Formation of plate boundaries: The role of mantle volatilization

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Abstract

In the early Earth, convection occurred with the accumulation of thick crust over a weak boundary layer downwelling into the mantle (Davies, G.F., 1992. On the emergence of plate tectonics. Geology 20, 963–966.). This would have transitioned to stagnant-lid convection as the mantle cooled (Solomatov, V.S., Moresi, L.-N., 1997. Three regimes of mantle convection with non-Newtonian viscosity and stagnant lid convection on the terrestrial planets. Geophys. Res. Lett. 24, 1907–1910.) or back to a magma ocean as the mantle heated (Sleep, N., 2000. Evolution of the mode of convection within terrestrial planets. J. Geophys. Res. 105(E7): 17563–17578). Because plate tectonics began operating on the Earth, subduction must have been initiated, thus avoiding these shifts. Based on an analogy with the continental crust subducted beneath Hindu Kush and Burma, we propose that the lithosphere was hydrated and/or carbonated by H2O–CO2 vapors released from magmas generated in upwelling plumes and subsequently volatilized during underthrusting, resulting in lubrication of the thrust above, and subduction of the lithosphere along with the overlying thick crust. Once subduction had been initiated, serpentinized forearc mantle may have formed in a wedge-shaped body above a dehydrating slab. In relict arcs, suture zones, or rifted margins, any agent that warms and dehydrates the wedge would weaken the region surrounding it, and form various types of plate boundaries depending on the operating tectonic stress. Thus, once subduction is initiated, formation of plate boundaries might be facilitated by a major fundamental process: weakening due to the release of pressurized water from the warming serpentinized forearc mantle.

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1. Introduction

At present, plate tectonics operates on the Earth. This was probably not the case during the magma ocean stage (Warren, 1993; Sleep, 1999).
Paleomagnetic, geochemical, and isotope studies suggest that plate tectonics had started earlier than the Archean/Proterozoic boundary (Burke et al., 1976; Campbell and Griffiths, 1992). However, even if we accept that it had started by that time, it is not easy to describe how it began. Because of the strongly temperature-dependent viscosity of the mantle material (e.g. Kirby, 1983), the lithosphere is not easy to rupture. A number of studies show that subduction initiation requires high levels of stress to disrupt the lithosphere, which is a few tens of kmps thick (e.g. McKenzie, 1977; Clotetong et al., 1984; Mueller and Phillips, 1991). Meanwhile, the largest available plate tectonic forces are of the order of $10^{17}$ N/m (Molnar and Gray, 1979; Parsons and Richter, 1980; England and Molnar, 1991; Fleitout, 1991); these forces are insufficient to cause top-to-bottom disruption of a plate. Strain-weakening is thus required to form new plate boundaries (e.g. Bercovici, 1998; Moresi and Solomatov, 1998; Tackley, 1998; Regenauer-Lieb et al., 2001; Tommasi and Vauzech, 2001; Bercovici, 2003). Because plate tectonics does not operate on Venus or Mars, it is inferred that the weakening mechanisms should include effects of volatiles like H$_2$O and CO$_2$ (e.g. Bercovici, 1998; Regenauer-Lieb et al., 2001; Solomatov, 2004; O’Neill et al., 2007). Their effects on the thermo-mechanical properties of the lithosphere have been considered in numerical simulations of the initiation of plate boundaries (e.g. Bercovici, 1998; Regenauer-Lieb et al., 2001). They are, however, in general, not in the context of the formation of each type of plate boundary during the Earth’s tectonic history. Eclogitization of granulites induced by fluid injection might have played an important role in deforming the lithosphere (Austrheim et al., 1997; Bjornerud et al., 2002) and could be an example of the effects of volatiles. However, it is not yet clear how eclogitization played a role in the initiation of plate tectonics.

Another difficulty in initiating plate tectonics is the production of a large amount of basaltic crust at the surface by upwelling of the mantle. This occurred due to the high mantle potential temperature during the Archean (Sleep and Windley, 1982; Bickel, 1986; McKenzie and Bickle, 1988; Vlaar et al., 1994). The buoyant crust does not cool enough, which prevents subduction and leads to heating-up of the Earth's mantle and a shift back to a magma ocean (Sleep, 2000).

In this paper, we present a scenario where H$_2$O and CO$_2$ play a role in the initiation of plate boundaries in the context of realistic tectonic situations. We propose that subduction was initiated by volatilization of the underthrusting lithosphere that was once hydrated or carbonated by upwelling plumes. The ultramafic minerals that constitute the lithospheric mantle are affected by vapors rich in H$_2$O–CO$_2$ released from solidifying magmas in a plume head, and are altered to minerals such as amphibole, phlogopite, dolomite, magnesite, chloride, and serpentine (Menzies et al., 1987; Spera, 1987; Wylie, 1988). If such hydrated/carbonated lithosphere approaches a convergent zone and is under-thrust, volatilization from it would occur due to increases in temperature and pressure. The released volatiles would lubricate the thrust above, and make subduction initiation possible. This inference is obtained from the fact that subduction of the lithosphere with thick continental crust is now occurring beneath Hindu Kush and Burma according to intermediate-depth seismicity observed there (Seno and Rehman, 2011), although the nature of the crust is chemically different from that in the Archean. The lower portion of the lithosphere in these regions is likely to have been hydrated/carbonated by plumes when it passed over the Reunion and Kerguelen hotspots. This will be described later in more details.

Once subduction started, formation of serpentinized forearc mantle in a wedge-shaped body above a dehydrating slab followed. Kirby et al. (2003, 2013) proposed that the San Andreas fault system (SAF) was mobilized by weakening of the warming and dehydrating serpentinized forearc mantle, following the cessation of subduction of the Farallon plate. Serpentinized wedge mantle formed in a forearc may be found in certain tectonic settings, such as relict arcs, suture zones, and rifted margins. If the wedge is warped by any agent, the resultant released pressurized water will weaken the region surrounding the wedge, and will form various types of plate boundaries, depending on the stress regime. In the latter half of this paper, we show application of this weakening mechanism to the creation of each type of plate boundary during the stage after subduction had been initiated.

