the western part of the Africa–Eurasia plate boundary: from the Mid-Atlantic ridge to Algeria, *Earth Planet. Sci. Lett.*, *205*(3), 257–271, doi:10.1016/S0012-821X(02)01045-2.

Jiménez-Munt, I., and R. Sabadini (2002), The block-like behavior of Anatolia envisaged in the modeled and geodetic strain rates, *Geophys. Res. Lett.*, *29*(20), 1–4, doi:10.1029/2002GL015995.

Jiménez-Munt, I., M. Fernàndez, M. Torne, and P. Bird (2001a), The transition from linear to diffuse plate boundary in the Azores–Gibraltar region: results from a thin-sheet model, *Earth Planet. Sci. Lett.*, *192*(2), 175–189, doi:10.1016/S0012-821X(01)00442-3.

Jiménez-Munt, I., P. Bird, and M. Fernàndez (2001b), Thin-shell modeling of neotectonics in the Azores-Gibraltar Region, *Geophys. Res. Lett.*, *28*(6), 1083–1086, doi:10.1029/2000GL012319.

Jiménez-Munt, I., R. Sabadini, A. Gardi, and G. Bianco (2003), Active deformation in the Mediterranean from Gibraltar to Anatolia inferred from numerical modeling and geodetic and seismological data, *J. Geophys. Res.*, *108*(B1), 1–24, doi:10.1029/2001JB001544.

Jolivet, L. et al. (2013), Aegean tectonics: Strain localisation, slab tearing and trench retreat, *Tectonophysics*, *597-598*, 1–33, doi:10.1016/j.tecto.2012.06.011.

Jolivet, L., C. Faccenna, and C. Piromallo (2009), From mantle to crust: Stretching the Mediterranean, *Earth Planet. Sci. Lett.*, *285*(1-2), 198–209, doi:10.1016/j.epsl.2009.06.017.

Kaneko Y, Fialko Y, Sandwell DT, Tong X, and M. Furuya (2013) Interseismic deformation and creep along the central section of the North Anatolian fault (Turkey): InSAR observations and implications for rate-and-state friction properties, *J Geophys Res*, 118, 316–331, doi:10.1029/2012JB009661

Karato, S.-i., Jung, H., Katayama, I., and Skemer, P., (2008), Geodynamic significance of seismic anisotropy of the upper mantle: new insights from laboratory studies. Annu. Rev. Earth Planet. Sci. 36, 59–95. http://dx.doi.org/10.1146/annurev.earth.36.031207.124120.

Kastelic, V., and M. M. C. Carafa (2012), Fault slip rates for the active External Dinarides thrust-and-fold belt, *Tectonics, 31*, TC3019, doi:10.1029/2011TC003022.

Koçyiğit, A., and A. Beyhan (1998), A new intracontinental transcurrent structure: the Central Anatolian Fault Zone, Turkey, *Tectonophysics*, *284*(3), 317–336, doi:10.1016/S0040-1951(97)00176-5.

Koçyiğit, A., and A. Beyhan (1999), Reply to Rob Westaway's Comment on "A new intracontinental transcurrent structure: the Central Anatolian Fault Zone, Turkey," *Tectonophysics*, *314*(4), 481–496, doi:10.1016/S0040-1951(99)00227-9.

Kong, X., and P. Bird (1995), SHELLS: A thin-shell program for modeling neotectonics of regional or global lithosphere with faults, *J. Geophys. Res.*, *100*(B11), 22129, doi:10.1029/95JB02435.

Koulali, A., D. Ouazar, A. Tahayt, R. W. King, P. Vernant, R. E. Reilinger, S. McClusky, T. Mourabit, J. M. Davila, and N. Amraoui (2011), New GPS constraints on active deformation along the Africa–Iberia plate boundary, *Earth Planet. Sci. Lett.*, *308*(1), 211–217, doi:10.1016/j.epsl.2011.05.048.

Kreemer, C., G. Blewitt, E.C. Klein (2014), A geodetic plate motion and Global Strain Rate Model, *Geochemistry, Geophysics, Geosystems*, 15, 3849-3889, doi:10.1002/2014GC005407.

Kreemer, C., N. Chamot-Rooke, and X. Le Pichon (2004), Constraints on the evolution and vertical coherency of deformation in the Northern Aegean from a comparison of geodetic, geologic and seismologic data, *Earth Planet. Sci. Lett.*, *225*(3), 329–346, doi:10.1016/j.epsl.2004.06.018.

