Tectonophysics 512 (2011) 47–59

Contents lists available at SciVerse ScienceDirect

Tectonophysics

journal homepage: www.elsevier.com/locate/tecto

State of stress and age offsets at oceanic fracture zones and implications for the initiation of subduction

Mignon D. Johnston [a], Maureen D. Long [b,*], Paul G. Silver [c,1]

[a] Department of Geology, Mount Holyoke College, 50 College St., South Hadley, MA 01075, United States
[b] Department of Geology and Geophysics, Yale University, PO Box 208109, New Haven, CT 06410, United States
[c] Department of Terrestrial Magnetism, Carnegie Institution of Washington, 5241 Broad Branch Road, NW, Washington, DC 20015, United States

A R T I C L E I N F O

Article history:
Received 16 December 2010
Received in revised form 15 September 2011
Accepted 28 September 2011
Available online 8 October 2011

Keywords:
Subduction initiation
Oceanic lithosphere
Oceanic fracture zones
Lithospheric stress

A B S T R A C T

The recycling of oceanic lithosphere back into the Earth's interior through subduction is a central component of plate tectonics. The process by which new subduction zones initiate, however, remains poorly understood. Several different mechanisms for subduction initiation have been proposed, including passive margin collapse (aided by sediment loading and/or rheological weakening due to the presence of volatiles) and forced convergence across a zone of preexisting lithospheric weakness. In this paper we focus on the latter type of model, which identifies three conditions necessary for subduction initiation: a zone of weakness such as a fracture zone, an age (and therefore density) offset along the fracture zone, and significant normal compressive stress which leads to shortening. We identify regions on the present-day Earth which meet these conditions and which may correspond to regions of relatively likely subduction initiation in the near future. Using a digital seafloor age model, we have created a database of oceanic fractures and quantified the associated age offsets. We have evaluated the state of stress on these lithospheric weak zones using two different global stress models. We find that the conditions needed to initiate subduction via the forced convergence model are relatively rare on the present-day Earth, and that there is little indication of incipient subduction at regions identified as relatively likely for subduction initiation. Using the same technique, we have evaluated the state of stress and seafloor age offset at regions of inferred present-day incipient subduction, and find that most of these regions are not associated with both high far-field compressive stresses and large age (and thus density) offsets. Subduction has likely initiated via forced convergence across preexisting zones of lithospheric weakness in the past, but our results indicate that the conditions needed for this type of subduction initiation are rare on the present-day Earth.

1. Introduction

Subduction, the sinking of oceanic plates into the mantle, is a central component of plate tectonic theory (e.g., Stern, 2002); it drives plate motion and allows for the formation of mid-ocean ridges. Once subduction begins, it is self-sustaining as the negative buoyancy of the older, colder, and denser subducting slab continues to pull the plate downward, thereby creating an asymmetric mantle downwelling (Gurnis et al., 2004). A self-sustaining subduction zone can also give rise to new subduction zones through transference or polarity reversal (Stern, 2004). The process through which new subduction zones spontaneously initiate, however, remains poorly understood. In order for oceanic lithosphere to begin to subduct, the considerable strength of the plate must be overcome by breaking and/or bending it (McKenzie, 1977), perhaps taking advantage of preexisting rheologically weak zones. The process through which this occurs remains uncertain, although many models have been suggested (e.g., Baes et al., 2011; Burov and Cloetingh, 2010; Cloetingh et al., 1984; Kemp and Stevenson, 1996; Mueller and Phillips, 1991; Regenauer-Lieb et al., 2001; Solomatov, 2004; Toth and Gurnis, 1998; Ueda et al., 2008). Yet, at some point, subduction zones must have initiated on an Earth without active plate tectonics (e.g., Hansen, 2007), and the initiation of new subduction zones has been a common occurrence throughout Earth's history; for example, ten new subduction zones of various lengths have initiated in the Pacific domain during the Cretaceous (Gurnis et al., 2004). To highlight one particular example, there is strong evidence indicating that the Izu–Bonin–Mariana system is a subduction zone that initiated (during the Eocene) independently of any existing subduction zone (Gurnis et al., 2004; Stern, 2004). The process through which subduction initiates, therefore, remains one of the important outstanding questions in plate tectonic theory.

Most studies of the subduction initiation process have focused on using either analog experiments (e.g., Goren et al., 2008; Leroy et

* Corresponding author at: Department of Geology and Geophysics, Yale University, PO Box 208109, New Haven, CT 06520, United States. Tel.: +1 203 432 5031; fax: +1 203 432 3134.
E-mail address: maureen.long@yale.edu (M.D. Long).
[1] Deceased.

0040-1951/$ – see front matter © 2011 Elsevier B.V. All rights reserved.
doi:10.1016/j.tecto.2011.09.017

al., 2004; Mart et al., 2005) or numerical modeling techniques (e.g., Gurnis et al., 2004; Hall et al., 2003; Nikolaeva et al., 2011; Toth and Gurnis, 1998) to model the formation of subduction zones, beginning with carefully defined initial and boundary conditions, to determine the combination of conditions that produces self-sustaining subduction. These studies generally seek to model physical processes that include the collapse of passive continental margins, perhaps with the aid of sediment loading or rheological weakening of the lithosphere, or the nucleation of a new subduction zone at a preexisting zone of lithospheric weakness. One example of the latter type of model includes the work of Gurnis et al. (2004), Hall et al. (2003), and Toth and Gurnis (1998), who invoke forced compression across a weak zone with a preexisting density difference, such as might be expected for an oceanic fracture zone. Gurnis et al. (2004) demonstrated that self-sustaining subduction could, indeed, be produced by such a mechanism, at least in the context of a two-dimensional numerical model. It remains unclear, however, how often the conditions needed to initiate subduction in this way (a preexisting zone of weakness, a density difference across that zone, and significant compression and shortening) actually occur on the Earth.

Here, we explicitly test the predictions of the Gurnis et al. (2004) model to determine where on the present-day Earth subduction might soon initiate (or might be currently initiating) by this process. We identified oceanic fracture zones as zones of preexisting weakness and calculate the difference between the lithospheric ages on either side of the fracture zone using a digital seafloor age model. We used two global stress models (Ghosh, 2008; Lithgow-Bertelloni and Guynn, 2004) in our calculations of the magnitude of normal compressive stress on each fracture. From these calculations, we identified locations with significant age offsets and relatively high compressive stress where subduction initiation may be relatively likely on the present-day Earth. We then examined global earthquake catalogs for evidence of anomalous seismicity in these regions that might be consistent with incipient subduction, as intraplate seismicity can indicate shortening and, perhaps, subduction initiation (e.g., Meuller and Phillips, 1991; Okal et al., 1986). Finally, we repeated our age offset and stress calculations for five regions which have been identified as likely sites of present-day incipient subduction (the Gorringe Bank, the Hjort Trench, the Mussau Trench, the Owen Ridge, and a proto-trench stretching from western Samoa to the Caroline trench) to evaluate whether subduction is likely initiating in these regions via the process modeled by Gurnis et al. (2004) or by some other mechanism.

2. Subduction initiation: models and observations

Since the development of plate tectonic theory, many different models for the initiation of subduction have been proposed and debated. Cloetingh et al. (1984) hypothesized that subduction initiates by passive margin collapse due to sediment loading, provided that the oceanic crust is young enough (since strength increases with age in oceanic lithosphere). The concept of passive margin collapse is consistent with the conceptual framework of the Wilson cycle of supercontinent assembly and dispersal and the opening and closing of ocean basins. Subsequent numerical modeling work demonstrated that additional external tectonic forces are needed to overcome the strength of the lithosphere and nucleate subduction, leading to the suggestion that existing weak zones located within plates are more potentially suitable for incipient subduction (Cloetingh et al., 1989). However, it has since been suggested that passive margin collapse may remain a viable mechanism for subduction initiation (e.g., Nikolaeva et al., 2010, 2011) if the lithosphere is significantly weakened, perhaps by hydration (e.g., Regenauer-Lieb et al., 2001). In particular, it has been suggested that the eastern US passive margin may be currently undergoing hydration due to the dewatering of the Farallon plate at depth (van der Lee et al., 2008) that may facilitate future subduction initiation.

The idea that subduction may initiate at preexisting weak zones has been studied by several different workers. Korenaga (2007) suggested that thermal cracking and subsequent serpentinization might produce localized weakening of the oceanic lithosphere and facilitate spontaneous subduction initiation. Modeling work by Hall et al. (2003) demonstrated that a fracture zone could progress into a self-sustaining subduction zone, provided that fault zone stresses are below ~20 MPa, that the net convergence rate is sufficiently fast to prevent excessive warming of the slab, that the fault zone is weak, that the coefficient of friction is low, and that the plate is young enough to allow for decoupling. Several studies have investigated how lateral density differences in the crust can facilitate subduction initiation (Goren et al., 2008; Lebrun et al., 2003; Mart et al., 2005; Stern, 2004). The lateral density difference can be at the boundary between continental and oceanic crust (Goren et al., 2008; Mart et al., 2005) or at a fracture zone or transform fault separating oceanic lithosphere of different ages (Lebrun et al., 2003; Stern, 2004) and may reflect compositional/petrological differences in the oceanic crust (Niu et al., 2003). Forced convergence across a weak zone may be provided by far-field tectonic stresses (e.g., Gurnis et al., 2004) or by mantle upwelling (e.g., Hynes, 2005; Ueda et al., 2008), and the subsequent transition from forced to self-sustaining subduction may be aided by non-Newtonian mantle rheology (Billen and Hirth, 2005).