### 2. Subduction initiation

In this section, we treat the case where we consider that the initiation of subduction has not occurred. In uniform-viscosity convection, both convergent and divergent motions occur at the surface. This may mimic the convection that occurred during the early stages of the Earth with a weak boundary layer sinking into the mantle (Fig. 1a and b, Davies, 1992). However, in these conditions, crust thicker than the modern one would form over the lithosphere (Sleep and Windley, 1982; McKenzie and Bickle, 1988). Such crust results from primordial melt production of the mantle having a basaltic composition and a density of ~2900 kg/m$^3$ (McKenzie and Bickle, 1988). It is still prevented from subduction due to the plate being thin during the Archean (Davies, 1992; See also Cloos, 1993). At convergent margins, the crust is offscraped, thrust upward, and juxtaposed over crustal slivers, as in modern collision zones (Mattauer, 1986), with the mantle boundary layer downwelling alone, leaving the surface tectonics behaving unlike plate tectonics (Davies, 1992; Fig. 1a and b). In such a tectonic situation, if the average Earth’s surface heat flow is larger than the radioactive heat production, the mantle is cooled. If the boundary layer is too strong to be subducted, there occurs a shift to stagnant-lid convection like that on Venus (Solomatov and Moresi, 1996, 1997; Sleep, 2000). Conversely, if the average surface heat flow is smaller than the radioactive heat production, the mantle becomes hotter, and large-scale melting of the mantle (i.e. a magma ocean) resumes (Sleep, 2000). Therefore, for plate tectonics to start on the Earth, the lithosphere must be subducted along with its thick crust as a whole, avoiding the shift to stagnant-lid convection or to a magma ocean. However, it seems unlikely to occur in preference to the buoyant crust piling up at the surface.

#### 2.1. Subduction of the continental lithosphere beneath Hindu Kush and Burma

We propose a mechanism that describes the start of plate tectonics on the Earth. The locations where continental lithosphere with thick crust subducts into the modern Earth give us a clue. Hindu Kush and Burma provide evidence of the mechanism (See Rehman et al., 2011 and the references therein for the geologic history of this region, and see Seno and Rehman, 2011 for the tectonic setting). Hindu Kush is located within the Asian plate, southwest of Pamir and west of Karakorum. To the south, there is the Mesozoic Kohistan–Ladakh arc, which was welded to Asia during the late Mesozoic. India later collided with this arc in the early Tertiary. The geological terranes associated with the collision between India and Kohistan—the high Himalaya, the lesser Himalaya, and the Siwaliks—are similar to those in other Himalayan regions where India and Asia collided. However, the widths of the high and lesser Himalayas are only ~1/2 of those in other regions. This indicates that offscraping and accretion of the underthrusting Indian continental margin crust have had a smaller extent south of Kohistan.

The plate boundary where the Indian plate is currently underthrust is located south of Kohistan (Searle et al., 2001). To the north, intermediate-depth seismicity dips steeply northward to a depth of ~300 km beneath Hindu Kush and dips southward to a depth of ~150 km beneath Pamir (e.g. Billington et al., 1977; Searle et al., 2001). Studies of this seismicity using focal mechanisms and seismic tomography (e.g. Van der Voo et al., 1999; Pavlis and Das, 2000) show that the seismic zone beneath Hindu Kush is contiguous to that beneath Pamir. Because India has been colliding against Kohistan since the early Tertiary, no oceanic plate is recently subducting beneath this region, as pointed out by Searle et al. (2001). It is thus difficult to relate
the intermediate-depth seismicity, which is a phenomenon that has only occurred within the past several Ma, with subduction of an oceanic plate. The continental crust is thus unusually subducting in this region.

In western Burma, there is a belt of high-grade metamorphic rocks with Mesozoic ophiolites (e.g., Maurin and Rangin, 2009). This ophiolite belt was regarded as an eastward continuation of the Indus-Zangpo suture by Mitchell (1984). In Tibet, south of the ophiolite belt, the Tethys and high Himalayas, i.e., the Mesozoic and Paleozoic rocks offscraped from the Indian continental margin, display collision characteristics. However, these characteristics are absent in Burma. Instead, to the west of the ophiolite belt, there is the Indo-Burman Wedge, i.e., a Tertiary accretionary complex growing to the southwest (Maurin and Rangin, 2009). The Miocene-Quaternary calcalkaline volcanics are found to the east of the ophiolite belt (Mitchell, 1984). The Quaternary volcanic rocks show isotope and trace element patterns that indicate subduction of the Indian continental lithosphere (Zhou et al., 2012). There is intermediate-depth seismicity beneath Burma down to a depth of 200 km (e.g., Le Dain et al., 1984). These characteristics suggest that, in the Burman arc, subduction, rather than collision, has been occurring during the late Cenozoic. Because the Bay of Bengal and the Bengal Basin to the north have a crustal thickness $\leq 25$ km (Brune and Singh, 1986; Kaila et al., 1992), Burma is an anomalous place where continental crust is subducting with intermediate-depth seismicity, similar to Hindu Kush.

