Global warming of the mantle beneath continents back to the Archean

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Abstract

The Earth experienced global magmatic events correlated with the presence of supercontinents, suggesting that the distribution of continental pieces at the surface is fundamental. In most cases, the plume model is invoked to explain the presence of these large igneous provinces although some do not clearly show evidence for such mechanism. The aggregation of continent is known to have a significant effect on mantle convection and thermal evolution below continents, mostly enlarging the wave-length of convection and insulating the underlying mantle. Both processes tend to increase the temperature below the continental lithosphere, eventually triggering melting even without plume. This model, called mantle global warming, has been tested with convection models (Coltice et al., 2007). This paper questions the validity of this model in the past comparing predictions from numerical simulations to geological observations for continental flood basalts back to the Archean. The simulations predict the mantle global warming model could be an efficient mechanism to account for the peculiarities of magmatic provinces on Pangea, Rodinia and proposed Archean supercontinents like Kenorland and ZimVaalbara.

1. Introduction

The growth of the continental crust is an episodic process (Moorbath, 1978). One way to identify pulses of continent formation is through the record of ages of zircons that carry the informations about the generation of juvenile granites. In the available data, major growth events are dated at 2.7, 2.5, 1.9, 1.1, 0.48, 0.28 and 0.1Ga, corresponding to mantle magmatism and orogenic activity (Condie, 2004). Although no rock has been preserved from the Hadean, 4.1Gy old zircons of Jack Hills in the Ylgarn cratons suggest the presence of early continents (Harrison et al., 2005).

Some of the major events, correlating with mantle volcanism, occurred when supercontinents were aggregating or breaking up (Yale and Carpenter, 1998; Condie, 2004). Rodinia was breaking up at 1.1Ga, Gondwana at 0.48Ga and Pangea was stable between 0.28 and 0.2Ga before disrupting into the continents we see today. There are suggestions of Archean supercontinents especillay between 2750-2650Ga (Aspler and Chiarenzelli, 1998) in a period of rapid and worldwide formation of juvenile continental crust. The temporal coincidence between supercontinents and widespread magmatism questions role of the distribution of continents in trigerring large-scale mantle melting. Mantle convection studies have shown that continents can impose their wavelength to the flow (Guilou and Jaupart, 1995; Phillips and Bunge, 2005) which can lead to an increase in mantle subcontinental temperature (Grigné et al., 2007). Indeed changing the wavelength changes the efficiency of heat removal as well. Another effect is to insulate the subcontinental mantle and reduce the heat loss to very low values. Indeed, mantle heat flow can be as low as 12mW m^{-2} within cratons (Jaupart et al., 2007), which impedes the cooling of the underlying mantle. A consequence of insulation and wavelength change is that continental aggregation leads to an increase in subcontinental temperature even without mantle plumes. Numerical simulations of mantle convection suggest it could be as high as 100°C, triggering melting in the asthenosphere and eventually the lithosphere on a very widespread area, as it is called the mantle global warming model (Coltice et al., 2007).

However, when plume exist they tend to start and focus below continents because it is a naturally hot region (Gurnis, 1988; Zhong and Gurnis, 1993; Guilou and Jaupart, 1995; Phillips and Bunge, 2005; Phillips and Bunge, 2007). For now, only the (super)plume model as been invoked to explain the magmatic activity during continental growth pulses. In most cases, plume evidence is very limited and some ambiguous geological proxies are used (Courtillot et al., 1998). To account for a plume origin, several observations have to be made in the framework of

a plume paradigm: hotspot tracks, domal uplift, deep mantle isotopic fingerprint, restricted geometry of magmatism, temperature excess (Sleep, 1990). Although, convection simulations that take into account the dynamical effects of chemical heterogeneities in the source of plumes show that this paradigm could be revised and a wide variety of thermo-chemical plumes could exist within the mantle (Farnetani et al., 2005).

