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Invited review

A mantle convection perspective on global tectonics

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ABSTRACT

The concept of interplay between mantle convection and tectonics goes back to about a century ago, with the proposal that convection currents in the Earth's mantle drive continental drift and deformation (Holmes, 1931). Since this time, plate tectonic theory has established itself as the fundamental framework to study surface deformation, with the remarkable ability to encompass geological and geophysical observations. Mantle convection modeling has progressed to the point where connections with plate tectonics can be made, pushing the idea that tectonics is a surface expression of the global dynamics of one single system: the mantle-lithosphere system. Here, we present our perspective, as modelers, on the dynamics behind global tectonics with a focus on the importance of self-organisation. We first present an overview of the links between mantle convection and tectonics at the present-day, examining observations such as kinematics, stress and deformation. Despite the numerous achievements of geodynamic studies, this section sheds light on the lack of self-organisation of the models used, which precludes investigations of the feedbacks and evolution of the mantle-lithosphere system. Therefore, we review the modeling strategies, often focused on rheology, that aim at taking into account self-organisation. The fundamental objective is that plate-like behaviour emerges self-consistently in convection models. We then proceed with the presentation of studies of continental drift, seafloor spreading and plate tectonics in convection models allowing for feedbacks between surface tectonics and mantle dynamics. We discuss the approximation of the rheology of the lithosphere used in these models (pseudo-plastic rheology), for which empirical parameters differ from those obtained in experiments. In this section, we analyse in detail a state-of-the-art 3-D spherical convection calculation, which exhibits fundamental tectonic features (continental drift, one-sided subduction, trench and ridge evolution, transform shear zones, small-scale convection, and plume tectonics). This example leads to a discussion where we try to answer the following question: can mantle convection models transcend the limitations of plate tectonic theory?

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1. Introduction

The idea that the deformation of the lithosphere of the Earth is a consequence of large-scale forces driving the cooling of the mantle, goes back to almost a century ago (Holmes, 1931). Arthur Holmes proposed that convection of the mantle is the driving mechanism for continental drift: rifts represent regions of mantle upwelling, whereas mountain ranges form where rocks sink back into the interior of the planet. The concept of a connection between mantle convection and global tectonics was pushed further with the advent of plate tectonics (Davies, 1977; Morgan, 1972; Turcotte and Oxburgh, 1972). Subduction corresponds to the sinking cold instabilities of the upper boundary layer of the Earth, and hotspots are the trace of hot plumes rising from the deep mantle, hitting the surface. There is little doubt today that the tectonic activity of the Earth is, to a large extent, the surface expression of mantle convection (see for instance Bercovici et al., 2000). Continuum mechanics offers a theory for large deformations of a complex medium with a rheology that can be elastic, viscous and plastic. Importantly, the digital era propelled the development of numerical methods to generate simulators of deep Earth dynamics, supported by increasing computational resources.

Despite these efforts, models that produce in a self-organised manner the minimum aspects of both convection and Earth-like surface tectonics are at their infancy. By self-organisation, we mean that the global structure of mantle flow and surface tectonics emerges spontaneously from dynamic interactions prescribed at the local scale. The organisation comes from distributed feedbacks, and is not controlled by forces or structures forced from the outside, by the modeler for instance. Since the 1970s, geodynamicists have created models of simplified mantle flow coupled with rigid plates, in order to infer the relationships between surface kinematics, elevation changes and convective forces (Bunge and Grand, 2000; Hager and O'Connell, 1981a; Ricard et al., 1993, among others). These models have improved over the years. However, they mostly keep the concept of treating separately the mantle and the lithosphere with little to no feedback. In particular, self-organisation is not allowed. As a consequence, many first-order issues cannot be solved, such as the size of plates (Morra et al., 2013), the development of subduction, the interaction between plate boundaries, and the evolution of tectonic forces in mountain building. It is only at the end of the 1990s that an improved description of the rheology with pseudo-plasticity, was introduced in convection models (Moresi and Solomatov, 1998; Tackley, 1998; Trompert and Hansen, 1998). These models produce, in a self-consistent way, regions with little deformation: plates, and regions of localised deformation: plate boundaries. Such modeling opens a novel perspective on the exploration of self-organisation in the unified lithosphere-convective mantle system. However, the geodynamic community has not yet fully taken advantage of this approach to evaluate the predictive power and robustness of these models. Indeed, while numerical codes are benchmarked (King et al., 1992; Koglin et al., 2005; OzBench et al., 2008; Tosi et al., 2015), it is the simplicity of the physics and the demanding computational resources that hold back investigations. For instance, the yielding parameters required in the model are often different from those obtained from studies of rock rheology. Hence, taking into account microphysics of grain size evolution within large-scale 3-D models could be a new goal within reach (Bercovici and Ricard, 2014).

In this paper we review the state of play in the geodynamic community to propose a mantle convection perspective on global tectonics. We choose here to focus on the links between tectonic observations and geodynamic modeling of the mantle-lithosphere system, instead of the origin or onset of plate tectonics, which are already the subject of comprehensive reviews (Bercovici et al., 2015b; Korenaga, 2013). We begin by relating how large-scale tectonics is an expression of mantle dynamics. We follow with the development of convection models that result in a better description of the lithosphere, allowing for feedback between tectonics and deep mantle flow. We show that these models follow the trail of those developed for regional lithosphere dynamics. Then, we discuss the tectonic features produced by these models allowing self-organisation, analysing a 3-D spherical convection calculation pushed to its limits. We finish by discussing how the progress of these models could lead to improved descriptions of the tectonics of the Earth.

2. Global tectonics as an expression of mantle convection

This first section covers some fundamentals with an emphasis on the modeling studies that inform our understanding of the global-scale processes behind surface tectonics. We focus on topics that have traditionally been studied in analog and numerical models with an imposed lithospheric structure, but that can now technically be addressed in self-organised convection models with plates. In the context of this review, we favour global computations over regional ones when possible.

We first look into surface kinematics, as a first-order means of understanding the mutual dynamics of the mantle and the lithosphere. More specifically, we examine the reconstruction of plate velocities, the net rotation of the lithosphere, and trench motions, and discuss where their strengths and limitations lie. This leads us to consider how surface kinematics can be used to better understand the inner workings of mantle convection and what forces move and deform the plates. Lastly, we examine the topic of dynamic topography, which can provide first-order information about the role of mantle convection in shaping the Earth's surface and has garnered renewed attention in the recent years.

2.1. Surface kinematics

2.1.1. Determining relative and absolute plate motions

Plate motions provide first order insights into the interplay between the lithosphere and the mantle, but their determination in itself requires some significant knowledge about the dynamics of the planet. Relative motions between the tectonic plates over the last few million years can be determined using seafloor magnetic anomalies, fracture zones and transform fault strikes, and earthquakes slip vectors (e.g., Chase, 1972; DeMets et al., 1990; Minster et al., 1974). In addition, GPS data can be included to clarify the current motions of small plates deprived of oceanic ridge connections (DeMets et al., 2010). Spreading rates and transform faults permit relative plate motion reconstructions in the past (e.g., Müller et al., 2016; Seton et al., 2012). With the increasing scarcity of preserved seafloor as a function of age, one may use paleomagnetic data to constrain the absolute motion of continents with respect to the Earth's spin axis and geological data to constrain their longitude (e.g., Irving, 1977; Torsvik et al., 2008a,b). Nonetheless, resolving the evolution of vanished portions of oceanic lithosphere constitutes a challenge (Domeier and Torsvik, 2014).

Constructing absolute plate velocities from relative displacements requires the non-trivial determination of a fixed terrestrial reference frame. One approach is to constrain the position of the Earth's spin axis through paleomagnetic data, assuming that the geomagnetic dipole is aligned with it. In this case, one has to correct for true polar wander (rigid body rotation of the whole mantle relative to the Earth's spin axis following mass redistribution by convection currents) (e.g., Steinberger and Torsvik, 2008). Another approach is to consider that the collection of plumes represent features that are fixed relative to the whole mantle (e.g., Dietz and Holden, 1970; Doubrovine et al., 2012; Gripp and Gordon, 2002; Morgan, 1971; O'Neill et al., 2005). With this assumption, hotspot tracks can be used to define a mantle-based reference frame. Paleomagnetic and mantle-based reference frames need to be combined for reconstructions that extend beyond the Cenozoic, i.e. for times from which there is too little preserved seafloor to reconstruct relative plate motions (e.g., Torsvik et al., 2008a,b).

Because paleomagnetic data do not constrain the longitudinal position of the plates, Burke and Wilson (1972) and Hamilton (2002) investigated the possibility of building reference frames based on the identification of a plate that would remain fixed over time. Torsvik et al. (2014) proposed another fixed feature for the mantle in time as a reference: the large low shear-wave velocity provinces at the base of the lower mantle, with plumes ascending from their boundaries. However, this assumption is challenged by convection calculations (Tan et al., 2011). Assuming that slabs sink vertically into the mantle, van der Meer et al. (2010), Butterworth et al. (2014), and Domeier et al. (2016) explored the possibility of using seismic images to build a reference frame by identifying slab remnants. Assuming that absolute plate motions are aligned with azimuthal seismic anisotropy in the upper mantle, Kreemer (2009), Becker et al. (2015), and Williams et al. (2016) have proposed reference frames for plate motions at the present-day and in the past. A distinct class of reference frames are designed to preclude any net rotation of the surface (e.g., Argus and Gordon, 1991; Kreemer et al., 2003).

The validity of these reference frames rely on geodynamic assumptions that are sometimes controversial and that bear large uncertainties. For instance, hotspots are mobile relative to each other (Molnar and Stock, 1987), such that geodynamic computations need to be used to correct for their motions (e.g., O'Neill et al., 2005). A long-standing issue is that all proposed 'absolute' plate motion models indicate different amounts of contemporary lithospheric net rotation and trench motions (see, e.g., Fig. 1), two central features of surface kinematics that reflect plate-mantle interactions. Thus, the determination of a stable reference frame for plate motions, which is needed to make progress in all subdisciplines of Earth Sciences, demands better constraints on mantle dynamics to ensure its reliability. Nonetheless, a consistent outcome across reference frames is that contemporary absolute plate velocities "correlate negatively with total continental area (Minster et al., 1974)", and "positively with the fraction of plate boundary being subducted (Forsyth and Uyeda, 1975; Jordan and Minster, 1974)", as summarised early on by Minster and Jordan (1978).

2.1.2. The net rotation of the lithosphere

A Helmholtz decomposition of the present-day plate velocities in a hotspot reference frame indicates that the power of the poloidal and toroidal velocity components decreases with increasing degree, but their ratio remains equivalent across scales (e.g., Bercovici, 1995; Čadek and Ricard, 1992; O'Connell et al., 1991; Olson and Bercovici, 1991). This holds substantial information about the mutual dynamics of the convecting mantle and the lithosphere (see Bercovici et al., 2015, for a review). Indeed, the poloidal component corresponds to the vertical and horizontal movements driven by density anomalies in a viscously stratified body, whereas the toroidal component corresponds to strike-slip movements in the horizontal plane, which solely develop in a more complex viscosity structure such as on Earth, with lateral variations induced by plate boundaries and a relatively strong sub-continental mantle (e.g., Gable et al., 1991; O'Connell et al., 1991; Ricard et al., 1991). A toroidal to poloidal ratio that approaches one, such as that at the present day, therefore attests to the considerable role that lateral viscosity variations in the mantle hold in plate tectonics.

In a deep mantle reference frame, the degree one of the toroidal velocity components holds the most energy (e.g., Bercovici et al., 2015). It corresponds to a net spin of the surface as a whole with respect to the deep mantle and is called the net rotation of the lithosphere. A reliable estimate of the net rotation is crucial to determine a suitable absolute reference frame for surface motions but also, in turn, to improve our understanding of mantle convection with plate tectonics. As a component of the toroidal flow field, the net rotation exists because of lateral viscosity variations in the upper mantle, such as cratonic keels (Ricard et al., 1991; Rudolph and Zhong, 2014; Zhong, 2001), a heterogeneous asthenosphere distribution (Becker, 2006; Gérault et al., 2012), subducting slabs (Harper, 1986; Zhong, 2001; Becker and Faccenna, 2009; Gérault et al., 2012), and an asymmetric arrangement of the plates (Gérault et al., 2012). Those induce a differential coupling between certain plates and the mantle, which a purely radial viscosity structure would not. All absolute plate motion models predict some amount of net rotation and place the current pole in the southern Indian Ocean (Becker et al., 2015; Becker and Faccenna, 2009), a direction of rotation that has been stable for the past 30 Ma (Doubrovine et al., 2012; Torsvik et al., 2010). Discrepancies arise between tentative absolute reference frames in terms of amplitude: from about 0.13°/Myr (Torsvik et al., 2010) to 0.44°/Myr (Gripp and Gordon, 2002); that is, between 1.4 and 4.9 cm/year along the equator of the net rotation. Plate models that correct for the

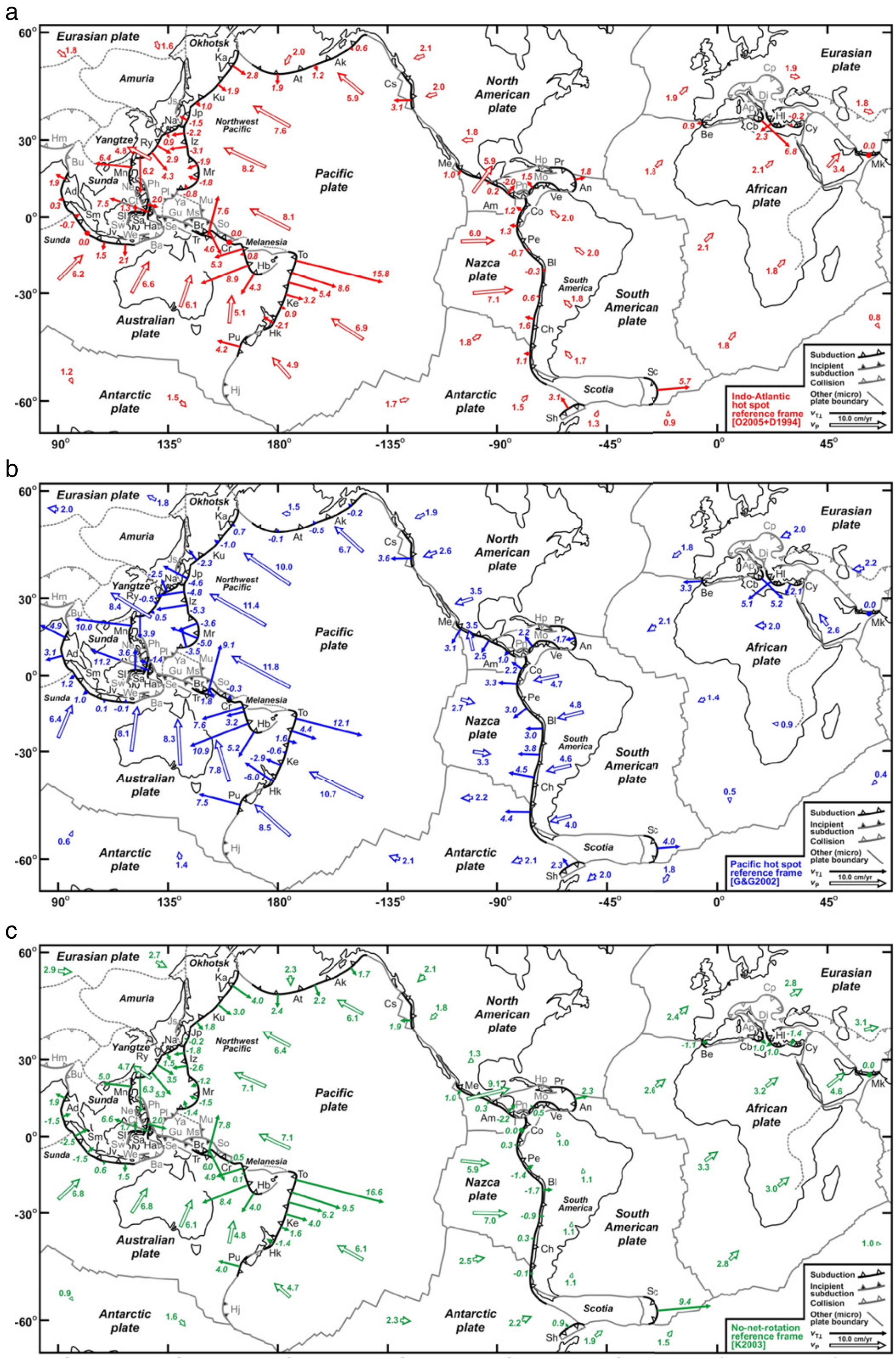

Fig. 1. Tectonic maps showing the major subduction zones on Earth, the velocities for the major plates (v_P) and the trench–perpendicular trench migration velocities ($v_{P\perp}$) in three global reference frames. (a) Indo-Atlantic hotspot reference frame of O'Neill et al. (2005) combined with the relative plate motion model of DeMets et al. (1994) (O2005+D1994); (b) Pacific hotspot reference frame of Gripp and Gordon (2002) (G&G2002), who use the relative plate motion model of DeMets et al. (1994); and (c) no-net-rotation reference frame of Kreemer et al. (2003) (K2003). Subduction zone abbreviations: Ad-Andaman, Ak-Alaska, Am-Central America, An-Lesser Antilles, At-Aleutian, Be-Betic Rif, Bl-Bolivia, Br-New Britain, Cb-Calabria, Ch-Chile, Co-Colombia, Cr-San Cristobal, Cs-Cascadia, Cy-Cyprus, Ha-Halmahera, Hb-New Hebrides, Hk-Hikurangi, Hl-Hellenic, Iz-Izu-Bonin, Jp-Japan, Jv-Java, Ka-Kamchatka, Ke-Kermadec, Ku-Kuril, Me-Mexico, Mk-Makran, Mn-Manila, Mr-Mariana, Na-Nankai, Pe-Peru, Pr-Puerto Rico, Pu-Puysegur, Ry-Ryukyu, Sa-Sangihe, Sc-Scotia, Sh-South Shetland, Sl-North Sulawesi, Sm-Sumatra, To-Tonga, Tr-Trobriand, Ve-Venezuela. Collision zones: Ap-Apennines, Ba-Banda, Bu-Burma, Cp-Carpathian, Di-Dinarides, Hp-Hispaniola, Se-Seram, So-Solomon. Incipient subduction zones: Ct-Cotobato, Gu-New Guinea, Hj-Hjort, Js-Japan Sea, Mo-Muertos, Ms-Manus, Mu-Mussau, Ne-Negros, Ph-Philippines, Pl-Palau, Pn-Panama, Sw-West Sulawesi, We-Wetar, Ya-Yap.
Source: Schellart et al. (2008)

motion of hotspots tend to predict amplitudes towards the mid- to low-end of the spectrum (e.g., Becker et al., 2015; Doubrovine et al., 2012; Steinberger et al., 2004; Torsvik et al., 2008a), whereas some others use shallow hotspots to find even greater amplitudes (e.g., Doglioni et al., 2015). Absolute plate motion reconstructions indicate that the amount of net rotation may have fluctuated between faster and slower magnitudes over time, although fast rates predating the Cenozoic could come primarily from uncertainties in the reconstructions (Domeier and Torsvik, 2014; Doubrovine et al., 2012; Rudolph and Zhong, 2014; Torsvik et al., 2010).