2.2. The Reunion and Kerguelen hotspots

The unusual subduction of the Indian continental lithosphere beneath Hindu Kush and Burma is in marked contrast to the collision in other parts of the Himalayas. Volcanics produced by the Reunion and
Kerguelen hotspots during the Mesozoic-Tertiary are distributed forming the traces of these hotspots on the Indian plate (Fig. 2). In accordance with the peculiarity mentioned above, Seno and Rehman (2011) found that the traces of the Reunion and Kerguelen hotspots on the Indian plate, formed at the time of ~100–126 Ma (Storey et al., 1989; Muller et al., 1993; Kent et al., 2002), are located in the vicinity of the intermediate-depth seismicity beneath Hindu Kush and Burma (Fig. 2). The Indian continental lithosphere in these regions is thus likely to have been affected by plumes upwelling beneath these hotspots. Following Wyllie (1988), we expect that upwelling plumes with partial melts involving H–O–C release these volatiles as vapors when magmas solidify near the base of the lithosphere (Fig. 3a, see also Figs. 45 and 46 of Wyllie, 1988). This would alter mantle minerals at depths < 100 km above the plate base by permeation through pores (Mibe et al., 1999), producing amphibole at T < 1050 °C, chlorite at T < 700–800 °C, and serpentine/talc at T < 600–700 °C (e.g. Poli and Schmidt, 1995; Ulmer and Trommsdorff, 1995; Wunder and Schreyer, 1997; Bose and Navrotsky, 1998; Iwamori, 1998; Bromiley and Pawley, 2003; Fumagalli and Poli, 2005; Till et al., 2012), in addition to other peridotite–CO$_2$ minerals at similar temperature–depth ranges (Brey et al., 1983; Wallace and Green, 1988; Falloon and Green, 1989; Canil and Scarfe, 1990; Dasgupta et al., 2007). As noted in the introduction...

![Figure 2](image-url)
section, we call these chemical alternation processes of lithospheric mantle minerals “hydration and/or carbonation” in this paper. If the lithosphere is in tensional tectonic stress, magmas, and vapors might further migrate upward by magma-hydrofracturing, inducing hydration/carbonation in the shallower portion (see Fig. 46 of Wyllie, 1988; Spera, 1987; Wilshire and Kirby, 1989). Fig. 3a schematically illustrates this hydration/carbonation, adapting Fig. 45 of Wyllie (1988). Because the solids at level 3 in Fig. 3a is assumed to be ~1300 °C (see Fig. 20 of Wyllie, 1988), the altered minerals are stable in the lithosphere when they are more than a few tens of kms above the solidus (green shaded in Fig. 3a). The lower portion of the Indian continental lithosphere, once located above the hotspots, is thus likely to have been hydrated/carbonated by plumes.

2.3. Hydration/carbonation of oceanic lithosphere

This kind of hydration/carbonation by plumes, albeit for oceanic lithosphere, has been invoked by Kirby (1995) and Seno and Yamanaka (1996) to explain the occurrence of double-seismic zones in the circum-Pacific subduction zones. Based on the devolatilization embrittlement mechanism for intraslab intermediate-depth earthquakes (Raleigh and Paterson, 1965; Nishiyama, 1992; Kirby et al., 1996; Seno and Yamanaka, 1996; Jung et al., 2004; Takahashi et al., 2009), they interpreted double seismic zones to represent volatilization of slabs due to increasing temperature and pressure during subduction. The upper plane of double seismic zones can be explained by devolatilization embrittlement of altered basalts in the subducting oceanic crust (Kirby et al., 1996; Peacock and Wang, 1999; Hacker et al., 2003; Yamasaki and Seno, 2003; Kita et al., 2006; Nakajima et al., 2009). The lower plane requires the deeper half of the oceanic lithosphere to be hydrated/carbonated prior to subduction. Kirby (1995) and Seno and Yamanaka (1996) attributed this to be the carbonatization and hydration, respectively, similar to that depicted in Fig. 3a, by H2O–CO2 vapors released from magmas in plumes rising up in the Pacific Ocean. Although direct evidence for this metamorphism is scarce, Seno and Yamanaka (1996), Hacker et al. (2003), and
Yamasaki and Seno (2003) showed that the dehydration locus of serpentinites formed in the lower portion of subducting lithosphere is consistent with the observed geometry of the double seismic zone; on the contrary, the locus formed by infiltration of seawater in the upper half of oceanic lithosphere by normal faulting (Peacock, 2001; Ranero et al., 2005; Faccenda et al., 2009) is not consistent.

2.4. Collision versus subduction

It is generally believed that, when continental lithosphere approaches a convergent boundary, subduction is hindered due to the thick buoyant crust, and therefore, collision generally occurs (e.g. McKenzie, 1969; Molnar and Gray, 1979; van den Beukel, 1992; Cloos, 1993). However, Seno (2008) showed that buoyancy is not an essential factor in preventing subduction, but absence of dehydration from a slab is rather important. Continental crust does not contain metamorphic minerals that dehydrate at intermediate-depths (Ernst et al., 1998), which makes both pore fluid pressure ratio in the thrust and intermediate-depth seismicity low (Seno and Yamasaki, 2003; Seno, 2007), resulting in offscraping and piling up of the crust in collision zones (Seno, 2008). Conversely, in subduction zones, dehydration of the slab crust and mantle (Anderson et al., 1976; Peacock, 1993; Poli and Schmidt, 1995; Okamoto and Maruyama, 1999; Peacock and Wang, 1999; Seno et al., 2001; Hacker et al., 2003; Fumagalli and Poli, 2005; Kameda et al., 2011; Kuwatani et al., 2011; Tull et al., 2012) results in pore fluid pressure ratios as high as 0.90–0.98 (Wang and He, 1999; Wang and Suyehiro, 1999; Seno, 2009; Kimura et al., 2012), which render smooth subduction with very low friction at the thrust.

The unusual subduction of the continental crust beneath Hindu Kush and Burma should be closely related to the existence of intermediate-depth seismicity in these regions. As mentioned before, the lower portion of the Indian lithosphere could be hydrated/carbonated by plumes upwelling beneath the Reunion and Kerguelen hotspots, similar to oceanic lithosphere with a double seismic zone. After traveling over thousands of kms to the north, the hydrated/carbonated lithosphere approaches and is underthrust beneath Asia (Fig. 2), and volatilization occurs from the lithosphere due to increases in temperature and pressure. This would result in intermediate-depth seismicity by devolatilization embrittlement and the released volatiles would lubricate the thrust above, enabling subduction under Asia (Fig. 3b). The intermediate-depth seismicity beneath these regions is essentially the same as that in the western Pacific (e.g. Billington et al., 1977; see other references cited in Seno and Rehman, 2011), which implies that the extent of hydration/carbonation of the slab is similar to that of subducting slabs in the circum-Pacific. Because the diameter of the temperature anomaly and related volcanics of an upwelling plume is generally of the order of 1000 km (White and McKenzie, 1989; see also the spatial dimension of the volcanics related to the Reunion and Kerguelen hotspots shown in Fig. 2), the induced subduction is not spot-like, but can be in a zone extending over several hundreds of km. Although direct evidence of subduction of hydrated/carbonated continental lithosphere in these regions is not yet available, Sumino et al. (2011) noted OIB signatures in the $^{3}$He/$^{4}$He ratio of diamond exhumed in the ultra-high pressure meta-morphic belt of northern Kazakhstan, as an example of such subduction. The Quaternary volcanic rocks in Burma also show isotope and trace element patterns indicating subduction of the Indian continental lithosphere there (Zhou et al., 2012), as mentioned before.

