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Review

Supercontinent–superplume coupling, true polar wander and plume mobility: Plate dominance in whole-mantle tectonics

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ABSTRACT

Seismic tomography has illustrated convincingly the whole-mantle nature of mantle convection, and the lower mantle origin of the African and Pacific superplumes. However, questions remain regarding how tectonic plates, mantle superplumes and the convective mantle interplay with each other. Is the formation of mantle superplumes related to plate dynamics? Are mantle plumes and superplumes fixed relative to the Earth's rotation axis? Answers to these questions are fundamental for our understanding of the inner workings of the Earth's dynamic system. In this paper we review recent progresses in relevant fields and suggest that the Earth's history may have been dominated by cycles of supercontinent assembly and breakup, accompanied by superplume events. It has been speculated that circum–supercontinent subduction leads to the formation of antipodal superplumes corresponding to the positions of the supercontinents. The superplumes could bring themselves and the coupled supercontinents to equatorial positions through true polar wander events, and eventually lead to the breakup of the supercontinents.
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Contents

1. Introduction

The plate tectonic theory developed nearly half a century ago enables us to see the Earth as a dynamic planet, with tectonic plates colliding to form mountain belts and breaking apart to create new oceans. Plate tectonics is an integral part of convective processes in the Earth's mantle with tectonic plates as the top thermal boundary layers (Davies, 1999). Popular mechanisms for driving plate motion include oceanic ridge push, slab pull, and dragging force of the convective mantle (Forsyth and Uyeda, 1975), but recent work places greater emphasis on slab pull (Hager and Oconnell, 1981; Ricard et al., 1993; Zhong and Gurnis, 1995; Lithgow-Bertelloni and Richards,

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1998; Becker et al., 1999; Conrad and Lithgow-Bertelloni, 2004) and slab suction (Conrad and Lithgow-Bertelloni, 2004). Increasingly higher resolution seismic tomographic images since the 1990s show that subducted slabs, after firstly stagnating at the mantle transition zone, can go down all the way to the core–mantle boundary (CMB) (e.g., Van der Hilst et al., 1997; Van der Voo et al., 1999; Fukao et al., 2001). Recognizing the importance of slab subduction in driving mantle convection and plate motion, Anderson (2001) coined the term "top-down tectonics" though his model is limited to the upper mantle.

A relevant debate is about whether mantle plumes exist and the nature of such plumes (e.g., Anderson and Natland, 2005; Davies, 2005; Campbell and Kerr, 2007). One of the main arguments against the plume hypothesis is that there is no material exchange between the upper and the lower mantle, and hot spots are upper mantle features only (e.g., Anderson and Natland, 2005). However, seismic topography has convincingly proved that subduction does go all the way to the CMB (e.g., Van der Hilst et al., 1997), and that at least some plumes originate from the lower mantle (e.g., Nolet et al., 2007). Seismic tomography also showed us that there are presently two dominating low-velocity structures in the lower mantle below Africa and the Pacific, commonly known as the African superplume and the Pacific superplume, respectively (e.g., Dziewonski, 1984; Ritsema et al., 1999; Masters et al., 2000; Zhao, 2001; Romanowicz and Gung, 2002; Romanowicz, 2008) (Fig. 1a). The African and Pacific superplume regions are also associated with anomalous topographic highs or superswells (McNutt and Judge, 1990; Davies and Pribac, 1993; Nyblade and Robinson, 1994; Lithgow-Bertelloni and Silver, 1998), positive geoid anomalies (Anderson, 1982; Hager et al., 1985; Hager and Richards, 1989), the majority of hotspot volcanism (Anderson, 1982; Hager et al., 1985; Duncan and Richards, 1991; Courtillot et al., 2003; Jellinek and Manga, 2004) and large igneous provinces (Burke and Torsvik, 2004; Burke et al., 2008).

Davies and Richards (1992) summarized the dynamics of tectonic plates and mantle plumes as two distinct modes of mantle convection: plate mode and plume mode. While plate mode convection is evidently essential in the geodynamic system, the nature of plume mode convection and its relation to the plate mode are more uncertain. Plumes and superplumes are considered by some as spontaneous instabilities of thermal boundary layers from the CMB, as such, plume or superplume activities are independent factors in the Earth's geodynamic system (e.g., Hill et al., 1992; Davaille, 1999; Campbell and Kerr, 2007). Assuming their relative stability in relation to the mantle and the lithosphere, mantle plumes (hot spots) have thus commonly been used as a relatively stable reference frame for estimating plate movements (e.g., Besse and Courtillot, 2002; Steinberger and Torsvik, 2008). Morgan and others (e.g., Morgan, 1971; Storey, 1995) proposed that the line-up of hotspots (representing mantle plumes) "produced currents in the asthenosphere which caused the continental breakup leading to the formation of the Atlantic" (Morgan, 1971), thus pointing to a more active and driving role of plumes in plate dynamics. On the other hand, tectonic plates are suggested to have important controls on the dynamics of mantle plumes including their origin, location, and evolution (e.g., Molnar and Stock, 1987; Lenardic and Kaula, 1994; Steinberger and O'Connell, 1998; Zhong et al., 2000; Tan et al., 2002; Gonnermann et al., 2004).

In this paper we will review some recent findings and ideas relating to global geodynamics, particularly in terms of supercontinent cycles, supercontinent–superplume coupling, true polar wander events related to superplume events, and speculate on interplay mechanisms between global plate dynamics and superplume activities. Extensive reviews of mantle plumes can be found in Jellinek and Manga (2004), Davies (2005), Sleep (2006), and Campbell and Kerr (2007). Here we focus on the relationships between mantle plumes, superplumes, and supercontinent events throughout geological history.

2. Supercontinent cycles and supercontinent–superplume coupling

2.1. What are superplumes?

Mantle plumes are thought to result from thermal boundary layer instabilities at the base of the mantle (Morgan, 1971; Griffiths and Campbell, 1990). Mantle plumes were first proposed to account for hotspot volcanism such as in Hawaii (Wilson, 1963; Morgan, 1971). It was also suggested that plume heads cause flood basalts or large igneous provinces (i.e., LIPs) (Morgan, 1981; Duncan and Richards, 1991; Richards et al., 1991; Hill et al., 1992). A fully developed mantle plume is suggested to have a diameter of several 100 km or less and an excess temperature of $<\sim$300 K in the upper mantle (Loper and Stacey, 1983; Griffiths and Campbell, 1990; Davies, 1999; Zhong and Watts, 2002). Mantle plumes may be responsible for a few Terawatts of heat transfer in the mantle (Davies, 1988; Sleep, 1990). In addition, mantle plumes may also play an important role in cooling the Earth's core (Davies, 1988; Sleep, 1990). Again, we refer readers to Jellinek and Manga (2004), Davies (2005), Sleep (2006), and Campbell and Kerr (2007) for more extensive reviews on mantle plumes.

Mesozoic-Cenozoic hotspot volcanism (i.e., the surface manifestation of mantle plumes) and LIPs preferentially occur in the two major seismically slow regions away from subduction zones (i.e., Africa and the central Pacific) (Anderson, 1982; Hager et al., 1985; Weinstein and Olson, 1989; Duncan and Richards, 1991; Romanowicz and Gung, 2002; Courtillot et al., 2003; Burke and Torsvik, 2004; Burke et al., 2008). This suggests a close association between mantle plumes and plate tectonics, contrary to an earlier view that mantle plumes operate largely independent of plate tectonic processes (e.g., Hill et al., 1992). The large-scale (>6000 km) seismically slow anomalies below Africa and the Pacific led to suggestions of African and Pacific superplumes (e.g., Romanowicz and Gung, 2002). However, seismic studies also suggest that the African and Pacific seismic anomalies or superplumes have not only thermal but also chemical/compositional origins at least near the CMB (Su and Dziewonski, 1997; Masters et al., 2000; Wen et al., 2001; Ni et al., 2002; Wang and Wen, 2004; Garnero et al., 2007).