Perhaps as a result of these ambiguities, certain plate kinematic models discard the net rotation entirely and use a no-net rotation reference frame, for example in order to display GPS velocities (e.g., Argus et al., 2011; Kreemer et al., 2006) (see also Fig. 1c). A recent trend in plate motion models (Domeier and Torsvik, 2014; Torsvik et al., 2010; Williams et al., 2015) and geodynamic simulations (Alisic et al., 2012) is to adopt the lowest end-member value of 0.13°/Myr (Torsvik et al., 2010), relying on their assumption that the net rotation should be as small as possible. The studies of Alisic et al. (2012) and Williams et al. (2015) use this value as means of quantifying rheological parameters and assessing the plausibility of various plate models, respectively. However, azimuthal anisotropy from surface waves (Becker et al., 2014a, 2008) and shear wave splitting measurements (Conrad and Behn, 2010; Kreemer, 2009) suggest an amplitude up to 0.2 to 0.3°/Myr. 3-D spherical computations of mantle flow find rates below 0.2°/Myr, with a rotation pole close to that of plate models (Becker, 2006; Rudolph and Zhong, 2014; Zhong, 2001). Using global computations with imposed plate motions, Alpert et al. (2010) find that the highest net rotation estimate of Gripp and Gordon (2002) provides the best fit to the deep stress state of slabs as indicated by earthquake focal mechanisms. High-resolution cylindrical computations indicate that uncertainties in slab and sub-lithospheric viscosities are such that it is possible to predict any of the proposed rates within those uncertainty bounds (Gérault et al., 2012).

The net rotation of the lithosphere with respect to the mantle is arguably the most prominent aspect of plate kinematics that remains to be unraveled. We suggest that self-organised convection computations, such as those described in Section 4 of this review, could provide additional constraints on this phenomenon.

2.1.3. Trench kinematics

The subduction of oceanic lithosphere into the mantle results in the formation of deep oceanic trenches that are mobile. Subduction occurs through trenchward motion of the oceanic plate as well as trench retreat, the balance of which is referred to as subduction partitioning (e.g., Schellart et al., 2008). Trench migration patterns have important implications for surface tectonics as well as mantle convection, in that they are a direct product of the interactions between viscous slabs and the convecting mantle. Trench migration affect lithospheric stresses, mixing, and heat flow through slab dip angles, slab penetration into the lower mantle, and toroidal/poloidal mantle flow partitioning (e.g., Christensen, 1996; Garfunkel et al., 1986; Griffiths et al., 1995; Kincaid and Griffith, 2003; Piromallo et al., 2006; Zhong and Gurnis, 1995a). Similarly to the net rotation, trench behaviour can be used to draw inferences about slab rheology (Alisic et al., 2012; Funiciello et al., 2008; Gérault et al., 2012) and to assess the realism of absolute plate motion models (Williams et al., 2015).

Absolute trench motions depend on the choice of reference frame in the same way as plate motions do (Funiciello et al., 2008; Schellart et al., 2008) (Fig. 1). The properties of the lithosphere and the mantle are such that both advance and retreat appear to be mechanically feasible (e.g., Capitanio, 2013; Di Giuseppe et al., 2009; Faccenna et al., 2007; Funiciello et al., 2003; Zhong and Gurnis, 1995a), but their relative importance depends on the choice of reference frame and the specifics of the analysis (e.g., Heuret and Lallemand, 2005; Schellart et al., 2008). South America offers a striking example where trenches are either predicted to retreat at a rate of several cm/year (e.g., Becker et al., 2015), or to advance slowly towards the overriding plate (O'Neill et al., 2005). A significant number of oceanic trenches is somewhat perpendicular to the equator of the net rotation, which crosses the Pacific domain from southeast to northwest. Consequently, plate motion and geodynamic models that show large net rotation amplitudes also indicate a westward motion of the trenches surrounding the Pacific Ocean (Becker et al., 2015; Gripp and Gordon, 2002; Heuret and Lallemand, 2005). Net rotation and trench motions are also physically connected through their mutual dependence on slab rheology (Alisic et al., 2012; Funiciello et al., 2008; Gérault et al., 2012).

Studies that subscribe to a 'westward drift' of the Pacific domain contend that on average, the western Pacific lithosphere is older, and therefore thicker and stiffer, than its counterpart on the eastern (Nazca) side, preventing it from unbending easily at depth (Di Giuseppe et al., 2008; Lallemand et al., 2008). Following that line of argument, western Pacific trenches advance and slabs tend to subduct more steeply in the upper mantle, resulting in a more efficient slab pull and faster plate velocities (Di Giuseppe et al., 2009; Faccenna et al., 2009) than on the Nazca side. This provides a geodynamic framework for the apparent dichotomy in trench motions, slab dip angles, and upper-plate deformation that exists on the opposite sides of the Pacific basin (Doglioni et al., 2015; Lallemand et al., 2005; Molnar and Atwater, 1978; Uyeda and Kanamori, 1979). However, westward Pacific trench motions are not observed if the reference frame is the Indo-Atlantic hotspot reference frame (O'Neill et al., 2005). Instead, for instance, slow trench advance is predicted in the central Andes, which may offer a better explanation for the on-going deformation (Schellart, 2008). An alternative framework explains trench motions in this Indo-Atlantic reference frame by considering regional dynamics: trenches retreat rapidly when located in the vicinity of lateral slab edges (narrow slabs, typically exhibiting convex trench curvature), while trenches that are far from slab edges move more slowly and are more likely to advance (wide slabs, with concave trench curvature) (Bellahsen et al., 2005; Schellart et al., 2011, 2007; Stegman et al., 2006). There is currently no consensus on how to reconcile these conflicting findings or dismiss any of them. However, self-consistent convection models such as those by Crameri and Tackley (2014) show a promising level of realism for trench morphology and slab geometry, and should help solving some of these issues.

2.2. Tectonic forces

In the previous section, we have introduced some fundamental characteristics of plate motions and described how these provide first-order insights into the properties of the plate-mantle system. In this section, we show how surface velocities can be further analysed to understand the workings of plate tectonics, usually by means of geodynamic modeling.

2.2.1. Plate driving forces

Geological observations on Earth are best explained by the interactions of a rigid, fragmented rocky shell moving on top of a mantle in solid-state convection. A fundamental question is whether plate motions are governed by forces that stem from within the lithosphere, acting on the edges of the plates, or from the mantle below. The early studies approached this problem by calculating torque balance on each plate from parameterised forces (e.g., Chapple and Tullis, 1977; Forsyth and Uyeda, 1975; Solomon et al., 1975). Their results, with which subsequent generations of physical models concur, suggest that viscous dissipation at the base of and within the lithosphere, as well as shear along thrust faults at collision zones resist plate motions, whereas buoyancy anomalies within the crust, the lithosphere, and the mantle act as the principal drivers. This result was parameterised by Conrad and Hager (1999). Becker and O'Connell (2001) estimate that among the density anomalies responsible for plate motions, ~70% originate from the mantle and ~30% from within the lithosphere, while Lithgow-Bertelloni and Richards (1995, 1998) argue that the ratio is more ~90% to ~10% when looking at plate motion history during the Cenozoic and the Mesozoic. These models include passive upwellings from subduction, but no active, buoyant upwellings. In an internally-heated planet with temperature-dependent viscosity, the top boundary layer is expected to be the primary source of buoyancy-driven flow, while rising plumes are likely to be small and distributed in ways that have little effect on plate motions (Hager and O'Connell, 1981b; McKenzie et al., 1974). However, there is evidence that shear tractions from mantle flow may have driven rapid continental movements in the past, through the interactions between hot plumes, large scale mantle flow, and cratonic keels (Becker and Faccenna, 2011; Cande and Stegman, 2011; Gurnis and Torsvik, 1994). The models of Crowley and O'Connell (2012) indicate that the dynamics of a viscous plate moving on top of a convecting mantle with depth-dependent viscosity are complex, such that multiple factors control whether basal tractions from mantle flow play a driving or a resistive role in plate motions.

The plate velocities and the intraplate stress field provide the strongest constraints on the forces acting on plates. Plates bordered by oceanic trenches move ~8 times faster than those bordered by spreading ridges (e.g., Gordon et al., 1978), which supports the idea that the process of subduction largely dominate others in its ability to move the plates. Through more or less sophisticated modeling approaches, global geodynamic models of mantle flow that account for slabs pull, and to a lesser extent 'ridge push', are able to predict geologically-current plate velocities fairly accurately (e.g., Becker and O'Connell, 2001; Bird et al., 2008; Conrad and Lithgow-Bertelloni, 2002; Stadler et al., 2010; Zhong, 2001). Similar conclusions can be drawn from plate motions as far back as during the late Mesozoic (Conrad and Lithgow-Bertelloni, 2004; Lithgow-Bertelloni and Richards, 1995; Zahirovic et al., 2015).

The parameterised edge-forces models of Conrad and Lithgow-Bertelloni (2002) show that slab anomalies contribute to a different style of flow depending on whether they are situated in the upper or the lower mantle: slab pull versus slab suction, respectively. Slab pull stems from the direct link between the subducted lithosphere and the plate it is attached to, the slab acting as a stress guide (Elsasser, 1969). It induces fast subducting plate motion, relative to slow overriding plate motion. Slab suction operates through shear tractions at the base of the plates: while sinking, a lower-mantle fragment of dense slab induces symmetrical mantle flow that draws the subducting plate and the upper plate equally fast towards it. The best fit to plate motions is obtained for a slab pull contribution of 40% to 75% of the total slab force, depending on the viscosity structure and the time period, while slab suction accounts for the rest (Conrad and Lithgow-Bertelloni, 2004). Importantly, the correct ratio of oceanic over continental plate velocities cannot be matched without invoking some amount of direct, physical pull from the slab onto the oceanic plate, which can be tricky to achieve in global geodynamic models that rely solely on tomography to infer buoyancy forces.

Other characteristics of the plate-mantle system are instrumental in shaping plate motions. For instance, viscosity differences between the sub-oceanic and the sub-continental mantles contribute to a proper ratio of oceanic to continental plate velocities, although studies using different parameterisations obtain slightly different results (Becker, 2006; Conrad and Lithgow-Bertelloni, 2006; Summeren et al., 2012; Zhong, 2001). The fit to current plate motions is also improved when slab density anomalies from the lower mantle are accounted for (Becker and O'Connell, 2001; Lithgow-Bertelloni and Richards, 1998). Gravitational potential energy variations within the crust and the lithosphere are delicate to include in global computations (e.g., Chapple and Tullis, 1977; Ghosh et al., 2013), but their contribution as a plate-driving force is expected to be as significant as that of mid-oceanic ridges (Artyushkov, 1973; Fleitout and Froidevaux, 1983; Frank, 1972; Ghosh et al., 2006; Husson et al., 2008; Meade and Conrad, 2008).

Modeling studies that explore the sources of lithospheric stress tend to suggest that slabs have an impact relatively similar to that from ridge push, gravitational potential energy variations within the lithosphere, and viscous drag (e.g., Ghosh and Holt, 2012; Richardson et al., 1979). Coupling viscous mantle flow models, driven by loads derived from seismic imaging, with an elastic lithospheric shell that accounts for variations in crustal thickness and plate geometry, produces intraplate stresses, plate velocities and geoid anomalies comparable to the Earth (e.g., Bai et al., 1992; Bird, 1998; Lithgow-Bertelloni and Guynn, 2004; Steinberger et al., 2001). In spite of recent methodological improvements (e.g., Ghosh et al., 2013; Stadler et al., 2010), quantifying the relative importance of plate boundary forces, intraplate forces, and basal drag from mantle flow for lithospheric stresses remains a challenge, in part because of the sensitivity of the problem to uncertainties in (1) the lithospheric structure (e.g., Naliboff et al., 2012), and (2) the most suitable treatment of plate boundaries (e.g., Alisic et al., 2012; Stadler et al., 2010).

2.2.2. Trench dynamics, slab geometry and lithospheric deformation

The forces that drive the plates also provide energy to deform them. Whereas the theory of plate tectonics entails that the plates move as rigid bodies, deforming exclusively along their boundaries within narrow regions, models of the current strain rate field show that the crust deforms within broad regions adjacent to plate boundaries (Kreemer et al., 2014, see Fig. 2). Here, we take a closer look at regions affected by subduction processes, and discuss how trench motions, slab geometry and deformation in the overriding plate are interconnected. We refer the reader to existing review papers on the subject of slab dynamics in the upper and lower mantle (e.g., Billen, 2008; Becker and Faccenna, 2009; Wada and King, 2015, and references therein) so as to focus on surface tectonics. Although we typically favour global-scale models in the context of this review, the studies that explore the dynamics of trench motions and upper-plate deformation are predominantly regional, which is reflected in the literature cited below. Unfortunately, many of these regional analog and numerical studies have a rigid bottom boundary in place of a lower mantle, which is likely to have an effect on trench migration (e.g., Christensen, 1996; Zhong and Gurnis, 1995a). The spherical, instantaneous studies of Stadler et al. (2010) and Alisic et al. (2012) suggest that regional effects have the largest influence on the state of stress in the lithosphere.

Trench mobility is a key variable in the balance of forces in a subduction system. At a regional scale, numerical modeling (e.g., Čížková and Bina, 2015; Enns et al., 2005; Holt et al., 2015a; Ribe, 2010; Sharples et al., 2014; Stegman et al., 2010) and analog modeling (e.g., Bellahsen et al., 2005; Funiciello et al., 2004; Kincaid and Olson, 1987; Meyer and Schellart, 2013; Schellart, 2004) in 2-D and in 3-D suggest that trenches retreat more easily than they advance, as long as a thin region of decoupling exists between the subducting and the overriding plates (e.g., Stadler et al., 2010; Zhong and Gurnis, 1995a). This is particularly true for young, weak slabs, which are easier to deform (e.g., Di Giuseppe et al., 2009) and whose behaviour is primarily governed by buoyancy as opposed to viscosity (Garfunkel et al., 1986). Slab shapes that reflect continuous trench advance in modeling studies, such as roll-over slabs (e.g., Bellahsen et al., 2005), are not found among seismically-observed slab geometries (e.g., Li and Van Der Hilst, 2010). However, old slabs in the western Pacific are steeper and seem to indicate trench advance in all reference frames. Trench migration is associated with significant toroidal flow (e.g., Becker and Faccenna, 2009; Bellahsen et al., 2005; Crameri and Tackley, 2014; Kincaid and Griffith, 2003; Piromallo et al., 2006; Stegman et al., 2006). In 2-D, mantle flow is purely poloidal and models show a greater propensity for trench advance

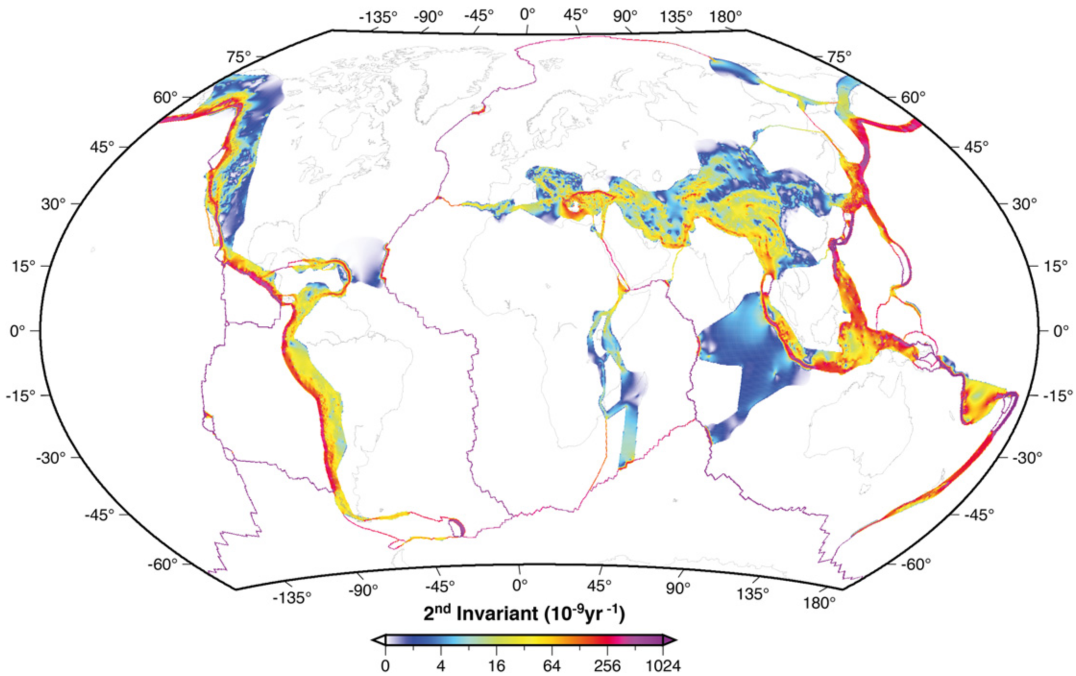

Fig. 2. Dilatational component of the strain rates in the geodetic model from Kreemer et al. (2014). Extension is positive, compression is negative. Highest values are saturated. Source: Kreemer et al. (2014)

and overriding plate compression, whereas in 3-D, toroidal flow and extension dominate near lateral slab edges (e.g., Chen et al., 2016; Duarte et al., 2013; Holt et al., 2015a; Meyer and Schellart, 2013; Schellart and Moresi, 2013).

In nature, the subduction interface is made of heavily faulted and damaged crustal rocks, unconsolidated sediments, and water, which remains soft in spite of plate convergence. In numerical models, this interface is typically modeled either as a thrust fault or as a zone of lower viscosity (e.g., Zhong and Gurnis, 1995a,b). Dynamic models that do not account for some weakness at the trench are faced with the difficulty of allowing for trench retreat and back-arc opening. Some circumvent this issue by dismissing the overriding plate and letting the low-viscosity mantle extend all the way up to the surface above the slab, but studies have shown that it substantially modifies subduction dynamics (e.g., van Dinther et al., 2010; Meyer and Schellart, 2013; Sharples et al., 2014; Holt et al., 2015a). A buoyant or neutrally buoyant (i.e., continental) upper plate facilitates trench rollback and promotes upper-plate extension (Garel et al., 2014; Sharples et al., 2014; Holt et al., 2015a). In nature, slab dip angles are on average 20° steeper beneath oceanic plates than beneath continental plates, which is the most convincing correlation between slab morphology and any another subduction property (Lallemand et al., 2005). In time-evolving numerical simulations, shallower slab dips form beneath thicker, colder upper-plate regions (Manea et al., 2012; Taramón et al., 2015; Holt et al., 2015b; Hu et al., 2016; Rodríguez-González et al., 2016). Buoyancy variations within the upper plate may affect subduction kinematics through interplate coupling. For instance, changes in Nazca plate velocity during the Cenozoic may be linked to changes in gravitational potential energy within the upper plate, due to the formation of the Andes (Husson et al., 2008; Iaffaldano et al., 2006; Meade and Conrad, 2008). Along the same lines, Austermann and Iaffaldano (2013) connect the recent slow-down of the Arabia-Eurasia collision to the Zagros orogeny. Although the presence of an overriding plate does not appear to have a substantial effect on oceanic plate velocity, trench mobility decreases substantially with increasing overriding plate thickness (e.g., Butterworth et al., 2012; Capitanio et al., 2010; Garel et al., 2014; Holt et al., 2015a; Meyer and Schellart, 2013; Yamato et al., 2009).