Kukkonen, I. ., and R. Lahtinen (2001), Variation of radiogenic heat production rate in 2.8–1.8 Ga old rocks in the central Fennoscandian shield, *Phys. Earth Planet. Inter.*, *126*(3), 279–294, doi:10.1016/S0031-9201(01)00261-8.

Lavecchia, G., F. Ferrarini, and R. De Nardis (2007), Active thrusting as a possible seismogenic source in Sicily (Southern Italy): Some insights from integrated structural–kinematic and seismological data, *Tectonophysics*, *445*(3-4), 145–167, doi:10.1016/j.tecto.2007.07.007.

Le Pichon, X., and C. Kreemer (2010), The Miocene-to-Present Kinematic Evolution of the Eastern Mediterranean and Middle East and Its Implications for Dynamics, *Annu. Rev. Earth Planet. Sci.*, *38*(1), 323–351, doi:10.1146/annurev-earth-040809-152419.

Lechmann S.M., D.A. May, B.J.P. Kaus, S.M. Schmalholz (2011), Comparing thin-sheet models with 3D multilayer models for continental collision. Geophysical Journal International. 187, 10-33. doi: 10.1111/j.1365-246X.2011.05164.x

Leeds, A. R., L. Knopoff, and E. G. Kausel (1974), Variations of Upper Mantle Structure under the Pacific Ocean, *Science*, *186*(4159), 141–143, doi:10.1126/science.186.4159.141.

Lenkey, L., P. Dövényi, F. Horváth, and S. A. P. L. Cloetingh (2002), Geothermics of the Pannonian basin and its bearing on the neotectonics, *Stephan Mueller Spec. Publ. Ser.*, *3*.

Lister, C. R. B., J. G. Sclater, S. Nagihara, E. E. Davis, and H. Villinger (1990), Heat flow maintained in ocean basins of great age - Investigations in the north-equatorial West Pacific, *Geophys. J. Int.*, *102*, 603–630, doi:10.1111/j.1365-246X.1990.tb04586.x.

Liu, Z., and P. Bird (2002a), Finite element modeling of neotectonics in New Zealand, *J. Geophys. Res.*, *107*(B12), 1–18, doi:10.1029/2001JB001075.

Liu, Z., and P. Bird (2002b), North America plate is driven westward by lower mantle flow, *Geophys. Res. Lett.*, *29*(24), 2–5, doi:10.1029/2002GL016002.

Louvari, E., A. Kiratzi, and B. Papazachos (1999), The Cephalonia transform fault and its extension to western Lefkada Island (Greece), *Tectonophysics*, *308*, 223–236, doi:10.1016/S0040-1951(99)00078-5.

Makropoulos, K., G. Kaviris, and V. Kouskouna (2012), An updated and extended earthquake catalogue for Greece and adjacent areas since 1900, *Nat. Hazards Earth Syst. Sci.*, *12*(5), 1425–1430, doi:10.5194/nhess-12-1425-2012.

Malinverno, A. (2012), Evolution of the Tyrrhenian Sea-Calabrian Arc system: The past and the present, *Rend. online Soc. Geol. It.*, *21*, 11–15.

Malservisi, R., U. Hugentobler, R. Wonnacott, and M. Hackl (2013), How rigid is a rigid plate? Geodetic constraint from the TrigNet CGPS network, South Africa, *Geophys. J. Int.*, *192*, 918–928, doi:10.1093/gji/ggs081.

Mansinha, L., and D. E. Smylie (1967), Effect of earthquakes on the Chandler wobble and the secular polar shift, *J. Geophys. Res.*, *72*(18), 4731–4743, doi:10.1029/JZ072i018p04731.

Mansinha, L., and D. E. Smylie (1971), The displacement fields of inclined faults, *Bull. Seismol. Soc. Am.*, *61*(5), 1433–1440.

Martínez-Díaz J. J. , E Masana, M Ortuño (2012), Active tectonics of the Alhama de Murcia fault, Betic Cordillera, Spain, *Journal of Iberian Geology*, 38 (1), 253-271.

Mattia, M., V. Bruno, F. Cannavò, and M. Palano (2012), Evidences of a contractional pattern along the northern rim of the Hyblean Plateau (Sicily, Italy) from GPS data, *Geol. Acta*, 10, 1–9, doi:10.1344/105000001705.