In addition to modeling studies that attempt to re-create the subduction initiation process in a numerical or analog context, there have been many studies that have sought to characterize present-day incipient subduction through geological or geophysical indicators. For example, Meckel et al. (2003) used marine swath bathymetry/reflectivity, seismic reflection, gravity, magnetic, and teleseismic data to determine that self-sustaining subduction is not yet occurring at the Hjort Trench (Australian-Pacific plate boundary). However, anomalously deep earthquakes are present, indicating lithospheric deformation and underthrusting, which suggest subduction initiation (Meckel et al., 2003). Similarly, in studying the morphostructure of the Puysegur Ridge and Trench, Collot et al. (1995) found that incipient subduction is characterized by crustal shortening, ridge formation, and the development of an incipient trench. In a study of this same area, trench parallel normal faults and reactivated fracture zones were found to be characteristic of the region surrounding the incipient subduction zone (Lebrun et al., 2003), consistent with subduction initiation at a preexisting zone of lithospheric weakness.

In this study, we focus on the "forced convergence" mechanism for intraoceanic subduction initiation explored by Gurnis et al. (2004), Hall et al. (2003), and Toth and Gurnis (1998) through two-dimensional numerical modeling. This scenario invokes a preexisting zone of weakness in the oceanic lithosphere, a contrast across the weak zone in lithospheric age (and therefore density) that is needed to produce an asymmetric downwelling and to force the denser, more negatively buoyant plate into the mantle, and a significant amount of compression and shortening in the direction perpendicular to the strike of the weak zone. Oceanic fracture zones are thought to be a prime candidate for the nucleation of incipient subduction, because there are lateral density gradients associated with plate age offsets (Gurnis et al., 2004; Stern and Bloomer, 1992) and serpentinization may significantly weaken the oceanic lithosphere (e.g., Hilairet et al., 2007; Saleeby, 1984). Gurnis et al. (2004) performed a suite of numerical models exploring forced compression of oceanic fracture zones (in addition to former spreading centers and homogeneous plates) and found that significant shortening (~100–150 km) across a fracture zone can result in a transition from forced to self-sustaining subduction over a period of a few million years.

3. Data and methods

Digital isochron data from the EarthByte database (Müller et al., 1997) were used to identify locations of fracture zones with associated age offsets in the oceanic crust. A digital map of the seafloor age (Fig. 1) was used to create a database of the endpoints of short, roughly linear segments of oceanic faults and fracture zones with an appreciable offset in lithospheric age. This approach produced a dataset of 1499 fault segments and their associated age offsets; each longer fault or fracture is made up of multiple fault segments. The complete dataset of 1499 sets of fault endpoints and age offsets can be found in the online supplementary information (see Appendix A).

In order to determine the magnitude of normal stresses on each fault segment in the database, we used two different global stress models, by Ghosh (2008) and Lithgow-Bertelloni and Guynn (2004) (Figs. 2 and 3). The Ghosh (2008) model (see also Ghosh et al., 2009) was developed by adding the deviatoric stress field associated with horizontal basal tractions arising from density-driven flow in the mantle computed from viscosity variations in the lithosphere to the deviatoric stresses generated by lateral variations in gravitational potential energy per unit area. Many different models were tested by Ghosh (2008) by varying the depth up to which the lateral viscosity contrasts occurred as well as the type of lateral viscosity variations in lithosphere and/or asthenosphere. We used the model that best matches the observed plate motions and predicts deviatoric stresses that best fit strain rates from the Global Strain Rate Map (GSRM). This model combines lateral strength variations due to continental cratons, old oceanic lithosphere, normal continental lithosphere, and normal oceanic lithosphere with a thick (~300 km) but relatively weak (10^{19} Pa·s) and laterally uniform asthenosphere (Ghosh, 2008).

The second stress model we used, model LVC+TD0 of Lithgow-Bertelloni and Guynn (2004), is derived from a finite element modeling scheme that calculates the lithospheric stresses induced by mantle flow, crustal heterogeneity, and topography. This model combines the stresses produced by mantle tractions and those produced by crustal contributions and was found to produce the best match to the world stress map (WSM) database (e.g., Heidbach et al., 2009) out of the suite of models developed by Lithgow-Bertelloni and Guynn (2004). Mantle density heterogeneity used to drive flow and calculate mantle tractions was derived by considering the history of subduction; the model included a lower mantle viscosity increase of an order of magnitude over the upper mantle and a low-viscosity channel between 100 and 200 km depth with a viscosity reduction of two orders of magnitude. Crustal contributions to the stress field were calculated taking into account topography, crustal thickness, and density structure (Lithgow-Bertelloni and Guynn, 2004).

Both Ghosh (2008) and Lithgow-Bertelloni and Guynn (2004) compared their lithospheric stress models to the observations of the stress field contained in the WSM, which should increase our confidence in the robustness of each model. Ghosh (2008) also compared their model predictions to observations of the strain rate tensor near plate boundaries contained in the Global Strain Rate Map (GSRM) of Kreemer et al. (2003). Because most of the stress measurements in the WSM (as well as in the GSRM) are confined to continents and mid-ocean ridges (due to the difficulty of measuring stress on the ocean floor), however, a comparison between models and observations has not been carried out in the regions of the model that we are utilizing in this study. A comparison of the two models (Figs. 2 and 3) indicates that the first-order features of the two models are generally similar, although they often differ in regional details and in a few regions even the first-order pattern is different.

For each stress model considered here, we calculated the magnitude of the normal stress on each fracture zone segment using the relationship $\sigma_N = \frac{\sigma_1+\sigma_2}{2} - \frac{\sigma_1-\sigma_2}{2}\cos 2\theta$, where σ_1 and σ_2 are the horizontal principal stresses and θ is the angle between the direction of maximum stress (σ_1) and the strike of the segment. In order to compare the models directly, we computed both an absolute and a normalized value of normal stress on each segment; for each of the two stress models, we identified the largest value of σ_N in the global fracture segment dataset and normalized the calculated stresses by this value. This produced a normalized normal stress value between −1 and 1 for each segment for each stress model (Fig. 4), with negative values indicating extensional stress and high positive values indicating relatively high compressive stress and, therefore, relatively higher likelihood of the eventual shortening needed to initiate subduction. The age offset along each fracture zone segment was calculated using values of seafloor age immediately adjacent to the fracture on either side (Fig. 5). Gurnis et al. (2004) used an age offset of 30 Myr for most of their models of forced convergence across a fracture zone; there are very few fracture zone segments on the present-day Earth with an age offset of greater than 30 Myr.

4. Results: oceanic fracture zones

The normal stress calculations shown in Fig. 4 indicate that several regions exist where there are relatively large compressive stresses oriented normally to zones of weakness in the oceanic lithosphere. The two stress models used tend to produce similar results for most regions, although there are some variations in the details of the

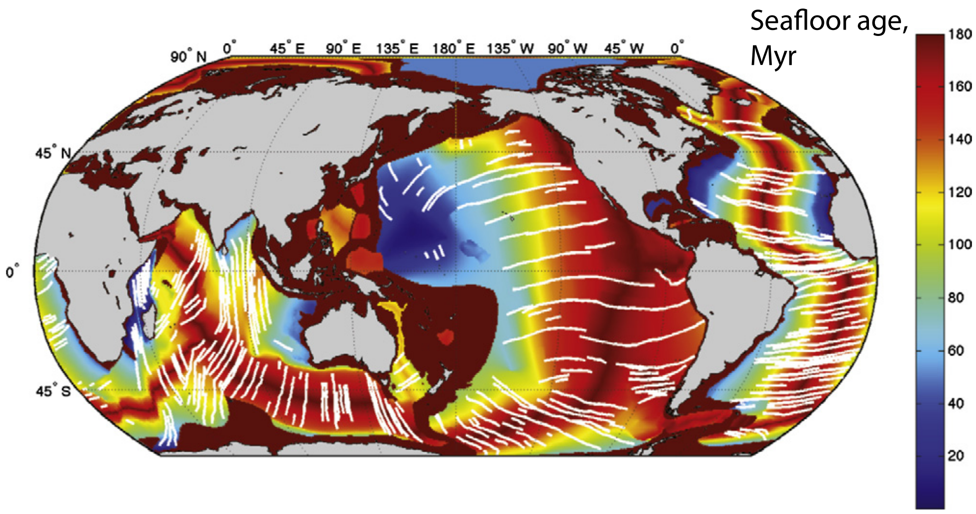


Fig. 1. Map showing digital seafloor age data from the EarthByte database (Müller et al., 1997). Young seafloor is shown in red, with colors progressing toward blue for older seafloor. Fracture zone segments with appreciable age offsets are shown as white lines; the endpoints of these segments along with their associated age offsets can be found in the online supplementary material.