2.5. Subduction initiation in the Archean

We propose that subduction initiation in the Archean occurred in a similar manner, although, in this case, the thick crust that prevented subduction is basaltic. Hydration/carbonation of the lithosphere by plumes, as shown in Fig. 3a, would have occurred more frequently during the Archean, because plume activity was much higher at that time (Campbell and Griffiths, 1992; Kroner and Layer, 1992). Conversely, the thickness and thermal state of the lithosphere were not much different from those at present (Bickel, 1978; Burke and Kidd, 1978; Jarvis and Campbell, 1983; England and Bickle, 1984; Bickle, 1986), due to the efficient removal of heat from the Earth’s interior by the high plate creation rate at that time. When such lithosphere approached and was underthrust in the Archean convergent zones, as in Hindu Kush and Burma, increases in temperature and pressure caused by underthrusting would have induced volatilization from the hydrated/carbonated lithosphere, and the volatiles thus released would have lubricated the thrust (Fig. 3b). If the subducted crust was transformed to denser eclogite as shown by Austrheim et al. (1997), it would have aided further underthrusting of the lithosphere (Ringwood and Green, 1966; Ahrens and Schubert, 1975). Eclogitization of the lower crust in the vicinity of the thrust zone also enhanced its weakness by the interplay between fluid injection, chemical reaction, and deformation (Austrheim et al., 1997; Bjornrud et al., 2002) and made subduction easier by overcoming the energy dissipation associated with plate bending (Conrad and Hager, 1999). These together would have made such lithosphere with thick crust enter the deeper mantle, resulting in subduction initiation (Figs. 1d and 3b).

3. Serpentinized forearc mantle wedge and its role in the formation and mobilization of plate boundaries

In uniform-viscosity convection, divergence can occur by ductile extension of a boundary layer, without any mid-ocean ridge (MOR) like feature (Davies, 1992). Toroidal motions, if any, can be accommodated by diffuse deformation between the loci of convergence and divergence (Fig.1a and c; see also Fig. 1 of Bercovici, 1996), rather than by transform faults (Fig.1f). As an exception, partial melting in the divergence zone may weaken the boundary layer and form a linear divergent boundary with a MOR-like feature (Fig. 1a and c, Sleep, 2000). Other types of plate boundaries, however, do not form easily in the lithosphere as it cools and becomes strong. In this section, we propose the following mechanism to cause strain localization, which leads to the formation of new plate boundaries in such strong lithosphere (Fig. 1e and f).

Initiation of subduction discussed in the previous section may form serpentinized forearc mantle in subduction zones. In the modern Earth, recent geophysical evidence shows that serpentinized mantle occurs in cold, shallow, wedge-shaped regions of forearcs above dehydrating slabs in several well-studied subduction zones (Kamiya and Kobayashi, 2000; Bostok et al., 2002; Brocher et al., 2003; Hyndman and Peacock, 2003; Blakely et al., 2005; Matsubara et al., 2005; Seno, 2005; Wang et al., 2006). Shallow serpentinized forearc mantle forms because water is released into the forearc from dowoggings, warming, and thus dehydrating slab crust and mantle, and also because the descending slab cools the forearc mantle nearest the trench and slab, resulting in a large region where serpentinite is stable (e.g. Hyndman and Peacock, 2003). The volume of water potentially stored in the forearc is huge, about 200 km$^3$/km of margin or more, depending on the degree of serpentinization of the forearc mantle (Kirby et al., 2013). The inboard and outboard wedge boundaries of the serpentinized mantle are controlled by the thermal structure of the forearc, the plate interface, and the forearc crust–mantle boundary.

If the subducting slab gets progressively warmer (e.g. as a MOR approaches the trench, or the rate of subduction slows) or subduction stops, the forearc mantle will begin to warm and the region of serpentine stability will shrink, releasing water along the wedge boundaries. Kirby et al. (2013) calculated the release rate of water for the post-subduction case pertinent to the California Coast Ranges. Most of the $10^{14}$ kg/km water stored in the wedge mantle is released as free water by 10–25 M.y. after subduction ceases; its longevity depends on specific model assumptions on the wedge size and the warming mechanism. In particular, these calculations were conducted for the subduction zone where the warm Farallon slab was subducting. In cold subduction zones, a volume of serpentinized forearc mantle could be
much larger, and consequently the water release time could be longer. During the period of dehydration of such serpentinized forearc mantle, the released pressurized water weakens its boundary by the mechanisms described below in more detail. If a regional shearing tectonic stress operates, the dehydrating boundary could become a transform fault (Fig. 1e and f), as suggested for the Coast Ranges by Kirby et al. (2013).