McKenzie, D. P. (1972), Active tectonics of the Mediterranean region, *Geophys. J. R. Astron. Soc.*, *30*, 109–185, doi:10.1111/j.1365-246X.1972.tb02351.x.

McKenzie, D. P. (1970), Plate tectonics of the Mediterranean region., *Nature*, *226*(5242), 239–43, doi:10.1038/226239a0.

Meghraoui, M., and F. Doumaz (1996), Earthquake-induced flooding and paleoseismicity of the El Asnam, Algeria, fault-related fold, *J. Geophys. Res.*, *101*(B8), 17617, doi:10.1029/96JB00650.

Milivojević, M. G. (1993), Geothermal model of Earth's crust and lithosphere for the territory of Yugoslavia: Some tectonic implications, *Stud. Geophys. Geod.*, *37*(3), 265–278, doi:10.1007/BF01624600.

Molinari, I., and A. Morelli (2011), EPcrust: a reference crustal model for the European Plate, *Geophys. J. Int.*, *185*(1), 352–364, doi:10.1111/j.1365-246X.2011.04940.x.

Molinari, I., V. Raileanu, and A. Morelli (2012), A Crustal Model for the Eastern Alps Region and a New Moho Map in Southeastern Europe, *Pure Appl. Geophys.*, *169*, 1575–1588, doi:10.1007/s00024-011-0431-y.

Müller, B., O. Heidbach, M. Negut, B. Sperner, and T. Buchmann (2010), Attached or not attached—evidence from crustal stress observations for a weak coupling of the Vrancea slab in Romania, *Tectonophysics*, *482*(1), 139–149, doi:10.1029/2001JB000743.

Müller, R. D., M. Sdrolias, C. Gaina, and W. R. Roest (2008), Age, spreading rates, and spreading asymmetry of the world's ocean crust, *Geochemistry, Geophys. Geosystems*, *9*(4), 1–19, doi:10.1029/2007GC001743.

Müller, R. D., W. R. Roest, J.-Y. Royer, L. M. Gahagan, and J. G. Sclater (1997), Digital isochrons of the world's ocean floor, *J. Geophys. Res. Solid Earth*, *102*(B2), 3211–3214, doi:10.1029/96JB01781.

Musumeci, C., Scarfì, L., Palano, M.,and D. Patanè (2014), Foreland segmentation along an active convergent margin: New constraints in southeastern Sicily (Italy) from seismic and geodetic observations, *Tectonophysics*, 630, 137–149, 10.1016/j.tecto.2014.05.017

Negredo, A. M., I. Jiménez-Munt, and A. Villaseñor (2004), Evidence for eastward mantle flow beneath the Caribbean plate from neotectonic modeling, *Geophys. Res. Lett.*, *31*(6), doi:10.1029/2003GL019315.

Negredo, A. M., P. Bird, C. Sanz de Galdeano, and E. Buforn (2002), Neotectonic modeling of the Ibero-Maghrebian region, *J. Geophys. Res.*, *107*(B11), doi:10.1029/2001JB000743.

Nocquet, J.-M. (2012), Present-day kinematics of the Mediterranean: A comprehensive overview of GPS results, *Tectonophysics*, *579*, 220–242, doi:10.1016/j.tecto.2012.03.037.

Nyst, M., and W. Thatcher (2004), New constraints on the active tectonic deformation of the Aegean, *J. Geophys. Res.*, *109*(B11), B11406, doi:10.1029/2003JB002830.

Ocakoğlu, F. (2007), A re-evaluation of the Eskişehir Fault Zone as a recent extensional structure in NW Turkey, *J. Asian Earth Sci.*, *31*(2), 91–103, doi:10.1016/j.jseaes.2007.05.002.

Oglesby, D. D., P. M. Mai, K. Atakan, and S. Pucci (2008), Dynamic models of earthquakes on the North Anatolian fault zone under the Sea of Marmara: Effect of hypocenter location, Geophys. Res. Lett., 35, L18302, doi:10.1029/2008GL035037.

Özeren, M., and W. Holt (2010), The dynamics of the eastern Mediterranean and eastern Turkey, *Geophys. J. Int.*, 1165–1184, doi:10.1111/j.1365-246X.2010.04819.x.

Özsayin, E., and K. Dirik (2009), Quaternary Activity of the Cihanbeyli and Yeniceoba Fault Zones: İnönü-Eskişehir Fault System, Central Anatolia, *Turkish J. Earth Sci.*, *16*(4), 471–492.