Many have proposed direct connections between subduction kinematics, lithospheric properties and deformation of the back-arc region (e.g., Faccenna et al., 2013; Hager et al., 1983; Jarrard, 1986). Slab age, slab width, slab dip angle, and large-scale mantle circulation may all cause overriding plate deformation. Regrettably, the issues associated with the choice of reference frame persist when trying to extract some systematics from natural examples. In the Pacific hotspot reference frame HS-3 (Gripp and Gordon, 2002), back-arc compression tends to be associated with the upper-plate moving towards the trench, and back-arc extension tends to be associated with the upper-plate moving away from the trench (Arcay et al., 2008; Heuret and Lallemand, 2005; Lallemand et al., 2008), whereas plate and trench motions in the Indo-Atlantic reference frame indicate the opposite (Schellart et al., 2008).

Molnar and Atwater (1978) note a correlation between style of overriding plate deformation and age of the subducting seafloor: surrounding the Pacific basin, subduction of younger oceanic lithosphere appears to be associated with mountain building, whereas older lithosphere is associated with the opening of back-arc basins. Molnar and Atwater (1978) invoke age-related differences in slab buoyancy to explain those opposite tectonic behaviours. Capitanio et al. (2011) uses regional numerical models to demonstrate that seafloor age distribution on the Nazca plate, older beneath Bolivia than on either sides, could be responsible for concave trench curvature and relief distribution in western South America. Alternatively, Schellart (2008) argue that compressive stresses should arise in the lithosphere above the center of wide slabs, because this is where slab-induced flow is mainly poloidal, trench motion should be the slowest, and a stagnation point could form Russo and Silver (e.g., 1994). Although their models do not include an overriding plate (see issues above), Schellart (2008) apply this line of reasoning to shortening in the Central Andes at the present-day. Schellart et al. (2010) suggest that it also applies to the tectonic evolution of the western United States during the subduction of the Farallon plate, from the Laramide orogeny to the formation of the Basin and Range.

In addition to seafloor age and slab width, slab dip angle is often invoked to explain overriding plate compression and extension in the back-arc region (e.g., Barazangi and Isacks, 1976; Bird, 1988; Jarrard, 1986; Lallemand et al., 2005; Lee et al., 2011; Uyeda and Kanamori, 1979; Wdowinski et al., 1989). Isacks (1988) connects the current Andes topography to past variations in slab geometry that would have weakened the crust. At the present-day, the plates overlying steep slabs show a greater propensity for back-arc extension, whereas compression is more commonly observed above low-angle subduction (Jarrard, 1986; Lallemand et al., 2005). The end-member case of flat-slab subduction is often used as means of explaining the formation of the American Cordilleras (e.g., Barazangi and Isacks, 1976; Bird and Baumgardner, 1984; Coney and Reynolds, 1977; Dickinson and Snyder, 1978; Jordan et al., 1983). O'Driscoll et al. (2009) and Jones et al. (2011) attribute the Laramide orogeny to fluid-flow interactions between the Farallon slab and a nearby continental keel. In the regional instantaneous models of Jadamec et al. (2013), flat-slab subduction induces stresses in the upper plate that are consistent with present-day surface deformation patterns in Alaska. The formation of a flat slab may require some combination of rapid trenchward motion of the overriding plate (Espurt et al., 2008; van Hunen et al., 2000, 2004; Liu and Currie, 2016; Silver et al., 1998), buoyancy anomalies on the subducting plate (Antonijevic et al., 2015; Hu et al., 2016; van Hunen et al., 2004, 2002; Liu and Currie, 2016), a weak mantle wedge above the slab (Bello et al., 2015; van Hunen et al., 2002; Manea and Gurnis, 2007), and a nearby protruding craton (Hu et al., 2016; Jones et al., 2011; Manea et al., 2012; O'Driscoll et al., 2009, 2012; Taramón et al., 2015) or a thick and strong overriding plate (Sharples et al., 2014). Once underplated, the oceanic lithosphere is susceptible of modifying the structure of the overriding plate chemically and mechanically, causing large-scale continental uplift (e.g., Bird and Baumgardner, 1984; Humphreys et al., 2003; James and Sacks, 1999; Jones et al., 2015; Spencer, 1996). Subduction margins have a history of episodic phases of extension and compression that are associated with changes in plate convergence rates at the trench (e.g., Capitanio, 2013; Clark et al., 2008; Martinod et al., 2010). The 2-D numerical models of Lee et al. (2011) show that slab bukling at the upper-lower mantle boundary induces periodic variations in slab dip angles and convergence rates that could explain successive phases of extension and compression in the overriding plate.

According to the analysis of Schellart (2008), which is tied to the Indo-Atlantic hotspots reference frame, no single parameter — among overriding plate velocity, subducting plate velocity, trench velocity, convergence velocity, subduction velocity, subduction zone accretion rate, subducting plate age, subduction polarity, shallow slab dip, deep slab dip, lateral slab edge proximity, and subducting ridge proximity — completely correlates with the type of deformation in the upper plate. In this analysis, trench motions and subduction velocity do best, but they may also be a consequence of what yields upper-plate deformation, instead of a cause (e.g., Faccenna et al., 2009; Iaffaldano et al., 2006; Meade and Conrad, 2008). Alternatively, the Indo-Atlantic hotspot reference frame may not be well suited for the description of the motion of trenches, which are mostly located in the Pacific. More importantly, it appears that a number of feedbacks exist between subducting and overriding plates, which are not well understood and not fully captured by current geodynamic simulations. Because both local and global torques appear to control these interactions over time, global, self-consistent and time-dependent mantle convection models that include a proper treatment of the lithosphere and plate boundaries are required to improve our understanding regional tectonics driven by mantle convection.

2.3. Topography from mantle flow

2.3.1. Terminology

Striving to understand the topography of the Earth is one of the oldest endeavours in the Geological Sciences. The topography of this planet reflects the influence of mantle processes in two ways: first, tectonic forces (see above) provide energy to deform the lithosphere and the crust, and the topography must evolve to compensate for any modification of the density structure of the plates. Second, mantle convection in itself involves the rise and fall of density currents that can affect the topography of the Earth's surface and the core-mantle boundary. The topography owing to buoyancy variations within the lithosphere is typically referred to as isostatic topography, whereas the topography sustained by mantle convection processes beneath the lithosphere is referred to as dynamic topography. A third term, residual topography, is defined here as the result of subtracting an isostatic signal from the full topography. If the isostatic component is well constrained, which demands a complete knowledge of the density structure of the crust and the lithosphere, residual and dynamic topography are identical. In practice, as we discuss below, it is rarely the case. Dynamic topography occurs in response to vertical stresses induced by mantle upwellings and downwellings (e.g., Braun, 2010; Hager et al., 1985), so to some extent, it contributes to topography all over the globe, including at passive margins (e.g., Conrad and Husson, 2009; Moucha et al., 2008). The magnitude of this contribution has been debated for decades. Part of the controversy comes from the challenge of employing a consistent, practical definition across disciplines that use distinct methods for their estimates. In the following, we review the various lines of evidence for dynamic topography, as well as the procedures and uncertainties involved in separating dynamic from isostatic contributions to the topography.

2.3.2. Geophysical evidence

Evidence for mantle flow-induced topography includes sedimentary records (e.g., Dávila et al., 2007; Gurnis, 1993; Hoggard et al., 2016; Mitrovica et al., 1989; Painter and Carrapa, 2013; Pang and Nummedal, 1995; White et al., 1997), geoid anomalies (e.g., Hager and Clayton, 1989; Hager et al., 1985; Panasyuk and Hager, 2000), free-air gravity anomalies (e.g., Molnar et al., 2015; Morgan, 1965a,b; Parsons and Daly, 1983), and deviations from an isostatic lithospheric model (e.g., Becker et al., 2014b; Faccenna et al., 2014b; Levandowski et al., 2014; Miller and Becker, 2014; Spencer, 1996).

The sedimentary record provides a means of calculating anomalous elevation in both continental and oceanic domains, at the present-day and in the past. On the continents, episodes of subsidence caused by mantle dynamics can result in large-scale flooding of continental interiors (e.g., Gurnis, 1993; Lithgow-Bertelloni and Gurnis, 1997; Mitrovica et al., 1989), whereas episodes of uplift result in identifiable erosion and/or deposition patterns (e.g., Braun et al., 2013; Burgess et al., 1997; Liu and Nummedal, 2004; White et al., 1997). The wavelengths of sedimentary deposits allow to distinguish between flexural and dynamic support of the topography (e.g., Dávila et al., 2007; Liu and Nummedal, 2004; Liu et al., 2014; Painter and Carrapa, 2013; Pang and Nummedal, 1995), those greater than a few hundreds of meters suggesting a mantle influence. Sedimentary records may be used to trace the geological evolution of a particular region. For instance, in western North America, the location of the depocenters followed the migration of the Farallon slab during the Late Mesozoic–Early Cenozoic (e.g., Jones et al., 2011; Liu et al., 2011; Roberts and Kirschbaum, 1995).

In the oceans, correcting for plate cooling (e.g., Crosby and McKenzie, 2009; Marty and Cazenave, 1989; Stein and Stein, 1992), water and sediment loading (e.g., Crough, 1983), flexural bending (e.g., Bry and White, 2007; Levitt and Sandwell, 1995), and seamounts, oceanic islands and plateaux (e.g., Hillier and Watts, 2005; Zhong et al., 2007a) provides residual depth measurements that can be interpreted in terms of dynamic support (e.g., Hoggard et al., 2016; Müller et al., 2008; Winterbourne et al., 2014; Zhong et al., 2007a). The recent compilation of Hoggard et al. (2016) gathers information from seismic reflection profiles, wide-angle experiments, and refraction experiments worldwide to assemble a map of oceanic residual depth, which they combine with mid-wavelength

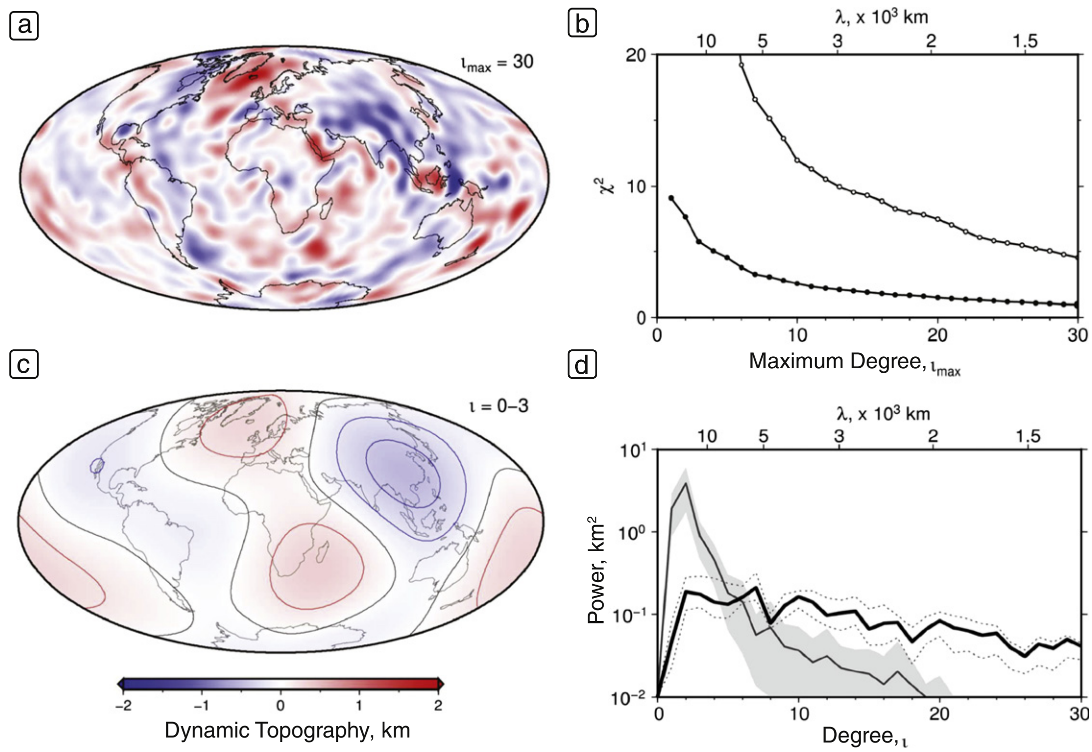

Fig. 3. Observed dynamic topography, misfits and power spectra. (a) Spherical harmonic model up to degree $l = 30$ of dynamic topography estimates using residual seafloor depth measurements in the oceans and free-air anomalies in the continents. (b) Residual misfit, χ^2, plotted as function of maximum degree, $l_m ax$, used to fit observations. Solid circles: χ^2 values from fitting combined offshore/onshore global database of dynamic topographic observations; open circles: χ^2 values from fitting spot measurements from oceanic realm alone. Corresponding wavelengths, λ, are shown at the top. (c) Spherical harmonic model up to degree $l = 3$. Contour lines every 200 m; black line = zero. (d) Power spectra of observed versus predicted dynamic topography. Black solid line = power spectrum obtained by regularised least squares inversion of global database of dynamic topography observations; dotted lines = uncertainty resulting from range of gradient regularisation and amplitude damping coefficients ($10^{-0.5}$–$10^{0.5}$ and 10^2–10^3, respectively); thin line with grey bounds = mean and standard deviation of power spectra for five predictive models of dynamic topography (Conrad and Husson, 2009; Flament et al., 2013; Ricard et al., 1993; Spasojevic and Gurnis, 2012; Steinberger, 2007).
Source: Hoggard et al. (2016)

free-air anomalies in the continents (Fig. 3a and c). Their results indicate amplitudes of ±1 km on average at wavelengths of the order of ~1000 km, and up to 2 km around the hotspots underneath Island, Hawaii, the Azores and the Red Sea at wavelength greater than 2000 km. Interpreting sedimentary records in the oceans is complicated by the need for a model that specifies 'normal' plate subsidence with age, of which there are a few. There are significant differences in the predicted residual topography in terms of amplitudes and patterns when using the classic half-space cooling model versus other age-depth relationships in order to distinguish between static and dynamic support (Adam et al., 2015; Davies and Pribac, 1993; Panasyuk and Hager, 2000; Yang and Gurnis, 2016; Zhong et al., 2007a).

The variations in the geoid and its radial derivative, free-air gravity, are frequently used to estimate where deep mass anomalies are not statically compensated. Free-air gravity anomalies contain signals from the topography at small wavelengths, whereas most of the geoid's power is contained in the long wavelengths, especially the degree two. In the early studies, the problem of dynamic topography is considered globally, in concert with deformation at the core-mantle boundary, the geoid, and the distribution of large-scale mass anomalies within the Earth's mantle (e.g., Hager et al., 1985; Ricard et al., 1984; Richards and Hager, 1984). Richards and Hager (1984) find that the surface deforms over time scales that are short in comparison with changes in convective viscous flow. In addition, they find that the density anomalies closest to the surface may be masked in the geoid signal because of the corresponding changes in topography that they induce.

Molnar et al. (2015) discuss some of the drawbacks to using free-air anomalies in order to appraise mantle support, such as the fact that elastic and isostatic signals cannot be completely removed by low-pass filtering. Nonetheless, Hoggard et al. (2016) find a good agreement between their oceanic residual depth estimates and free-air anomalies at wavelengths greater than 700 km. In the western United States and in northern Mexico, Stephenson et al. (2014) combine free-air anomalies at wavelengths between 800 and 2500 km, and estimates of uplift rates from longitudinal river profiles to find that this province has been uplifted dynamically by 2–3 km since 40 Ma, finding up to 1 km of dynamic support at present. Molnar et al. (2015) examine isostatic and free-air gravity anomaly maps from Balmino et al. (2012) at wavelengths between 500 and 6000 km to infer that dynamic topography does not exceed a few hundred of meters. However, their analysis relies on an analytical solution for flow in an isoviscous, infinite half-space as well as the assumption that admittance, the ratio of surface topography to free-air gravity anomalies, is constant for wavelength over 500 km and equal to roughly 50 mGal/km (McKenzie, 1968, 1977). It is common to rely on this uniform scaling for mid-wavelength free-air anomalies (e.g., Hoggard et al., 2016; Jones et al., 2012; McKenzie, 2010; Stephenson et al., 2014), despite the fact that using such a constant admittance for the geoid does not hold up (e.g., Cazenave et al., 1989; Liu and Zhong, 2016). Several studies now suggest that this scaling is incorrect (Colli et al., 2016; Liu and Zhong, 2016; Yang and Gurnis, 2016), and that at wavelengths greater than several thousands of kilometers, dynamic topography can be large while free-air gravity anomalies are small.

In principle, the most straightforward way of computing dynamic topography is to know the geometry and the density of the crust and the lithosphere. It should enable one to compute the isostatic contribution to the topography and, by subtracting it from the actual topography, to recover a residual topography equal to the dynamic topography. In practice, the applicability of this method is entirely dependent upon our ability to solve for seismic, magnetotelluric and gravimetric inverse problems accurately, and to interpret the results in terms of rock properties, and crustal and lithospheric structure (Guerri et al., 2016). The studies of Faccenna and Becker (2010), Becker and Faccenna (2011) and Faccenna et al. (2014b) produce regional residual topography maps by correcting the Earth's topography for (1) the isostatic contribution of the crust (using CRUST2.0), (2) the average regional topography, and (3) the short wavelengths. They find amplitudes of ± 2 km, similar to the predictions of their global mantle flow calculations focused on the Mediterranean. Global crustal data-sets include CRUST1.0 (Laske et al., 2013) and its predecessors, CRUST2.0 and CRUST5.1 (Bassin et al., 2000; Mooney et al., 1998), which rely on a compilation of refraction profiles. However, the details of the compilation are not known, so it is difficult to estimate the assumptions and the errors involved. The highest resolution model currently available ($0.5^\circ \times 0.5^\circ$), GEMMA2012C (Reguzzoni and Sampietro, 2012), combines gravimetric data from the European Space Agency mission GOCE with CRUST 2.0. The authors report a standard deviation of up to 7 km when comparing their results with CRUST2.0 and other regional and global data-sets. As new data-sets and methodologies keep being published (e.g., Cadio et al., 2016), the coming years should see increasingly accurate estimates of the isostatic support of topography. Although the crust is at least twice as effective as the lithosphere in affecting surface topography (e.g., Faccenna et al., 2014b), the uncertainties in terms of lithospheric structure are much greater (e.g., Griffin et al., 2009). Consequently, even at a regional scale, the discrepancies in the residual topography predicted by various studies that use different crustal and lithospheric models are substantial, as illustrated for instance in the Mediterranean (e.g., Boschi et al., 2010; Faccenna et al., 2014a). In the western United States, Levandowski et al. (2014) and Becker et al. (2014b) find residual topography amplitudes that vary from about -2 to $+1$ and -2 to $+2$ km, respectively. The differences originate mainly from the difficulty of converting mantle seismic velocities to density.