Other evidence also suggests that plume and plate modes of convection are related. It has long been recognized that plate tectonic motion affects motion of mantle plumes (Molnar and Stock, 1987; Steinberger and O'Connell, 1998; Gonnermann et al., 2004). Subducted slabs, by cooling the mantle and inducing sub-adiabatic temperatures in the lower mantle, promote plume formation at the CMB (Lenardic and Kaula, 1994). Subducted slabs may also cause plumes to preferentially form near the slabs above the CMB (Tan et al., 2002). However, most early geodynamic models were done in simplified geometries of a 2-D or 3-D box, and the geometrical constraints of the models tend to force plumes or a sheet of hot upwelling to be located below the spreading centers, making it difficult to use these models to study the dynamics of multiple plumes or superplumes. The relation between mantle plumes and superplumes in the framework of plate tectonics has only been explored recently in 3-D models with more realistic plate configurations with two views that are not necessarily mutually exclusive to each other.

In the first view, a superplume region represents a cluster of mantle plumes originated from the CMB (Schubert et al., 2004) or a thermochemical pile at the CMB (Davaille, 1999; Kellogg et al., 1999; McNamara and Zhong, 2005a; Tan and Gurnis, 2005; Bull et al., 2009). Based on global mantle convection models with tectonic plates, Zhong et al. (2000) found that mantle plumes tend to

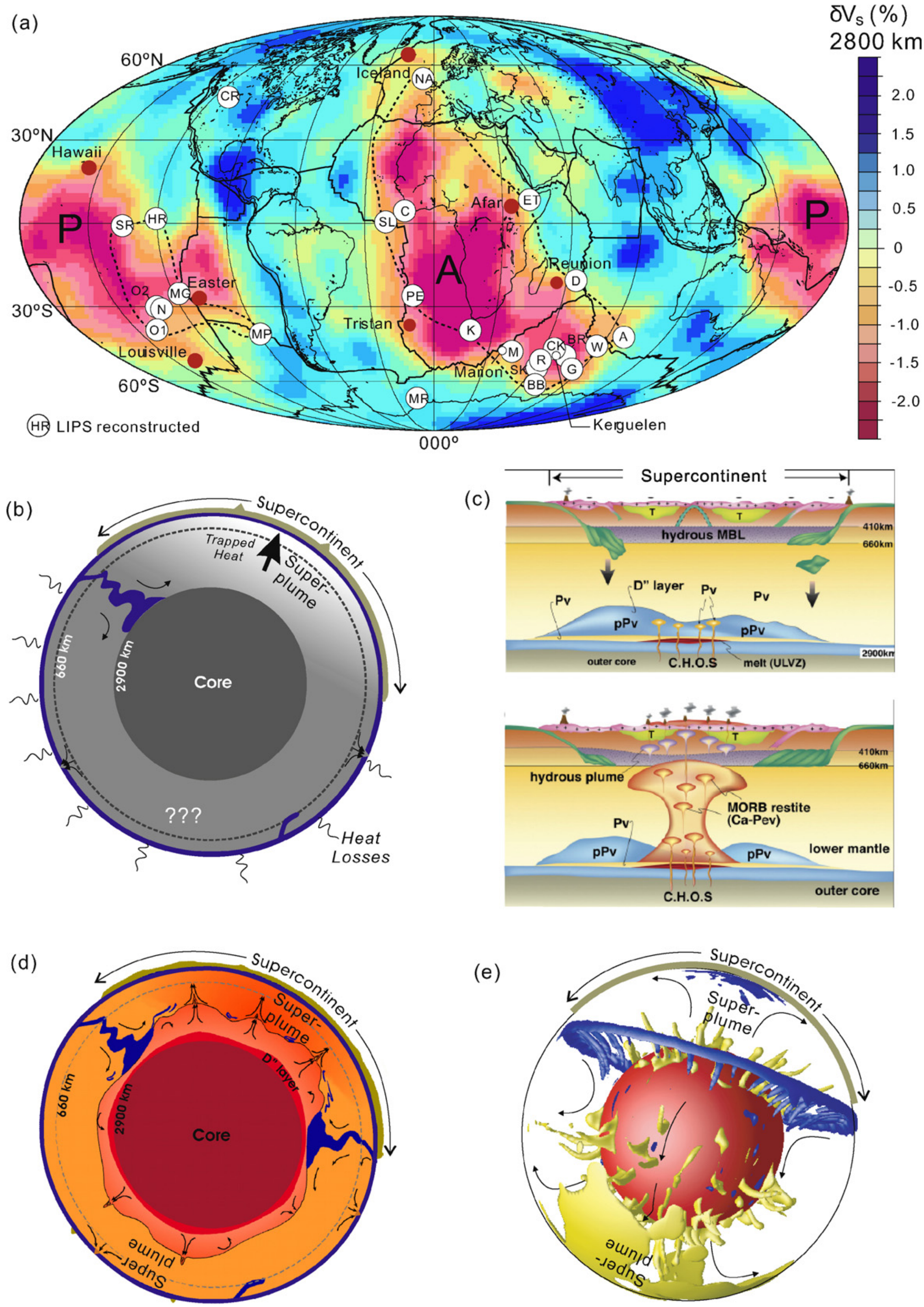

Fig. 1. (a) SMEAN shear wave velocity anomalies near the core–mantle boundary (Becker and Boschi, 2002), illustrating the location and lateral extent of the present African (A) and Pacific (P) superplumes. White circles show 201–15 Ma large igneous provinces restored to where they were formed (Burke and Torsvik, 2004) with letters referring to the names of the large igneous provinces: C, CAMP; K, Karroo; A, Argo margin; SR, Shatsky Rise; MG, Magellan Rise; G, Gascoyne; PE, Parana–Etendeka; BB, Banbury Baslats; MP, Manihiki Plateau; O1, Ontong Java 1; R, Rajmahal Traps; SK, Southern Kerguelen; N, Nauru; CK, Central Kerguelen; HR, Hess Rise; W, Wallaby Plateau; BR, Broken Ridge; O2, Ontong Java 2; M, Madagascar; SL, S. Leone Rise; MR, Maud Rise; D, Deccan Traps; NA, North Atlantic; ET, Ethiopia; CR, Columbia River. Red dots show hot spots regarded as having a deep origin by Courtillot et al. (2003). The figure is modified from Burke and Torsvik (2004). Figures (b)–(e) are cartoons showing mechanisms for the formation of mantle superplumes, with (b) being the thermal insulation model (e.g., Anderson, 1982; Coltice et al., 2007; Evans, 2003b; Gurnis, 1988; Zhong and Gurnis, 1993), (c) the supercontinent slab graveyard-turned "fuel" model (e.g., Maruyama et al., 2007), (d) the circum-supercontinent slab avalanche model (Li et al., 2004, 2008; Maruyama, 1994), and (e) degree-2 planform mantle convection model with sub-supercontinent return flow in response to circum–supercontinent subduction (Zhong et al., 2007). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of the article.)

form in stagnation regions of the CMB that are commonly below the central regions of major tectonic plates and are controlled by the distribution of subducted slabs, consistent with the statistical analysis of hot spot distribution (Weinstein and Olson, 1989) and seismic tomography imaging (e.g., Nolet et al., 2007). Schubert et al. (2004) further suggested that superplumes may be just clusters of mantle plumes organized together by plate motion. McNamara and Zhong (2005a) showed that the African and Pacific seismically slow anomalies at the CMB are better explained as thermochemical piles organized by plate motion history in the last 120 Ma. This is confirmed by Bull et al. (2009) who, using more rigorous comparisons between geodynamic model predictions and seismic models at the same resolution level, showed that the African and Pacific anomalies near the CMB are better explained as thermochemical piles than clusters of purely thermal plumes.