Some of the largest CFB on Earth do not display evidence for the plume model, even taking into account some of the complexities of the thermo-chemical plumes. Off course, for most of the geological record, the only traces of CFBs left are giant dikes or large-scale basaltic eruptions from which recognition of a plume pattern is very difficult (Ernst and Buchan, 2002). It is only fwith rocks of ages younger than 200My that the emplacement of CFBs can be more accurately documented. But even in the recent record, some CFBs hardly match any plume criteria. One of the best examples is the Central Atlantic Magmatic Province (CAMP): there is no hotspot track, no geochemical fingerprint for deep mantle sources, no domal uplift, and it is so large-scale (>10^6 km^2) that a plume origin cannot be accounted for. In this case, the global warming model fits well the observations (Coltice et al., 2007) and suggests that other CFBs in the past could be too.

In this paper, we investigate the role of the mantle global warming model in generating continental flood basalts on supercontinents over the Earth's history. We first explicit the model and show 3D spherical simulations describing how the temperature changes in the subcontinental mantle for variable continental covers that correspond to situations in the past. These results allow us to propose some continental flood basalts over Earth's history that could be explained by this model instead of plume models.

2. The model

2.1. Physics and phenomenology

There is a strong feedback between mantle convection and the evolution of the continental distribution at the surface of the Earth. Cold downwellings draw continents by focusing convergence while hot upwellings push them away. At the same time, the strength and insulating power of continents strongly impede the presence of downwellings below them so they tend to create hot regions. It has led Anderson (1982) to suggest that the geoid high within the Atlantic is a remnant of hot mantle generated below Pangea for instance. Hence continental rafts play an

important role in the convective organization since they constitute mechanical and thermal heterogeneities within the top boundary layer that drives most of the dynamics since heat from the core probably does not exceed 25% of the global heat budget (Hernlund and Labrosse, 2007).

Aggregation and dispersal of continents are then fundamental in modifying the temperature field in the mantle. A supercontinent has a pronounced insulating power and can force the flow towards longer wavelength than with separate smaller continents. As a consequence, aggregated continents have hotter sublithospheric temperature than when they are dispersed theoretically, and this even if there is no heat coming from the core i.e. without hot plumes. In principle, a mantle only heated from within (cooling and radiogenic heat) can display subcontinental thermal excursions when continents aggregate and disperse. Coltice et al. (2007) used 2D and 3D mantle convection models with continental rafts to show the existence of such mechanism. In their framework, they showed continental aggregation would generate a temperature increase greater than 70°C in 100My when continents aggregate (with moving continents in 2D) in the absence of heat from the core. However, the simulations contain simplifications like rigidity of continents, linear rheology, and fixed continents for the 3D case. It is extremely challenging today to use realistic rheologies (temperature, depth and stress dependent) in spherical models that generate self-consistently a plate like behaviour (Tackley, 2000). The viscosity used in numerical models could be crucial since there is a feedback between temperature, stress and the flow through viscosity. In the following, we explore in simple spherical models the effects of the size and distribution of continents to investigate how the mantle global warming model could have worked in the past.

2.2. Numerical models

Modelling the thermal evolution of the convective mantle with continents involves several ingredients. First, a physical and numerical model of mantle convection has to be used. Then a physical formulation of continents has to be inserted consistently with the physical framework. Interaction between mantle flow and continents has to be implemented without violating the conditions of free convection in which self-organization is crucial. Hence, we solve the Navier-Stokes equations for incompressible material with the code TERRA in 3D spherical geometry (Bunge and Baumgardner, 1995). The viscosity is laterally constant but increases by a factor of

30 at the 660km boundary, as suggested by geoid and post-glacial rebound studies (Ricard et al., 1993). A temperature and stress-dependent rheology is beyond the capabilities of present-day numerical models for highly vigorous convection in spherical geometry. It is not worth to use only temperature-dependence in the models since it would freeze the cold surface and generate a stagnant lid that is not realistic to study the Earth (Christensen, 1984). The calculations involve 85% of heat from within (radioactivity) and 15% from the core. The Rayleigh number based on the upper mantle viscosity is 10^7 here, which is supposed to be close to present-day convective vigor.

The continents are simulated by rigid lids. Their thickness is 220km thick, consistently with the thickness of stable continental lithosphere determined by geophysical observations (Artemieva and Mooney, 2001). In this study, the lids are fixed and hence have a zero velocity, acting like undeformable lithosphere. It is a rather crude model but more advanced representation of deformable continental lithosphere involves numerical complexities (Lenardic et al., 2003) which are sufficiently difficult that no spherical 3D model of deformable continent has been published up to now. However, rigid continents simulate the first-order effects of thermal insulation and mechanical strength on mantle convection, as in many other studies (Gurnis, 1988; Guilou and Jaupart, 1995, Lowman and Jarvis, 1999).