Because the early studies of dynamic topography were interested in the geoid predicted by mantle flow models (e.g., Hager et al., 1985), they brought attention to the long-wavelength component of the signal. There is now growing evidence that a significant fraction of the total dynamic topography is made of shorter wavelengths, which can largely be attributed to mantle-lithosphere interactions in the shallow upper mantle. Geological records indicate that temporal changes in slab geometry adjacent to continental areas translate into major topographic changes (e.g., Fan and Carrapa, 2014; Gurnis, 1993; Mitrovica et al., 1989). Indeed, subducting slabs exert a suction in the mantle wedge that can cause vertical motion in the overriding plate. The amplitude depends on the slab dip angle and the subduction velocity (Tovish and Schubert, 1978): the shallower the dip angle and the greater the velocity, the greatest the suction, and reciprocally. Variations in one or both of these parameters can result in a successive periods of uplift and subsidence (e.g., Dávila and Lithgow-Bertelloni, 2015; Husson et al., 2014). Billen and Gurnis (2001, 2003) show that the viscosity of the mantle wedge exerts a considerable control on dynamic topography near the trench. Regional numerical modeling studies of slab-induced dynamic topography cover most of the world's subduction zones, such as (1) the western coast of the Americas: in South America (Dávila and Lithgow-Bertelloni, 2013, 2015; Dávila et al., 2010; Eakin et al., 2014; Flament et al., 2015; Shephard et al., 2010), in southwestern Mexico (Gérault et al., 2015), in the western United States (Gurnis, 1993; Heller and Liu, 2016; Liu and Gurnis, 2008; Mitrovica et al., 1989), and in Alaska (Jadamec et al., 2013); (2) the Mediterranean (Becker and Faccenna, 2011; Boschi et al., 2010; Faccenna et al., 2014b,; Husson, 2006); (3) Tibet (Husson et al., 2014); and (4) the western Pacific: in Indonesia (Lithgow-Bertelloni and Gurnis, 1997) and Australia (Gurnis et al., 1998; Lithgow-Bertelloni and Gurnis, 1997; Müller et al., 2016). Billen and Gurnis (2001, 2003) model the Tonga-Kermadec and Central Aleutian subduction zones. All of these studies report dynamic topography amplitudes of ± 1 km or more.

2.3.3. *Ongoing controversies*

In spite of the substantial disparities in methodology between geological and geophysical analyses of the dynamic topography (see also, e.g., discussions in Flament et al., 2013; Steinberger, 2016), convection computations tend to predict amplitudes that are greater than those derived from field-based analyses (Fig. 3d), which gives rise to considerable debate in the solid Earth community (e.g., Flament et al., 2013; Forte et al., 1993; Gurnis, 1990; Hoggard et al., 2016; Steinberger, 2016; Thoraval et al., 1995; Wen and Anderson, 1997; Yang and Gurnis, 2016, Guerri et al., 2016). The actual amplitude differences, as reported by Hoggard et al. (2016), are the following: "The rms amplitude for $l = 0$–3 is 170 m with peaks of -515 m and $+373$ m. A typical predictive model has an rms amplitude of 648 m with peaks of -1658 m and $+1664$ m for $l = 0$–3 (Flament et al., 2013)." So there is roughly 500 m of disagreement on average, and 1000 m at the extremums. It is worth noting that the small amplitudes reported by Molnar et al. (2015), and Hoggard et al. (2016) within the continents, are too small for mantle convection models to fit the geoid (Hager et al., 1985). There is increasing evidence that a constant scaling relationship between free-air gravity anomalies and topography is invalid for a viscously stratified, spherical Earth, i.e., that admittance kernels are depth-dependent (Colli et al., 2016; Liu and Zhong, 2016; Yang and Gurnis, 2016). For instance, following the approach adopted by Richards and Hager (1984) for the geoid, Colli et al. (2016) find that a free-air anomaly of less than 10 mGal can be associated with a dynamic topography of over 1 km, in a model where the upper mantle is less viscous than the lithosphere and the lower mantle by two orders of magnitude.

There is also major disagreement as to what wavelengths dominate dynamic topography. Spherical harmonic functions are fitted to the residual topography estimates of Hoggard et al. (2016) using a regularised least-squares inversion, and compared with the predictions of geodynamic computations (Fig. 3d). Hoggard et al. (2016) find that their model requires fitting up to degree $l = 30$, whereas geodynamic models predict that most of the power is contained within the degrees 1–3 (Fig. 3b and d). This difference can be partially explained by the buoyancy anomalies that are and are not accounted for in geodynamic models, and the scale of the analysis (e.g., the geoid is hardly sensitive to the small wavelengths). Often, mantle flow is driven by a density field derived from a tomographic model that excludes the first 300 km or so (e.g., Conrad and Lithgow-Bertelloni, 2004; Flament et al., 2015; Lithgow-Bertelloni and Richards, 1998; Thoraval et al., 1995; Yang and Gurnis, 2016). This is the simplest way of excluding the crust and the lithosphere from the calculation of mantle-driven topography in a geodynamic model, with the major caveat that it dismisses the anomalies that are the closest to the lithosphere and the most likely to induce small-scale dynamic topography. The anomalies from slabs can, however, be included using a separate slab model the builds upon tomography and seismicity. Typically, global computations cited herein do not include a crust and sometimes no lithosphere per se. Thus, the full topography cannot be computed and compared with the Earth. Although calculations of the geoid and the plate velocities, for instance, support the assumptions made in these models, the uncertainties are such that other conversion factors may be acceptable and offer a better fit to the topography. At a time when geodynamic inversions are becoming more common (e.g., Baumann and Kaus, 2015; Baumann et al., 2014; Levandowski et al., 2015; Yang and Gurnis, 2016), it is likely that fitting the full topography should improve the match between observations and predictions of dynamic topography. Beyond the controversy, improving the consistency between the predictions of dynamic topography predictions is an opportunity to understand more closely how the crust, the lithosphere and the mantle interact.

3. The feedbacks between the lithosphere and the mantle

In this section, we review the progress of models that couple the evolution of the lithosphere and that of the mantle. A primary goal of these models is to allow for the self-consistent formation of strong plates at the surface (mantle lithosphere) with limited deformation, and narrow regions with localised, large deformation. Despite tremendous efforts over the past 40 years (see reviews by Bercovici, 2003; Bercovici et al., 2000; Lowman, 2011; Tackley, 2000c), convection models with complete Earth-like surface behaviour have yet to be developed (see Bercovici, 1993, 1995; Tackley, 1998, 2000b; Van Heck and Tackley, 2008). In the following, we point out the successes and limitations of existing convection models. In addition, we draw attention to emerging theories that have been developed based on microscale physics.

3.1. Models with prescribed plate structures

Although the plates form an integrated convective system together with the mantle, the simple fluid dynamic convection theory fails in explaining plate generation and evolution. Hence, the earliest models investigating coupled plate-mantle system prescribe ad-hoc structures at the surface.

3.1.1. Weak zones as plate boundaries

In a first approximation, the Earth's lithosphere can be divided into large blocks whose kinematics can be described using rigid body rotations. Lithospheric deformation is then concentrated in narrow zones of plate boundaries. In the top few kilometers (up to a depth of around 10 km), faulting in the lithosphere (that is itself about 100 km thick) occurs by brittle failure (see reviews by Kohlstedt et al., 1995 and Burov, 2011), whereas at greater depth, convective stresses cause ductile shearing (see Fig. 4). The transition from brittle to viscous behaviour happens over several tens of kilometers with combined brittle-ductile deformation (Kohlstedt et al., 1995).

Using simple viscous flow for generating plates is by itself problematic because a viscous description implies finite deformation, whereas plates behave like perfectly rigid pieces. Moreover, a viscous flow description is continuous whereas plate boundaries are, in principle, discontinuities. Thus, many models prescribe explicitly the plate boundaries. The so-called weak zones and faults of finite extent and low viscosity are artificially prescribed at the top of the convecting layer (Davies, 1989; Gurnis, 1989; Gurnis and Hager, 1988; King and Hager, 1990, 1994; Kopitzke, 1979; Puster et al., 1995; Zhong and Gurnis, 1992, 1995, 1996; Zhong et al., 1998, 2000). Weak zones, which are typically up to hundreds of kilometers wide, simulate plate margins with imposed deformation and high shearing distributed over entire weak domains. Faults are prescribed as planes (or lines) between plates across which tangential velocities are discontinuous (Zhong and Gurnis, 1995a,b, 1996). Such faulting represents lithospheric deformation as observed at convergent margins where most deformation is accommodated through thrust faults and shear zones. Incorporating faults rather than weak zones produces more realistic plate boundary features at subduction zones. In particular, it leads to a more realistic topography at the trench and in the back-arc region (Zhong et al., 1998). The stiff plate interior is then simulated either through highly nonlinear temperature-viscosity dependence (Davies, 1989; Gurnis, 1989; King and Hager, 1994; Zhong et al., 2000)

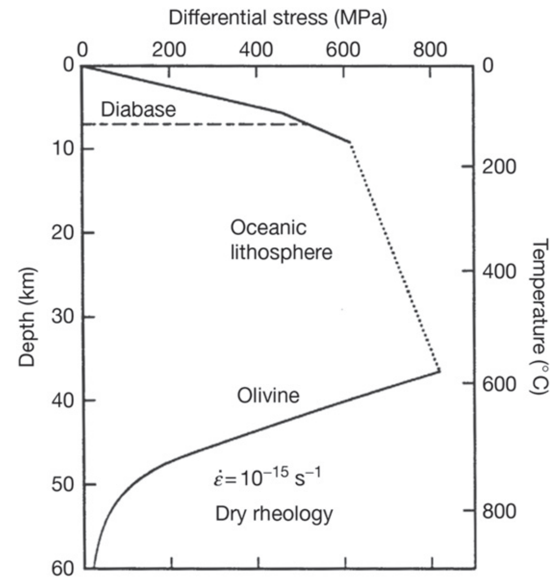

Fig. 4. Strength profile of the oceanic lithosphere. Linear stress increase with depth corresponds to brittle failure. At intermediate depths, the rock exerts brittle-ductile behaviour. At greater depths, the oceanic lithosphere deforms as ductile medium. Source: Kohlstedt et al. (1995)

or by explicitly prescribing strong regions (King and Hager, 1990; Koglin et al., 2005; Weinstein and Olson, 1992).

2-D Cartesian (Weinstein and Olson, 1992) and annulus with cylindrical coordinates (Puster et al., 1995) studies propose that the number and size of plates is in accordance with the size of convective cells, and that the plates organise mantle flow. These results are confirmed in full 3-D spherical calculations (Zhong et al., 2000). Olson and Corcos (1980), King and Hager (1990), and Puster et al. (1995) study how plate velocities emerge from buoyancy distribution in the convective interior. Olson and Corcos (1980) present an expression for the plate speed as a function of its size, mantle viscosity and surface heat flux. They show that plates can move at 4–5 cm year^{-1} in whole mantle convection, consistently with observations. Puster et al. (1995) confront the obtained plate velocities with the plate tectonic record for the past 120 My in order to evaluate convection models using plate reconstructions. They find that a 30-fold viscosity increase in the lower mantle provides the best fit to observations, in agreement with the early geoid studies of Hager et al. (1985).

The drawback of this approach is that weak lithospheric structures are imposed by hand, and thus it cannot explain how plate boundaries form and evolve.

3.1.2. Force balance method

The motion of a perfectly rigid plate can be determined from the balance of forces (Brandenburg et al., 2008; Brandenburg and van Keken, 2007; Davies, 1989; Gable et al., 1991; Gait and Lowman, 2007a,b; Gurnis and Hager, 1988; King et al., 2002; Lowman et al., 2001, 2003, 2004, 2011; Monnereau and Quéré, 2001; Ricard and Vigny, 1989; Stein and Lowman, 2010) by considering that, at each time step, the buoyancy force and the viscous forces at the base of plates must be in equilibrium. Resulting plate velocities are then discontinuous at limits between the plates. Plate boundaries are considered either fixed through a calculation (Gait and Lowman, 2007a; Lowman et al., 2001) or evolving with time (Gait and Lowman, 2007b; Gait et al., 2008; Lowman et al., 2011; Stein and Lowman, 2010).

Gable et al. (1991) present 3-D Cartesian calculations of mantle dynamics with the force balance method. They conclude that plates could be driven by thermal convection alone. In addition, they investigate the ratio of toroidal to poloidal components of the flow.

Although on Earth nearly half of the kinetic energy of the surface is represented by the toroidal flow (cf. Section 2.1 above), the first 3-D convection models yield only to relatively low values of toroidal motion (Christensen and Harder, 1991). Indeed, Gable et al. (1991) demonstrate that plates at the surface introduce a strong toroidal component into the flow field in agreement with the work of Ricard and Vigny (1989).

Monnereau and Quéré (2001) show that the simple introduction of plate conditions in 3-D spherical models, generates features such as subduction, mid-oceanic ridges and hotspots similar to Earth. Imposing plate geometry results in the development of strong downwellings along the plate boundaries that drive mantle circulation. However, Monnereau and Quéré (2001) consider fixed plate boundaries, only allowing for pure rigid body rotation. Including mobile margins would require to add rules for boundary motions, similarly as was done in 2-D models by Combes et al. (2012).

Combes et al. (2012) investigate how plate tectonics influence the thermal structure and thermal history of the mantle and test how surface processes affect mantle and plate dynamics in a 2-D cylindrical geometry. They calculate plate velocities using force balance, considering driving forces such as slab pull, ridge push, mantle drag, slab suction and slab bending. In order to include surface kinematics such as subduction initiation, trench migration, continental breakup or plate suturing, they use semi-empirical laws. Although these laws are simple, they can be improved in the future by, for example, considering more complex rheology for the plate margins. They obtain Earth-like plate dynamics: the synthetic velocities that they calculate are in a good agreement with the observed Earth plate motions. They found that for each individual plate, its velocity is mainly controlled by the size of the plate and mantle viscosity, or in other words by slab pull and mantle drag. They also show that mantle viscosity variations influence the thermal regime mainly on short time scales, and have a minor effect on Earth's long-term thermal evolution. This suggests that the Earth's cooling mechanisms are firstly controlled by surface processes, rather than being a direct consequence of mantle rheology.

The presence of mobile boundaries, thus allowing for consumption and breaking of the plates, has an important effect on oceanic heat flux. Combes et al. (2012) find that the heat loss decreases as average plate length increases. This had already been shown by Stein and Lowman (2010) in a 3-D Cartesian geometry. Stein and Lowman (2010) use the response of plates to triple junction movement to include evolving plate geometry, and assume that triple junctions move with a velocity equal to the area-weighted mean of the three neighbouring plates. In addition, evolving plate boundaries cause greater amplitude heat flux variations compared to the fixed plate models (Combes et al., 2012; Gait and Lowman, 2007b; Stein and Lowman, 2010).

Models that use weak zones and force balance, which have been constructed independently, have been confronted. Both methodologies give comparable results in terms of heat transfer, plate velocities and global kinematic energies, although differences might be found in plate deformation (King et al., 1992; Koglin et al., 2005). This shows that the two different approaches are robust and pertinent for the Earth system although they have significant limitations for the investigation of plate boundary dynamics.

3.2. Rheological methods producing plate-like behaviour

The models described above suggest that convection drives plate motion and in turn, subducting plates drive convection. The scale of convection is then mainly controlled by the size of the surface plates. With increasing plate size, the wavelength of convection increases. This has important implications for the thermal structure and heat loss of the mantle over time. In turn, mantle convective flow dictates plate velocities. In such models, the size of plates is prescribed, so that it is not a result of the self-organisation of the convective system. However, plate boundaries move with velocities comparable to those of plates. Convective models that feature dynamic evolution of plates and plate boundaries are thus necessary, and are described below.

3.2.1. Non-linear rheologies with temperature dependent viscosity

The question that arises is as follows: what is the simplest relevant rheology that produces a plate-like behaviour self-consistently, within a mantle convection framework? Plates are stiff and resistant with little deformation and can be approximated by rigid blocks with very high viscosity. This is consistent with the fact that the viscosity is activated by the temperature, with a strong dependence. Additionally, most of the deformation of the surface is confined in weaker plate boundaries, where the viscosity is significantly reduced. Strain localisation and stress dependence of the viscosity indicate that non-linear rheology of the flow is required.

It has been shown that even highly nonlinear power law rheologies with a strain rate that depends on the stress to a power exponent up to ~20, do not lead to a satisfactory plate-like flow (Bercovici, 1993, 1995; Christensen and Harder, 1991; Weinstein and Olson, 1992; Zhong et al., 1998). In particular, such rheologies do not generate toroidal to poloidal ratios comparable to the Earth, unless the power law exponent is −1 (so-called stick-slip rheology, Bercovici, 1995). In addition, large diffuse zones are observed at the surface with slowly changing velocities, which does not conform to sharp velocity gradients observed at the Earth's surface (Christensen and Harder, 1991; Zhong et al., 2000). Depending on the viscosity variations in the system, different convective styles are observed for temperature dependent rheology only (Christensen, 1984; Solomatov, 1995). For a weak temperature dependence of viscosity, the convection is akin to isoviscous flow. For moderately temperature dependent viscosity (viscosity contrast across the convecting body lower than 10^4), convection develops a sluggish mobile lithosphere with increased wavelength of convection cells. Finally, strong dependence of the viscosity on temperature locks up the surface and stagnant-lid convection regime develops. The top boundary layer becomes frozen and immobile (see Fig. 5). Considering temperature dependence of the viscosity only, the Earth (viscosity contrast between 10^7 and 10^{16} assuming viscosity of the interior of 10^{21} Pa s and Rayleigh number 10^8 based on the core-mantle boundary viscosity) should be in the stagnant-lid regime (Bercovici et al., 2015; Solomatov, 1995).

3.2.2. Pseudo-plasticity

In order for the stiff lid in the convective models to yield, some softening behaviour is necessary. The simplest method to achieve mobilisation of the upper cold boundary while keeping large rigid areas is to limit the stress and to maintain a linear relationship

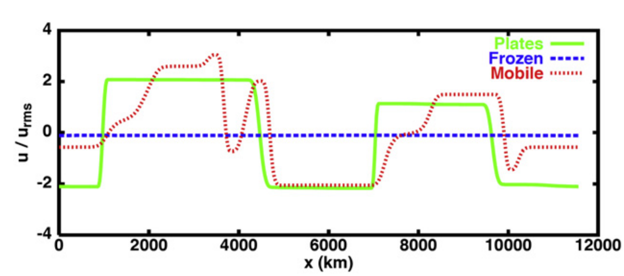

Fig. 5. Normalized horizontal velocity profiles at the surface of the 2-D convecting Cartesian box. Stagnant lid regime (blue dashed line) occurs for the fluid with strongly nonlinear temperature dependent rheology. In this case, lithosphere does not participate in convective movements. In mobile lid regime (red dotted line) continuous lithospheric deformation occurs. Plate like regime (green solid line) results in nearly piecewise linear step velocity profile. Rigid blocks move with constant velocity while high deformation is localised in narrow regions representing the plate boundaries.
Source: Richards et al. (2001)

between stress and strain rate (Moresi and Solomatov, 1998, see Fig. 6). This results in reducing the viscosity in narrow zones representing failure of the lid, when a yield stress is reached. Such rheology is called pseudo-plastic. A special effort has been done in the geodynamical community to benchmark codes that incorporate visco-plastic yielding. Tosi et al. (2015) propose several benchmark exercises that test linear and nonlinear rheological behaviour and confront 11 different convection codes. They match the required diagnostics such as average temperature, bottom and top Nusselt numbers, root mean square velocity over the whole domain, maximum and minimum viscosities in the system, average rate of work done against gravity and average rate of viscous dissipation with a maximum of 3% difference.