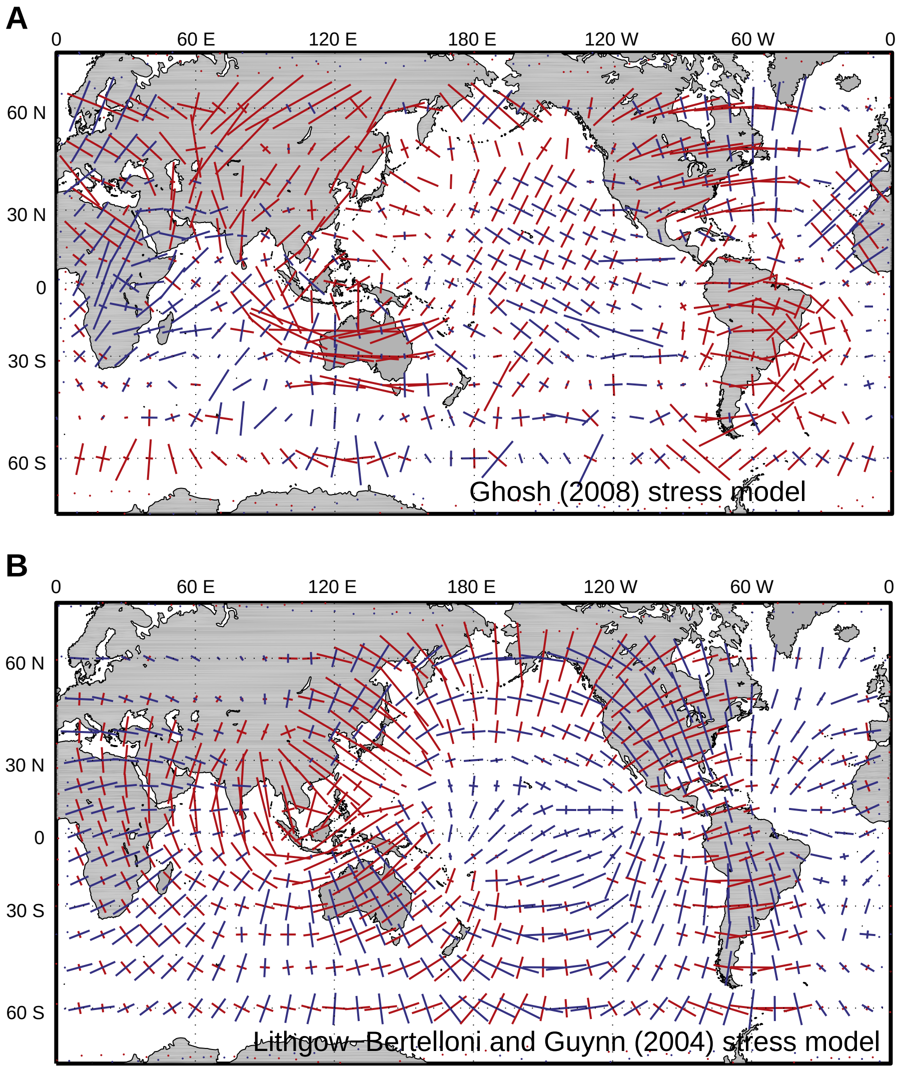


Fig. 2. Maps of the (a) Ghosh (2008) and (b) Lithgow-Bertelloni and Guynn (2004) stress models. The local principal horizontal stresses are plotted at 10° intervals as oriented bars; compressive stresses are plotted in red, while extensional stresses are plotted in blue. Note that the scales are different between the two models; bars are scaled to the maximum global stress value in each model (see Fig. 3).

calculated stress values. Both models indicate that fracture zone segments are currently subjected to relatively large compressive normal stresses in several regions, including parts of the northern Pacific Ocean, the eastern Indian Ocean, and the Southern Ocean between Australia and Antarctica (Fig. 4). Additionally, the Ghosh (2008) stress model predicts relatively high values of normal compressive stress for fault segments located off the eastern coast of South America, southwest of the southern tip of South America, and off the west coast of central Africa. The Lithgow-Bertelloni and Guynn (2004) stress model predicts relatively high compressive stresses for a region of the northern Atlantic around the mid-Atlantic Ridge and the western Indian Ocean.

The calculated age offsets show considerably more variation among adjacent fracture zone segments than the normal stress values (Fig. 5). However, we were able to identify ten regions associated with fracture zone segments that exhibit age offsets of 30 Myr or more, which is equal to the age offset value used in many of the Gurnis et al. (2004) models (see, e.g., Fig. 12 of Gurnis et al.). These regions include off the southern coast of Africa, the eastern Indian Ocean, the northern Pacific Ocean, the southern Pacific Ocean, off the southeastern and northeastern coasts of South America, off the southeastern coast of West Africa, off the coasts of Spain and Portugal, and in the northern Atlantic Ocean. We have qualitatively compared the normal stress maps in Fig. 4 and the age offset map in Fig. 5, to identify regions that are associated with both large age offsets and significant compressive normal stresses. Of the regions for which both stress models predict significant compressive normal stresses, the eastern Indian Ocean and the northern Pacific both exhibit large age offsets, but the region between Australia and Antarctica does not. Of the regions with the most striking age offsets, several are associated with generally modest normal stresses, such as the southern Pacific (Fig. 4), and others exhibit large compressive stresses for one stress model but not the other, such as the northern Atlantic.

Fig. 6A shows a map of the regions that we have identified as relatively likely sites for ongoing or (near) future subduction initiation. Of the sites shown on Fig. 6A, we show more detailed analysis in Fig. 6B–D for three regions. One of these, the eastern Indian Ocean, is perhaps the most likely site for subduction initiation via forced compression, as both stress models predict large compressive normal stresses and there are large age offsets. The northern Pacific is also highlighted, because large portions of the fractures here have large age offsets and both models predict compressive normal stresses, although for the Ghosh (2008) model these stresses are fairly modest. We also show a region of the south Central Pacific which has large age offsets. Although the compressive normal stress predicted by the Lithgow-Bertelloni and Guynn (2004) model is relatively modest and the resolved normal stress

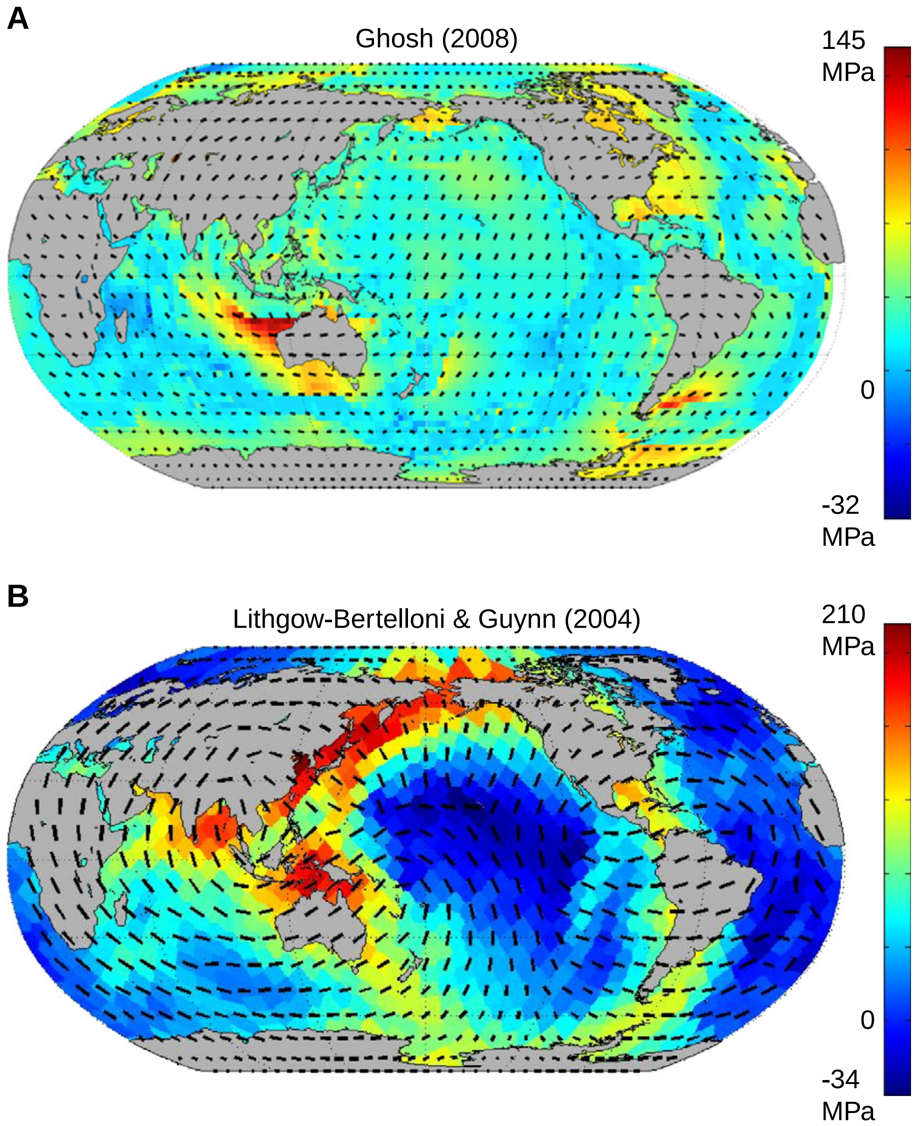


Fig. 3. Maps showing the maximum compressive horizontal stresses in the global stress models of (A) Ghosh (2008) and (B) Lithgow-Bertelloni and Guynn (2004). Background colors show the local value of maximum compressive stress, from most compressive (red) to least compressive (blue); in some cases the maximum compressive stress is actually extensional. Bars show the local orientation of the maximum compressive stress.

for the Ghosh (2008) model is actually extensional, there is significant intraplate seismicity in this region, as discussed below. Given the significant compression normal to the strike of the fracture zone the regions shown in Fig. 6 are predicted to undergo, they might, given enough time and sufficient weakening of the fracture interface (perhaps due to serpentinization), yield sufficient shortening to initiate subduction. All of these regions exhibit both a significant contrast in age (and therefore density) of the oceanic lithosphere as well as large compressive normal stresses for at least one stress model.

To evaluate further the possibility of ongoing (or near-future) subduction initiation in the most likely regions, we analyzed the seismicity in each region by searching the Global CMT and USGS earthquake catalogs for earthquake hypocenter locations and (where applicable) moment tensors, which are shown in Fig. 6. All three regions of relatively likely subduction initiation shown in Fig. 6B–D are associated with some intraplate seismicity, but it is unclear whether this seismicity might be associated with reverse faulting and shortening in the lithosphere, as there are very few constraints on earthquake mechanisms. In the Indian Ocean region, there are a few well-constrained focal mechanism solutions (Fig. 6E), but only one is clearly consistent with thrust faulting along a fault plane oriented N–S, as might be predicted if subduction is initiating along a N–S directed weak zone. We note, additionally, that the region that we identify as a zone of relatively likely future subduction initiation in the Indian Ocean is not co-located with the region of the central Indian Ocean lithosphere that has inferred to be undergoing buckling due to compression and shortening (e.g., Bull and Scrutton, 1992; Chamot-Rooke et al., 1993; Gerbault, 2000; Krishna et al., 2001). Of the three regions shown in Fig. 6, it is notable that the south central Pacific Ocean (Fig. 6C) is associated with very abundant intraplate seismicity and which may be undergoing significant intraplate deformation. However, the compressive normal stress predicted by the Lithgow-Bertelloni and Guynn (2004) model is quite modest and the normal stress predicted by the Ghosh (2008) model is actually extensional, so it is somewhat puzzling that this region is associated with such abundant seismicity.