2.3 Simulations

The goal of the calculations with fixed continents is to explore the potential of the mantle global warming model for a planet with smaller continents. Indeed, Archean and Proterozoic continents could have been smaller than today and melting caused by subcontinental warming after aggregation could be questioned. Models are run for several convective overturns in order to obtain statistically steady-state values for the subcontinental temperature. The total continental covers is varied between 10% and 30% of the Earth's surface with dispersed and aggregated lids.

The temperature beneath supercontinents is always larger than beneath dispersed continents whatever the total continental cover in the models. A widespread zone develops below the continent because of insulation and long wavelength of the flow (Fig.1). Even with 15% basal heating, plumes don't necessarily occur below the continents. We compute a temperature

difference between the supercontinent case and dispersed case and present a value which is normalized to the average temperature at the base of the boundary layer (i.e. temperature at the base of the oceanic lithosphere would be the equivalent). The temperature at statistically steady state beneath a supercontinent is 5% larger for small continental cover (10% of the Earth's surface), 10% for intermediate continental cover (20%) and up to 15% for a present-day continental cover (see Fig.2). The larger the continental cover, the larger the temperature difference between aggregated and dispersed case is. These numbers are not very dependent on the Rayleigh number in our calculations. Increasing the size of continents increases both insulation and flow wavelength which consequently increase the temperature. In our models, the "oceanic" area (area without continent) is not influenced by the distribution of the continents since the temperature at the base of the boundary layer is similar in the aggregated and dispersed states.

Even if 5% temperature difference seems small, considering the temperature difference over the boundary layer being around 1500K during the Archean (Jaupart et al., 2007), the warming up would already be 75K which is significantly high to melt the mantle. With the continents growing while the mantle is slowly cooling, the expected temperature increase after continental aggregation increased sligthly with time and probably remained between 50 to 150K since the Archean according to our calculations.

2.2.1.Global warming provinces: case studies

The results of the models can be confronted to observations on various large igneous provinces in the past that can difficultly be explained by the plume mode.

2.3. Pangea provinces

Pangea provides the best example to test the global warming versus mantle plume models, because it is the last supercontinent, which assembly and break-up stories are relatively well constrained. Its final stages of assembly occurred between 320 and 250 Ma (Cawood and Buchan, 2007), while its dispersal progressively spread out from ca. 190 Ma (Sahabi et al., 2004) and is still in progress. The characteristics of the CFBs associated with the break-up of the Pangean supercontinent are better constrained, compared to those associated to previous

supercontinents, providing the best opportunity to test the global warming model as Pangea evolves from a supercontinent configuration to the present-day continental dispersal.

Six CFBs were emplaced after the complete aggregation of Pangea and are therefore directly linked to its break-up, recording a 170 My-long story of continental dismemberment: the Central Atlantic Magmatic Province (CAMP) at 200 Ma, the Karoo at 180 Ma, the Parana-Etendeka at 130 Ma, the Deccan at 65 Ma, the North Atlantic Province at 60 Ma and the Ethiopia-Yemen at 30 Ma (Courtillot et al., 1999). It is generally accepted that the four youngest CFBs (Parana-Etendeka, Deccan, North Atlantic, Ethiopia-Yemen) were emplaced in response to a mantle plume head activity, while Pangea was already more or less largely dismembered. On the other hand, the plume model has been questioned for the two oldest CFBs (CAMP and Karoo) emplaced while Pangea was still a supercontinent, corresponding to a geodynamic setting more favorable to the global warming model.

Hereafter, some key-points considered as critical for the plume versus global warming models will be (re)-examined for the CAMP and Karoo CFBs. These points concern mainly the size and shape of the CFB province and the volume of erupted magmas, the presence or absence of a hot-spot track, the geometry of the dyke swarms, the timing of magmatism, the chemical composition of the magmas.