Such formation of plate boundaries may not be restricted to active or relict forearcs. In some cases, serpentinized forearc mantle that overrider active margins may be amalgamated into a collision zone (Burke et al., 1977; Dewey, 1977). Utilizing the result that water release occurs over a time scale of tens of Myr, some of this serpentinized mantle may survive and be stabilized into suture zones, if the period of orogeny is short enough. Furthermore, since the Proterozoic, rifts disrupting a supercontinent often occurred along suture zones (e.g. Burke et al., 2003). Disrupter smaller continents amalgamated again to form a new supercontinent. This is called a Wilson cycle or a supercontinent cycle, after Wilson (1966) pointed out that a proto-Atlantic (Lapetus Ocean) closed during the Middle–Upper Paleozoic and the Atlantic re-opened in the early Mesozoic. It is now known that such cycles have occurred a few times since the Proterozoic (Mckerron and Ziegler, 1972; Dalziel, 1995; Rogers, 1996). For example, West Gondwana collided with Laurasia at ~750, 490, and 370 Ma since the dispersal of Rodinia (Dalziel, 1995). Following breakup of a supercontinent, rifted margins surrounded the ocean that had opened. Some of the serpentinized forearc mantle may survive and remain in such rifted margins. Relict arcs, suture zones, and rifted margins are thus possible loci for storage of serpentinized mantle. Any agents warming them would release water and weaken the dehydrating boundaries. This strain localization, with sufficient tectonic stress, would create new plate boundaries at these localities.

Possible mechanisms for the released water to weaken the lithosphere are (1) reduction of fault strength by increased pore fluid pressure (Hubbert and Rubey, 1959), (2) hydrofracturing, and (3) reaction-enhanced ductility. The effects of increased pore fluid pressure are similar to those for active faults in the crust. Fluid production rate arising from the interplay between dehydration, permeability of the host rock, pore creation by faulting, pore compaction by ambient stresses, and shear heating controls the temporal change in the pore pressure and thus, the strength of the dehydrating boundary (e.g. Sleep and Blanpied, 1992; Segall and Rice, 1995; Sibson, 1996). However, application of this weakening mechanism to realistic serpentinized forearc mantle awaits future studies. Hydrofracturing occurs under the following four conditions for the dehydrating material: positive volume change ($\Delta V > 0$) through the dehydration reaction, low permeability, low strain rate (Nishiyama, 1989), and small differential stress with $\sigma_{\text{min}} < 0$ (Secor, 1965; Etheridge, 1983). For antigorite, $\Delta V > 0$ is satisfied by dehydration that occurs at depths $< 75$–125 km (Ulmer and Tromsdorff, 1995; Wunder and Schreyer, 1997; Bromiley and Pawley, 2003; see also Fig. 4b of Yamasaki and Seno, 2003), which is the depth range for the serpentinized mantle of our interest. Enhancement of ductility of the lower crust by transformation to eclogite from granulites through fluid injection (Austrheim et al., 1997; Bjørnerud et al., 2002) may be in the category of reaction-enhanced ductility, although there is evidence for the existence of brittle deformation because pseudotachylites are observed simultaneously for this transformation (Austrheim and Boundy, 1994). In the following subsections, we depict the processes of formation for each type of plate boundary in terms of these weakening mechanisms.

3.1. Divergent boundaries

In this subsection, we consider the formation mechanism of a divergent boundary within thick lithosphere. Generally, it is difficult for slab pull or suction forces to disrupt continental lithosphere. Previous studies suggested that plumes play important roles in the breakup of a continent or a supercontinent. Morgan (1971) pointed out temporal coincidence between flood basalt and continental breakup and suggested a connection between plume activity and the onset of seafloor spreading. Duncan (1984) also noted that hotspots have been overridden by spreading ridges in the central Atlantic. White and McKenzie (1989) showed that huge volcanic provinces are located in rift zones, which may be associated with uprising plumes (i.e. volcanic rifted margins). Courtillot et al. (1999) further advanced the Morgan’s (1971) idea on the temporal coincidence between plume activities and breakup. Bott (1982) further proposed that a rising plume warms and thins the lithosphere, which may help disrupt it under slab suction forces. Even if, however, the mechanical thickness of the lithosphere is halved, its strength is still an order larger than the magnitude of plate tectonic forces. Therefore, pre-existing heterogeneities in the lithospheric structure are necessary for breakup, in addition to plate tectonic forces and thinning by plumes (Dunbar and Sawyer, 1989; Courtillot et al., 1999; Tommasi and Vauchez, 2001). In fact, breakup often occurs along sutures (e.g. Burke et al., 2003), and Tommasi and Vauchez (2001) attributed the weakness of suture zones to lattice preferred orientation of olivine. However, reduction in strength produced by this mechanism is less than two thirds of magnitude (see Fig. 5 of Tommasi and Vauchez, 2001) and is still short of that needed for plate tectonic forces to cause breakup.

In Fig. 4a, we assume that 200-km-thick cratonic lithosphere has a temperature of ~1200 °C at the base (adapting values of Figs. 7 and 8 of Wyllie, 1988), serpentinized forearc mantle is embedded in the suture zone, and subduction occurs from both sides of the continent (Bott, 1982). Antigorite is stable at depths $< 150$ km in such lithosphere (see Fig. 4b of Yamasaki and Seno, 2003). Even if the temperature at the base is ~100 °C higher during the Archean (Jarvis and Campbell, 1983), antigorite is still stable at a similar depth range. We now assume that plumes rise beneath some locations in the suture. Thermal anomalies in the regions adjacent to the plumes would significantly reduce the strength of the lithosphere by dehydration (Fig. 4b). This mechanism, together with thinning of the lithosphere (Bott, 1982) and weakening due to lattice preferred orientation (Tommasi and Vauchez, 2001), might lead to breakup of a supercontinent, which is then succeeded by oceanfloor spreading (Fig. 4c). In Fig. 1e, we assume a different tectonic situation where a supercontinent in a divergent zone, without subduction from both sides, overriders an upwelling plume. In this case, basal drag forces, with a dehydration weakening mechanism similar to that shown in Fig. 4b, may lead to breakup of a continent.

Supporting this inference, serpentinized mantle is often found in non-volcanic rifted margins in the Atlantic (Fig. 5, Billot et al., 1989; Whitmarsh et al., 1993; Reid, 1994; Chian et al., 1995; O’Reilly et al., 1996). This may be comprised of relics of serpentinized forearc mantle embedded in suture zones, whose dehydration boundaries acted as weak zones for continental breakup. In volcanic margins, we do not observe such serpentinized mantle, but high velocity layers in the thickened lower crust, which are interpreted as igneous material added at the time of rifting (White et al., 1987). We speculate that serpentine is dried out in such volcanic margins beneath which plumes of high temperature rose up (White and McKenzie, 1989).