Papadopoulos, G.A (1989), Seismic and volcanic activities and aseismic movements as plate motion components in the Aegean area, *Tectonophysics*, 167, 31-39, 1989.

Papadopoulos, G.A., D. Kondopoulou, G.-A. Leventakis and S. Pavlides. Seismotectonics of the Aegean region. Tectonophysics, 124, 67-84, 1986.

Pasquale, V., M. Verdoya, and P. Chiozzi (1996), Heat flux and timing of the drifting stage in the Ligurian-Provençal basin (northwestern Mediterranean), *J. Geodyn.*, *21*(3), 205–222, doi:10.1016/0264-3707(95)00035-6.

Pérouse, E., N. Chamot-Rooke, A. Rabaute, P. Briole, F. Jouanne, I. Georgiev, and D. Dimitrov (2012), Bridging onshore and offshore present-day kinematics of central and eastern Mediterranean: Implications for crustal dynamics and mantle flow: Central to East Mediterranean kinematics, *Geochemistry, Geophys. Geosystems*, *13*(9), doi:10.1029/2012GC004289.

Petricca, P., M. M. C. Carafa, S. Barba, and E. Carminati (2013), Local, regional, and plate scale sources for the stress field in the Adriatic and Periadriatic region, *Mar. Pet. Geol.*, *42*, 160–181, doi:10.1016/j.marpetgeo.2012.08.005.

Piccardi, L., Y. Gaudemer, P. Tapponnier, and M. Boccaletti (1999), Active oblique extension in the central Apennines (Italy): evidence from the Fucino region, *Geophys. J. Int.*, *139*(2), 499–530, doi:10.1046/j.1365-246x.1999.00955.x.

Platt, J. P., W. M. Behr, K. Johanesen, and J. R. Williams (2013), The Betic-Rif Arc and Its Orogenic Hinterland: A Review, *Annu. Rev. Earth Planet. Sci.*, *41*(1), 313–357, doi:10.1146/annurev-earth-050212-123951.

Pondrelli, S., C. Piromallo, and E. Serpelloni (2004), Convergence vs. retreat in Southern Tyrrhenian Sea: Insights from kinematics, *Geophys. Res. Lett.*, *31*, 2–5, doi:10.1029/2003GL019223.

Pondrelli S., Salimbeni S., Morelli A., Ekström G., Postpischl L., Vannucci G. and Boschi E. (2011) European-Mediterranean Regional Centroid Moment Tensor Catalog: solutions for 2005-2008, *Phys. Earth Planet. Int.*, 185, 3-4, doi :10.1016/j.pepi.2011.01.007.

Presti, D., A. Billi, B. Orecchio, and C. Totaro (2013), Earthquake focal mechanisms, seismogenic stress, and seismotectonics of the Calabrian Arc, Italy, *Tectonophysics*, *602*, 153–175, doi:10.1016/j.tecto.2013.01.030.

Provost, A., J. Chéry, and R. Hassani (2003), 3D mechanical modeling of the GPS velocity field along the North Anatolian fault, *Earth Planet. Sci. Lett.*, *209*, 361–377, doi:10.1016/S0012-821X(03)00099-2.

Radulov, A. (2007), Active faults in the Plovdiv Depression and their long-term slip rate, *Geol. Balc.*, *36*, 51–56.

Reilinger, R. et al. (2006), GPS constraints on continental deformation in the Africa-Arabia-Eurasia continental collision zone and implications for the dynamics of plate interactions, *J. Geophys. Res.*, *111*(B5), B05411, doi:10.1029/2005JB004051.

Ritsema, J., A. Deuss, H. J. van Heijst, and J. H. Woodhouse (2011), S40RTS: a degree-40 shear-velocity model for the mantle from new Rayleigh wave dispersion, teleseismic traveltime and normal-mode splitting function measurements, *Geophys. J. Int.*, *184*, 1223–1236, doi:10.1111/j.1365-246X.2010.04884.x.

Ritsema, J., and H. J. Van Heijst (2000), Seismic imaging of structural heterogeneity in Earth’s mantle: evidence for large-scale mantle flow, *Sci. Prog.*, *83*(3), 243–259.