In the second view, the mantle below major tectonic plates may be hot due to incomplete homogenization of its thermal state (Davies, 1999; King et al., 2002; Huang and Zhong, 2005; Hoink and Lenardic, 2008) or due to thermal insulation by the tectonic plates themselves (Anderson, 1982; Gurnis, 1988; Zhong and Gurnis, 1993; Lowman and Jarvis, 1995, 1996). This may lead to broad-scale seismically slow anomalies, especially at relatively shallow depths, resembling the African and Pacific anomalies. However, localized occurrence of hotspot volcanism and rapidly forming and short-lived LIPs may still require mantle plumes and thermal boundary instabilities near the CMB, and the broad scales of the African and Pacific seismic anomalies above the CMB may still require the presence of thermochemical piles (Bull et al., 2009). It should be pointed out that these two different processes (i.e., the incomplete heat homogenization and generation of mantle plumes above the thermochemical piles) may occur simultaneously in the Earth's mantle (e.g., Davies, 1999; Jellinek and Manga, 2004).

Therefore, the African and Pacific superplumes are best characterized by clusters of mantle plumes originating from or near broad-scale thermochemical piles immediately above the CMB (Davaille, 1999; Torsvik et al., 2006). The locations of these superplumes, including the thermochemical piles are controlled by subduction zones (McNamara and Zhong, 2005a). Thermochemical piles may have important influences on plume dynamics (Jellinek and Manga, 2002). However, in our characterization of superplumes, thermochemical piles are only demanded from seismic observations of the CMB regions (Masters et al., 2000; Wen et al., 2001; Ni et al., 2002; Wang and Wen, 2004). Surface expressions of the superplumes may include anomalous topographic highs or superswells (McNutt and Judge, 1990; Davies and Pribac, 1993; Nyblade and Robinson, 1994), positive geoid anomalies (Anderson, 1982; Hager et al., 1985), hotspot volcanism (Anderson, 1982; Hager et al., 1985; Duncan and Richards, 1991; Courtillot et al., 2003; Jellinek and Manga, 2004) and large igneous provinces (Larson, 1991b; Burke and Torsvik, 2004; Burke et al., 2008).

Four different mechanisms for formation of superplumes (or clusters of mantle plumes) have been proposed, all of which are somewhat related to the dynamics of tectonic plates, contrary to the early idea that plumes, entirely as products of thermal instabilities at or close to CMB, are independent of plate dynamics (e.g., Hill et al., 1992).

(1) Superplumes form by thermal insulation of supercontinents (e.g., Anderson, 1982; Gurnis, 1988; Zhong and Gurnis, 1993; Evans, 2003b; Coltice et al., 2007) (Fig. 1b). Major drawbacks of this model include insufficient heat build-up for a superplume (Korenaga, 2007) and the formation of the Pacific superplume without the insulation of a supercontinent (e.g., Zhong et al., 2007).
(2) Melting of slab graveyard (the "fuel") underneath a newly formed supercontinent by heat conducted from the core (e.g., Maruyama et al., 2007) (Fig. 1c). Authors of this model suggest that the Pacific superplume is the residual of the superplume formed beneath Rodinia at ca. 800 Ma. However, this does not explain why such an aged superplume is going as strong as the much younger African (Pangean) superplume, why there is no active superplume inherited from older supercontinents such as Columbia (Rogers and Santosh, 2002; Zhao et al., 2004), and the antipodal nature of the African and Pacific superplumes.
(3) Pushup effects of circum–supercontinent slab avalanches (Maruyama, 1994; Li et al., 2004; Li et al., 2008) (Fig. 1d). This mechanism, emphasizing the dominant role of slab avalanches (e.g., Tackley et al., 1993) in the formation of superplumes, explains the antipodal distribution of the present Pacific and African superplumes (as residuals of the antipodal Paleo-Pacific and Pangean superplumes) without necessarily involving much of the lower mantle materials in the whole-mantle convection. It is in line with the "lava lamp" layered mantle model (Kellogg et al., 1999).
(4) Dynamically self-consistent formation of degree-1 planform (where there is a major upwelling system in one hemisphere and a major downwelling system in the other hemisphere) or degree-2 planform (where there are two major, antipodal upwelling systems) mantle convection during supercontinent cycles (Zhong et al., 2007) (Fig. 1e). A major difference of this model from that of model (3) is that in this model the lower mantle is very much involved in the convection, and that the superplume in the oceanic realm may have been active before the formation of the sub-supercontinent superplume.

A common feature in all these models is the coupling of superplumes with supercontinent cycles. Is such an assumption born out by the Earth's geodynamic record? In the sections below we will briefly review superplume events throughout geological history and their possible links to supercontinent cycles.

2.2. Superplume record during the Pangean cycle

Pangea is by far the best known supercontinent that encompassed almost all known continents on Earth (Wegener, 1966). The bulk of Pangea formed through the collision of Laurentia (North America with Greenland) and Gondwanaland at around 320 Ma, joined by the Siberian craton and central Asian terranes by the Early Permian (e.g., Li and Powell, 2001; Scotese, 2004; Torsvik and Cocks, 2004; Veevers, 2004). It is important to note that Pangea was largely surrounded by subduction zones (i.e., circum–supercontinent subduction) (Fig. 2), a feature that seems to be true for other supercontinents and is likely to have important geodynamic implications. Pangea started to break up at ~185 Ma (Veevers, 2004).

It has been widely accepted that the present African superplume started in the Mesozoic or earlier beneath the supercontinent Pangea (e.g., Anderson, 1982; Burke and Torsvik, 2004). Plume breakout started no later than ca. 200 Ma (i.e., the Central Altantic Magmatic Province; Marzoli et al., 1999; Hames et al., 2000), and possibly earlier (Veevers and Tewari, 1995; Doblas et al., 1998; Torsvik and Cocks, 2004) with documented plume records include the ca. 250 Ma Siberian traps (e.g., Renne and Basu, 1991; Courtillot and Renne, 2003; Reichow et al., 2008), the ca. 260 Ma Emeishan flood basalts (e.g., Chung and Jahn, 1995; He et al., 2007), the ca. 275 Ma Bachu LIP (Zhang et al., 2008), and the ca. 300 Ma Skagerrak-Centered LIP in NW Europe (Torsvik et al., 2008a). The time lag of ca. 20–120 My between the assembly of Pangea by ca. 320 Ma (e.g., Veevers, 2004) and the starting time of the Pangean (African) superplume has important implications for the dynamics of supercontinents and superplumes. Plumes related to the Pangean superplume, either primary or secondary (Courtillot et al., 2003),