5. Results: incipient subduction zones

The calculations described above identify several regions of relatively likely intraoceanic subduction initiation given the conditions

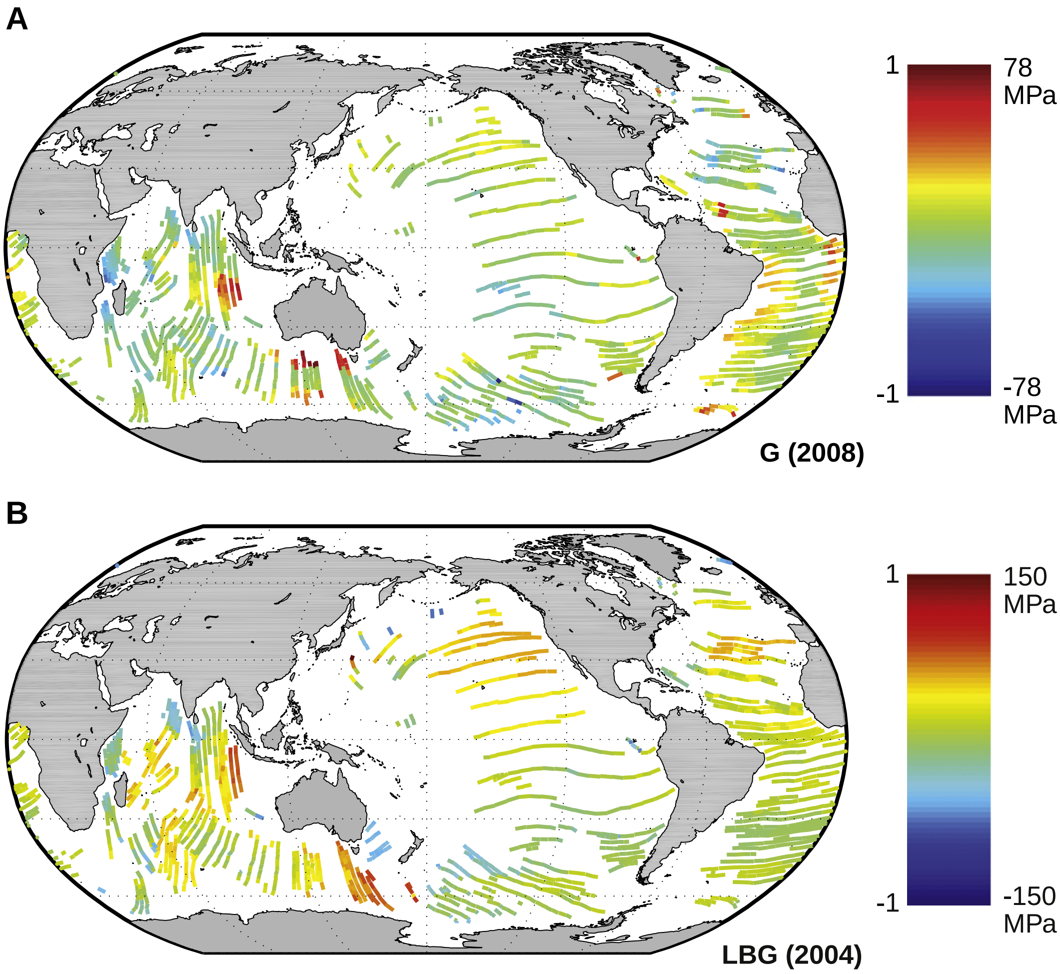


Fig. 4. Maps of fracture zone locations with the magnitude of the normal stress on each segment indicated by color, using the Ghosh (2008) (A) and Lithgow-Bertelloni and Guynn (2004) (B) stress models. The color scale goes from blue (most extensional) to red (most compressive) and represent both absolute and normalized (with values between −1 and 1) stress values (see color bar at right).

that are prescribed in numerical models of forced convergence across a fracture zone (Gurnis et al., 2004). It is instructive to compare these regions with regions that have been identified as incipient subduction zones on the present-day Earth. We have repeated our calculations of stress state and age offset for five inferred incipient subduction zones, namely the Gorringe Bank, the Hjort Trench, the Mussau Trench, the

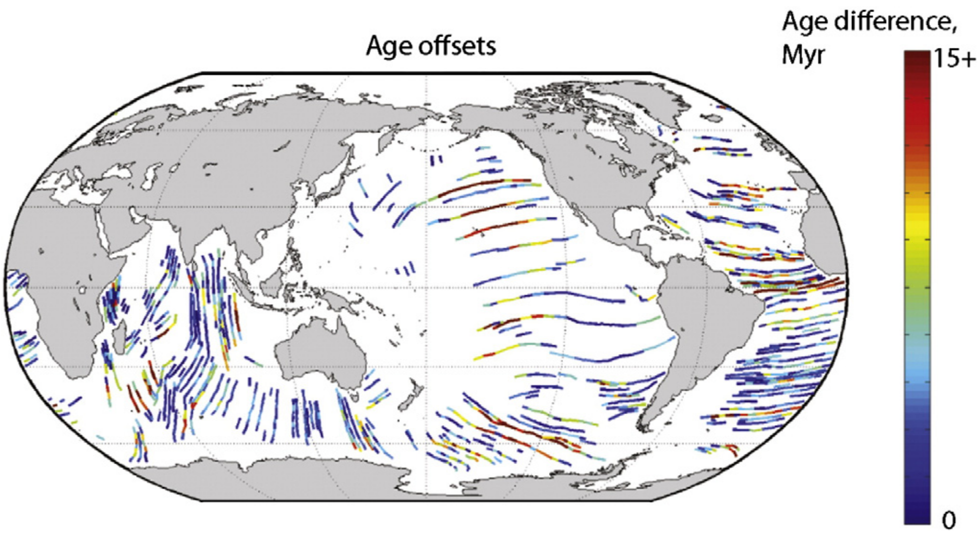


Fig. 5. Map of fracture zone locations with the magnitude of age offsets at each segment shown in color. The color scale saturates at an age offset of 15 Myr, so the color scale goes from blue (small age offset) to red (offset of 15 Myr or more).

Owen Ridge, and the Samoa–Caroline proto-trench (Fig. 7). Several of these are inferred to have nucleated off a fracture zone segment; Gurnis et al. (2004) classifies the Mussau Trench, Gorringe Bank, and Owen Ridge as "potential incipient boundaries," while the Hjort trench is inferred to be undergoing a transition from forced to self-sustaining subduction. The Gorringe Bank is seafloor morphological feature in the Atlantic Ocean off the coast of Portugal (e.g., Borges et al., 2001) and is associated with significant seismicity, but it is not parallel to the fracture zone segments identified in this study (Figs. 4 and 7); its origin remains controversial. The Hjort Trench is located to the south of New Zealand (e.g., Lebrun et al., 2003) and is thought to be undergoing incipient subduction, although the age of initiation is poorly known (Gurnis et al., 2004). It is characterized by transpressive deformation, short wavelength geoid anomalies,