Rovida, A., R. Camassi, P. Gasperini, and M. Stucchi (2011), CPTI11, the 2011 version of the Parametric Catalogue of Italian Earthquakes, , doi:10.6092/INGV.IT-CPTI11. Available from: http://emidius.mi.ingv.it/CPTI

Rutter, E. H., R. H. Maddock, S. H. Hall, and S. H. White (1986),Comparative microstructures of natural and experimentally produced clay-bearing fault gouges, Pure Appl. Geophys., 124, 3 – 30.

Sartori, R. (2004), The Tyrrhenian back-arc basin and subduction of the Ionian lithosphere, *Episodes*, *26*(3), 217–221.

Scisciani, V., and F. Calamita (2009), Active intraplate deformation within Adria: Examples from the Adriatic region, *Tectonophysics*, *476*(1-2), 57–72, doi:10.1016/j.tecto.2008.10.030.

Scuderi, M., Niemeijerc, A. R., Collettini, C. and C. Marone (2013), Frictional properties and slip stability of active faults within carbonate–evaporite sequences: The role of dolomite and anhydrite, *Earth Planet. Sci. Lett.*, 369, 220–232.

Serpelloni, E., G. Vannucci, S. Pondrelli, A. Argnani, G. Casula, M. Anzidei, P. Baldi, and P. Gasperini (2007), Kinematics of the western Africa-Eurasia plate boundary from focal mechanisms and GPS data, *Geophys. J. Int.*, doi:10.1111/j.1365-246X.2007.03367.x.

Sesetyan, K., M. Demircioglu, A. Rovida, P. Albini, and M. Stucchi (2013), SHARE-CET, the SHARE earthquake catalogue for Central and Eastern Turkey complementing the SHARE European Earthquake Catalogue (SHEEC), Available from: http://www.emidius.eu/SHEEC/

Solum, J.G. and B.A. van der Pluijm (2009) Quantification of fabrics in clay gouge from the Carboneras fault, Spain and implications for fault behavior *Tectonophysics*, 475 (3–4), 554–562.

Shaw, B., N. N. Ambraseys, P. C. England, M. A. Floyd, G. J. Gorman, T. F. G. Higham, J. A. Jackson, J.-M. Nocquet, C. C. Pain, and M. D. Piggott (2008), Eastern Mediterranean tectonics and tsunami hazard inferred from the AD 365 earthquake, *Nat. Geosci.*, *1*(4), 268–276, doi:10.1038/ngeo151.

Stein, C. A., and S. Stein (1992), A model for the global variation in oceanic depth and heat flow with lithospheric age, *Nature*, *359*(6391), 123–129, doi:10.1038/359123a0.

Stucchi, M. et al. (2013), The SHARE European Earthquake Catalogue (SHEEC) 1000–1899, *J. Seismol.*, *17*(2), 523–544, doi:10.1007/s10950-012-9335-2.

Stucchi, M., C. Meletti, V. Montaldo, H. Crowley, G. M. Calvi, and E. Boschi (2011), Seismic Hazard Assessment (2003-2009) for the Italian Building Code, *Bull. Seismol. Soc. Am.*, *101*(4), 1885–1911, doi:10.1785/0120100130.

Stucchi, M., R. Camassi, A. Rovida, M. Locati, E. Ercolani, C. Meletti, P. Migliavacca, F. Bernardini, and R. Azzaro (Editors) (2007). DBMI04, il database delle osservazioni macrosismiche dei terremoti italiani utilizzate per la compilazione del catalogo parametrico CPTI04 , Quaderni di Geofisica, Istituto Nazionale di Geofisica eVulcanologia, Rome, Vol. 49, 38 pp., doi:10.6092/INGV.IT-DBMI04,http://emidius.mi.ingv.it/DBMI04/ (last accessed January 2014).

Ten Veen, J. H., J. M. Woodside, T. A. . Zitter, J. F. Dumont, J. Mascle, and A. Volkonskaia (2004), Neotectonic evolution of the Anaximander Mountains at the junction of the Hellenic and Cyprus arcs, *Tectonophysics*, *391*, 35–65, doi:10.1016/j.tecto.2004.07.007.

Tesei, T., Collettini, C., Carpenter, B. M., Viti, C., Marone C. (2012), Frictional strength and healing behavior of phyllosilicate-rich faults *J. Geophys. Res.*, 117, B09402 http://dx.doi.org/10.1029/2012JB009204.