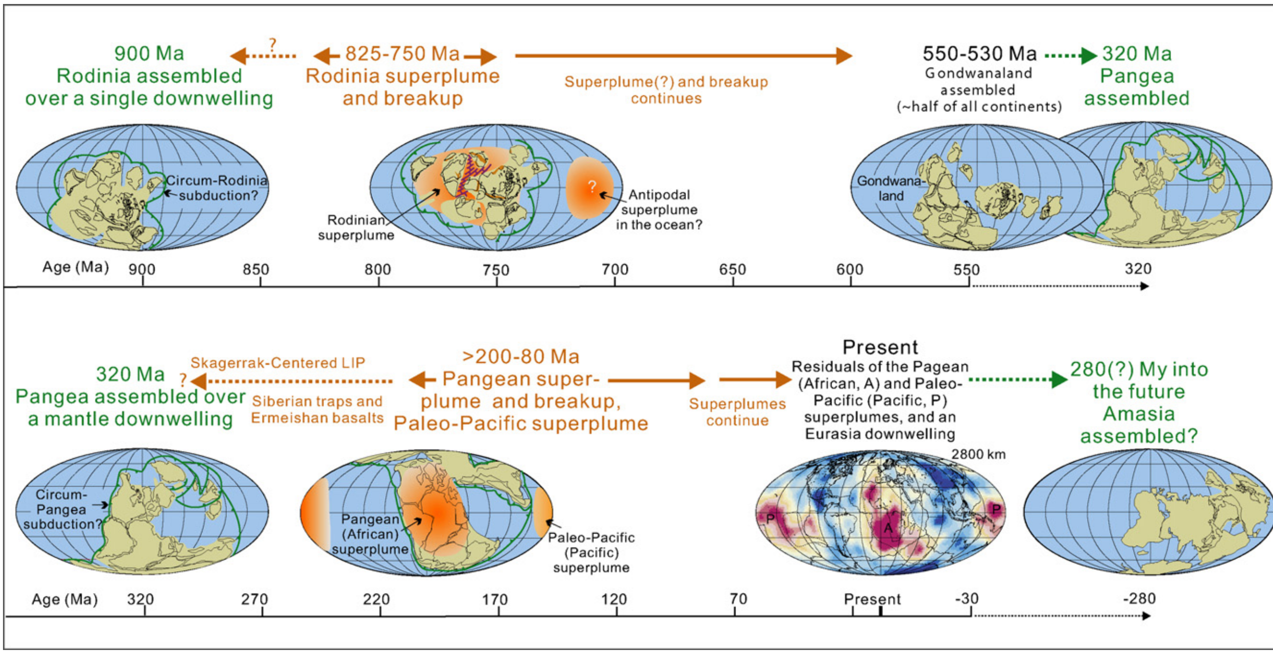

Fig. 2. Illustration of supercontinent–superplume coupling in the past 1000 Ma, and the possible presence of a ∼600 My supercontinent–superplume cycle. Sources for (paleo-) geographic maps: 900–320 Ma (Li et al., 1993, 2008), 200 Ma (Scotese, 2004), present with SMEAN shear wave velocity anomalies near the core–mantle boundary (Becker and Boschi, 2002; Burke and Torsvik, 2004), and 280 My into the future (S. Pisarevsky, personal communication).

are believed by some to have caused the breakup of Pangea (e.g., Morgan, 1971; Storey, 1995).

The antipodal Paleo-Pacific superplume started no later than ca. 125 Ma (e.g., Larson, 1991b). Due to the lack of pre-170 Ma oceanic record, the starting time for the Pacific superplume is difficult to determine directly. However, there are geological observations interpreted to indicate the Triassic subduction of a >250 Ma oceanic plateau along southeastern South China, reflecting >250 Ma plume activities in the Paleo-Pacific realm (Li and Li, 2007). In addition, if Torsvik et al.'s (2008b) model, in which South China is placed above the Paleo-Pacific superplume, is correct, the starting time of the Paleo-Pacific superplume could be as early as ca. 260 Ma as suggested by the Emeishan LIP. It is therefore possible that the starting times for the African and Pacific superplumes are comparable.

2.3. Superplume record during the Rodinian cycle

Until the 1990s, the only well-accepted supercontinent consisting of almost all known continents was the Pangea supercontinent. The well-known "supercontinent" Gondwanaland (or Gondwana), existed since the Cambrian (ca. 530 Ma; e.g., Meert and Van Der Voo, 1997; Li and Powell, 2001; Collins and Pisarevsky, 2005) until its amalgamation with Laurentia in the mid-Carboniferous to become part of Pangea, consisted of little more than half of the known continents and is thus not a supercontinent of the same magnitude as Pangea.

Landmark work in 1991 (Dalziel, 1991; Hoffman, 1991; Moores, 1991) led to a wide recognition of the possible existence of a Pangea-sized supercontinent Rodinia in the late Precambrian (McMenamin and McMenamin, 1990). Although most earlier workers believed Rodinia existed before 1000 Ma, subsequent work demonstrated that the assembly of Rodinia was likely not completed until ca. 900 Ma (see review by Li et al., 2008 and references therein). The precise configuration of Rodinia is still controversial (e.g., Hoffman, 1991; Weil et al., 1998; Pisarevsky et al., 2003; Torsvik, 2003; Li et al., 2008). Here we adapt the IGCP 440 configuration as in Li et al. (2008) (Figs. 2 and 3).

Like the case for Pangea, a mantle superplume has also been invoked to account for a series of tectono-thermal events in the lead-up to the breakup of Rodinia by ca. 750 Ma (see Li et al., 2008 and references therein; Figs. 2 and 3). Key evidence for plume activities includes continental-scale syn-magmatic doming in an anorogenic setting that was followed closely by rifting (e.g., Li et al., 1999, 2002, 2003b; Wang and Li, 2003), radiating mafic dyke swarms and remnants of large igneous provinces (Zhao et al., 1994; Fetter and Goldberg, 1995; Park et al., 1995; Wingate et al., 1998; Frimmel et al., 2001; Harlan et al., 2003; Li et al., 2008; Ernst et al., 2008; Wang et al., 2008), widespread and largely synchronous anorogenic magmatism that requires a large and sustained heat source for a region >6000 km in diameter over 100 My (e.g., Li et al., 2003a, 2003b), and petrological evidence for >1500 °C mantle melts accompanying some of the large igneous events (Wang et al., 2007). The time lag between the final assembly of Rodinia by ca. 900 Ma and the first major episode of plume breakout at ca. 825 Ma was ca. 75 My. Li et al. (2003b, 2008) demonstrated that there were two broad peaks of plume-induced magmatism before the breakup of Rodinia: one at ca. 825–800 Ma, and the other at ca. 780–750 Ma. Plume activities likely continued during the protracted process of Rodinia breakup (e.g., the ca. 720 Ma Franklin igneous event; Heaman et al., 1992; Li et al., 2008). We note that the plume interpretation for some of the Neoproterozoic magmatism is not universally accepted (e.g., Munteanu and Wilson, 2008).

2.4. Supercontinent–superplume coupling and geodynamic implications

As shown in Fig. 2, the two better known supercontinents since 1000 Ma, Pangea and Rodinia, experienced parallel events in terms of the development of a superplume beneath each supercontinent: (1) In each case a superplume appeared under the supercontinent (i.e. widespread plume breakout over the supercontinent) some 20–120 My after the completion of the supercontinent assembly; (2) The sizes of both the Rodinian and the Pangean superplumes are >6000 km in dimension; (3) Each superplume lasted at least a few hundred million years with peak activities lasting ∼100 My

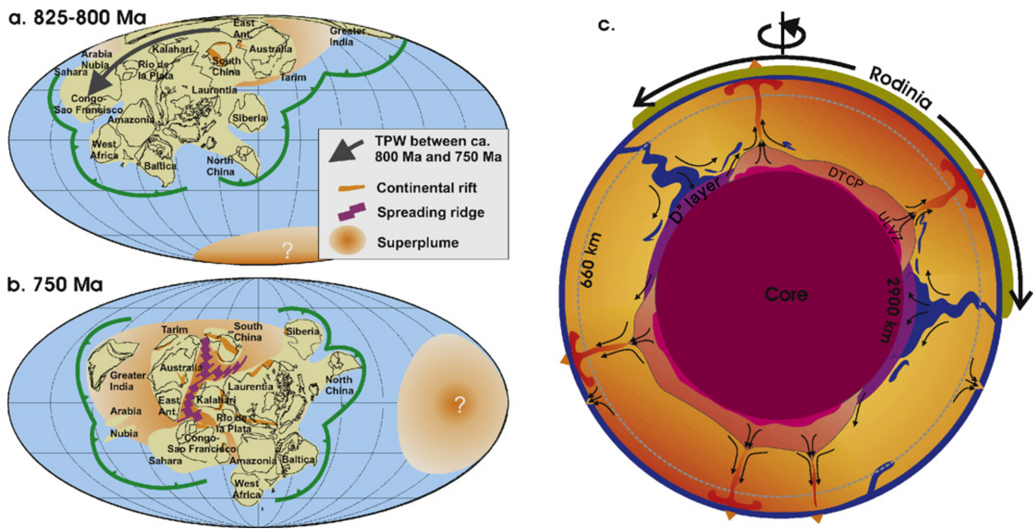

Fig. 3. An ~90° true polar wander (TPW) event between (a) 825–800 Ma and (b) 750 Ma, and (c) a schematic geodynamic model indicating circum–supercontinent subduction causing the formation of superplumes and plumes (after Li et al., 2004, 2008). DTCP = dense thermal–chemical piles; ULVZ = ultra-low velocity zone.