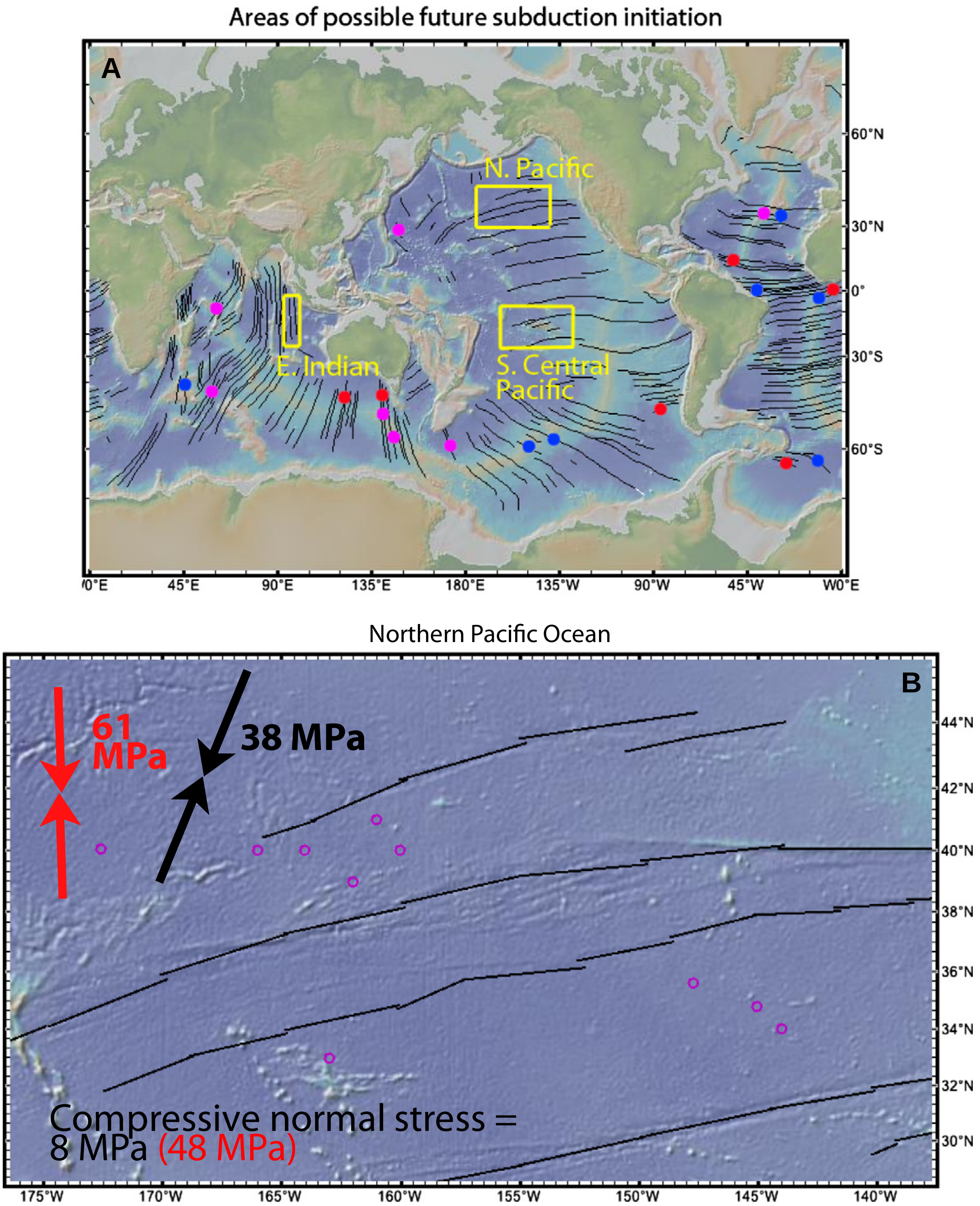


Fig. 6. Regional maps of those areas that are identified as most likely for subduction initiation (either ongoing or in the near future); these regions are associated with both significant compressional normal stress along fracture zone segments and with an age offset of ~15 Myr or more. (A) Global map showing the location of each region. The three regions which we identify as most likely sites for subduction initiation are marked with yellow boxes. Also marked are additional regions with particularly high normal compressive stresses in the Ghosh model (red dots) or the Lithgow-Bertelloni and Guynn model (pink dots), or that have large age offsets (blue dots). (B–D) Regional maps of the three regions we have identified as most likely to undergo subduction initiation via forced compression in the near future. Background colors show bathymetry; the location of fracture zone segments is shown as black lines. Earthquake hypocenter locations from the ISC catalog are plotted as pink dots. Available focal mechanisms from the global CMT catalog are shown in red and white. The orientation and magnitude of the representative regional maximum compressive stress from the Ghosh (2008) (large black arrows) and Lithgow-Bertelloni and Guynn (2004) models (large red arrows) are shown, along with the resolved average compressive normal stress for each model on the fracture segments shown. (Note that these are average regional values and do not reflect variation along individual segments.)

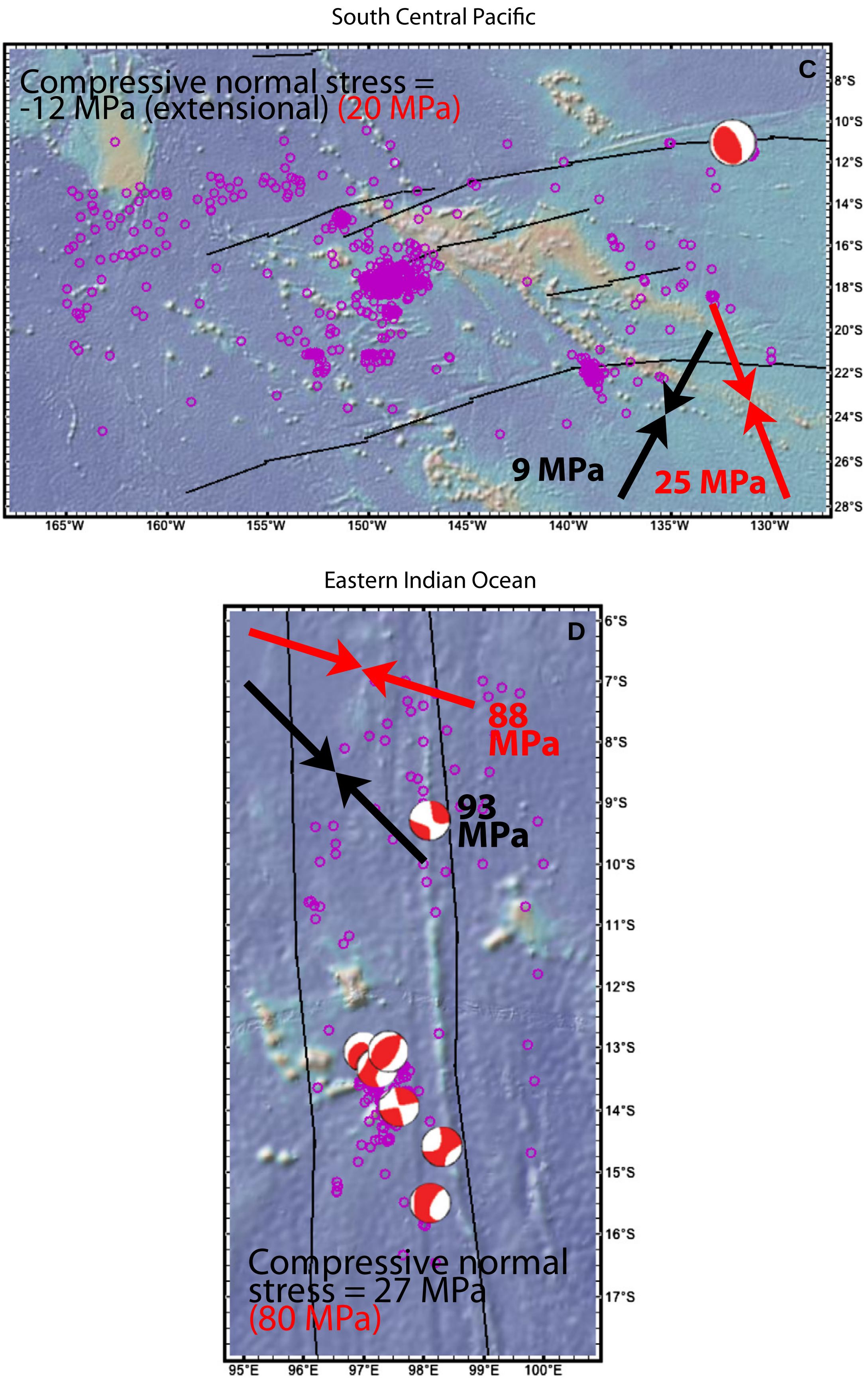


Fig. 6 (*continued*).

and paired ridge/trench topography (Ruff and Cazenave, 1985; Ruff et al., 1989). The Mussau Trench is located on the eastern boundary of the Caroline Plate and is viewed as a juvenile (<1 Myr old) subduction zone (Gurnis et al., 2004; Hegarty et al., 1982). The Owen Ridge is located in the Arabian Sea, off the coasts of Yemen and Oman, and has been suggested to be an incipient subduction zone associated with a preexisting fracture (Gurnis et al., 2004). The Samoa–Caroline proto-trench extends from Samoa to the Caroline Islands and is associated with significant intraplate deformation. Okal et al. (1986) speculated that the region represents an incipient plate boundary and that present-day subduction along the Vanuatu and Solomon trenches to the south may be in the process of being transferred to the Samoa–Caroline proto-trench.

We find that none of these inferred incipient subduction zones is associated with a significant offset in lithospheric age; the largest age offset is ~0.2 Myr at the Hjort Trench, which is much smaller than the age offsets of 30+ Myr observed elsewhere on the present-day Earth (Fig. 5). We also find that the proto-trenches associated with the supposed incipient subduction zones are not always associated with unusually large compressive normal stresses relative to our global database of fracture zone segments (Fig. 4). Calculations using both stress models indicate that these sites are generally undergoing some degree of compressive normal stress (Fig. 7), but it is often small compared to the largest stresses in the global database. For example, calculations with the Ghosh (2008) stress model indicate moderately high normal compressive stress at the Gorringe Bank