This explanation for the serpentinized mantle found in non-volcanic rifted margins is an alternative to downward infiltration of seawater through brittle fractures of the entire crust proposed by others (O’Reilly et al., 1996; Perez-Gussinye and Reston, 2001). Note that such infiltration is only possible after extension of the rifted crust is largely accomplished and already thinned (Perez-Gussinye and Reston, 2001). The fact that serpentinized mantle is observed at continent–ocean boundaries, not beneath centers of abyssal plains (Billot et al., 1989; White et al., 1990; Pinheiro et al., 1992; Whitmarsh et al., 1993; Reid, 1994; Chian et al., 1995) seems to rule out such infiltration. Isotope studies by Skelton and Valley (2000) in the Iberia Abyssal Plain also show that a massive volume of the mantle was serpentinized before it was exhumed. The mass of water stored in the serpentine beneath the
Rockall Trough amounts to $0.6 \times 10^{14}$ kg/km (O'Reilly et al., 1996), which is on the same order of the mass of water estimated for the California Coast Ranges (Kirby et al., 2013), suggesting that our interpretation is plausible.

3.2. Convergent boundaries

Since the initiation of plate tectonics, a number of subduction zones must have been newly created. Stern (2004) classified these into induced nucleation subduction zones (INSZ) and spontaneous nucleation subduction zones (SNSZ). In INSZ, transference and polarity reversal types are distinguished. Transference INSZ move the new subduction zone outboard of the failed one. The on-going development of a plate boundary in the Indian Ocean, southeast of India, in response to the Indo–Asian collision (Wiens et al., 1985; Neprochnov et al., 1988; Stein et al., 1989) is a typical example of transference INSZ processes. Polarity reversal INSZ also follow collision, but in these cases, a new subduction zone forms behind the magmatic arc. As discussed in the Introduction, a force one order larger than plate tectonic forces is required to initiate a new subduction zone. This is the same for INSZ. A low pore fluid pressure ratio at a collision zone thrust due to lesser dehydration from underthrusting continental crust might produce such a large tectonic force (Seno, 2007, 2008). The fact that no feature like a subduction zone has yet developed in the Indian Ocean (e.g. Bull and Scrutton, 1990; Van Orman et al., 1995) indicates, however, that the magnitude of such a collision force would not be overwhelming.

SNSZ results from gravitational instability of oceanic lithosphere, either at a passive margin or at a transform fault/fracture zone. Cloetingh et al. (1984), however, showed that spontaneous collapse of old lithosphere at a passive margin is unlikely due to its strength. In fact, there is no example of SNSZ at passive margins during the Cenozoic (Stern, 2004). Therefore, collapse under a tensional or strike–slip tectonic setting has been invoked (Turcotte et al., 1977; Erickson, 1993; Kemp and Stevenson, 1996). Accordingly, numerical simulations show that lithosphere in contact with another one along a transform fault/fracture zone could be unstable if there is enough difference in age (e.g. Fujimoto and Tomoda, 1985; Hall et al., 2003), resulting in SNSZ. Not many such examples are, however, known in geologic history. The Izu–Bonin–Mariana arc is often referred to as a typical example of SNSZ formed by such collapse along a transform fault (Hilde et al., 1977; Hawkins et al., 1984; Wells, 1989; Stern and Bloomer, 1992). Seno and Maruyama (1984), Seno (1988), Hall et al. (1995),

Fig. 4. Possible mechanism of continental breakup (modified from Fig. 1 of Bott, 1982). Subduction occurs from both sides of a supercontinent. $F_{su} =$ trench suction, $F_{rp} =$ ridge push. These forces are not enough large to cause continental breakup. (a) In this case, we assume that a suture zone was formed in the middle of the supercontinent by the closure of an intervening ocean basin, and that serpentinized forearc mantle, formed during this earlier phase of subduction, is embedded in the suture zone. Note that antigorite is stable in the upper 3/4 of the lithosphere if the temperature at the base is ~1200 °C (Wyllie, 1988). (b) In the succeeding stage, we assume that a plume begins rising beneath the suture (White and McKenzie, 1989). Dehydration of the serpentinized forearc mantle occurs as it is affected by such a plume, making the dehydration boundary very weak. (c) Along with non-volcanic margins heated by the plume, this would make it easier for the continent to rift and break up. The above tectonic situation is different from Fig. 1e (left), but the proposed mechanism of formation of a divergent boundary is the same.
Deschamps and Lallemant (2002), and Taylor and Goodlife (2004), however, discounted such an origin, because the Bonin arc has been active at least since 48 Ma, with an arc trend oblique to the hypothesized transform fault.

Therefore, although it has been generally conceived that a passive margin or a transform fault/fracture zone turns into SNSZ, the mechanism is still disputed. Similarly, the amount of tectonic force required for nucleation of INSZ is yet unresolved. Even if thermal cracking weakens oceanic lithosphere (Korenaga, 2007), the lower half of the lithosphere might remain strong, as seen by the dominant reverse fault-type earthquakes beneath mid-oceans, outer-rises, and passive margins (Stein et al., 1979; Forsyth, 1982; Wiens and Stein, 1985; Seno and Yamanaka, 1996).

We propose, similar to the formation of divergent boundaries, that dehydration weakening of serpentined forearc mantle may have played an important role in initiating subduction zones in the Earth's history after plate tectonics had started. As stated before, serpentined forearc mantle may be embedded in non-volcanic rifted margins or in relict arcs where subduction has stopped (Fig. 1e). If any thermal event occurs in its vicinity under a regional compressional stress, the released pressurized water will weaken the dehydration boundaries of the mantle, and possibly induce INSZ or SNSZ. Upwelling plumes may be responsible for such thermal events, and may behave as a control factor of subduction zone nucleation. This would be much more efficient in weakening lithosphere along a rifted margin or a relict arc than a simple interaction between lithosphere and plumes (Burov and Cloetingh, 2010) or sediment loading (Cloetingh et al., 1984; Erickson, 1993). An extraordinary amount of collision forces is not required for INSZ in this case.