(from >825 Ma to <750 Ma for the Rodinian superplume and from >200 Ma to <80 Ma for the Pangean and Pacific superplumes); (4) Both superplumes are thought to have led to the breakup of the supercontinents. We no longer have an in situ geological record for a possible superplume antipodal to the Rodinian superplume in the middle of the pan-Rodinia ocean, as we have for the Paleo-Pacific superplume (e.g., Larson, 1991a) antipodal to the Pangean (the present African) superplume (Fig. 2), but any remnant of oceanic plateaus or sea-mounts produced by such a superplume, or record of their subduction, would be located in late-Neoproterozoic to Early Paleozoic orogens (i.e., evidence for subduction and delamination of oceanic plateaus; Coney and Reynolds, 1977; Li and Li, 2007).

The time delay between the final assembly of the supercontinents and the appearances of the superplumes under the supercontinents hints at a possible causal relationship between the formation of supercontinents and the generation of superplumes (see Sections 2.1, 3 and 4 regarding possible mechanisms). Is there a geological record for similar coupling of supercontinent–superplume events in the pre-1000 Ma geological history?

A number of older supercontinents have been proposed for geological time before Rodinia, but there is so far little consensus regarding the existence of such supercontinents or their configurations. There is nonetheless a consistent correlation between proposed supercontinent cycles and intensity changes in mantle plume activities (Fig. 4). The hypothesized Pangea-sized supercontinent immediately before Rodinia has variously been called Nuna (Evans, 2003a; Hoffman, 1997; Bleeker, 2003; Brown, 2008) and Columbia (Rogers and Santosh, 2002; Zhao et al., 2004). Although geologists argue for the existence of Nuna (Columbia) between 1.8 and 1.3 Ga (Zhao et al., 2004), present paleomagnetic data indicate a <1.77 Ga formation age (Meert, 2002). The breakup of Nuna (Columbia) is also believed to have been linked to a flare up of plume activities (Zhao et al., 2004; Ernst et al., 2008).

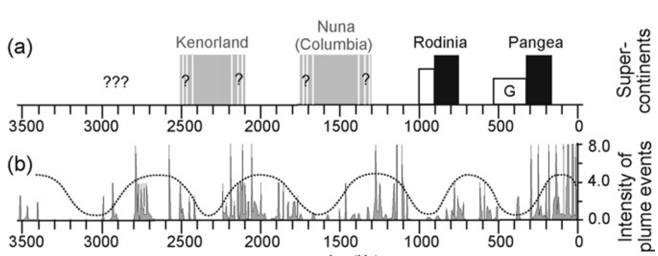

Fig. 4. (a) Time distribution of supercontinents in Earth's history and (b) probability plot of 154 large igneous province events since 3500 Ma (AB50 data; Prokoph et al., 2004) demonstrating the possible cyclic nature of mantle plume activity with ca. 750–550 My wavelengths (Prokoph et al., 2004). The intensified plume activities in the Mesozoic–Cenozoic partly reflect the preservation of oceanic plateaus since the mid-Mesozoic. G = Gondwanaland.

An even more speculative supercontinent, Kenorland (Williams et al., 1991), is supposed to have existed from >2.5 Ga until 2.1 Ga (Aspler and Chiarenzelli, 1998), and its breakup has also been linked with widespread plume activities (Heaman, 1997; Aspler and Chiarenzelli, 1998).

Fig. 4 shows the time distributions of both known/speculated supercontinents (a), and proxy of mantle plume events (e.g., LIPs, dyke swarms etc.) as compiled by Prokoph et al. (2004) (b). Prokoph et al. (2004) reported the presence of a 730–550 My periodicities in mantle plume activities since Archean time (Fig. 4b). Remarkably, such periodicities appear to largely mimic that of supercontinent cycles beyond the above-discussed Pangean and Rodinian cycles (Fig. 4a). In all cases the new cycles of plume activity appear to start during the lifespan of the corresponding supercontinents, whereas peaks of plume activity appear to coincide with the breakup of the supercontinents. We note that the suggested lifespan for both Nuna (Columbia) and Kenorland are significantly longer than that of Pangea and Rodinia. The starting times of the two elder (and more speculative) supercontinents also appear to correspond to the waning stages of the previous superplume cycles, rather than close to the troughs of the plume activity as is the case for Pangea and Rodinia. This could reflect the poor knowledge we have about those two potential older supercontinents. These possible ~600 Myr cycles are broadly similar to, but slightly longer than, the 400–500 My cycles suggested by Nance et al. (1988).

If superplume events were indeed coupled with supercontinent events in Earth's history, there are significant geodynamic implications:

(1) The formation of superplumes might be related to the formation of supercontinents, plate subduction and related mantle convection, rather than spontaneous thermal boundary instabilities derived from the CMB.
(2) The position of a superplume (whether bipolar or not) is linked to the position of the supercontinent. Unless supercontinents

always form and stay at the same or its antipodal position in a geographic reference frame, which is not the case in view of the rapid drift of Rodinia (Li et al., 2004, 2008), superplumes do not form or stay at the same place as a rule, contradicting speculations by Maruyama (1994), Burke et al. (2008) and Torsvik et al. (2008b).

(3) Superplumes or clusters of plumes lead to the breakup of supercontinents (e.g., Gurnis, 1988; Li et al., 2003b).

(4) The lifespan of superplumes is likely linked to the time-span of related supercontinent cycles, with the Pangean superplume starting between 250 and 200 Ma and lasting to sometime into the future, and the Rodinian superplume starting between 860 and 820 Ma and lasting to at least ca. 600 Ma (Ernst et al., 2008; Li et al., 2008) (Fig. 2). This again contradicts speculated long-life superplumes by some (Maruyama, 1994; Torsvik et al., 2008b). The lifespan of the Pacific superplume is more uncertain and it may be of the same age as the Pangean superplume (Li et al., 2008) or older (Zhong et al., 2007; Torsvik et al., 2008b).

(5) Although the Earth's thermal gradient and lithospheric/crustal thickness may have undergone secular changes since the Archean (e.g., Moores, 2002; Brown, 2007), the first order geodynamic patterns, i.e., cyclic and coupled supercontinent and superplume events, do not seem to have changed significantly.

We note the differences between the supercontinent–superplume links discussed here and that of Condie (1998) in which slab avalanche and plume bombardment occur during supercontinent assembly rather than after the assembly.

3. Paleomagnetism and true polar wander: critical evidence for supercontinent–superplume coupling, and a case for a whole-mantle top-down geodynamic model

Paleomagnetic data from supercontinents sitting directly above superplumes provide a way to independently verify the proposed dynamic interplay between supercontinents and superplumes. The Earth's geomagnetic field is generated by the geodynamo in the core (e.g., Glatzmaier and Roberts, 1995), with the time-averaged geomagnetic pole position coinciding with the Earth's rotation axis (e.g., Merrill et al., 1996). This coincidence is believed to be applicable for most of the past two billion years (Evans, 2006), making paleomagnetism an effective tool for measuring both the movements of individual tectonic plates, and that of the silicate Earth (above the CMB) as a whole, relative to its rotation axis. If the positions of the past superplumes were fixed relative to the Earth's core and its rotation axis, they (and their secondary plumes) are expected to be relatively stationary regardless of the movements of the supercontinents. If they have been coupled with the supercontinents, they are then expected to have moved with the supercontinents.

3.1. Definition of true polar wander

Theoreticians have long suggested that redistribution of mass in the Earth's mantle or its surface could make the whole Earth move relative to its rotation axis, motions termed "true polar wander (TPW)" (Gold, 1955; Goldreich and Toomre, 1969; Richards et al., 1997; Steinberger and O'Connell, 1997; Tsai and Stevenson, 2007). TPW occurs because the minimization of rotational energy of the Earth tends to align the axis of the Earth's maximum moment of inertia with its rotation axis, placing long-wavelength (i.e., spherical harmonic degree 2) positive mass anomalies on the equator. Kirschvink et al. (1997) and Evans (1998) defined a variant of TPW, inertial interchange true polar wander (IITPW), which occurs when magnitudes of the intermediate and maximum moment of inertia cross, causing a discrete burst of TPW up to 90°.