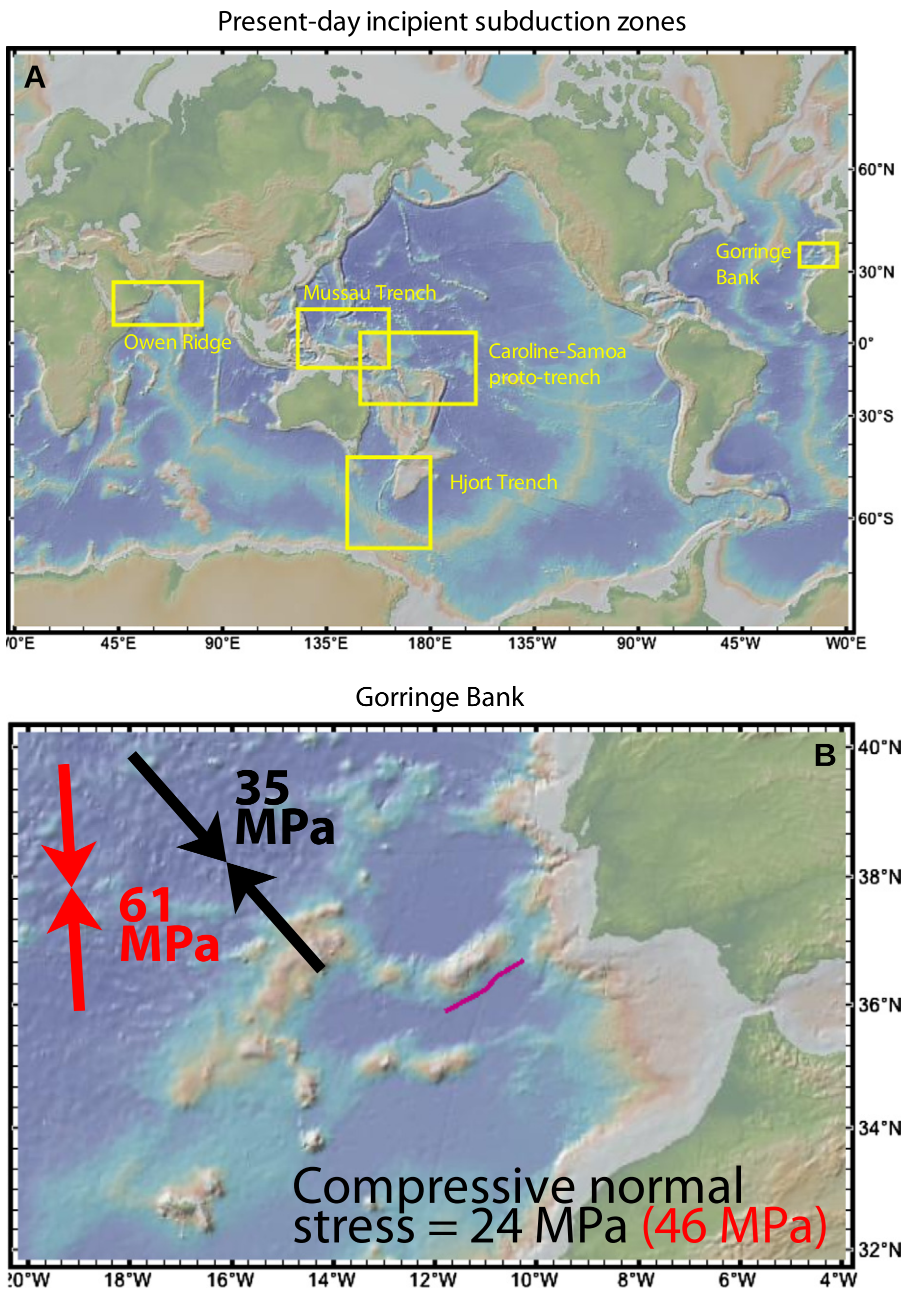


Fig. 7. Regional maps of those areas that have been identified as sites of likely ongoing incipient subduction by previous workers. (A) Global map showing locations of each incipient subduction zone. (B–F) Regional maps for the Gorringe Bank, Hjort Trench, Mussau Trench, Owen Ridge, and the Samoa–Caroline proto-trench — incipient trench locations are shown with pink lines. The orientation and magnitude of the average regional maximum compressive stress from the Ghosh (2008) (large black arrows) and Lithgow-Bertelloni and Guynn (2004) (large red arrows) models are shown, along with the resolved compressive normal stress for each model on the proto-trench. (Note that these are average regional values and do not reflect variation along individual segments.)

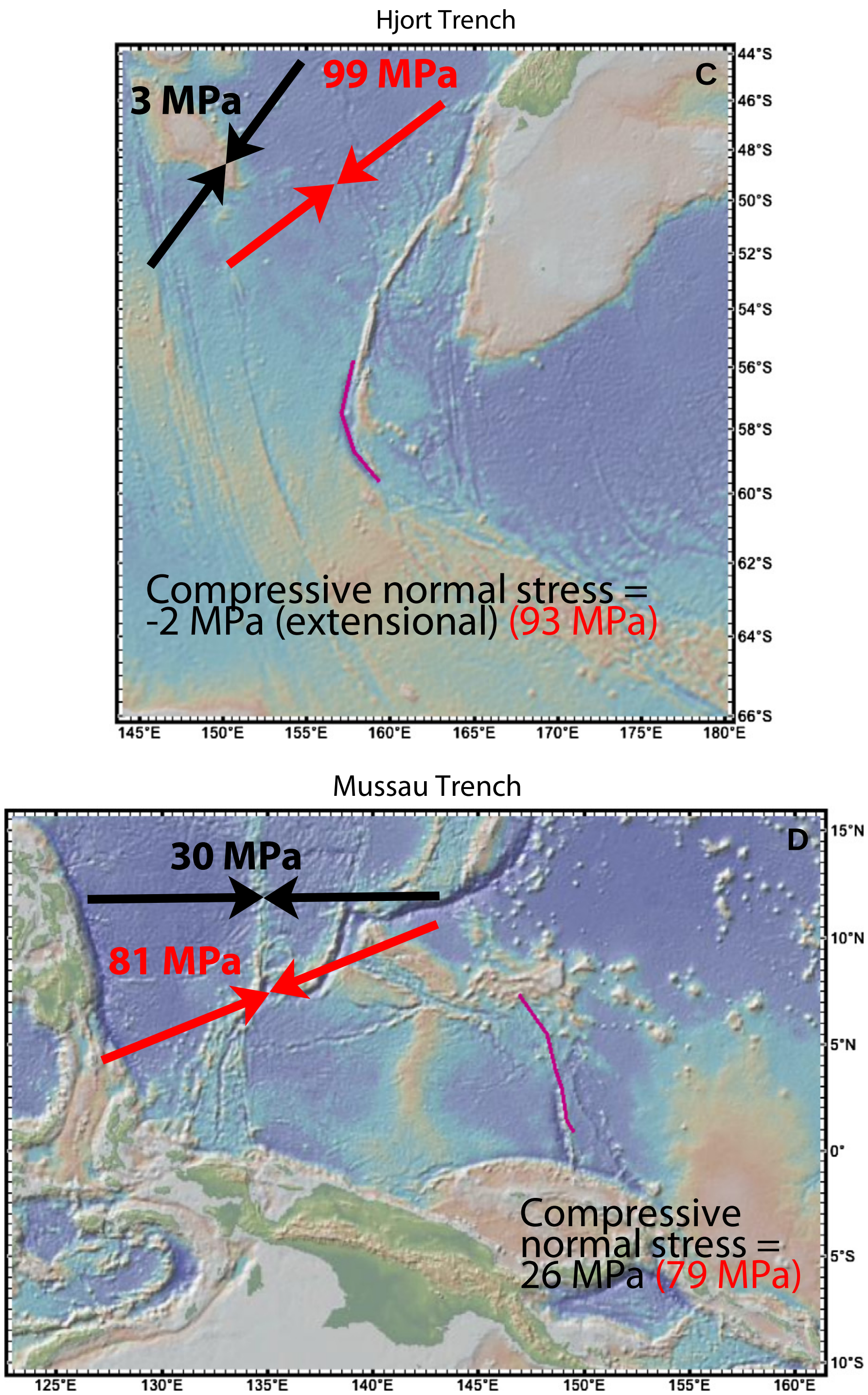
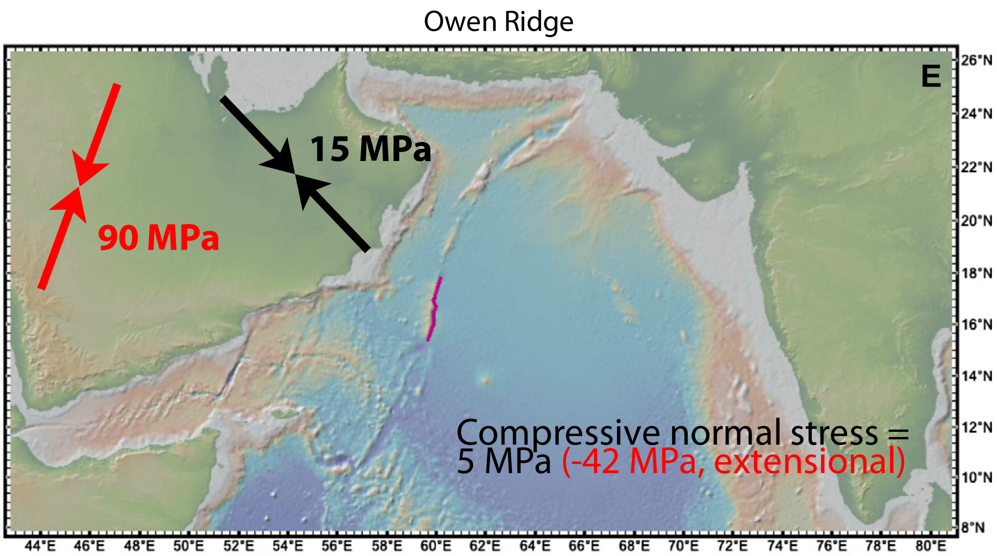


Fig. 7 (*continued*).

and Mussau Trench, but relatively low compressive stresses at the Owen Ridge and Samoa–Caroline proto-trench, with extension at the Hjort Trench. The Lithgow-Bertelloni and Guynn (2004) model suggests a high degree of compressive stress at the Hjort and Mussau Trenches, moderately high compressive stresses at Gorringe Bank, and extension at Owen and Samoa–Caroline. There is, surprisingly, no correlation between the regions identified as potential incipient subduction boundaries by Gurnis et al. (2004) and sites identified by our analysis as relatively likely for subduction initiation according to the dual criteria of large compressive stresses and large age offsets at oceanic fracture zone segments.

6. Discussion

Our approach toward evaluating the subduction initiation potential on fracture zone segments involves estimates of the state of stress, which is only an indirect indicator of the potential for compressive strain. The rheology of fracture zone segments is not well known, although if they are, as a rule, extensively serpentinized, then presumably they are weak enough to localize significant strain (e.g., Hilairet et al., 2007). From a modeling perspective, however, the amount of compressive stress necessary to produce the amounts of shortening (~100–150 km) needed to produce self-sustaining subduction in the



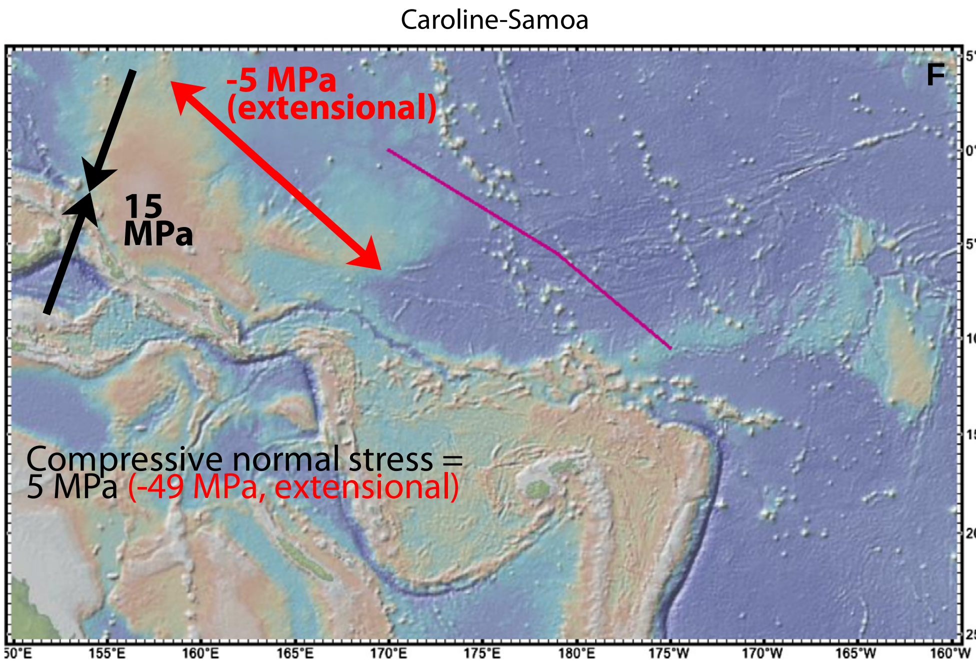


Fig. 7 (*continued*).