Incorporating a yield stress together with highly temperature dependent viscosity produces plate-like tectonics: subduction of the stiff layer, plate-like surface motion, continuously evolving plate boundaries that persist over long periods of time, high toroidal-poloidal ratio of the surface velocity field, presence of passive spreading centers and long wavelength of mantle convection (Loddoch et al., 2006; Tackley, 2000d; Trompert and Hansen, 1998). Examples of 3-D Cartesian numerical calculations for different yield stress values are depicted in Fig. 7. Different convective modes are observed depending on the value of the yield threshold: from sluggish lid convection with diffuse spreading centers for low yield strength to plate-like surface for some optimal value interval of the yielding. Further increasing the yield stress leads to episodic lid mobility. Finally, a permanent rigid lid is formed at high yield strength values. Additionally, plate-like behaviour is enhanced by a low viscosity asthenosphere that helps localising the deformation by decoupling the plates from the underlying mantle (Tackley, 2000d; Zhong and Gurnis, 1996; Zhong et al., 1998). Great effect on localisation of deformation is also obtained when melting is included (Tackley, 2000d). Tackley (2000d) introduces a viscosity reduction where the temperature exceeds a depth dependent solidus resulting in low-viscosity regions underneath the spreadings. Such lubrication of spreading centers results in stronger plates and interconnected network of plate limits improving the plate like behaviour. Extension to full 3-D spherical geometry favours a more Earth-like torodidal-poloidal ratio compared to simulations in Cartesian box (Foley and Becker, 2009; Van Heck and Tackley, 2008).

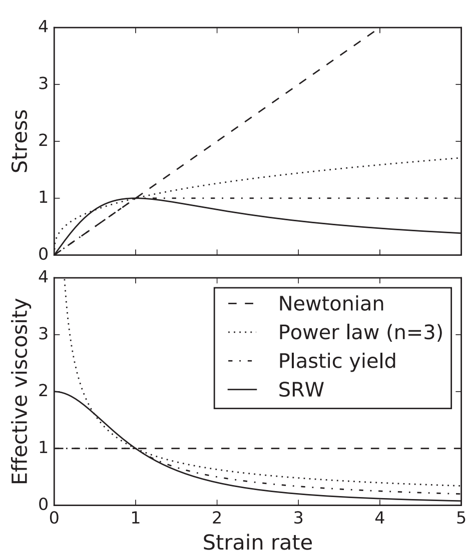

Fig. 6. (Top) Stress and (bottom) effective viscosity dependence on the strain rate for different rheologies: linear Newtonian, power law rheology with $n = 3$, visco-plastic rheology with yield stress threshold and strain rate weakening (SRW) rheology. Source: Tackley (2000b)

Although pseudo-plasticity successfully achieves first-order tectonic features, it still fails to produce accurate plate boundaries, including sharp, pure strike-slip margins, which are intrinsically 3-D features associated with toroidal motion. These are often rare or absent in such models. Also, adjacent plates participate in downwelling currents at convergent boundaries, producing symmetric subduction zones. This is non-conform to terrestrial intra-oceanic subduction zones that are always single sided. Importantly, imposing a single yield threshold does not allow to explore how lithosphere rheology evolves from strong plate interior to weak plate boundaries (Zhong et al., 1998).

The proposed values of yield stress for which a mobile plate regime exists are rather restricted (Stein et al., 2004; Tackley, 2000d), and do not match with experimental (Demouchy et al., 2013; Evans and Goetze, 1979; Hirth and Kohlstedt, 2003) and field based studies (Zhong and Watts, 2013). The experimental estimates are about 10 times larger than the geodynamic ones. This disagreement points towards a lack of realism in the rheology used in models. However, this straightforward comparison is misleading because the conditions required to do it are not ensured. Indeed, comparing the dimensional value of the yield stress in a geodynamic model to that of an experiment requires that (1) all of the other parameters of the model are realistic already, (2) the transfer of the scale from the experiment to that of the model is correct. Neither conditions can yet be matched. A realistic model for the Earth's mantle in terms of parameters is still out of reach in our computations, and the approximations in the physics might be too crude anyway. Therefore, the non-dimensional ratio of the yield stress to a measure of viscous stresses is relevant. Viscous stresses increase themselves with Rayleigh number.

In the laboratory experiments, the time scales and the length scales are by no means comparable to those needed in models (times larger than 10,000 years, and scales larger than 1 km). Also, the dominating deformation mechanism at low temperature ($<$1300 K), is still questioned, as well as the behaviour of slip systems (Tielke et al., 2016). In situ tensile stress measurements on micron scale samples open the way to study ductile deformation in more appropriate conditions, but the experiments are performed only under strain-hardened conditions (Idrissi et al., 2016). They suggest that plasticity occurs at lower stresses compared to former experiments. The transfer of scale from the crystal or aggregate to the 1 km scale is a difficult task that needs to account for the presence of petrological heterogeneities and fluids.

3.2.3. Self-softening rheologies

In order to overcome the limitations of pseudo-plastic rheology, models with additional ingredients have been proposed with the intention to better capture localisation of deformation at plate boundaries. Indeed, many processes weaken the lithosphere to a greater extent than plastic yielding, including for example dynamic recrystallisation, shear heating or migration of weak material into fault zones.

Regarding these issues, regional modeling plays a guiding role. In particular, lithosphere studies show that self-lubricating rheology, such as strain and strain-rate weakening, is necessary to produce transform plate boundaries (Bercovici, 1993, 1995; Gerya, 2010, 2013). The regional model of Gerya (2013) assumes realistic temperature and strain rate dependent viscosity law computed according to experimentally determined flow laws (Ranalli, 1995), although thermally induced stresses are neglected. For long-term global models, this remains out of reach because of computational limitations. Therefore, mantle convection calculations use instead parametric laws with tuned parameters that capture essential self-softening properties.

Unlike pseudo-plastic behaviour, strain rate softening mechanisms decrease both the viscosity and stress with increasing strain rate (Bercovici, 1995; Whitehead and Gans, 1974, see Fig. 6). Such rheology implemented in 3-D Cartesian calculations show that

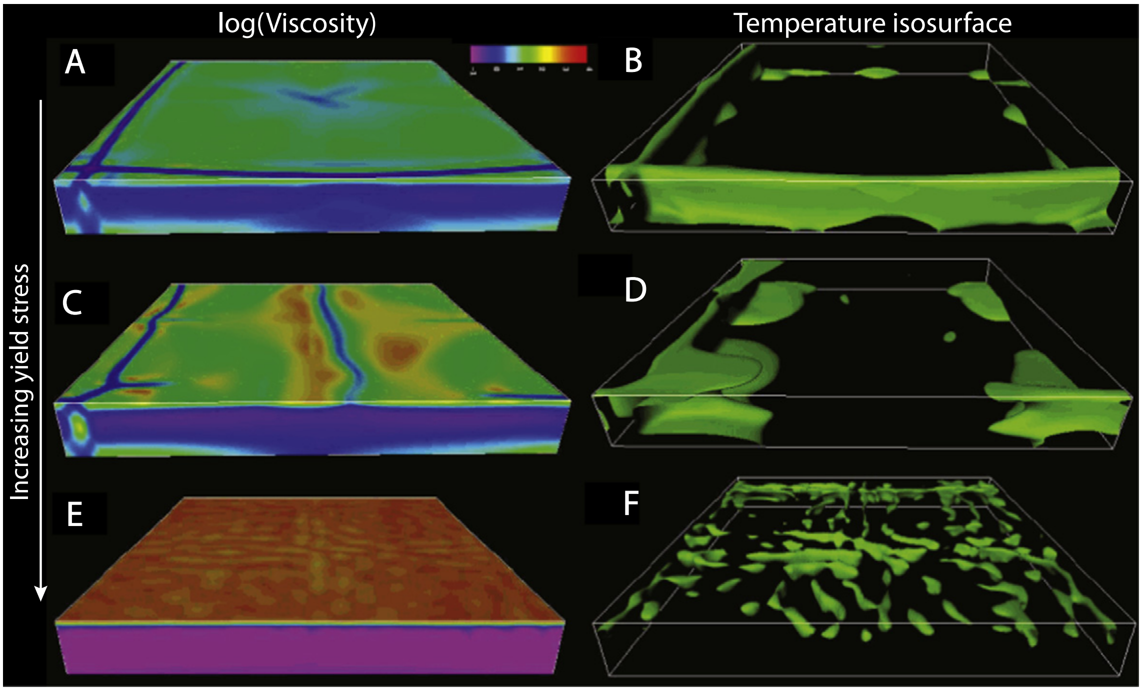

Fig. 7. The effect of yield stress on convective regime in 3-D numerical simulation. (Left) Nondimensional viscosity varying from 10^{-1} (pink) to 10^4 (red). (Right) Isosurface representing cold downwellings where the temperature is 0.1 lower than the geotherm. For increasing yield threshold, the convection changes from sluggish mobile lid (top row) to plate-like regime (middle row). For inefficient yielding, the top boundary locks up and stagnant lid develops (bottom row).
Source: Tackley (2000a)

using strain rate softening is more effective in producing plate like behaviour than plastic yielding (Tackley, 1998, 2000d). A leap forward has been achieved in weak zone generation. Using strain rate weakening rheology produces more localised deformation zones above down- and upwellings, but also in regions that connect them (Tackley, 1998). Regions with significant shearing are then characterised by an extreme viscosity drop compared to the inner parts of the plates. Although strain rate weakening improves strain localisation at divergent boundaries, it can produce a more complex network of spreading centers leading to more fragmented plates and highly episodic downwellings (Tackley, 2000d).

On Earth, it has been acknowledged that lithospheric deformation at the plate boundaries has a memory of previous yielding. Hence, large-scale faults and shear zones reflect the deformation history rather than instantaneous stress (Bercovici and Ricard, 2003, 2005, 2012, 2013, 2014, 2016; Gurnis et al., 2010; Zhong et al., 1998). This is not captured by any rheology described so far. In order to include memory in the system and incorporate the pre-existing weak structures, the rheology must be stress history dependent. A first order approach includes a memory of the crust and lithosphere by introducing a damage parameter that describes to which degree the local materials are damaged (Ogawa, 2003; Tackley, 2000d; Yoshida and Ogawa, 2005). The more the material has been subjected to deformation, the more it is damaged. The viscosity drops with increasing damage. In this case, the damage parameter does not stand for a specific microscopic weakening mechanism, but rather represents a general weakening. Damage then follows an evolutionary equation that is proportional to the work done and depends on the healing rate. Healing is itself dependent on temperature, enforcing long memory of cold lithosphere, while allowing for rapid healing at mantle temperatures. In a steady state limit, such description simplifies into an instantaneous strain-rate weakening rheology as described above, ensuring compatibility with previous models.

Such a rheology has been employed in 2-D convection models by Ogawa (2003) and in 3-D by Tackley (2000b) and Yoshida and Ogawa (2004). Tackley (2000b) considers that the damage parameter reduces viscosity linearly. He concludes that the time dependent damage rheology and instantaneous strain rate weakening give very similar results with two main differences: damage rheology results in less fragmentation of spreading regions and there are less episodic downwellings. However, such rheologies generally lead to weaker, more localised spreading centers, a higher effective viscosity of the plates and a more realistic vorticity distribution (higher vorticity along plate boundaries). Ogawa (2003) assumes nonlinear damage equation that results in a stress-viscosity hysteresis with weak and strong branches: the intact branch at low stresses, and the damaged branch at high stresses. The hysteresis induces a plate-like regime with fragmented lithosphere that consists in large rigid moving blocks that are on the intact branch, separated by narrow mechanically weak zones that are on the damaged branch. In addition, Ogawa (2003) shows that in order to obtain a plate-like regime, the viscosity must be strongly temperature dependent (viscosity contrast between the surface and bottom boundaries larger than 10^6) and there must be also a strong dependence of viscosity on damage (see Fig. 8). Yoshida and Ogawa (2005) further explore the influence of the internal heating showing that in such model the plate-like regime occurs only for low and moderate heating rates.

3.3. Directions for improved plate-like behaviour

3.3.1. Grain damage and pinning rheology

Damage, as invoked in Section 3.2.3, introduces memory in the rheology. However, a shortcoming of the models mentioned above is that the damage parameter does not represent any particular physical mechanism. A theory based on damaged grain size-dependence aims at improving convection models by treating the physics that reflects micro scale processes (Auth et al., 2003; Bercovici, 1998; Bercovici and Ricard, 2003, 2005; Bercovici et al., 2001a,b; Landuyt and Bercovici, 2009; Landuyt et al., 2008; Ricard and Bercovici, 2003, 2009; Rozel et al., 2011). The observation of grain structure of deformed rocks pilot this approach. In particular, grain reduction in mylonites (rocks that formed by the accumulation of large shear

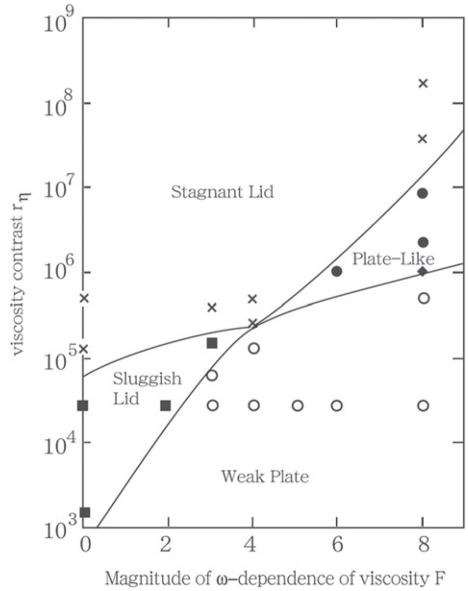

Fig. 8. The regime diagram for a convecting fluid with temperature, pressure and stress history dependent rheology. Plate like behaviour develops for high viscosity contrasts between the surface and the bottom boundary together with the strong dependence of viscosity on damage parameter ω. Symbols in the graph represent where numerical calculations were conducted.
Source: Ogawa (2003)

stresses in ductile fault zones) indicates that grain size dependent rheology controls shear localisation. In the grain dependent damage model, the self weakening positive feedback comes through the interaction of the grain size dependent-rheology and grain reduction driven by dynamic recrystallisation. However, several barriers exist in this model in order to successfully explain shear localisation and weakening. Firstly, grain size reduction by recrystallisation happens in deformation regime that is independent of grain size (dislocation creep) and that differs from the deformation regime where softening by grain reduction occurs. Thus, the necessary lubricating feedback is not straightforward. Secondly, healing of the damaged zones due to grain-growth by coarsening is too fast to ensure long-lived weak zones.

To overcome these problems, a more complex continuum theory has been proposed (Bercovici and Ricard, 2012, 2013, 2014, 2016). The earliest theoretical models suppose that the lithosphere is composed of single phase mineral assemblages, while Earth's shallow parts consist of olivine (about 60%) and pyroxene as the second major component. Taking into account both phases proves to be critical since it considerably slows down grain coarsening, as pyroxene grains act as an obstruction for grain boundary migration. This effect, known as Zener pinning, is key for plate generation and shear localisation. In fact, it allows for damage and grain size reduction to occur simultaneously with weakening. Moreover, interfacial barriers significantly reduce healing rates, leading potentially to long-lived dormant weak zones that take several 100 My or more to vanish (Bercovici and Ricard, 2012, see Fig. 9). Although these results are ultimately encouraging, such rheological model is too complex to be easily injected into global convection models yet, and extensive benchmarks of this theory against natural and experimental observations are required.

3.3.2. Models with a deformable upper boundary

Although the aforementioned models have been successful in obtaining Earth-like features such as localised deformation, there has been little progress in reproducing robust stable evolving pure strike-slip margins. In addition, none of the models produce an Earth-like subduction that is one sided, with the overriding plate staying on top of the subducting plate. Indeed, in global convection models both plates actively participate at convergent boundaries and sink deep into the mantle. This can be overcome by introducing continental lithosphere that eventually breaks the symmetry. However on Earth, one sided ocean-ocean subduction exists too.

It has been argued that the topography at subduction zones might affect how plates dive into the mantle. Indeed, on Earth, significant topography lows are associated with deep trenches where subducting plates bend. Global convection calculations usually neglect elevation change, imposing the upper surface to be flat with zero shear stress and finite normal stresses, supposedly proportional to what a topography would be if the surface was allowed to deform. Increasing computational power, together with advanced numerical techniques (Crameri et al., 2012; Schmeling et al., 2008) now allows for a new type of global convection models with a free surface that can deform. Indeed, Crameri et al. (2012) and Crameri and Tackley (2014) show that a free deformable upper boundary condition leads to single sided subduction and hence to more realistic slab dip, stress state inside the slab, trench retreat rates, plate velocities, slab-induced mantle flow, and back-arc spreading. Additionally, as suggested by earlier regional models (Gerya et al., 2008), the presence of weak hydrated layer atop subducting slabs enhances the asymmetry of subduction, as it acts as a lubricant between the sinking and overriding plate (see Fig. 10). However, the convection is often episodic in these models and single sided-subduction is temporally unstable. In order to make it long-lived, an extreme viscosity contrast between plates and sublithospheric mantle is necessary as it helps the sinking part of the plate and its surface to behave as one connected unit (Crameri and Tackley, 2015).

Nevertheless these types of models are still not common in the geodynamical community, as they are computationally demanding partly because they require a high resolution (of the order of kilometers). In the meantime, new numerical methods have been recently developed to facilitate the inclusion of a deformable surface at lower computational costs in global convection codes (Duretz et al., 2016). However, such techniques are yet to be implemented in global covection codes.

3.3.3. Visco-elasto-pseudo-plasticity

Geological processes associated with mantle convection happen over long timescales of the order of 10 My and more. On the other hand, for shorter time windows, the elastic response of lithosphere is relevant too (e.g., the mantle is seen like elastic body for seismic waves or flexure). The characteristic time that determines rheological behaviour is given by the ratio of effective viscosity η_{eff} and elastic shear modulus μ: the Maxwell time $\tau = \eta_{\mathrm{eff}}/\mu$. Maxwell characteristic time is several orders of magnitude higher for the lithosphere compared to the sub-lithospheric mantle, due to viscosity increase with decreasing temperature. For a shear modulus of 10^{11} Pa and an effective viscosity of 10^{21} Pa s and 10^{26} Pa s for the mantle and the plates, the Maxwell time is 300 years and 30 My, respectively. As a consequence, neglecting elasticity for mantle convection studies featuring plates is not straightforward. Therefore, many authors have investigated the role of elasticity for primarily lithospheric processes such as slab bending or lithospheric instability growth. Solving for the viscous and the elastic deformations in one model is challenging, since these two occur on different timescales. Therefore, Moresi et al. (2003) develop a numerical treatment that integrates the effects of elasticity in long-term convection modeling.