3.1.1 The CAMP case (200Ma)

The CAMP is the largest CFB on Earth, now extending on four continents (Europe, Africa, North America and South America) and covering ca. 10^7 km^2 (Fig.3). It consists of a) huge sills and some layered intrusions, mainly developed in West Africa and Brazil (Deckart et al., 1997, Marzoli et al., 1999), b) elongated isolated dykes (up to 800 km long) mainly occurring in North Africa, Europe and Canada (Bertrand, 1991; McHone et al., 2005) and dense dyke swarms, along Eastern North America margin, in West Africa and Guyana (Dupuy et al., 1988; Deckart et al., 1997; Verati et al., 2005; McHone et al., 2005), c) some lava flows remnants preserved in Triassic basins in Portugal, Morocco, Eastern North America, Brazil and Bolivia (Bertrand et al., 1982; Puffer, 1992; Marzoli et al., 1999, 2004; Bertrand et al., 2005; Verati et al., 2007). The lava piles are 10 to 450 meters thick. The CAMP CFB mainly extends over the peri-cratonic panafrican to hercynian belts and intra-cratonic basins (brazilian et west-african cratons).

All the volcanics and intrusive bodies are basaltic to gabbroic, respectively, except a few ultramafic cumulates. The rocks are dominantly low-Ti tholeiites (Bertrand, 1991), except a few

high-Ti tholeiites restricted to a limited area in Liberia, Guyana and Brazil (Dupuy et al., 1988; De Min et al., 2003; Deckart et al., 2005).

The CAMP is the oldest CFB which postdates by several tens million years the ultimate aggregation of Pangea. It is associated with the earliest disruption of the supercontinent, initiating the central Atlantic Ocean as far as ~190 Ma ago (Sahabi et al. 2004). The peak igneous activity, established by recent $^{40}Ar/^{39}Ar$ dating (~70 reliable plateau ages), is around 199-200 Ma and coincides with the Triassic-Jurassic boundary (Marzoli et al., 1999, 2004; Hames et al., 2000; Knight et al., 2004; Verati et al., 2007), taking into account ~1% bias between $^{40}Ar/^{39}Ar$ and U/Pb chronometers (Schaltegger et al., 2008). Late minor activity seems to persist up to 190 Ma (Deckart et al., 1997; Marzoli et al., 1999; Nomade et al., 2007).

The origin of the CAMP has been assigned by several authors to a mantle plume head impingement beneath the lithosphere (Hill, 1991; Wilson, 1997; Oyarzun et al., 1997; Leitch et al., 1998; Courtillot et al., 1999; Janney and Castillo, 2001; Ernst and Buchan, 2002), yet this model has been refuted by others (McHone, 2000; De Min et al., 2003; Beutel et al., 2005; Deckart et al., 2005; McHone et al., 2005; Verati et al., 2005) who favor a shallow mantle origin. Several lines of evidence challenge the plume model:

- No hotspot tracks have been recorded on the Atlantic ocean floor as it was opening (McHone, 2000; McHone et al., 2005).

- The geometry of the CAMP is not consistent with the plume head model which predicts a radial spreading over an equant area 2000-2500 km across (Campbell and Griffiths, 1990), whereas the CAMP is much more extended (Fig…) along a ~8000 km area elongated from Brittany, France (Jourdan et al., 2003) to Bolivia (Bertrand et al., 2005). This distribution is more likely controlled by pre-existing lithospheric heterogeneities and weakness zones.

- To account for the elongated geometry, a northward channeling of the plume head has been invoked, starting from the center of the hypothetical plume head inferred to be located close to the Blake Plateau (Wilson, 1997; Oyarzun et al., 1997), yet this scenario is invalidated by the absence of northward age propagation. On the contrary, if any migration of the magmatic activity exists, it would rather propagate southward (Baksi, 2003; Nomade et al., 2007).

- Contrasting to its wide surface, the CAMP is characterized by a relatively low rate of magma supply (mean thickness of volcanic sequences of 100-300 m) compared to the much thicker lava piles observed in plume-related CFBs, such as Deccan or Ethiopian traps. The preservation of

thin lava units interstratified into sedimentary levels precludes that important volumes of lavas might have been removed by erosion.

- The area near the Blake Plateau does not show evidence of uplift that would be expected if a plume head had impinged the lithosphere (McBride, 1991; McHone, 2000).