3.3. The San Andreas fault system (SAF)

The SAF was formed by disappearance of the Farallon plate and succeeding contact of the Pacific plate with the North American plate (Atwater, 1970). Because the fault system developed a few hundred km east of the relict trench axis (Fig. 6a, McCulloch, 1989), we cannot apply, in a straightforward manner, Atwater's (1970) idea (Fig. 6b) for the initiation of the SAF. As already mentioned, Kirby et al. (2003, 2013) proposed a model suggesting that the mobility of the SAF was enabled by dehydration weakening of the serpentined forearc mantle beneath the Coast Ranges. As a margin-parallel shear stress was applied along the mantle wedge by the Pacific–North American relative motion (Fig. 6c), the weak zones in the dehydrating mantle would have faulted aseismically, and the crust located above would have deformed by strike–slip faulting (see Kirby et al. (2013) for more details). Warming occurred in this case by a slab gap (Dickinson and Snyder, 1979; Furlong et al., 1989) or by a stalled slab (Bhattacharya and Parsons, 1995; ten Brink et al., 1999) after the cessation of subduction of the Farallon plate. Fulton and Saffer (2009) supported this idea with a numerical simulation that showed that such a release of water could maintain a pore fluid pressure high enough to mobilize the SAF. In other regions, similar release of pressurized water could occur along a relict arc or a rifted margin, if they contain serpentined forearc mantle warmed by some agents. If a shearing tectonic stress operates as in the Coast Ranges, it will induce transient current motion along the dehydrating boundary, which might lead to initiation of a transform fault (Fig. 1e and f).

3.4. Other continental and oceanic transform faults

A spectacular example of formation of major strike–slip faults in collision zones is the North Anatolian fault (NAF) of northern Turkey. The NAF, whose activity was initiated by the collision of Arabia with Eurasia (Sengor, 1979), is located along a suture zone formed by the earlier collision between Laurasia and Gondwana (the Pontide–Sakarya collision during the Late Cretaceous) (e.g. Sengor, 1979; Yilmaz et al., 1995). Three metamorphosed units are observed in the western NAF, between the two branches (north and south); their protoliths are the Sakarya continent, ophiolites, and the Pontide fragment from south to north (Yilmaz et al., 1995). Yilmaz et al. (1995) regarded the metamorphosed ophiolites as the fragments of the Neotethys. We suggest an alternative interpretation that these rocks represent the serpentined forearc mantle formed by the subduction of the Neotethys beneath the Sakarya continent, because the serpentined peridotites occur within the Sakarya continent (Figs. 2, 3, 6, and 7 of Yilmaz et al., 1995). If this is the case, its dehydration due to warming in recent times after the cessation of subduction of the Neotethys might have induced the strike–slip motion of the NAF. This, however, remains speculative at present, and evidence for this should be sought from geological data in this region.

The above mechanism may also allow strike–slip faulting to occur at low stresses far into continental interiors. There are many geological examples of multiple suture zones in East Asia and North America, where fragments of continents and island-arc were amalgamated during the Mesozoic–early Tertiary (e.g. Ben-Avraham et al., 1981; Klimetz, 1983; Silver and Smith, 1983; Davis and Pfafer, 1986; Scholl et al., 1986). These terrane collisions and accretions, in association with the seaward stepping of trenches, suggest the possible existence of serpentined forearc mantle in such suture zones. Some of them may have survived and mobilized huge continental strike–slip faults by later thermal events. The strike–slip Denali fault north of the Alaska Trench and the Altn Tagh, Kunlun, and Red River faults in interior Asia might be such examples, although further investigations on their origins are necessary.

Ridge–ridge transform faults developing in mid-oceans following continental breakup or trench–trench transform faults along fracture zones may not be relevant to the dehydration weakening proposed in this study. In these transform faults, infiltration of seawater along the faults may form serpentine (Calvert and Potts, 1985; Macdonald and Fyfe, 1985; Potts et al., 1986), whose low amount of friction (Reinen et al., 1991; Moore et al., 1997) may play an important role in reducing the shear strength of those faults (e.g. Escartin et al., 1997). Even for ridge–ridge transform faults in their initiation stage, however, there may be embayments within the serpentined forearc mantle, because continental breakup often occurs along suture zones (e.g. Burke et al., 2003). Trench–trench transform faults may also originate from arcs that might contain serpentinated forearc mantle. In such cases, dehydration of the warming serpentined forearc mantle may provide an important clue to elucidate the evolution of this type of transform fault.

4. Discussion

Initiation of plate tectonics by the formation of plate boundaries is a difficult, but nevertheless important, problem in Earth sciences. The difficulty arises from two contradictory aspects of plate tectonics: a plate (lithosphere) must be strong and, at the same time, it must be mobile in order to extract heat efficiently from the Earth's interior. Moving plates require weak plate boundaries, but the strength of plates makes the formation of these boundaries difficult. The thick crust formed at the Earth's surface also prevents subduction. The problem is diverse and vast, covering chemical alteration, rheology of the surface and interior of the Earth, and its tectonic history. As means to address this diverse problem, we propose that volatilization of an underthrusting hydrated/carbonated slab and dehydration of serpentined forearc mantle are major agents for strain localization and weakening of lithosphere to initiate plate boundary formation.