Gurnis et al. (2004) models is not well constrained. In the absence of precise rheological constraints, it is difficult to place a lower bound on the amount of compressive normal stress that may produce significant shortening, but we can infer that fracture zones with higher normal compressive stresses are relatively more likely to undergo shortening and are the most likely sites for future subduction initiation. Our approach also does not explicitly take into account either the timescale needed to initiate subduction via forced convergence or the timescale over which far-field lithospheric stresses are likely to change. The models of Gurnis et al. (2004) suggest that ~6 Myr is needed for a newly initiated subduction zone to accomplish the transition to self-sustaining subduction, while the lithospheric stress models used in our calculations only represent the present-day stress field.

It is also important to keep in mind that our estimates of the stress state on each fracture segment are derived from lithospheric stress models, which may have significant limitations. We are using global stress models for region-specific calculations, and this is a potentially severe limitation of our study. The global stress models used here make predictions that are generally consistent with observations of stress contained in the World Stress Map (Ghosh, 2008; Lithgow-Bertelloni and Guynn, 2004), but the paucity of stress observations in ocean basins means that the models have not been validated in these regions. While the two models used here are, to first order, similar, they do differ in their details in many regions, and our estimates for normal stresses differ as well. For example, the Lithgow-Bertelloni and Guynn (2004) model predicts modest compressional stresses on fracture zones in the northern Atlantic, while the Ghosh (2008) model predicts a neutral or mildly extensional normal stress regime here (Figs. 2 and 4).

Keeping these limitations on our study in mind, we can consider the implications of our analysis for the forced compression model for subduction initiation. We have identified a few regions of relatively high likelihood for present or (near) future subduction initiation that are associated with both high compressive normal stresses and large age (and therefore density) offsets. Our analysis has, however, demonstrated that the conditions that are needed to induce incipient subduction via the forced compression model of Gurnis et al. (2004) are relatively rare on the present-day Earth. There are few fracture zone segments that support an age offset of 30 Myr or more (Fig. 5) and very few segments that are associated with both a large age offset and relatively high compressive normal stresses. In the regions we have identified as associated with relatively high subduction initiation likelihood, evidence for actual intraplate compression and shortening is ambiguous. All three regions shown in Fig. 6B–D are associated with some intraplate seismicity, which may indicate deformation associated with shortening, but the limited number of well-resolved focal mechanisms hampers our ability to determine whether the Ghosh (2008) and Lithgow-Bertelloni and Guynn (2004) stress models accurately represent the stress field, and whether these regions are, in fact, undergoing reverse faulting that might indicate shortening. It is also notable that of the three regions shown in Fig. 6B–D, the region with the most abundant intraplate seismicity, the south central Pacific, actually exhibits relatively modest compressive stress for the Lithgow-Bertelloni and Guynn (2004) model and is in extension for the Ghosh (2008) model. Additionally, our analysis of five inferred incipient subduction zones indicates that these regions do not generally meet the two criteria that we have been evaluating. None of the five regions of inferred incipient subduction is associated with an age offset of larger than 0.2 Myr, and we also find that the models for far-field stresses in this region do not generally predict particularly large compressive normal stresses, at least in the context of our global database of fracture zone segment stress states.

What are the implications of our observations for models of subduction initiation? First, we note that taken at face value, our results imply that if the five regions shown in Fig. 7 are, indeed, associated with incipient subduction, subduction may be initiating via a process other than the forced convergence modeled by Gurnis et al. (2004). Alternatively, it is possible that shortening and underthrusting which leads to self-sustaining subduction can be produced even in the absence of a large density offset and with relatively low normal compressive stresses. If this were the case, however, then we would expect intraoceanic subduction initiation to be a very common process on the present-day Earth, as many fracture zone segments are under compressive stresses larger than those experienced by the regions of inferred incipient subduction (Figs. 4 and 6). A third possibility is that the stress models of Ghosh (2008) and Lithgow-Bertelloni and Guynn (2004) do not accurately represent the far-field stresses in these regions, or that the stress regimes have changed over relatively short timescales. The suite of forward models carried out by Gurnis et al. (2004) suggests that ~6 Myr is required to accomplish the transition from forced to self-sustaining convergence and complete the subduction initiation process. If the state of stress in incipient subduction regions has changed relatively rapidly, due to large-scale plate motion reorganization or some other process, then we can reconcile our estimates of relatively low compressive stresses with the idea that subduction is initiating via forced compression. If these regions are no longer being subjected to significant compression, however, then these incipient subduction zones may never accomplish the transition to self-sustaining subduction, and subduction initiation may be in the process of stalling.

Our inference that regions of present-day subduction initiation may be undergoing a process other than forced convergence across large-offset fracture zones and our observation that the conditions needed to produce this mode of incipient subduction are rare on the present-day Earth do not necessarily mean that this process is not a viable model for the initiation of subduction in general. There is some geological evidence that subduction zones have nucleated off large-offset fracture zones in the past (e.g., Choi et al., 2008; Pearce et al., 1992), although for some present-day subduction zones the pre-subduction tectonic regime is under debate. [For example, the Marianas subduction zone has often been inferred to have nucleated off a fracture zone (e.g., Gurnis et al., 2004; Hall et al., 2003; Pearce et al., 1992), but there has been some debate as to whether the proto-trench actually cut across preexisting tectonic structures (Taylor and Goodliffe, 2004).] The conditions needed to produce subduction initiation along a large-offset fracture zone do, however, appear to be fairly rare on the present-day Earth. Looking at the history of subduction initiation throughout the Cretaceous, this observation is perhaps unsurprising. At least five subduction zones have made the transition from forced to self-sustaining subduction during the Cretaceous (Gurnis et al., 2004), but there are long intervals of time over this period during which no subduction zones successfully initiated (e.g., between the initiation of South Scotia subduction at 45 Myr and that of the Manila subduction zone at 23 Myr). This punctuated nature of subduction initiation over the past ~65 Myr is consistent with the idea that the conditions needed to produce intraoceanic subduction initiation are relatively rare, and that while we observe several incipient subduction margins on the present day Earth, it is possible that few or none of them will actually develop into full-fledged self-sustaining subduction zones.

Our observations can be reconciled with the forced compression model in light of the intermittent history of subduction initiation during the Cretaceous (Gurnis et al., 2004). Another possibility, however, is that the forced compression model for subduction initiation does not describe the subduction initiation process well in general. The fact that fracture zones on the present-day Earth are generally parallel to plate motion (Fig. 1), while most subduction zones are approximately perpendicular to plate motion, tends to argue against the idea that most subduction zones initiate via forced compression across a fracture zone. Our finding that the conditions needed to initiate subduction via this mechanism are rare on the present-day Earth and that many of the known regions of incipient subduction do not seem to meet the criteria of the forced compression model may argue against the viability of this model for subduction initiation.

7. Summary

We have compiled a database of oceanic fracture zone segment endpoints and evaluated the age offsets and state of stress along each segment using two different global models for lithospheric stress. We then assessed these results in the context of intraoceanic subduction zone initiation models, which invoke compression and shortening across fracture zones, that act as preexisting zones of weakness and that involve a contrast in age (and therefore density) across the weak zone. We find that the conditions required for subduction initiation via the forced convergence model are relatively rare on the present-day Earth, but we have highlighted three regions where present-day or near-future subduction initiation is most likely: the northern Pacific Ocean, the southern central Pacific Ocean, and the eastern Indian Ocean. We also evaluated the age offsets and state of stress at sites of inferred incipient subduction on the present-day Earth, and find that these regions are not generally associated with large age offsets or particularly large compressive normal stresses. These observations suggest that regions of present-day subduction initiation may be undergoing a process other than forced convergence across large-offset fracture zones. Our analysis suggests that the conditions needed to produce incipient subduction via forced convergence across a preexisting weak zone in the oceanic lithosphere are rare on the present-day Earth, which is generally consistent with the punctuated nature of subduction initiation over the past ~65 Myr.

Acknowledgments

We thank the participants in many stimulating discussions on subduction initiation as part of the geophysics journal club at the Department of Terrestrial Magnetism, Carnegie Institution of Washington. We thank Attreyee Ghosh and Carolina Lithgow-Bertelloni for providing their stress models, and Bill Holt and Mike Gurnis for useful discussions. We are grateful to Katie Cooper for insightful comments on a draft of this manuscript. We thank two anonymous reviewers who provided thoughtful, detailed, and constructive critiques that improved the paper substantially. This work was supported by the Carnegie Institution of Washington Summer Intern Program in Geoscience (funded through the NSF-EAR Research Experiences for Undergraduates program), by Mount Holyoke College, and by Yale University.

Appendix A. Supplementary data

The complete database of fracture zone segment endpoints and their associated ageoffsets can be found in table form in the online supplementary material. Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.tecto.2011.09.017.

References

Baes, M., Govers, R., Wortel, R., 2011. Subduction initiation along the inherited weakness zone at the edge of a slab: insights from numerical models. Geophysical Journal International 184, 991–1008.

Billen, M.I., Hirth, G., 2005. Newtonian versus non-Newtonian upper mantle viscosity: implications for subduction initiation. Geophysical Research Letters 32, L19304. doi:10.1029/2005GL023457.