- The apparent radial pattern of feeder dyke swarms first shown by May (1971), that would account for the impingement of a plume head (Hill, 1991; Wilson, 1997; Leitch et al., 1998; Ernst and Buchan, 2002), is a misleading oversimplification that ignores the regional geology:

a) Some dykes do not exist, e.g. the ENE-WSW dyke swarm repeatedly reported in SE Mauritania (Hill, 1991; Leitch et al., 1998; Ernst and Buchan, 2002) corresponds to faults, not to dykes (1:1000000 geological map of Mauritania, 1975). In addition, the subsurface dyke swarm displaying the same trend in Senegal is hypothesized from magnetic survey, yet its existence and "a fortiori" its age are unknown. Therefore, this ENE-WSW to E-W branch of the radial dyke pattern is not constrained.

b) Some dykes that would fit the radial pattern (e.g. NE-SW dykes in northern CAMP) display inherited trends reactivated during the CAMP event rather than neo-formed trends imposed by a plume head impact. Other dyke swarms (e.g. in Guyana-Surinam) include both Proterozoic and CAMP dykes (Deckart et al., 1997, 2005). On the other hand, the dyke swarm intruding the Reguibat shield in NW Mauritania is poorly dated, but provided so far only Proterozoic ages (Dosso, 1975).

c) Some dyke swarms are much more complex than shown on the "radial pattern". For exemple, the Taoudenni dyke swarm in northern Mali comprises more than 800 dykes among which only a few fit the NE-SW trend that would be expected in this area in a radial pattern (Verati et al., 2005). Similarly, the eastern North America coastal dykes cannot be reduced to those which fit the radial pattern, as they show multiple orientations and cross-cutting relationships which reflect a changing regional stress field (Beutel et al., 2005; McHone et al., 2005)

- The chemical and isotopic compositions of CAMP basalts are diagnostic of shallow-mantle sources and do not bear a deep plume composition. The predominating low-Ti group is characterized by negative Nb-Ta anomalies, enrichment in large ion lithophile elements (LILE) relative to high field strength elements (HFSE) and Nd-Sr isotopic compositions diagnostic of lithospheric sources formerly enriched by ancient subduction processes (Bertrand et al., 1982; Bertrand 1991; Pegram 1990; Puffer, 2001; Cebria et al., 2003; De Min et al., 2003; Deckart et al., 2005; Verati et al., 2005). The subordinate high-Ti group, chemically and isotopically

distinct, indicates the contribution of the asthenospheric mantle (Dupuy et al., 1988; Deckart et al., 2005).

None of the CAMP areas on which Nd-Sr-Pb isotopes were investigated so far (Pegram, 1990; Jourdan et al., 2003; Cebria et al., 2003; Deckart et al., 2005) bear the HIMU signature diagnostic of the mantle plumes (Cape Verde, Fernando do Noronha, Asuncion) which might have been in a favorable position in late Triassic times to trigger the CAMP magmatism (Morgan, 1983).

3.1.2 The Karoo case (180Ma)

The Karoo CFB extends over more than $3x10^6$ km^2, from southern South Africa to Malawi and from western Namibia to Mozambique, with minor outcrops in Antarctica (Fig.4). It postdated the CAMP by about 20 My and is associated to the second phase of the Pangea fragmentation, corresponding to the break-up of southern Gondwana leading to the opening of the SW Indian Ocean and Southern Ocean. The Karoo CFB consists of a vast cover of lava flows and sills, giant dyke swarms and more localized intrusive centers, intruding the Archean Kaapvaal-Zimbabwe cratons, the Proterozoic Limpopo belt and the Permian-Jurassic Karoo sedimentary basins.

Karoo lavas remnants consist of: a) tholeiitic basalt lava piles in Lesotho (Marsh et al., 1997), in Botswana and adjacent Zambia-Zimbabwe (Wigley, 1995; Jones et al., 2001; Jourdan et al., 2005) and in central Namibia (Duncan et al., 1997); b) the ~ 10 km-thick Lebombo lava pile (Cox, 1992; Watkeys, 2002) comprising an upward succession of nephelinites, picrites, tholeiitic basalts and rhyolites (Sweeney et al., 1994). Picrites and basalts also occur in the Tuli and Mwenezi basins (Cox, 1992). Several gabbroic to granitic intrusive complexes also intrude the latter basin (Jourdan et al., 2007a). Vast tholeiitic dolerite sills are intruded at various stratigraphic levels in several Karoo basins, particularly in South Africa (Jourdan et al., 2008), but also in south-eastern Namibia, in southern Botswana and in Mozambique. The main dyke systems are the N110° Okavango dyke swarm, the N70° Save-Limpopo dyke swarm and the N-S Lebombo and Rooi Rand dyke swarms, forming, together with the associated rifts, a pseudo-radiating system (Fig.), i.e. the so-called Karoo triple junction (Burke and Dewey, 1972; Campbell and Griffiths, 1990; Ernst and Buchan, 2002).