There are two distinct views regarding what reference frame to use for determining TPW. The commonly held view refers TPW to the relative motion between a paleomagnetically determined global common polar wander and an assumed fixed hotspot reference frame (e.g., Gordon, 1987; Besse and Courtillot, 2002). However, in view of increasing evidence for the non-steady nature of the hotspot reference frame and mantle plumes (e.g., Molnar and Stock, 1987; Steinberger and O'Connell, 1998; Torsvik et al., 2002; Tarduno et al., 2003; O'Neill et al., 2005; Torsvik et al., 2008c), this approach is inappropriate for detecting potential whole-mantle motion relative to the rotation axis on a large time-scale. An alternative view refers TPW to the common components of the apparent polar wander shared by all plates (Gold, 1955; Van der Voo, 1994; Kirschvink et al., 1997; Evans, 2003b), which is what we adopt here. Readers are referred to Evans (2003b) and Steinberger and Torsvik (2008) for how to extract the TWP component from the paleomagnetic record of individual plates.

3.2. True polar wander events in the geological record

Besse and Courtillot (2002) demonstrated that under the hypothesis of fixed Atlantic and Indian hot spots, the Earth has experienced TPW of less than 30° since ca. 200 Ma. A more recent analysis by Steinberger and Torsvik (2008) using only the global paleomagnetic record, illustrated that although the Earth underwent episodic coherent rotations (TPW) around an equatorial axis near the centre of mass of all continents since the formation of Pangea at ca. 320 Ma, the maximum deviation from the present Earth was just over 20° and the net rotation since 320 Ma was almost zero. In other words, TPW has been insignificant during Pangean time.

Multiple TPW events were proposed for the time of Pangea assembly, but they are often accompanied by controversies. Van der Voo (1994) identified possible TPW events during the Late Ordovician to the Devonian. Kirschvink et al. (1997) invoked IITPW to explain an episode of rapid polar wander during the Early Cambrian, but was challenged by others (Torsvik et al., 1998) for the lack of enough reliable data.

More recently, TWP events have been reported for the Neoproterozoic during the breakup of the supercontinent Rodinia. Li et al. (2004) reported that paleomagnetic results from South China, the Congo craton, Australia and India suggest a possible near-90° rotation of the supercontinent Rodinia between ca. 800 Ma and ca. 750 Ma (Fig. 3), interpreted as representing a TPW event. Maloof et al. (2006) identified a similar yet somewhat more complex TPW record from rocks in East Svalbard, thus supporting the occurrence of TPW events during Rodinian time.

For geological time beyond 800 Ma, Evans (2003b) speculated the possible occurrence of a number of IITPW events, but, as the data used were predominantly from Laurentia, independent variations from other continents are required to establish those cases.

3.3. Superplume to true polar wander: a case for and a possible mechanism of supercontinent–superplume coupling

The reported TPW events during Rodinian time (Li et al., 2004; Maloof et al., 2006), coinciding with the timing of the Rodinian superplume (Li et al., 2003b, 2008), are most intriguing: (1) the first major episode of superplume breakout in Rodinia lagged in time by ca. 70 My from the final assembly of Rodinia; (2) the timing of the TPW event between ca. 800 and 750 Ma coincides with the prime time interval for the proposed Rodinian superplume (825–750 Ma), and (3) Rodinia and the superplume beneath it appear to have traveled from a moderate to high-latitude position at ca. 800 Ma to a dominantly equatorial position by ca. 750 Ma (Fig. 3). Li et al. (2003b, 2008) thus proposed the following causal relationships:

1. Circum–Rodinia avalanches of stagnated oceanic slabs at the mantle transition zone, plus thermal insulation of the supercontinent, drove the formation of the Rodinian superplume and possibly an antipodal superplume in the pan-Rodinian ocean;
2. Mass anomalies caused by the antipodal superplumes and dynamic topography (Hager et al., 1985; Evans, 2003b) led to the 800–750 Ma TPW event(s) above the CMB. This implies that both superplumes and secondary plumes associated with superplumes (Fig. 3c; Courtillot et al., 2003) could rotate rapidly relative to the Earth's rotation axis;
3. Weathering of plume-induced flood basalts (e.g., Godderis et al., 2003), the low-latitude positions of all the continents (Kirschvink, 1992) brought about by the TPW event(s), and the enhanced silicate weathering and organic carbon burial over a breaking-apart supercontinent at an equatorial position (Worsley and Kidder, 1991; Young, 1991), may all have contributed to the extremely cold condition (a snowball Earth?) after ca. 750 Ma (Hoffman and Schrag, 2002).

The schematic model in Fig. 3c, modified after Li et al. (2008), shows how the circum–supercontinent slab avalanches drive whole-mantle convection and the formation of bipolar superplumes, although the piles of chemically distinct mantle materials immediately above the CMB in the superplume regions (Masters et al., 2000; Wen et al., 2001; Ni et al., 2002) may not get much involved in such convection. Superplume is a key component of the model by Li et al. (2004) for TPW during Rodinia time.

The model in Fig. 3c predicts that the superplume under the supercontinent would start stronger than the antipodal superplume in the oceanic realm because it is closer to the ring of slab graveyard (the present total area of oceanic crust is about twice as big as that of the continents). This might explain a possible time lag in peak plume activities between the Pangean superplume (represented by the ca. 200 Ma Central Atlantic Magmatic Province?) and the Paleo-Pacific superplume (~120 Ma? Larson, 1991b). We note that the starting time for the Paleo-Pacific superplume is uncertain as discussed in section 2.2, and that the Paleo-Pacific superplume was proposed by some to be older than the African superplume based on geodynamic considerations (Zhong et al., 2007, and Section 4) and other geological speculations (Maruyama et al., 2007; Torsvik et al., 2008b).

Once the supercontinent starts to breakup, the circum–supercontinent subduction zone will begin to retreat. This would predict the intensity of the sub-supercontinent superplume to gradually decrease whereas that of the sub-oceanic superplume to gradually increase after the breakup of the supercontinent. These trends would continue until the ring-shaped subduction system eventually gives way to multiple and more randomly distributed subduction systems.

It should be pointed out that the dynamics of TPW are not always considered as being controlled by superplumes. For example, Richards et al. (1997) suggested that the present-day Earth's geoid including a degree-2 component is controlled by subducted slabs in the mantle, and that the lack of TPW in the last 65 Ma resulted from the relatively small change in subduction configuration. Since the Earth's plate motion and subduction history can only be reconstructed relatively reliably for the last 120 Ma, Richards et al. (1997) did not discuss the cause for the lack of significant TPW since Pangea formation at 320 Ma. However, if one takes the view that the degree-2 geoid is controlled by superplumes (e.g., Hager et al., 1985; Zhong et al., 2007), the lack of significant TPW since Pangean time can be explained as a consequence of the formation of the supercontinent Pangea, and thus the Pangean and Pacific superplumes, near the paleo-equator (Zhong et al., 2007; Torsvik et al., 2008a). Zhong et al. (2007) also offered an explanation for likely TPW events during the assembly of Pangea (Van der Voo, 1994) in terms of long-wavelength mantle convection (see Section 4), but whether or not TPW events occurred during the assembly of Rodinia still awaits further studies.