Borges, J.F., Fitas, A.J.S., Bezzeghoud, M., Teves-Costa, P., 2001. Seismotectonics of Portugal and its adjacent Atlantic area. Tectonophysics 337, 373–387.

Bull, J.M., Scrutton, R.A., 1992. Seismic reflection images of intraplate deformation, central Indian Ocean, and their tectonic significance. Journal of the Geological Society of London 149, 955–966.

Burov, E., Cloetingh, S., 2010. Plume-like upper mantle instabilities drive subduction initiation. Geophysical Research Letters 37, L03309. doi:10.1029/2009GL041535.

Chamot-Rooke, N., Jestin, F., de Voogd, B., 1993. Intraplate shortening in the central Indian Ocean determined from a 2100-km-long north–south deep seismic reflection profile. Geology 21, 1043–1046.

Choi, S.H., Mukasa, S.B., Shervais, J.W., 2008. Initiation of Franciscan subduction along a large-offset fracture zone: evidence from mantle peridotites, Stonyford, California. Geology 36, 595–598.

Cloetingh, S.A.P.L., Wortel, M.J.R., Vlaar, N.J., 1984. Passive margin evolution, initiation of subduction and the Wilson cycle. Tectonophysics 109, 147–163.

Cloetingh, S., Wortel, R., Vlaar, N.J., 1989. On the initiation of subduction zones. Pure and Applied Geophysics 129, 7–25.

Collot, J.-Y., Larmarch, G., Wood, R.A., Delteil, J., Sosson, M., Lebrun, J.-F., Coffin, M.F., 1995. Morphostructure of an incipient subduction zone along a transform plate boundary: Puysegur Ridge and Trench. Geology 23, 519–522.

Gerbault, M., 2000. At what stress level is the central Indian Ocean lithosphere buckling? Earth and Planetary Science Letters 178, 165–181.

Ghosh, A., 2008. Elucidating lithosphere-mantle coupling by modeling the lithospheric stress field and predicting plate motions. Ph.D. thesis, Stony Brook University.

Ghosh, A., Holt, W.E., Flesch, L.M., 2009. Contribution of gravitational potential energy differences to the global stress field. Geophysical Journal International 179, 787–812.

Goren, L., Aharonov, E., Mulugeta, G., Koyi, H.A., Mart, Y., 2008. Ductile deformation of passive margins: a new mechanism for subduction initiation. Journal of Geophysical Research 113, B08411. doi:10.1029/2005J13004179.

Gurnis, M., Hall, C., Lavier, L., 2004. Evolving force balance during incipient subduction. Geochemistry, Geophysics, Geosystems 5, Q07001. doi:10.1029/2003GC000681.

Hall, C.E., Gurnis, M., Sdrolias, M., Lavier, L.L., Müller, R.D., 2003. Catastrophic initiation of subduction following forced convergence across fracture zones. Earth and Planetary Science Letters 212, 15–30.

Hansen, V.L., 2007. Subduction origin on early Earth: a hypothesis. Geology 35, 1059–1062.

Hegarty, K.A., Weissel, J.K., Hayes, D.E., 1982. Convergence at the Caroline-Pacific Plate boundary: collision and subduction. In: Hayes, D.E. (Ed.), The Tectonic and Geological Evolution of Southeast Asian Seas and Islands, Part 2, Geophys. Monogr. Ser., vol. 27. AGU, Washington, DC, pp. 326–348.

Heidbach, O., Tingay, M., Barth, A., Reinecker, J., Kurfeß, D., Müller, B., 2009. The World Stress Map based on the database release 2008, equatorial scale 1:46,000,000, Commission for the Geological Map of the World, Paris, doi:10.1594/GFZ.WSM.Map2009.

Hilairet, N., Reynard, B., Wang, Y., Daniel, I., Merkel, S., Nishiyama, N., Petitregard, S., 2007. High-pressure creep of serpentine, interseismic deformation, and initiation of subduction. Science 318, 1910–1913.

Hynes, A., 2005. Buoyancy of the oceanic lithosphere. International Geology Review 47, 938–951.

Kemp, D.V., Stevenson, D.J., 1996. A tensile, flexural model for the initiation of subduction. Geophysical Journal International 125, 73–95.

Korenaga, J., 2007. Thermal cracking and the deep hydration of oceanic lithosphere: a key to the generation of plate tectonics? Journal of Geophysical Research 112, B05408. doi:10.1029/2006JB004502.

Kreemer, C., Holt, W.E., Haines, J.E., 2003. An integrated global model of present-day plate motions and plate boundary deformation. Geophysical Journal International 154, 8–34.

Krishna, K.S., Bull, J.M., Scrutton, R.A., 2001. Evidence for multiphase folding of the central Indian Ocean lithosphere. Geology 29 (8), 715–718.

Lebrun, J.-F., Lamarche, G., Collot, J.-Y., 2003. Subduction initiation at a strike-slip plate boundary: the Cenozoic Pacific-Australian plate boundary, south of New Zealand. Journal of Geophysical Research 108, 2453. doi:10.1029/2002JB002041.

Leroy, M., Dauteuil, O., Cobbold, P.R., 2004. Incipient shortening of a passive margin: the mechanical roles of continental and oceanic lithospheres. Geophysical Journal International 159, 400–411.

Lithgow-Bertelloni, C., Guynn, J.H., 2004. Origin of the lithospheric stress field. Journal of Geophysical Research 109, B01408. doi:10.1029/2003JB002467.

Mart, Y., Aharonov, E., Mulugeta, G., Ryan, W., Tentler, T., Goren, L., 2005. Analogue modeling of the initiation of subduction. Geophysical Journal International 160, 1081–1091.

McKenzie, D.P., 1977. The initiation of trenches: a finite amplitude instability. In: Talwani, M., Pitman III, W.C. (Eds.), Island Arcs, Deep Sea Trenches, and Backarc Basins, Maurice Ewing Ser., vol. 1. AGU, Washington, DC, pp. 57–61.

Meckel, T.A., Coffin, M.F., Mosher, S., Symonds, P., Bernardel, G., Mann, P., 2003. Underthrusting at the Hjort Trench, Australian-Pacific plate boundary: incipient subduction? Geochemistry, Geophysics, Geosystems 4, 1099. doi:10.1029/2002GC000498.

Mueller, S., Phillips, R.J., 1991. On the initiation of subduction. Journal of Geophysical Research 96, 651–665.

Müller, R.D., Roest, W.R., Royer, J.-Y., Gahagan, L.M., Sclater, J.G., 1997. Digital isochrons of the world's ocean floor. Journal of Geophysical Research 102, 3211–3214.

Nikolaeva, K., Gerya, T.V., Marques, F.O., 2010. Subduction initiation at passive margins: numerical modeling. Journal of Geophysical Research 115, B03406. doi:10.1029/2009JB006543.

Nikolaeva, K., Gerya, T.V., Marques, F.O., 2011. Numerical analysis of subduction initiation risk along the Atlantic American passive margins. Geology 39, 463–466.

Niu, Y., O'Hara, M.J., Pearce, J.A., 2003. Initiation of subduction zones as a consequence of lateral compositional buoyancy contrast within the lithosphere: a petrological perspective. Journal of Petrology 44, 851–866.

Okal, E.A., Woods, D.F., Lay, T., 1986. Intraplate deformation in the Samoa–Gilbert–Ralik area: a prelude to a change of plate boundaries in the Southwest Pacific? Tectonophysics 132, 69–77.

Pearce, J.A., van der Laan, S.R., Arculus, R.J., Murton, B.J., Ishii, T., Peate, D.W., Parkinson, I.J., 1992. Boninite and harzburgite from Leg 125 (Bonin–Mariana Forearc): a case study of magma genesis during the initial stages of subduction. Proceedings of the Ocean Drilling Program, Part B: Scientific Results 125, 623–659.

Regenauer-Lieb, K., Yuen, D.A., Branlund, J., 2001. The initiation of subduction: criticality by addition of water? Science 294, 578–580.

Ruff, L., Cazenave, A., 1985. SEASAT geoid anomalies and the Macquarie Ridge complex. Physics of the Earth and Planetary Interiors 38, 59–69.

Ruff, L.J., Given, J.W., Sanders, C.O., Sperber, C.M., 1989. Large earthquakes in the Macquarie Ridge Complex: transitional tectonics and subduction initiation. Pure and Applied Geophysics 128, 72–129.

Saleeby, J.B., 1984. Tectonic significance of serpentinite mobility and ophiolite melange. In: Raymond, L.A. (Ed.), Melanges: Their Nature, Origin, and Significance, Geol. Soc. Am. Special Paper., vol. 198, pp. 153–168.

Solomatov, V.S., 2004. Initiation of subduction by small-scale convection. Journal of Geophysical Research 109, B01412. doi:10.1029/2003JB002628.

Stern, R.J., 2002. Subduction zones. Reviews of Geophysics 40, 1012. doi:10.1029/2001RG00108.

Stern, R.J., 2004. Subduction initiation: spontaneous and induced. Earth and Planetary Science Letters 226, 275–292.

Stern, R.J., Bloomer, S.H., 1992. Subduction zone infancy: examples from the Eocene Izu–Bonin–Mariana and Jurassic California. Geological Society of America Bulletin 104, 1621–1636.

Taylor, B., Goodliffe, A.M., 2004. The West Philippine Basin and the initiation of subduction, revisited. Geophysical Research Letters 31, L12602. doi:10.1029/2004GL020136.

Toth, J., Gurnis, M., 1998. Dynamics of subduction initiation at preexisting fault zones. Journal of Geophysical Research 103, 18053–18067.

Ueda, K., Gerya, T., Sobolev, S.V., 2008. Subduction initiation by thermal–chemical plumes: numerical studies. Physics of the Earth and Planetary Interiors 171, 296–312.

van der Lee, S., Regenauer-Lieb, K., Yuen, D.A., 2008. The role of water in connecting past and future episodes of subduction. Earth and Planetary Science Letters 273, 15–27.