A comprehensive $^{40}Ar/^{39}Ar$ dating campaign (~90 reliable plateau ages) shows that the main volume of the basaltic sequence was emplaced over 3 to 4.5 Ma around 180 Ma, whereas the

entire province sustained activity over a total of ~10 Ma, from 184 to 174 Ma (Jourdan et al., 2005, 2007a, 2007b). The increase of magmatic activity seems to coincide with the Pliensbachian-Toarcian minor biotic crisis (Jourdan et al., 2008).

Since the pioneering work of Burke and Dewey (1972), the Karoo CFB is repeatedly referred to as an example of plume related CFB (e.g. Campbell and Griffiths, 1990; Courtillot et al., 1999; Ernst and Buchan, 2002). One of the main argument supporting this plume model was the presence of the so-called triple junction supposed to have been triggered by the impact of a mantle plume head beneath the southern Africa lithosphere in Jurassic times, despite the lack of dating on the related dyke swarms. Recent $^{40}Ar/^{39}Ar$ dating performed on the N110° Okavango, N70° Save-Limpopo and N-S Lebombo dyke swarms (forming the "triple junction"), combined to geochemical analyses, reveal that these dyke swarms unambiguously include Proterozoic dykes (Jourdan et al., 2004, 2006). For instance, the N110° Okavango dyke swarm includes 12% of Proterozoic dykes (Jourdan et al., 2004). In addition, a structural analysis strongly suggests that Karoo dyke orientations are largely controlled by pre-existing structures that also controlled emplacement of Precambrian dykes (Jourdan et al., 2006). The apparent triple junction geometry was not induced by the arrival of a mantle Karoo plume head but is inherited from previous history of the Kaapvaal and Zimbabwe cratons. Therefore, this "triple junction" should no longer be used as an argument for demonstrating (although it does not exclude) the existence of a Karoo mantle plume.

Another reason why the Karoo mantle plume may be questioned is the lack of volcanic track linking the Karoo CFB with a present day hot spot, which position vary (Crozet, Marion, Bouvet …) depending of the authors (Morgan, 1981; Richards et al., 1989; Hawkesworth et al., 1999).

The involvement of a deep mantle plume in the genesis of Karoo CFB is also questioned on the basis of geochemical arguments. Only a few geochemical studies conclude that a mantle plume does contribute to the chemical composition of the Karoo magmas (Ellam and Cox, 1992). Based on isotopes and trace elements patterns (LILE/HFSE enrichment), most geochemical investigations argue for melting of heterogeneous old sub-continental lithospheric mantle (SCLM) (Duncan et al., 1984; Hawkesworth et al., 1984; Ellam and Cox, 1989; Sweeney and Watkeys, 1990; Elburg and Goldberg, 2000) or mixing between lithospheric and asthenospheric mantle sources (Sweeney et al., 1991; Ellam and Cox, 1991; Sweeney et al., 1994). More recently, two scenarios have been proposed, involving either the polybaric melting of SCLM or

mixing between SCLM and an asthenospheric or deeper OIB-like mantle plume (Jourdan et al. 2007c). Regardless of which of the two scenarios is invoked, the spatial distribution of the low- and high-Ti magmas matches the relative positioning of the cratons and the Limpopo belt in such a way that strong control of the lithosphere on magma composition and distribution is pre-requisite of any petrogenetic model applied to the Karoo CFB (Sweeney and Watkeys, 1990; Jourdan et al., 2007c).