Tortorici, L., C. Monaco, C. Tansi, and O. Cocina (1995), Recent and active tectonics in the Calabrian arc (Southern Italy), *Tectonophysics*, *243*(1-2), 37–55, doi:10.1016/0040-1951(94)00190-K.

Vanneste, K., A. Radulov, P. De Martini, G. Nikolov, T. Petermans, K. Verbeeck, T. Camelbeeck, D. Pantosti, D. Dimitrov, and S. Shanov (2006), Paleoseismologic investigation of the fault rupture of the 14 April 1928 Chirpan earthquake (*M* 6.8), southern Bulgaria, *J. Geophys. Res.*, *111*(B1), doi:10.1029/2005JB003814.

Vergés, J., and M. Fernàndez (2012), Tethys–Atlantic interaction along the Iberia–Africa plate boundary: The Betic–Rif orogenic system, *Tectonophysics*, *579*, 144–172, doi:10.1016/j.tecto.2012.08.032.

Vernant, P., A. Fadil, T. Mourabit, D. Ouazar, A. Koulali, J. M. Davila, J. Garate, S. McClusky, and R. Reilinger (2010), Geodetic constraints on active tectonics of the Western Mediterranean: Implications for the kinematics and dynamics of the Nubia-Eurasia plate boundary zone, *J. Geodyn.*, *49*(3), 123–129, doi:10.1016/j.jog.2009.10.007.

Villani, F., and S. Pierdominici, (2010), Late Quaternary tectonics of the Vallo di Diano basin (southern Apennines, Italy). Quaternary Science Reviews, 29, 3167-3183, doi: 10.1016/j.quascirev.2010.07.003

Westaway, R. (1999), Comment on “A new intracontinental transcurrent structure: the Central Anatolian Fault Zone, Turkey” by A. Koçyiğit and A. Beyhan, *Tectonophysics*, *314*(4), 469–479, doi:10.1016/S0040-1951(99)00223-1.

Yielding, G., M. Ouyed, G. C. P. King, and D. Hatzfeld (1989), Active tectonics of the Algerian Atlas Mountains-evidence from aftershocks of the 1980 El Asnam earthquake, *Geophys. J. Int.*, *99*(3), 761–788, doi:10.1111/j.1365-246X.1989.tb02057.x.

Zitellini, N., E. Gràcia, L. Matias, P. Terrinha, M. A. Abreu, G. DeAlteriis, J. P. Henriet, J. J. Dañobeitia, D. G. Masson, and T. Mulder (2009), The quest for the Africa–Eurasia plate boundary west of the Strait of Gibraltar, *Earth Planet. Sci. Lett.*, *280*(1-4), 13–50, doi:10.1016/j.epsl.2008.12.005.

Table captions

Table 1. Euler poles used in this work. The Euler pole of the Ionian Sea microplate is deduced by using the GPS stations of Nocquet et al. [2012] falling inside this microplate. We used the code vplate [Malservisi et al., 2013] sharply approximating the GPS measurements to not be affected by elastic accumulation.

Table 2. Earthquakes per century in the study region. Observed: counted within the SHEEC catalog; Predicted: determined within the best model. (*) The seismic coupling factor is 0.38 [Howe & Bird, 2010]. (#) Completeness areas from Stucchi et al. [2013]. Region outlines not included in Stucchi et al. [2013] paper are avaibles as Data Set S9 in Supporting Information.

Figure captions

Figure 1. Topography of the Mediterranean region. In red: the plate boundaries of the MP2014 plate model used in this work. NG: Northern Greece; CG: Central Greece. Focal mechanisms from European-Mediterranean Regional Centroid Moment Tensor [Pondrelli et al, 2011 and reference therein], available at http://www.bo.ingv.it/RCMT/ (last accessed, 31st January 2015)

Figure 2. Grid of the 2-D spherical-triangle finite elements used in this study. Gray lines represent edges of continuum elements; typical triangles are 55 km wide. Red lines are faults along plate boundaries or in orogens. Tick marks indicate the assigned dip: filled triangle 18° (seismogenic portion of a subduction zone), open triangle 30° (other thrust faults), box 45°, straight 55°, none 90°. Contour bands: ETOPO1 relief model [Amante & Eakins, 2009]. Mercator projection. NG: Northern Greece; CG: Central Greece; MM: Marmara. Numbers in the map refer to the position of active faults discussed in Section 2 and inserted in the finite element grid. Numbers increase toward the east. In

alphabetic order: Alhama de Murcia fault (1); Apennines external thrust (3); Central Anatolian fault (14); Corinth Rift (8); Inonu-Eskisehir fault (11); Ionian subduction (6); Kefalonia fault (7); Kutahya fault (10); North Anatolia fault zone (12); Sicilian basal thrust (5); South-Alpine thrust (2); and Southern Tyrrhenian thrust (4).