Thielmann et al. (2015) shows that including elastic rheology into convection models significantly alters the stress distribution in the lithosphere. It is thus important to have a realistic stress distribution in the upper most mantle. Most convection models use free slip upper boundary that might give non-realistic stresses due to the imposed flat surface. Indeed, employing a free surface results

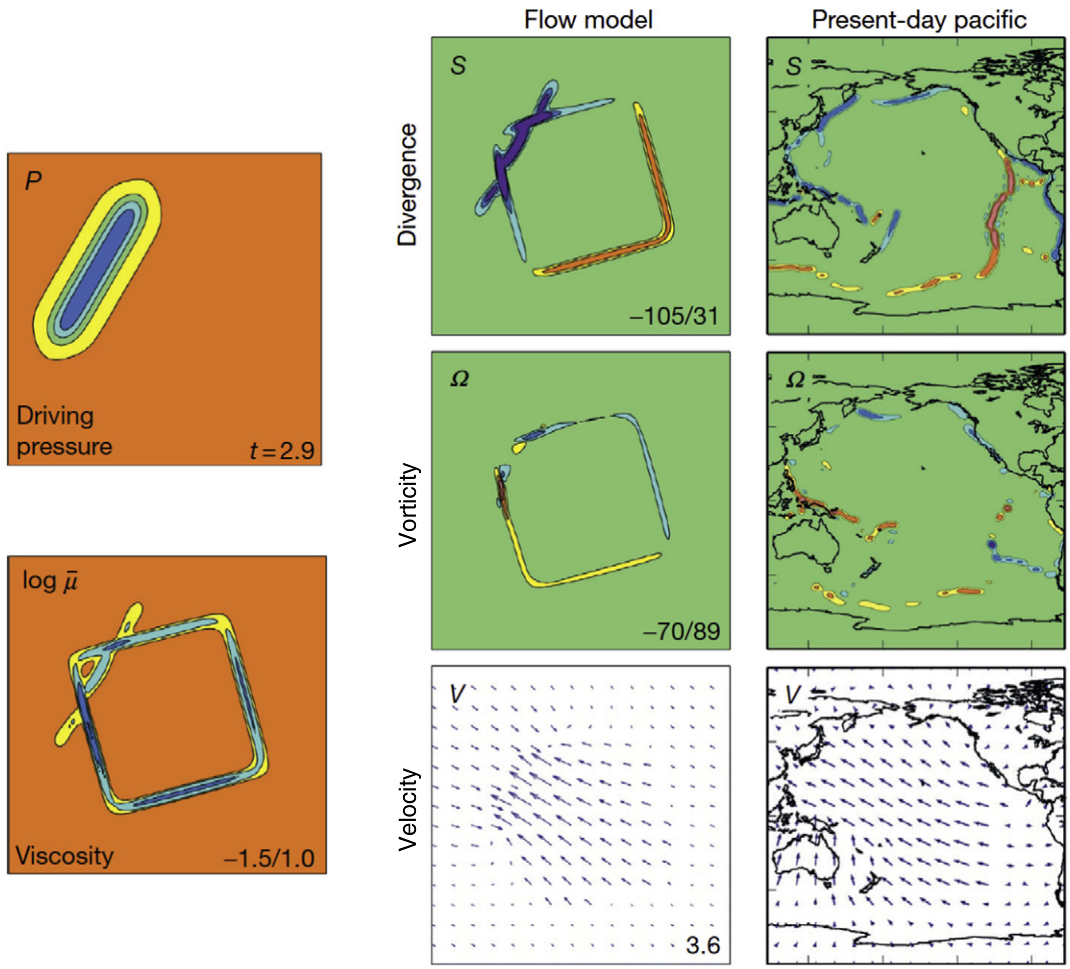

Fig. 9. Lithospheric flow model with complex rheology. The model assumes instantaneous idealized subduction driven flow (top left) with rheology that includes grain evolution and damage mechanism and is consistent with field and laboratory observations on rock deformation. After abrupt rotation is applied, the new configuration retains initial driving configuration that is visible in the divergence and vorticity fields (middle column) as well as in the viscosity structure (bottom left). The sudden change in direction corresponds to reorganisation of the Pacific plate motion as is indicated by the Hawaiian chain. For comparison, the present day Pacific configuration is depicted in the third column.
Source: Bercovici and Ricard (2014)

in increased stresses in the stagnant lid caused by bending, at both shallower and deeper levels (up to two magnitudes larger than in the free slip case). Elasticity then counteracts the bending effect and significantly reduces the surface stresses, but distributes stresses over a thicker layer than in the case of viscous rheology. Both the combined effect of the free surface and elasticity thus increase stresses at greater depths in the lithosphere. This might create deep shear zones that are ultimately needed for subduction initiation. However, the models of Thielmann et al. (2015) suggest that even increased, the stresses still remains too small for the rock to yield at depth and additional weakening such as grain-size reduction is necessary.

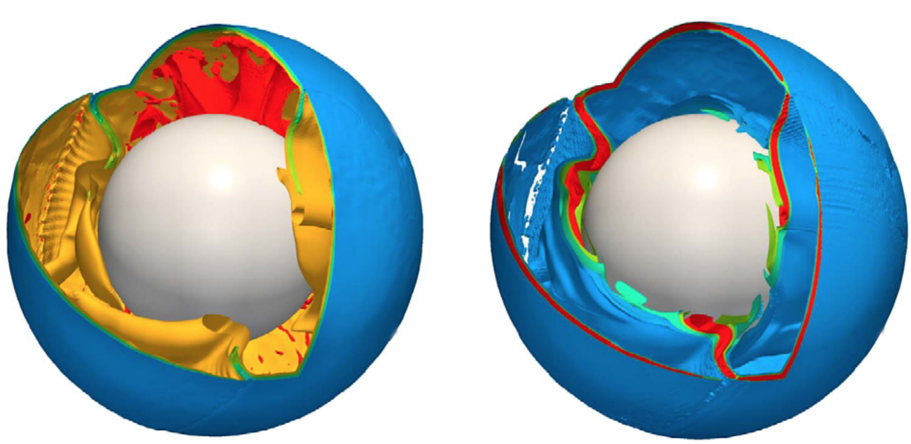

Fig. 10. 3-D fully spherical numerical simulation with deformable surface featuring one-sided subduction. (Left) Temperature and (right) viscosity isosurfaces depicting hot upwellings and cold downwellings.
Source: Crameri et al. (2012)

4. Global tectonics of convection models with self-consistent plate generation

The connection between the surface of the Earth and the dynamics of its interior is now widely accepted in the community. In order to study the feedbacks between tectonics and convection, geodynamicists have developed advanced models in terms of numerical techniques and physics (see Section 3). Importantly, these models feature tectonics and convection in a self-organised evolving system. While their primary goals remain to identify the physical requirements that lead to the existence of plate tectonics, very few of them investigate the physics behind observed tectonics. Below we first describe these studies. We then choose one particular convection solution that is carefully analysed in terms of global tectonics. The realistic tectonic features produced show such models can be used to understand fundamental mechanisms behind plate tectonic organisation and evolution on Earth.

4.1. Tectonics and self-organisation in convection models

4.1.1. Continental drift

Continents drift at the Earth surface, in response to a dynamic feedback between convective flow, and unsinkable and conductive continents. The coupling between continental lithosphere and the rest of the mantle is both thermal and viscous. Early models of convection with viscous rafts as continental lithosphere analogs, show that a downwelling in between two continents draws them to aggregate, while downwellings away from a supercontinent break it up (Gurnis and Hager, 1988; Trubitsyn et al., 1993). Subsequent studies show that large scale flow tends to produce continental drift and aggregation, while small-scale flow (plumes being among the small scale structures) tends to prevent them (Lowman and Jarvis, 1993, 1996; Phillips and Bunge, 2005, 2007; Zhang et al., 2009; Zhong and Gurnis, 1993). These works highlight that imposing a continent atop a convecting layer increases the wavelength of the flow (Guillou and Jaupart, 1995; Phillips and Bunge, 2005; Phillips and Coltice, 2010; Zhong and Gurnis, 1993). Similarly, the transition from aggregation to dispersal is viewed as a flow reversal that breaks apart a supercontinent in case of large-scale flow (Lowman and Jarvis, 1993, 1996). As a synthesis, Zhong et al. (2007b) and Zhong and Liu (2016) propose a conceptual model for the cyclic behaviour referred to as the Wilson cycle: an organisation with a dominating degree 1 in the temperature field leads all the continents to build up a supercontinent above the single subduction system, and following the subsequent flow reorganisation into a degree 2 configuration, the new ring of subduction away from the supercontinent drags pieces apart to cause break up of the supercontinent (Fig. 11). This idea is consistent with what is observed in the models cited here.

In the above modeling, only the works of Lowman and Jarvis (1993, 1996) employ models with a type of self-organised plate-like behaviour, using a time-dependent coupling between rigid surface plates and the underlying flow. It is only recently that numerical models have started to take into account the interactions between self consistent plate generation and continental rafts (Rolf and Tackley, 2011; Yoshida, 2010). A major effect of including pseudo-plasticity in these models is the increase of the wavelength of convection with the yield strength of the upper boundary layer (Foley and Becker, 2009; Van Heck and Tackley, 2008). In most of the plate-like regime, the dominating wavelength of convection is already around degree 2 without the presence of continents. There is a strong possibility that the strength of the lithosphere exerts a stronger control on the long wavelength structure of the flow, than continental rafts at the surface. However continents enhance fluctuations of the thermal structure (Rolf et al., 2012). When 2 continental rafts and plate-like behaviour are combined in 2-D spherical annulus geometry and 3-D spherical geometry, there is a range of values for the yield stress that produces a statistical cyclicity of aggregation and dispersal on times scales comparable to the observed timings on Earth (about 500–700 My for a cycle). When the yield stress is too low, convection is characterised by smaller scales, and continents do not aggregate frequently enough, while for a high yield stress convection is too large-scale, almost stuck to degree 1, to produce frequent supercontinent breakups (Rolf et al., 2014). Consistently with this result, Yoshida (2014) shows that an intermediate yield stress in the range of those producing a plate-like behaviour, breaks up and disperses the pieces of a supercontinent with the shape of Pangea. Such forecasting experiment, though limited by unknown initial conditions precluding accurate prediction (Bello et al., 2014), generates some continental motions in similar directions as on Earth (Yoshida and Hamano, 2015).

The drift of continents is a response to the dynamic coupling between the buoyant and viscous continental lithosphere with the underlying flow. Continents are advected at the surface towards a subduction zone until they reach the trench. At this point subduction could die out or reverse its polarity to pursue. The strength of the lithosphere produces overall a large scale flow that allows continents to aggregate together, and the reorganisation of the flow in a new large scale configuration breaks up the supercontinent. Although modeling may not be the limiting factor now, many first-order questions about continental drift remain: is there a speed

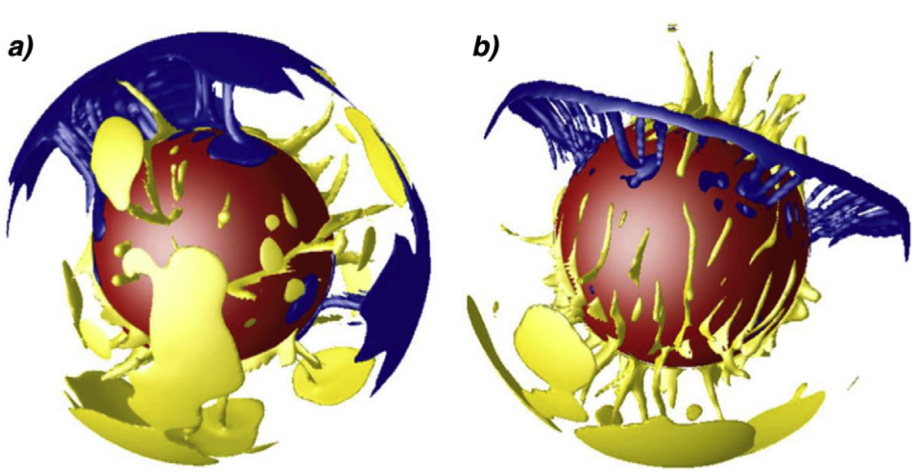

Fig. 11. Typical thermal structures in the numerical solutions of 3-D spherical convection of Zhong et al. (2007b), represented as isotherms (cold in blue and hot in yellow): a) a degree-1 planform leading to continental aggregation over the region of downwellings, and b) a degree-2 planform leading to dispersal of a supercontinent.
Source: Modified from Zhong et al. (2007b)

limit for continents as suggested by plate reconstructions (Zahirovic et al., 2015) and why do continents go slow or fast? Are there different mechanisms for aggregation, since some supercontinents form by closing up the oldest ocean basins (extroversion) and some the youngest ones (introversion) (Nance et al., 2014)? How do plumes and melting impact break up? What is the impact of small scale convection below continent on continental drift and stability of the continental lithosphere? Are there specific tectonic plate configurations (size, velocity, age) when continents aggregate and disperse? Zhong (2001) suggested that net rotation is influenced by continental blocks, but when plates are taken into account, how does this connection work? How is continental drift influenced by the presence of deep mantle chemical heterogeneities? Do continents localise subduction initiation at their boundaries?

4.1.2. Seafloor spreading

As seen above, models that could produce seafloor spreading emerged just before 2000 with the introduction of pseudo-plasticity or force balance with evolving plates, but they are mostly used to identify the key factors for the existence of plate tectonics on Earth and other planets (see Bercovici et al., 2015b), or to characterise the overall flow patterns of global convection (Gait and Lowman, 2007b, for instance). It turns out that average properties of flows models in the plate-like regime display similar dependence with Rayleigh number to that of isoviscous models (Grigné et al., 2005), and a small dependence on yielding values when in the plate-like regime.

Although numerous numerical solutions of convection with plate-like behaviour exist in the literature, very little has been made on characterising how seafloor spreading operates. Perhaps this is because convection models with one-sided subduction are still difficult to generate, and hence models with imperfect subduction style could make the study of seafloor spreading a hazardous project. Despite this problem, Husson et al. (2015), using models with limited self-organisation, propose that the spreading rate is modulated by forces acting on convergent boundaries of the plates, continental collisions or specific subduction dynamics curbing spreading. More complex models with plate-like behaviour and continental rafts reproduce the distribution of seafloor ages on Earth, uncovering why subduction affects oceanic lithosphere of all ages (Coltice et al., 2012, 2013). These two studies show that the shape of continents geometrically imposes that the age of subduction on the adjoining oceanic plate varies along the trench. Coltice et al. (2013) also show that these models display a significant time-dependence of the average spreading rate, which alter the age vs. area distribution of ocean basins. Increasing continental area enhances spreading rate and associated fluctuations (Coltice et al., 2014). Plate reorganisations produce spreading rate changes that can reach a factor of 2 (Coltice et al., 2013), along with variations of the length of mid ocean ridges, and changes in the shape of the area vs. seafloor age distribution (Fig. 12). These predictions by convection models are consistent with the plate reconstructions for the past 200 My. Yet, several problems arise from these works: in these calculations, seafloor age is computed from the heat flow, hence erroneous ages are generated close to subduction and where small-scale convection below oceanic plates operates; the models are simplified, having a lower vigor of convection than expected on Earth, and lacking viscosity change with depth or relatively smooth temperature dependence of viscosity among other complexities; plate reconstructions are restricted to the Mesozoic and Cenozoic, and uncertainties grow with geological time, hence it is difficult to compare evolutions of the models to the evolution of the Earth.

Seafloor spreading suffers from a lack of observation going back in time, compared to continental drift. Hence, convection models with plate-like behaviour have a strong potential to fill in gaps in the geologic record, and to improve significantly our knowledge on the dynamics of ocean basins. Very little has been attempted so far, therefore first-order questions remain that could be tackled with the models that already exist: how do spreading and subduction initiate and die out? How does spreading rate depend on the plate/subduction configuration? How does small-scale convection develop beneath oceanic plates in a self-organised convective system? What are the factors producing mid ocean ridge/trench length fluctuations, and can we put bounds on these fluctuations? How do plumes interact with mid ocean ridges and how do they modify (or not) spreading? What is the impact of deep chemical heterogeneities on spreading rate evolution? among other questions.

4.1.3. Plate dynamics

Numerical models of mantle convection produce continuous fields while plate tectonic theory is discontinuous. Bercovici and Wessel (1994) propose a continuous description of surface kinematics for plate tectonics on Earth. A complementary direction, which would transform the continuous fields of convection models to discontinuous plates, is yet to be done. Mallard et al. (2016) analysed a combination of information given by the viscosity, velocity and temperature fields to build plate layouts from convection models by hand. This work shows that the large plates on Earth are connected to the dominating length-scales of convection, while the smaller plates come from the fragmentation near trenches by stresses depending on subduction geometry. Hence smaller plates record the evolution of a subduction zone over 10s of million years, while larger plates record the onset and demise of slabs over 100s of million years (Fig. 13). Because studying the self-organisation of plate tectonics with convection models is at its infancy, many first-order questions can now be investigated and it is useless to try an inventory here.

4.2. A numerical solution of joint convection and tectonics

The purpose of this subsection is to describe a numerical solution of particular convection models that feature plate-like surface behaviour and continental rafts, pushing the models at their limit. We highlight their key structures and give insight on what they can be used for.

4.2.1. The model

We describe here a 3-D spherical numerical solution for a convection model exhibiting continental drift, plate-like behaviour and asymmetric subduction over 250 My. Such calculation requires significant computational power. Most of this time is spent to reach a statistical steady-state and find proper pseudo-plastic parameters. Exploring the parameter space could give access to more interesting solutions, but significant additional resources would be needed. Therefore, we present here a tectonic and convective evolution started from a statistical steady state obtained in the plate-like behaviour regime.