4. Geodynamic modelling: what is possible?

4.1. Models of supercontinent processes

Continents have been suggested to play an important role in the dynamics of the Earth's mantle (e.g., Elder, 1976). However, compared to the large number of geodynamic modelling studies on mantle and lithospheric processes, the number of studies on supercontinent processes is rather limited, possibly due to the complicated nature of the dynamics of continental deformation (e.g., Lenardic et al., 2005). Two types of models have been proposed for supercontinent cycles: stochastic and dynamic models. Tao and Jarvis (2002) proposed a stochastic model of supercontinent formation in which continental blocks with semi-random motions collide to form a supercontinent. Tao and Jarvis (2002) showed that when continental motions and collisions follow certain rules, such a stochastic model gives rise to reasonable formation time (~400–500 Ma) for supercontinents. In this model, the physical processes in the mantle are ignored. However, Zhang et al. (2009) indicated that the time-scales for supercontinent formation in the stochastic models may vary from several 100 Ma to more than 1 Ga, strongly dependent on the rules that must be assumed for the semi-random motions of continental blocks in this type of models, indicating the importance of dynamic modelling of convective processes with continents.

Most studies on supercontinent processes have been based on dynamic modelling of mantle convection with continents. Using 2-D convection models, Gurnis (1988) demonstrated that continents are drawn to convective downwellings and collide there to form a supercontinent, and that an upwelling develops below the supercontinent and eventually causes it to break up. Subsequent work focused on the relative roles in continental breakup by upwellings below and downwellings surrounding a supercontinent (Lowman and Jarvis, 1995, 1996), the periodicity of supercontinent cycles in 2-D (Zhong and Gurnis, 1993) and 3-D spherical geometry and the role of heating modes (Phillips and Bunge, 2005, 2007). Zhang et al. (2009) reported that the time-scales for supercontinent formation depend on convective planform or convective wavelengths from mantle convection. For mantle convection with inherently short-wavelengths (e.g., for models with homogeneous mantle viscosity), it may take ~1 Ga for continental blocks to merge to form a supercontinent (Zhang et al., 2009). This suggests that the planform of mantle convection is important for supercontinent processes (Evans, 2003b).

4.2. Dynamically self-consistent generation of long-wavelength mantle structures

Dynamic generation of long-wavelength convective planform and plate tectonics has been an important goal in mantle convection studies for two reasons. First, the present-day Earth's mantle structure is predominated by long-wavelength structures, particularly the strong degree-2 components associated with the African and Pacific superplumes and circum-Pacific subduction (Dziewonski, 1984; Van der Hilst et al., 1997; Ritsema et al., 1999; Masters et al., 2000; Grand, 2002; Romanowicz and Gung, 2002). This poses challenges to mantle convection models with uniform mantle viscosity that produces convective structures with much shorter wavelengths (Bercovici et al., 1989). Second, long-wavelength convective planform may be important in supercontinent formation. If a supercontinent forms above downwellings (e.g., Gurnis, 1988), then short-wavelength convection with a large number of downwellings may pose difficulties in forming the supercontinent, as demonstrated by Zhang et al. (2009).

The scales of tectonic plates and continents have important controls on mantle convection wavelengths (Hager and Oconnell, 1981; Davies, 1988; Gurnis, 1989; Zhong and Gurnis, 1993, 1994). Bunge et al. (1998) and Liu et al. (2008) showed that regional and global mantle downwelling structures can be reasonably reproduced in mantle convection models with plate motion and subduction history included as boundary conditions. McNamara and Zhong (2005a) showed that the African and Pacific thermochemical piles and superplumes can be reproduced in a similar way. These studies clearly demonstrate that the mantle structure is closely related to the surface plate motion. However, by imposing plate motion history as boundary conditions, these studies did not address the question of what processes control the length scales of tectonic plates. Unfortunately, it remains a challenge to formulate models of mantle convection in which plate tectonics emerges dynamically self-consistently, mainly because the complicated physics of deformation at the plate boundaries is poorly understood (e.g., Zhong et al., 1998; Tackley, 2000; Richards et al., 2001; Bercovici, 2003).

Bunge et al. (1996) showed that a factor of 30 increase in viscosity from the upper to lower mantle, as suggested by modelling the geoid anomalies (Hager and Richards, 1989), produces much longer wavelength mantle structures than that from a mantle with uniform viscosity. Bunge et al.'s finding is consistent with previous studies with radially stratified viscosity (e.g., Jaupart and Parsons, 1985; Zhang and Yuen, 1995). Tackley et al. (1993) reported that the endothermic phase change from spinel to pervoskite at the 670-km discontinuity may also increase the convective wavelengths. However, neither Bunge et al. (1996) nor Tackley et al. (1993) produced large enough convective wavelengths that are comparable with seismic observations. Furthermore, these models ignored temperature-dependent viscosity that likely has significant effects on phase change dynamics (Zhong and Gurnis, 1994; Davies, 1995) and convective planform (Zhong et al., 2000, 2007). These drawbacks make it difficult to apply the models of Tackley et al. (1993) and Bunge et al. (1996) to supercontinents and superplumes.

The effects of stratified mantle viscosity and temperature-dependent viscosity on convective wavelengths have been further explored in recent years (e.g., Richards et al., 2001; Lenardic et al., 2006; Zhong et al., 2000, 2007). Moderate temperature-dependent viscosity (3–4 orders of magnitude viscosity variation due to temperature change in the mantle) that gives rise to Earth-like mobile-lid convection helps increase convective wavelengths, compared to mantle convection with constant viscosity (Ratcliff et al., 1996; Tackley, 1996; Zhong et al., 2000). This is because the top thermal boundary layer is stabilized by its high viscosity due to its low temperature. Indeed, mantle convection with a top thermal boundary layer that has a relatively high viscosity (but still mobile) may yield the longest wavelength convective planform (i.e., spherical harmonic degree-1 convection) (e.g., Harder, 2000; McNamara and Zhong, 2005b; Yoshida and Kageyama, 2006). However, this type of long-wavelength convection only occurs at moderately high Rayleigh numbers or convective vigor (McNamara and Zhong, 2005b; Yoshida and Kageyama, 2006).

However, Zhong et al. (2007) showed that degree-1 convective planform develops independent of model parameters, including Rayleigh number, internal heating rate, and initial conditions, provided that the lithosphere (i.e., the top boundary layer) and lower mantle viscosities are ~200 times and 30 times larger than the upper mantle, respectively, which are consistent with observations of lithospheric deformation and the long-wavelength geoid (Hager and Richards, 1989; England and Molnar, 1997; Conrad and Hager, 1999) (see Fig. 5a–c for an example in which initially short wavelength convection evolves to degree-1 planform). Zhong et al. (2007) emphasized the role of combination of moderately strong lithosphere and lower mantle in increasing convective wavelengths, contrary to Bunge et al. (1996) who only focused on the role of viscosity contrast between the lower and upper mantle. For example, Zhong et al. (2007) showed that without moderately strong lithosphere, a factor of 30 increase in viscosity in the lower mantle may actually lead to decreased convective wavelengths in mantle convection with temperature-dependent viscosity. However, it should be pointed out that the mobile-lid convection in Zhong et al. (2007) does not accurately capture the localized deformation at plate boundaries as observed in plate tectonics. Also, Zhong et al. (2007) did not consider thermochemical piles. Nonetheless, we think that provided the total volume of thermochemical piles is significantly smaller than that of the mantle, as suggested by seismic observations (Wang and Wen, 2004), thermochemical piles may not affect the overall mantle dynamics significantly.

4.3. Implications of degree-1 mantle convection for superplumes, supercontinent cycles and TPW

Realizing that the single downwelling system in degree-1 convection naturally leads to supercontinent formation and that the single upwelling system in degree-1 convection represents a superplume, Zhong et al. (2007) further proposed a 1–2–1 model for mantle structure evolution and supercontinent cycles. The key components of this model are summarized as follows. First, when continents are sufficiently scattered, mantle convection tends to evolve to degree-1 planform that draws continental blocks to the downwelling hemisphere to form a supercontinent (Fig. 5b–c). Second, once a supercontinent is formed, circum–supercontinent subduction produces a return-flow below the supercontinent, and this return-flow gives rise to the second superplume below the supercontinent (Fig. 5d) and hence largely degree-2 structures. The sub-supercontinent superplume eventually causes the supercontinent to break.