Figure 3. Heat flow of the finite element grid (available as Data Set S1 in supporting Information), obtained with (1) model heat flow from Bird et al. [2008], Jaupart et al. [2007], and Lister et al. [1990] (based on seafloor ages from Muller et al. [2008]), (2) heat flow interpolated from data in the "International Heat Flow Commission database" (last revised January 12, 2011; http://www.heatflow.und.edu/index2.html) and (3) published maps of certain areas [Milivojević, 1993; Pasquale et al., 1996; Della Vedova et al., 2001; Lenkey et al., 2002].

Figure 4. Anomaly in the travel time of virtual vertically propagating S-waves from the S40RTS model [Ritsema et al., 2011], corrected for the crustal thickness to provide the upper mantle anomaly at depths no greater than 400 km.

Figure 5. Lithosphere thickness, including crustal thickness. In continental areas, the mantle-lithosphere thicknesses were scaled from travel-time anomalies in the upper mantle (Figure 4). In the oceans, an age-dependent model was adopted.

Figure 6. GPS residual velocities with respect to the Eurasia reference frame [Nocquet [2012]. Horizontal velocities falling inside the area delimited by blue-dotted lines are scaled to increase map readability.

Figure 7. Stress azimuths (SHmax) resulting from the interpolation of data [World Stress Map 2008, Heidbach et al., 2008] to a regular global grid after declustering [Carafa & Barba, 2013] using the web tool SHINE [Carafa et al., 2015], available at

shine.rm.ingv.it. Strategic parameters used in SHINE: search radius = 5 (181 Km), minimum cluster number = 3; 90% confidence bounds = 45°. The 90% confidence bounds are color coded.

Figure 8. Geodetic misfit m_1, stress discrepancy m_2, and combined misfits for the 480 candidate models. The combined misfit of each model is computed by the geometric mean ($\sqrt[2]{m_1 \cdot m_2}$). The preferred model misfits are marked with “x”.

Figure 9. Long-term horizontal velocity field. Black arrows and contour bands: velocity resulting from the preferred model. Velocities falling inside the area delimited by white lines are scaled to increase map readability.

Figure 10. Predicted long-term rates of fault heave from the preferred model. Rates falling inside the area delimited by black lines are scaled to increase map readability.

Figure 11. Scalar strain rates in the preferred model. Per construction, these are predicted long-term permanent strain rates (non elastic) and are equivalent to the non-rotational portions of the velocity gradients in Figure 9. Common logarithm of the size of the largest principal strain rate (s^{-1}). Symbols indicate the predicted type and orientation of conjugate microfaults. Rectangles: normal faults (extension); dumbbells: thrust faults (compression); Xs: strike-slip faults.

Figure 12. Predicted stress azimuths (SHmax), assumed to be parallel to the most compressive horizontal strain-rates, from the preferred model. Color indicates the stress regime: red – normal faulting; green – strike-slip faulting; blue – thrust faulting. In grey: the scoring dataset; see Figure 7 for details.

Figure 13. Fictitious point forces (3 per plate) that are statically equivalent to side-strength, basal-strength, and lithostatic-pressure torques from the preferred model. CG: Central Greece; IO: Ionian Sea; MM: Marmara; NG: Northern Greece.

Figure 14. Long-term seismicity prediction for magnitudes $M_w > 6$ for the central part of the model domain according to the SHIFT model. Colors represent the common logarithm of the rate of shallow earthquakes ($M_w > 6$) in SI units predicted by the preferred model after applying the SHIFT assumptions [Bird & Liu, 2007].

Figure 15. Observed versus predicted number of earthquakes per century for classes of magnitude from $M_w > 5$ to $M_w > 8$ (see Table 2 for $M_w > 6$). Crosses: large regions (available as Data Set S9 in Supporting Information); squares: completeness areas from Stucchi et al. [2013]. The logarithmic scale permits the inclusion of areas and earthquakes of largely varying sizes in the same plot. See Table 2 for the numeric discrepancies.