We solve for the equations of mass, momentum and energy conservation and advection of material properties. We use the code StagYY (Tackley, 2008) in 3-D spherical geometry with a Yin-Yang grid (Kageyama and Sato, 2004). StagYY handles viscosity contrasts up to 10^4 between adjacent nodes (Tackley, 2008). The resolution used here is 30 km on average, refined close to boundary layers in the vertical direction (10 km radial resolution close to the surface). Convection is incompressible, and the specificity of this calculation is to display a strongly variable viscosity, up to 9 orders of magnitude. Viscosity increases exponentially with depth by a factor of 20. We also impose a viscosity increase by a factor of 30 at 660 km. Viscosity is thermally activated, following

$$\eta = \eta(z)\ \exp\left(\frac{E_a}{RT}\right),$$

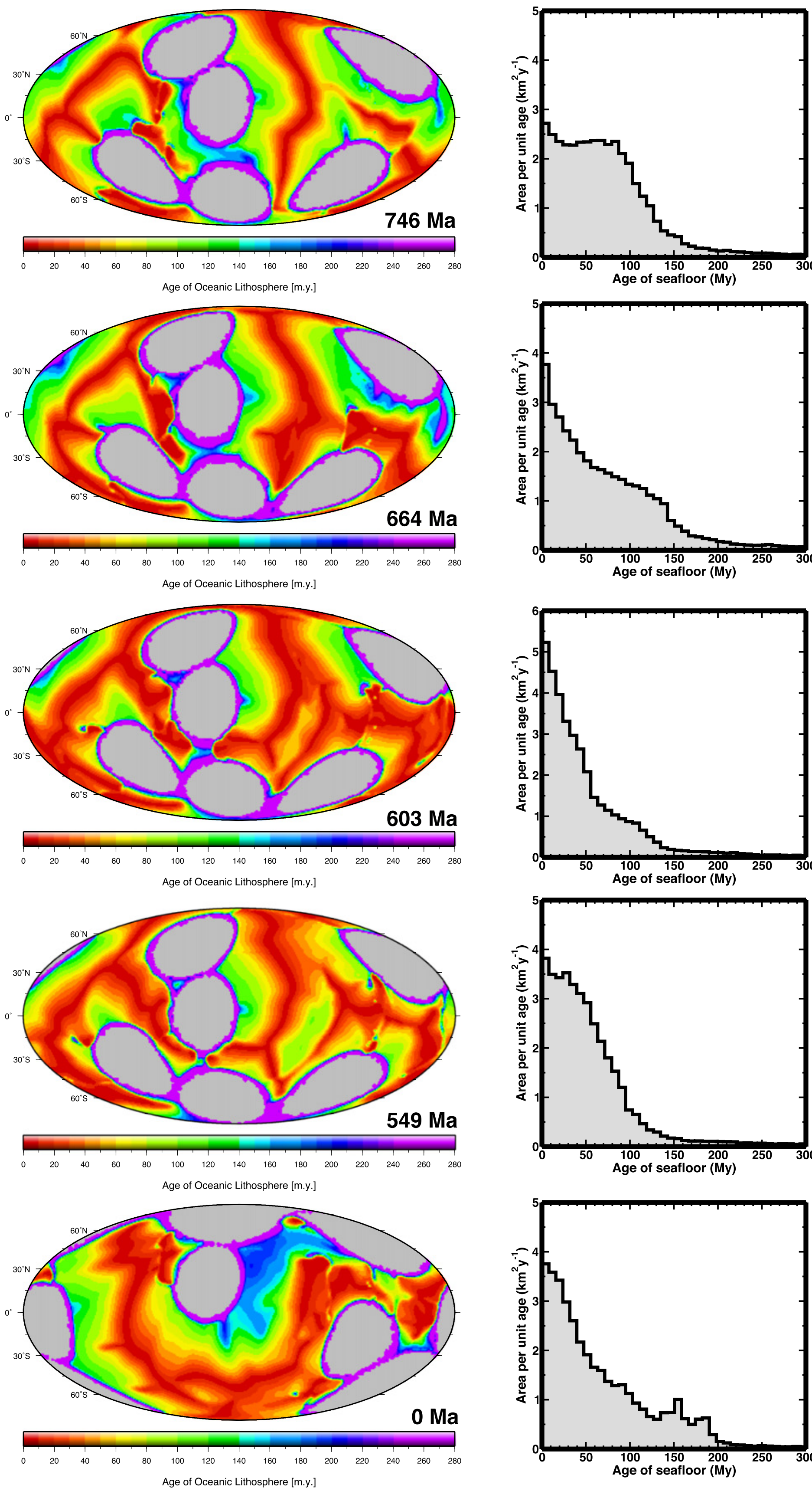

Fig. 12. Synthetic maps of seafloor ages and associated area-age distributions in a mantle convection model with 6 continental rafts (in grey). The selected results present 200 My of evolution in the first four rows and the situation 549 My after that 200 My evolution.
Source: Coltice et al. (2013)

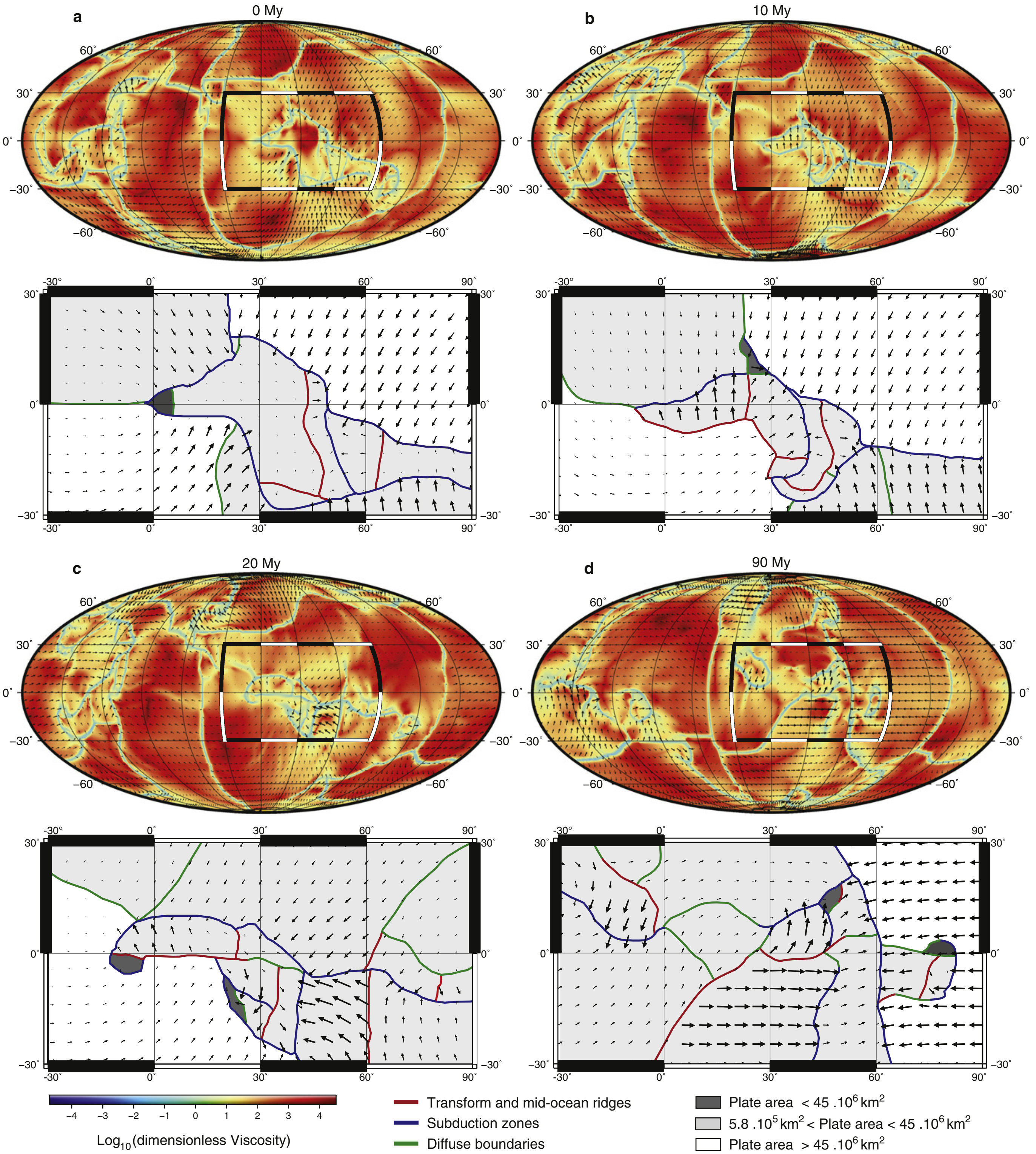

Fig. 13. Typical global viscosity maps and associated local velocity in a 3-D spherical convection model with pseudo-plasticity. Plate boundaries are interpreted from the viscosity, velocity and temperature fields. Ridges are in red, trenches in blue, and diffuse and transform boundaries are in green. a–c are separated by 10 Ma. The shape of large plates show very little changes, in contrary, the adjustment of small plates evolve quickly. d. 90 My after the first snapshot, the distribution of large plates and small plates has evolved significantly.
Source: Mallard et al. (2016)

where E_a is the activation energy being 210 kJ mol^{-1} (taking 2500 K for the dimensional measure of the temperature drop between the top and bottom of the system), R the gas constant and T the absolute dimensional temperature (being 300 K at the surface here). The viscosity contrast can reach 8 orders of magnitude variations at a given depth. We apply a cut-off for the viscosity, therefore viscosity is never higher than 10^{26} Pa s (taking a reference viscosity of $2\ 10^{22}$ Pa s that is contistent with our Rayleigh number).

Viscosity depends on stress as well, since we use pseudo-plasticity to produce self-consistently plate boundaries surrounding strong plate interiors. Finally, viscosity depends on the type of material tracked with markers: continental rafts are 100 times stiffer than ambient mantle, and a weak oceanic layer is 10 times less viscous than ambient mantle. The yielding parameters are also proportional to these viscosity ratios. Physical parameters are listed in Table 1.

We use three types of materials. Ambient mantle corresponds to the largest fraction of the spherical shell. We introduce continental rafts, which are 175 km thick. They are buoyant, their buoyancy number being −0.4 that corresponds to a density difference between continental and normal mantle of about −200 kg m^{-3}. We choose continents with Earth-like shapes, because any shape would be arbitrary anyway. We also impose at each time step a 15 km thick weak oceanic layer, which represents hydrated lithosphere. This layer is as dense as ambient mantle, but its weak rheological behaviour favours asymmetric to one-sided subduction as explained in Section 3. When such material reaches 300 km, it is transformed back into ambient mantle, as if dehydration is complete. One has to remember that the weak oceanic layer is close to be sub-scale. Although instantaneous flow solutions are stable with increased resolution, it is clear that improving resolution thanks to increasing computational power would improve the numerical treatment of the impact of the weak oceanic layer on the flow.

The system we solve for is heated mostly from within, but also from the core up to 25 % of the total surface heat flux. Both the surface and the bottom are isothermal, defining the temperature drop ΔT for the Rayleigh number (non-dimensional number measuring the vigor of convection) Ra of 5.910^6, based on the volume averaged viscosity $\bar{\eta}$:

$$\mathrm{Ra} = \frac{\rho_0 g \alpha \Delta T h^3}{\kappa \bar{\eta}},$$

ρ_0 being the reference density, g the gravitational acceleration, α the thermal expansivity, h the depth of the mantle, and κ the thermal diffusivity.

The average surface velocity obtained with these parameters, scaled with a thermal diffusivity of 10^{-6} m^2 s^{-1} is 1.2 cm $year^{-1}$. This is a factor of 3 lower than the Earth today, so the Rayleigh number used here is about a factor of 5 lower than expected (the velocity being proportional to $\mathrm{Ra}^{2/3}$). In the following, for practical comparison, the dimensional time scale we use is scaled to the Earth in a typical way. Hence, in the following, velocities are velocities in the model multiplied by 3, and consequently time corresponds to the time computed from the model divided by 3.

Table 1
Non-dimensional and dimensional parameters of the convection model, taking the following values for dimensionalisation producing the computed Rayleigh number: reference density is 4400 kg m^{-3}, thermal expansivity is 4.5 10^{-5} K^{-1}, thermal diffusivity is 10^{-6} m^2 s^{-1}, thermal conductivity is 4 W m^{-1} K^{-1}, reference viscosity is 2 10^{22} Pa s.

Parameter	Non-dimensional value	Dimensional value
Rayleigh number	5.910^6	
Heat production rate	20	5.810^{-12} W kg^{-1}
Top temperature	0.12	300 K
Basal temperature	1.12	2800 K
Viscosity jump factor at 660 km	30	
Activation energy	8	210 kJ mol^{-1}
Yield stress at the surface	210^4	47 MPa
Yield stress depth derivative	2.5 10^5	205 Pa m^{-1}
Continent viscosity and yielding factor	100	
Weak layer viscosity and yielding factor	0.1	
Maximum viscosity cutoff	10^4	210^{26} Pa s
Buoyancy number for continents	−0.4	

Starting from initial conditions obtained for a simpler model, we compute a dynamic evolution to reach the statistical steady-state, in two steps to minimise computing time. The first step uses a lower resolution (45 km) until a statistical steady state is reached. The second is to compute 500 My of evolution at high-resolution. We perform these two steps for several models varying the yield stress of mantle material such that we find the regime of plate-like behaviour with sufficiently large plates. The model presented here uses a non-dimensional yield stress at the surface of 210^4, which is equivalent to 47 MPa, and to an integrated value of 67 MPa over the 200 km boundary layer. It is important to note that the convective vigor in our model is lower than for Earth. The yield stress value that we take to obtain plate-like behaviour is relevant to our convective flow, for which stresses are lower than on our planet. Therefore, models with Earth-like convective vigor would require a higher value of the yield stress to obtain the same tectonic regime. Because of the velocity scaling mentioned above, we expect that the integrated value of the lithosphere should be 200 MPa, 3 times higher than the value we use here. For the same rheological parameters, several tectonic regimes can potentially be stable (Weller and Lenardic, 2012). Here, because the calculation is intensive, we could not investigate whether several tectonic regimes are possible. After the second step we obtain the initial condition for the calculation presented below.

4.2.2. Convection

The numerical solution is obtained for 250 My of evolution. We obtain a multi scale convective flow with hot plumes rising up and cold sheet-like downwellings. The temperature field is large-scale since degrees 1–4 are dominating as seen on Fig. 14. As discussed in previous work with pseudo-plasticity (Mallard et al., 2016; Van Heck and Tackley, 2008, Foley and Becker, 2009), a strong lithosphere enforces large-scale flow even without the presence of continental rafts or viscosity jump within the mantle. A large-scale flow is consistent with the spatial properties of seismic velocity anomalies in the Earth's mantle. A spectral analysis of tomographic models leads to the dominance of degrees 1–4 as well (Becker and Boschi, 2002).

However, this model also presents small-scale convection. At the base of the thermal boundary layer, below continental rafts and oceanic basins, plume-like instabilities sink before being thermally equilibrated before they reach the more viscous lower mantle (see Fig. 15). These cold plumes are less than 100 km in radius and are spaced by about 500 km. Such geometry is consistent with those obtained in 3-D calculations below plates with prescribed kinematics (Ballmer et al., 2007; van Hunen et al., 2005; Marquart, 2001). Small-scale convection below oceanic basins in the model onsets at variable ages, between 30 to 50 My. The instabilities form interconnected networks mostly parallel to the direction of the flow, with some variability we will discuss elsewhere. Small-scale convection is

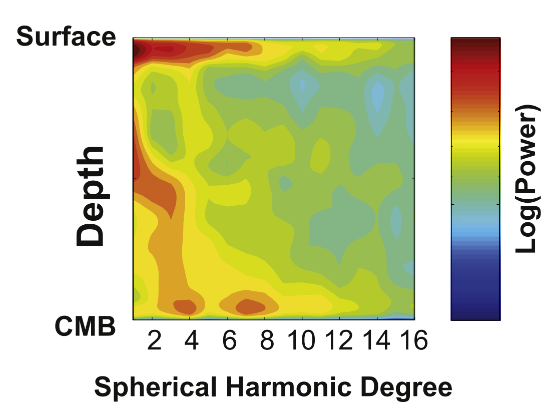

Fig. 14. Spectral heterogeneity map of the temperature field for the snapshot at 100My. All the snapshot have similar spectral heterogeneity maps.

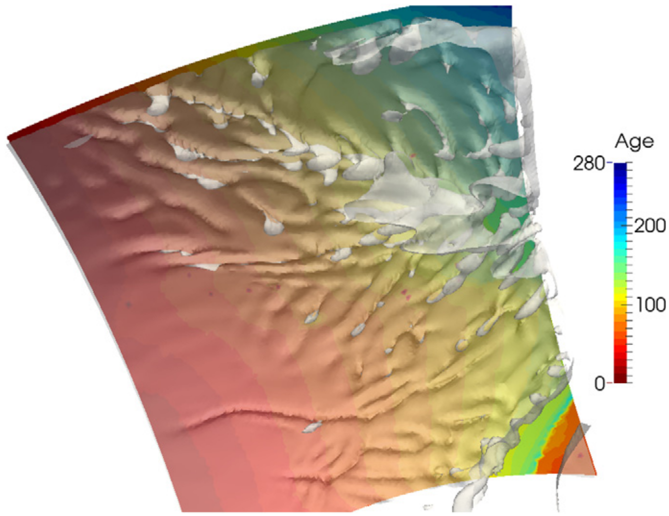

Fig. 15. Close up on the contour of the isotherm 0.75 representing the base of the upper boundary layer for the 100My step. The view is from the center of the model such that we can observe from below the small scale instabilities sinking. The age of the seafloor, in My, is seen throughout the transparent isotherm. The right side of the figure corresponds to a subducting slab we cut to observe better the small scale convection.

advocated to explain gravity anomalies on the seafloor of the Earth at length scales similar to smaller to those in the present numerical solution (Haxby and Weissel, 1986; Marquart et al., 1999). These dynamical perturbations could explain the flattening of lithosphere and higher heat flow than expected from a half-space cooling model (Parsons and McKenzie, 1978).

Hot plumes rising from the bottom boundary layer cross the whole shell without being swept away, and impact the base of the boundary layer, spreading mostly in the direction of the horizontal flow at the surface (see Fig. 16). There are constantly between 10 and 20 active plumes throughout the evolution computed here. Plumes are located below continents, oceanic basins, mostly intraplate but sometimes interact at ridges like West of South America in Fig. 16. Because they are hot and hence extremely less viscous than the surrounding, they rise 10 times faster than surrounding velocities. In the upper mantle, hot plumes rise at velocities between 30 to 120 cm year^{-1}, which is between 8 to 35 times faster than the rms surface velocity in the model. The persistence of plumes is variable here, since some exist for 20 My while others do for almost 200 My. Most of them persist for about 100 My.

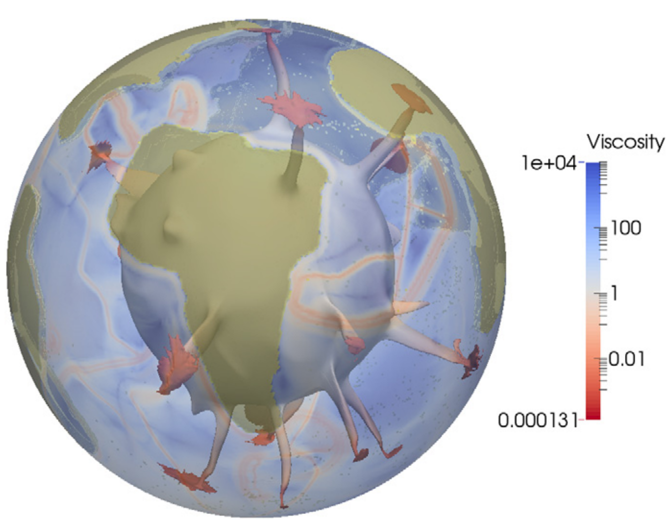

Fig. 16. Contour of the isotherm 0.9 to highlight plumes rising from the CMB for the 100 My step. A transparent surface layer shows the dimensionless viscosity field and hence divergent and convergent boundaries. Continental rafts are coloured in yellow.

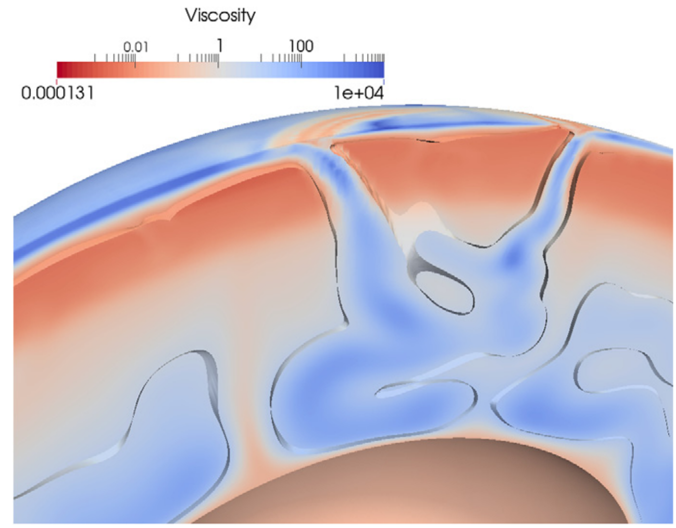

Fig. 17. Cross section through the two trenches south of South America for the 100My step. The dimensionless viscosity shows stiff slabs and zones of high stresses in the top boundary layer. The isotherm 0.7 is contoured.