While the basic notion that supercontinent cycles and mantle convection dynamically interact with each other is consistent with Gurnis (1988), Zhong et al. (2007) discussed three implications of their model. (1) After supercontinent formation, a superplume forms rapidly below the supercontinent as a result of return flow in response to circum–supercontinent subduction (rather than due to a much slower insulation process by the supercontinent). This is consistent with the formation of the African superplume not long after Pangea formation at ~320 Ma as recorded by volcanism (Veevers and Tewari, 1995; Doblas et al., 1998; Marzoli et al., 1999; Hames et al., 2000; Isley and Abbott, 2002; Torsvik et al., 2006). (2) The superplume that is antipodal to the downwelling in degree-1 convective planform is associated with the formation of a supercontinent and is therefore older than the supercontinent. This suggests that the Pacific superplume might be older than the Pangean (African) superplume. This scenario is somewhat different from the synchronous superplume formation presented in Figs. 1d and 3. Future work may test these two different models. (3) The present-day Earth's mantle is in a post-supercontinent state for which the seismically observed structures with a strong degree-2 component and two antipodal superplumes (i.e., African and Pacific) are expected.

By modelling the degree-2 geoid and TPW, Zhong et al. (2007) also propose that their degree 1–2–1 cycle of mantle structure evolution is consistent with the first order observations of TPW and latitudinal locations of Pangea and Rodinia during their breakup. They showed that after a sub-supercontinent superplume is formed

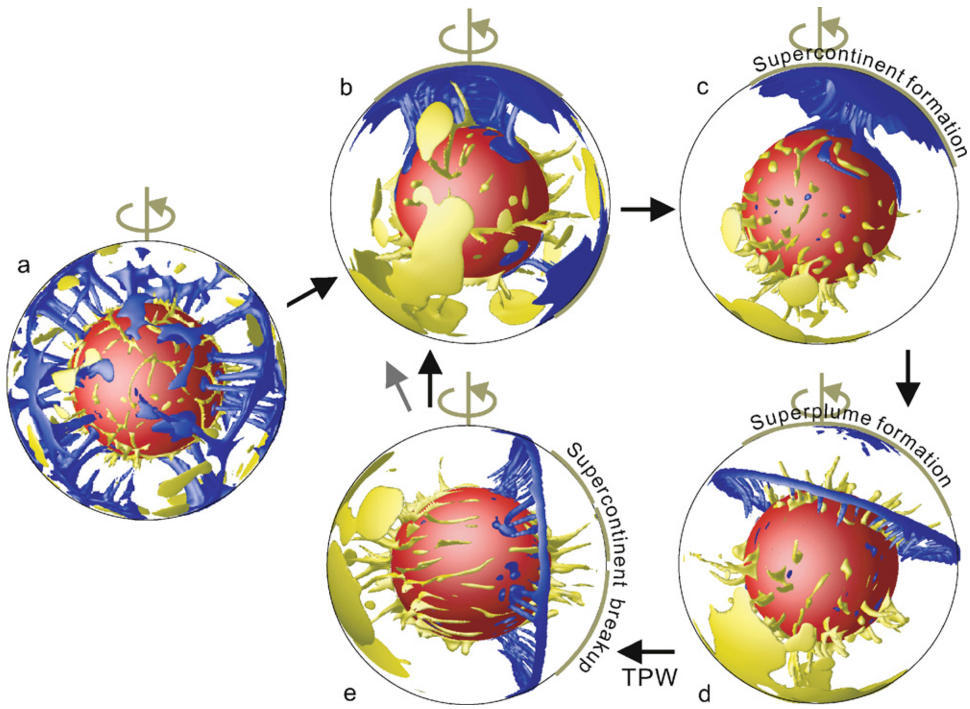

Fig. 5. Numerical modelling results of transitions from (a) a global state of small-scale convections to (b) an early stage of degree-1 planform (supercontinent assembling stage), (c) a stable degree-1 where a supercontinent forms above a super downwelling (or "cold plume"), (d) formation of a degree-2 plantform (early stage of superplume development beneath the supercontinent, and (e) and fully developed degree-2 planform (antipodal superplumes and breakup of the supercontinent; note the true polar wander event between (d) and (e)). Blue shows relatively cold mantle, yellow shows relatively hot mantle, and red shows the Earth's core. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of the article.)

and when the mantle has two antipodal superplumes, the degree-2 geoid highs over the two superplumes require that the superplumes and supercontinent be centered at the equator (Fig. 5e). This explains the lack of TPW for the last 250 Ma or longer after the Pangea formation as the consequence of formation and continuous existence of the African and Pacific superplumes near the equator (Zhong et al., 2007; Steinberger and Torsvik, 2008). This also suggests that major TPW is expected if the supercontinent is not at the equator after the sub-supercontinent superplume is formed (Fig. 5d–e), similar to the TPW event suggested for Rodinia (Li et al., 2004; Maloof et al., 2006) (Fig. 3).

More importantly, the 1–2–1 model by Zhong et al. (2007) predicts that TWP is expected to be more variable and likely large during the formation of a supercontinent when convective planform is predominantly degree-1. This is because TPW is sensitive only to degree-2 structure which is a secondary feature during supercontinent assembly if the assembly is associated with degree-1 mantle convection. This provides a simple explanation for the inferred TPW events during the assembly of Pangea (Van der Voo, 1994; Evans, 2003b). As a final remark, a number of recent studies also have examined the relationship between TPW and supercontinent processes (Nakada, 2008; Phillips et al., 2009).

5. Conclusions

Although our understanding of supercontinent history before Rodinia is still very sketchy, the available geological record, particularly the record since ca. 1000 Ma, appears to suggest a cyclic nature of supercontinent assembly and breakup accompanied by superplume events. The antipodal nature of the Pangean (Africa) and the Paleo-Pacific (Pacific) superplumes, the time-lags between supercontinent assembly and the breakout of the superplume beneath it for both Rodinia and Pangea, and the coincidence between the Rodinia superplume event and the TPW event between ca. 800–750 Ma that brought a disintegrating Rodinia to an equatorial position, suggest that the formation of the antipodal superplumes was likely related to circum–supercontinent subduction. We thus suggest that plate tectonic processes, including circum–supercontinent subduction, drive the formation of superplumes, which in turn cause the breakup of the supercontinents; They also cause TPW events if the supercontinents and coupled superplumes are not on the equator. This implies that the relatively stable nature of the Africa and Pacific superplumes and associated plumes since the Cretaceous in the geographic reference frame (Steinberger and Torsvik, 2008) may not be an intrinsic nature of the so-called "plume reference frame". In essence, the predominant effects of plate subduction on mantle structures, including plumes and superplumes, suggest whole-mantle, top-down tectonics.

Multidisciplinary studies, integrating geophysics, geochemistry and geology, are required to further advance our understanding of the Earth's geodynamic system. Key areas of research include, but are not limited to, the following. (1) We need to have a more solid understanding of the history of the supercontinent Rodinia, and the history of any supercontinent before Rodinia. This will involve international geological correlations, and high-quality paleomagnetic data coupled with precise geochronology. (2) There is a need for establishing a more robust record of plume activities throughout Earth's history through mapping-out and dating all remnants of large igneous provinces (including mafic dyke swarms), and evaluating geological records of oceanic plateau/seamount subduction which represent plume activities in the oceanic realms. Outcomes from (1) and (2) will enable us to establish temporal–spatial connections between supercontinent evolution, superplume events and true polar wander events. (3) Understanding the dynamics of plate tectonics in the framework of mantle convection is essential for improving our understanding of supercontinent dynamics. The key questions are how to improve our understanding of plate boundary processes such as rifting and subduction and to incorporate them in dynamic modelling. (4) Seismic tomographic, petrological, geochemical and geodynamic modelling studies, in combination with better knowledge of supercontinent–superplume history and TPW record, will provide a more robust understanding of the nature of mantle plumes and superplumes, and their relationships with plate subduction and mantle convection, thus making it possible to establish a more realistic 4-D geodynamic history of the Earth from the Present back through geological time.

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