Although previous studies have discussed various weakening mechanisms of lithosphere, including the effects of volatiles (Bercovici, 1998; Moresi and Solomatov, 1998; Tackley, 1998; Regenauer-Lieb et al., 2001; Bercovici, 2003; O'Neill et al., 2007), their treatments are all general. In this paper, we provide a practical hypothesis on the roles of volatiles in the formation processes of new plate boundaries such as initiation of subduction, rifting/spreading, and transform faults. Because the mechanisms we propose are put in the context of the Earth's history,
they are more easily tested than the mechanisms proposed in previous studies. Obviously, however, we present only a rough sketch of the mechanisms in this paper, and do not claim to have proven them to be correct. A lot of work remains to be done to examine these mechanisms quantitatively based on geological or geophysical data in each region.

Sleep (2000) presented a scheme of transition between a magma ocean, plate tectonics, and stagnant-lid convection for terrestrial planets. In some planets, transitions between these modes may have occurred several times. In this scheme, however, mechanisms of the transition to plate tectonics, except for the role of partial melting in the divergent zone, were not presented. In the future, the Earth will transition to stagnant-lid convection as it is cooled efficiently by plate tectonics. After this, it is expected that radioactive heat production will heat up the interior, and plate tectonics will resume (Sleep, 2000). For this to occur, however, mechanisms that enable the initiation of subduction (similar to those presented in this study) are necessary; otherwise the Earth will return to a magma ocean. It is noted that both the convective vigor and volatile extent of the Earth's interior change temporally, interacting with each other in a complicated manner. There is no guarantee that the condition for resumption of plate tectonics will be satisfied in the future. The weakening mechanisms we propose will, nevertheless, provide a useful key to conjecture about shifts between the modes of convection for terrestrial planets.

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**Fig. 5.** Examples of serpentinite bodies in non-volcanic rifted margins. (a) Map of the Labrador Sea, showing the zones of high-velocity (7.2–7.7 km/s) material on each margin (shaded), which is inferred to be serpentinitized peridotite (Chian et al., 1995). The grey lines are fracture zones and the solid lines are the lines of seismic experiments. (b) Map of the region offshore of the western continental margin of Iberia (contours of ocean depth in meters), showing zones of thin crust underlain by serpentinitized peridotite (shaded area, Whitmarsh et al., 1993). A.P denotes abyssal plain. The solid lines are the lines of seismic experiments. These maps show that the serpentinitized peridotites are located at the ocean–continent boundaries of the non-volcanic rifted margins, not at the center of the abyssal plains.

**Fig. 6.** (a) Tectonic setting of the California Coast Ranges: the San Andreas fault system (SAF) and various terranes (stippled area) (after McCulloch, 1989). The location of the SAF is 100–200 km landward of the relict trench in which the Farallon plate was subducting during the Tertiary. (b) The Farallon plate (FA) vanished beneath the North American plate (NA), and the contact of the Pacific plate (PA) to NA became the SAF along the relict trench, according to the theory of Atwater (1970). (c) Dehydration from the warming serpentinitized forearc mantle occurs (Kirby et al., 2003, 2013) at the boundaries of the wedge. The SAF could initiate 100–200 km landward of the relict trench along such a boundary. The forearc wedge behaves as part of PA after the formation of the SAF.
Our hypothesis also has implications for the formation and exhumation of high pressure (HP)/ultrahigh pressure (UHP) metamorphic rocks and emplacement of some ophiolites and ultramafic rocks in subduction zones. Belts of HP or UHP metamorphic rocks are found in forearcs of active subduction zones and collision zones. Exposure on the Earth’s surface of such coesite- and diamond-bearing HP or UHP rocks implies exhumation from a source in the mantle deeper than 100–150 km (Coleman and Wang, 1995; Ernest and Liou, 1999; Chopin, 2003). Although the origin of these rocks is still disputed (Brueckner, 1998; Terry and Robinson, 1999; Guillot et al., 2001), it is very likely that their exhumation involves tectonic movements not only at the mid-lower crustal level but also deep in the mantle. It is also episodic, i.e., not a steady-state process. In many cases, HP/UHP belts are sandwiched between lower metamorphic-grade units in fault contact (Worrall, 1981; Burg et al., 1984; Wheeler, 1991; Kaneko et al., 2003), which implies that they were brought to the surface largely by faulting or strain localization (Burchfiel and Rodeney, 1985; Maruyama et al., 1996). These features, i.e., episodicity, strain localization, and mantle-scale processes, are common to those of plate boundary formation discussed in this study, suggesting that a mechanism similar to dehydration weakening and mobilization of huge thrusts in relic subduction or collision zones might be operating in exhumation of UHP/HP rocks. Subduction of continental materials into the deep mantle may provide sources for UHP rocks via the process we propose for subduction of lithosphere with thick crust. This subject, however, will be treated in a separate paper.

5. Conclusions

In the early Earth, convection would have occurred with thick basaltic crust accumulating over a soft boundary layer downwelling into the mantle (Davies, 1992). This might have transitioned to stagnant-lid convection as the mantle cooled (Solomatov and Moresi, 1997) or back to a magma ocean as the mantle heated (Sleep, 2000). For plate tectonics to operate on the Earth, subduction of lithosphere with thick crust must have been initiated, thus avoiding these shifts. Based on an analogy with the continental plate subduction that is currently occurring in Hindu Kush and Burma, we propose that volatilization from underthrusting lithosphere, once hydrated/carbonated by H₂O–CO₂ rich vapors released from magmas in upwelling plumes, would have made it possible. Once subduction has been initiated, serpentinitized forearc mantle may form in a wedge-shaped body above a dehydrating slab. Therefore, in relic arcs, suture zones, or rifted margins, any agent that warms and dehydrates the serpentinitized mantle would weaken the region surrounding it, and form various types of plate boundaries depending on the operating tectonic stress.

Thus, once subduction initiates, formation of plate boundaries might be facilitated by a single fundamental process (except for the MOR formed by partial melting in the divergent zone): weakening due to the release of pressurized water from warming serpentinitized forearc mantle. The fundamental ideas presented in this paper are simple, and details of geological and geophysical facts in each appropriate region should be examined more quantitatively. However, we feel that the insights that we have obtained from a preliminary application of these concepts will have important implications for interpreting geological histories in mobile belts, and are sufficiently encouraging to report and put them into other people’s discussion.

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