The cold downwellings are the dominating features of the flow in the calculation. They sink throughout the whole shell as sheet-like structures, that fold and buckle in the more viscous lower mantle. Although in motion, they are long-lived features since we can track some of them through the 250 My of the calculation. They display a diversity of slab shapes. One-sided subduction exists in the two following cases: when a slabs sink at the ocean-continent boundary, or when a slab is rolling back fast enough as in Fig. 17. In this calculation, it seems that only these two cases generate asymmetry. Downwellings sinking from oceanic domains are often two-sided here in the absence of slab roll-back or fast trench motion. We also observe the presence of flat slabs around the Antarctic continental raft (Fig. 18).

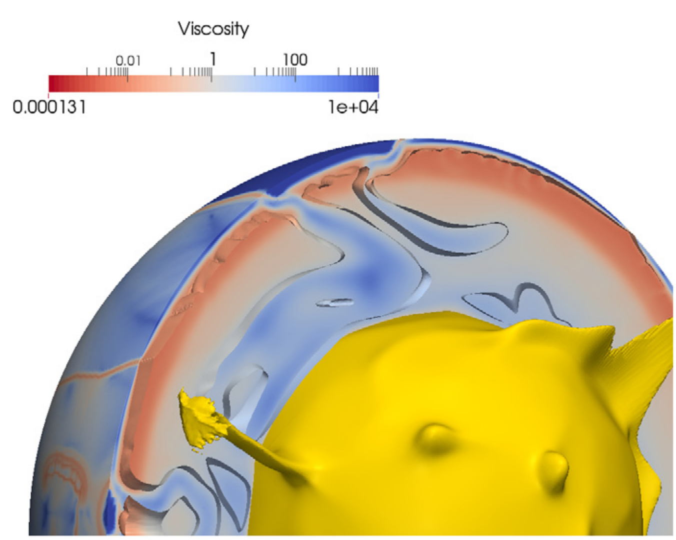

Fig. 18. Cross section through Antarctica for the 100My step. The dimensionless viscosity shows two flat slabs and zones of high stresses in the top boundary layer. The continent is represented as a very viscous raft in between the slabs here. The isotherm 0.7 is contoured.

4.2.3. Tectonics

The tectonic evolution produced in this model emerges self-consistently from the set of equations, and without prescribing any kinematic conditions or plate boundary at the surface. In the top boundary layer, low viscosity regions arise where yielding occurs. Continental lithosphere is more viscous than the mantle here, therefore internal deformation of continents is limited, especially when comparing to the Earth. Hereafter, we describe the calculation in the reference frame of the lithosphere (no-net-rotation of the surface). Continents drift relatively slowly at about 1 cm year^{-1}. As depicted in Fig. 19, Antarctica and Australia are aggregated from the start, drift South for 150 My before remaining stable. North America is moving South and slightly rotating clockwise, colliding Antarctica and Australia in the last snapshot. Africa, India and Eurasia are collectively moving North. South America is rotating counter clockwise, at a faster rate for the last 100 My.

Deformation is located in the ocean basins, in narrow zones of shearing where the viscosity drops because the plastic limit is reached. Convergent shear zones are located above downwellings. They exist within ocean basins and at some continent-ocean boundaries. The dominating downwelling ring in this calculation extends between Eurasia, North America and the aggregated Australia and Antarctica. After the first 100 My, subduction stops on the West side of North America, Australia and Antarctica, when intra-oceanic subduction close to the East of Eurasia progresses North and a new subduction zone develops in the middle of the basin where a ridge system existed before. East and South of South America, two subduction systems evolve throughout the whole time of the calculation. The highest velocities here are always reached within domains in contact with subduction zones. Trenches in the model are very often curved in the horizontal plane. The longest are the less curved ones. The fastest trench retreat rates are observed for short and curved trenches, as on Earth (Schellart et al., 2007). The trench at the South of South America starts as a circular shape before breaking up in one trench moving East and the other moving West. The one moving West gets anchored to the continent and the other splits again in one moving North towards the bigger one lying between Africa and South America, and the other is moving South. At the end they all interact to form a long subduction zone lying between Africa, South America and the Southeastern ocean.

Divergent shear zones are narrow and exist in two different ways. Long ridge-like systems are present in the center of wide oceanic basins, and back-arc ridge systems connected to trenches, in response to the stresses exerted by subduction (Mallard et al., 2016). The ridge system accommodating the subduction ring is first made of three ridges connected by a triple junction. The Eastward retreat on the West side towards the North stops the spreading of the Northern segment after 50 My. Between 150 My and 200 My, the trench moves close enough to the ridge system, stopping its activity. It generates a new subduction system oriented North-South. The spreading systems connecting Eurasia to South America is far from most trenches and remains stable throughout the calculation.

Transform shear zones develop in this calculation, connecting either segments of ridges, or ridges and trenches. They are not as sharp as on Earth, as explained in the previous section. Transform motion here is very often associated to divergent motion. Shear zones are about 150 km wide so transform motion between close ridge segments cannot exist. However, 1000-km transform shear zones exist. Taking the snapshot at 100 My, a transform shear zone lies West of South America and South of Eurasia at −50° longitude (see Figs. 19 and 20). Transform shear zones exist at smaller scale connecting the two curved trenches close to the tip of South America (velocities are parallel along the shear boundary but have different magnitudes). They are usually more diffuse than ridges and trenches in this model, probably because strain-rate softening and other softening mechanism is not used here.

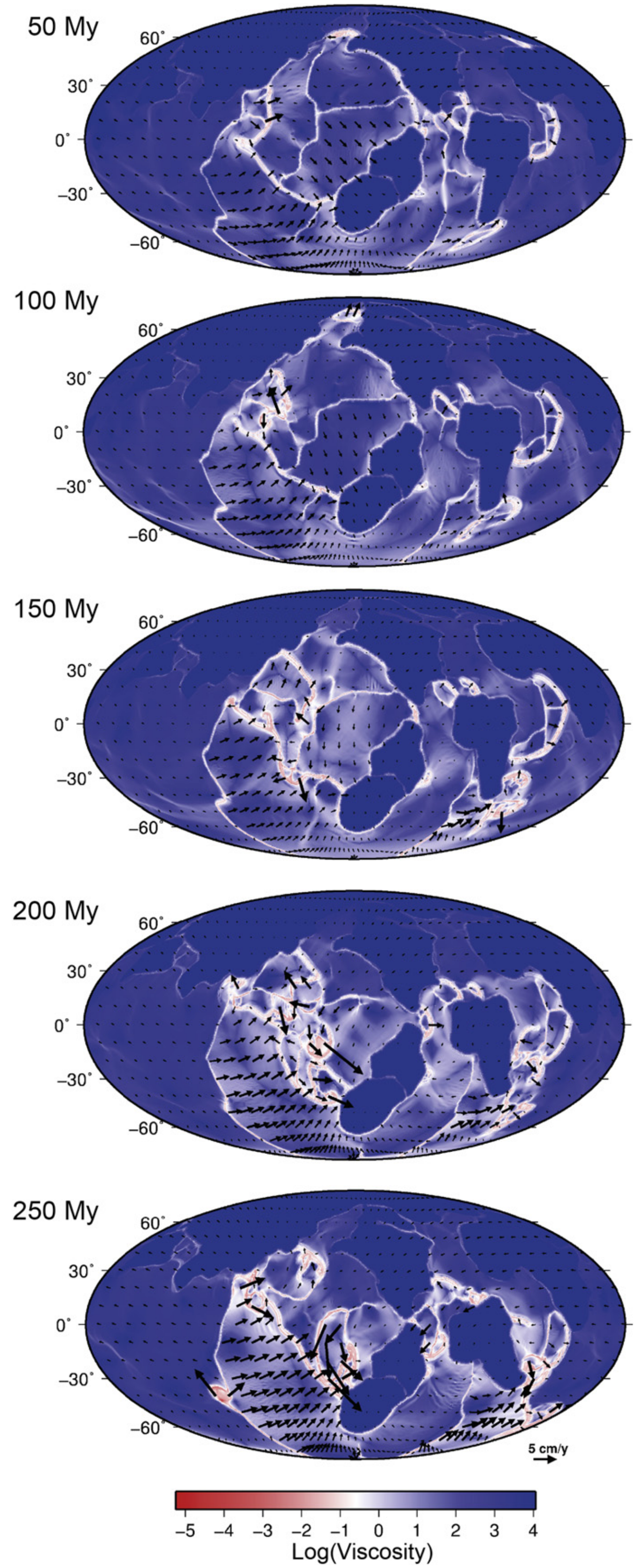

Fig. 19. Maps of the dimensionless viscosity and kinematics of the surface of the model for different times. Low viscosity represent boundaries between more viscous material.

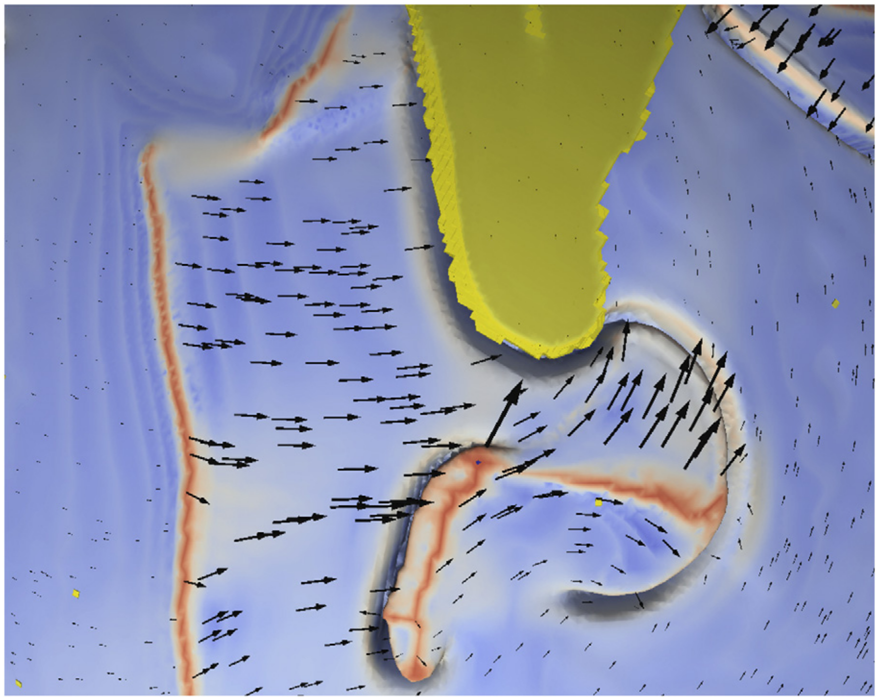

Fig. 20. Contour of the isotherm 0.4 for the 100My step, coloured with the dimensionless viscosity as in Fig. 17 and Fig. 18. The section is seen from the top south of South America, with associated kinematics. On the north-west side, a diffuse region localises transform motion. A fine low viscosity region connects the north tip of both trenches. This boundary presents transform character as well.

Although the calculation solves for continuous fields, the low viscosity shear zones are localised and delimit stiff plates. Diffuse boundaries also exist such that in a few places the velocity field changes without localised deformation (South of Antarctica). In this model, Africa and Eurasia are part of a very large plate comprising a large oceanic area as well. Large plates also exist at the South pole. In the middle of the subduction ring medium large plates develop, eventually fragmenting during the tectonic reorganisation. The fastest moving plates are often the smaller ones, which are in back-arc systems. The plate East of Australia and Antarctica experience a progressive shift in plate velocities over the first 150 My, while the plate East of Antarctic accelerates but its motion remains in the same direction.

In the snapshot at 250 My of Fig. 19, a peculiar tectonic regime happens at 45° longitude below Eurasia: a strong plume impacts the ridge and the transform shear zone, producing a strong temperature contrast with its surrounding. This temperature contrast is enough to initiate subduction at the border of the plume as showed by Fig. 21. This phenomenon was described by Gerya et al. (2015) as plume tectonics. The onset and first phase of retreat of the instability is very fast, involving fast velocities at the surface and the formation of small back-arc ridge systems.

5. Can convection models transcend the limitations of plate tectonic theory?

In Section 2, we show that geodynamic studies are successful at explaining critical tectonic expressions of deep mantle flow: surface kinematics, trench motions, lithospheric stress distribution and deformation, and dynamic topography. However, models in these works define the organisation and structure of either the lithosphere or the mantle a priori. For simplicity, both mantle and lithosphere structures tend to be imposed, and therefore these studies focus on tectonic features at present day. Such models miss fundamental information of the physics at play, and are at risk of discussing the possibility of a process or a situation that would never spontaneously emerge in a self-organised system. Therefore, a substantial amount of work is dedicated to building models of self-organisation of the lithosphere-mantle system (see Section 3). Assuming that plate-like behaviour arises naturally if the mechanics of peridotite is correctly taken into account, recent developments focus on introducing rheologies that generate plate like-behaviour self-consistently. The first milestone has been reached: plate-like behaviour does emerge with the use of simplified plastic behaviour, and produces first-order properties of the plate-mantle system we observe. The second milestone may be reached as well, but it is poorly documented (see Section 4): convection models with plate-like behaviour produce tectonic features relevant to study the physics behind plate tectonics on our planet. However, the community needs more convincing results, because of the simplicity of the rheology used and its relative gap with experimental values and field based studies. The pseudo-plastic rheology is often used in its simplest form which is instantaneous, without history dependence. Several authors challenge this view in convection modeling, proposing plate boundaries arise from regions of already high integrated strain (Bercovici and Ricard, 2003; Zhong and Gurnis, 1993). Many examples of reactivation of plate boundaries show tectonic inheritance is fundamental for continental rifting (for instance Wilson, 1966). However, the possible reactivation of a region of weakness depends on the orientation of the stress field relative to the mechanical heterogeneity at the lithospheric scale and microscale (Tommasi and Vauchez, 2001). In oceanic domains, the importance of tectonic inheritance is questioned since there is little evidence for it: the location and jumps of ridges, especially in the Pacific basin, are difficult to connect to pre-existing structures. Intra-oceanic trenches cannot be identified as pre-existing structures as well. Although tectonic inheritance for the initiation of subduction has been proposed in the past (Mueller and Phillips, 1991), evidence is still lacking and convection models with plate-like behaviour produce realistic plate tessellation, continental drift and seafloor spreading without history-dependent rheology in oceanic area. Therefore, more complexities in the physics are required to refine the role of pre-existing deformation on the

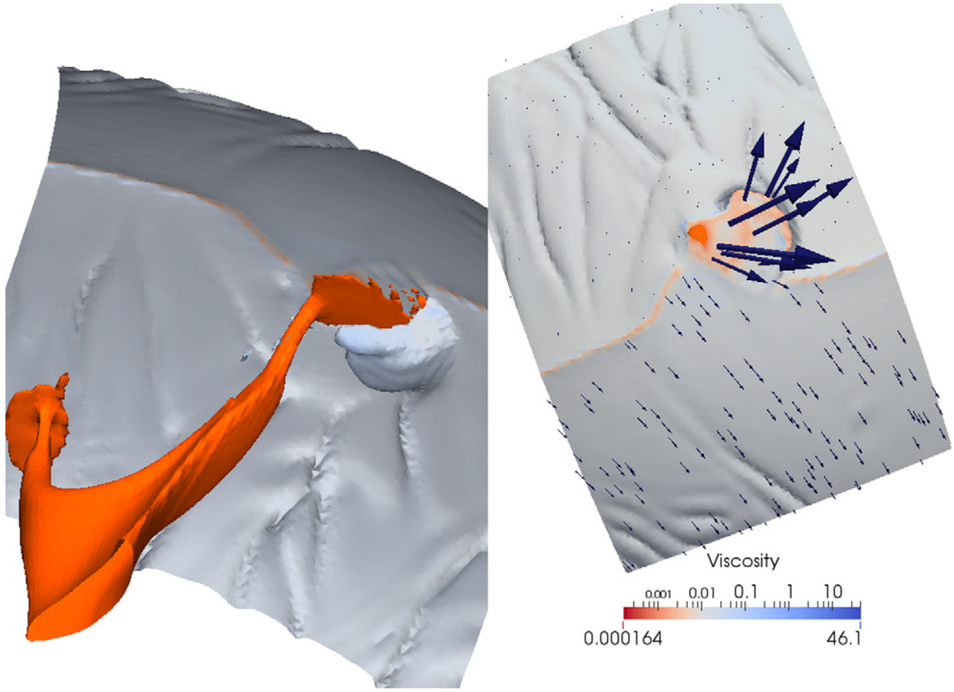

Fig. 21. Two views of the plume tectonic event at 240My. Contours of the isotherms 0.7 and 0.9 are shown. On the left side the view from below shows a starting subduction. On the right, a map view from the top with surface kinematics offers a complementary description of the snapshot.

organisation of tectonic plates, the wager being on macroscopic theories of grain size dynamics in polycrystalline aggregates. The assumption here is that microscale physics the key to modeling self-organisation of large scale mantle dynamics. The community is making progress in terms of theoretical developments and numerical methodologies in that regard (see Section 3). The modeling approaches should be confronted with experiments and field work with the caution expressed in Section 3.

Before we reach that stage, recent convection models express several key ingredients of the self-organisation of the mantle-lithosphere system. The basics of continental drift and seafloor spreading are understood, even the processes behind the organisation of the plate layout are beginning to emerge (see Section 4). The numerical results we present here show fundamental features of plate tectonics on Earth although the model requires improvement. Such models have the ability and resolution to study tectonics across the scales: regional tectonics and dynamics (small-scale convection, plume tectonics, slab dynamics, triple junction evolution) interact with larger scale processes. As long as computationally limitations exist, there is a scale limit below which regional models are more adapted and relevant.

Obtaining mantle convection models that are predictive enough (in the sense that a model does not have to be perfect but accurate enough for determined predictions) could be in sight. We could then ask the question of this concluding section: can convection theory transcend the limitations of plate tectonics? Indeed, mantle convection theory seems more complete than plate tectonic theory regarding certain issues: mantle convection is a dynamic theory (a state at a given time can in principle be predicted from knowing a state at another time), whereas plate tectonics is not because it is based on geometry, not forces; mantle convection can account for intraplate deformation and mountain building, whereas plate tectonics cannot because it assumes perfectly rigid plates; mantle convection links the surface to the deep mantle, whereas plate tectonics does not. Transcending plate tectonics would mean that convection models could produce tectonic histories for the Earth, with a higher degree of understanding of the physics and with a high level of accuracy. Hence it would necessary come out of inverse approaches, which find the best compromise between selected datasets and models. Since some geological observations provide informations at specific times in the past, the natural methodology would be data assimilation, such as what is done in oceanography (Talagrand, 1997, for an introduction) or geomagnetism (Fournier et al., 2010, for a review), where surface observations at different times combined with dynamic models provide estimates of ocean flow or outer core motions. A first attempt seems promising (Bocher et al., 2016). However, this direction in geodynamics requires an objective assessment of which datasets would be useful for the inversion, and which ones would be suitable to evaluate the quality of the inversion. Also, a lot of datasets that could be used for inversions are already models themselves, such as seismic tomography for instance. It requires a careful appraisal of the uncertainties, because the conditions for a compromise between a model and data is that (1) both are imperfect hence 'negotiation is possible' and (2) both contain enough information about the state of the system. Discussing these points is beyond the scope of this manuscript, but we emphasise the fact that bridging the gap between tectonics and mantle convection could be fundamental for making a leap forward in our understanding of the processes that control global tectonics. Another benefit would be to bring together observational and modeling communities for a common goal: jointly reconstructing tectonics and convection in the past.

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