2.4. Rodinia provinces

Similarly to Pangaea, the assembly and breakup of the Proterozoic mega continent Rodinia was accompanied by the emplacement of major LIPs and dyke swarms (e.g. Ernst and Buchan, 2001; Ernst et al., 2005, 2008). Most of these LIPs are only scarce remnant occurring as sills and dyke swarms as the lava flow components have been in most cases eroded. Furthermore, Proterozoic LIPs is challenging because of significant alteration and anchizone up to green shist fascies metamorphism affecting the samples.

From 1.3Ga until its final assembly (~900 Ma; Li et al., 2008) and subsequent breakup, Rodinia have been the witness of many LIP events (e.g. Ernst et al., 2008). Of particular interest, is a cluster of LIPs around 1.1 Ga (Fig. 5) as the surface covered by each of the provinces remnants is of several 10^6 km². This makes these provinces equivalent in size to the Pangean provinces mentioned previously.

Hereafter, we will focus on three major provinces (Umkondo, Keweenawan and Warakurna) emplaced sub-synchronously at ~1.1 Ga during the assembly of Rodinia (Fig. 5). According to recent paleocontinent reconstructions, Rodinia was not complete before ca 900 Ma (e.g. Li et al., 2008) and the 1.1-Ga configuration allow us to test our model further on a partially assembled mega continent (Fig. 5).

3.2.1. *The Umkondo case (1.1Ga)*

Paleomagnetical and geochronological (mainly zircons U/Pb analyses) investigations permit to attribute a common origin to Proterozoic tholeiitic rocks occurring throughout northern South Africa, Botswana, Zimbabwe and possibly Antarctica (Hanson et al., 2004). This province is now called Umkondo based on the name of Umkondo Zimbabwe dolerite formation of the same age. (Hanson et al., 1998). This province consists of tholeiitic mafic intrusions (sills and dykes swarms, e.g. Jourdan et al., submitted) and scarce remnants of eroded basaltic lava-flows

emplaced over an estimated paleo-surface of ~$2.5 \cdot 10^6$ km² (Hanson et al., 2004). Robust ages clustering around 1108 ± 3 Ma have been obtained using zircon and baddeleyite U/Pb TIMS technique suggesting a relatively short duration of the magmatism. Umkondo LIP is therefore synchronous with the Rodinia mega-continent formation (Hanson et al., 1998; Dalziel et al., 2000) and in particular the Laurentia and Kalahari craton collisions (Kibaran Grenville-Llano and Namaqua-Natal Orogenies; 1.4-1.0 Ga). The "Kibaran" (Namaqua-Natal) Mesoproterozoic belt that occurs over 3000 km long through central and southern Africa. Between 1.4 and 1.0 Ga, Kibaran metasedimentary and igneous rocks were involved into two compression events (Johnson & Oliver, 2000) including an active continental margin followed by a continental collision.

The majority of the rocks underwent green schist metamorphose and alteration, but their composition is not modified beyond the point they can be used to study petrogenetic processes (Munyanyiwa, 1999). The UIP dolerites are not well studied, but studies carried out so far shat that they mostly consist of low-Ti basaltic rocks, as commonly found in many LIPs. Umkondo dolerites have a strong lithospheric signature as shown by a strong Nb anomaly (Nb/Nb*=0.2-0.4) and low Ce/Pb (2 to 7) ratios (Jourdan et al., submitted). The province is not associated with any hot spot track. No evidence of magmatic contribution from a mantle plume is recognized in the samples analyzed so far (e.g. Hanson et al., 2006; Jourdan et al., submitted). Similar composition with the spatially overlapping 180-Ma Karoo province suggest that both provinces are likely to originate from a very similar mantle source. The source have been only slightly modified since the extraction of the Umkondo basalts and Jourdan et al. (submitted) proposed that the source should be "attached" to the African plate and therefore preferentially located in subduction-metasomatised sub-continental lithospheric mantle.

3.2.2 The Laurantian large igneous provinces (1.1-1.07 Ga)

The Laurantian provinces as defined hereafter includes the Keweenawan "mid-continental rift" (~2×10^6 km²; Cannon, 1992) and the South Western U.S. diabase (~0.4×10^6 km²; Ernst et al., 2008) provinces.

Both provinces have been emplaced sub-synchronously with still scarce U/Pb ages ranging from ca 1.11 to 1.09 Ga for the former (e.g. Vervoort et al., 2007) and from ca 1.10 Ga to 1.07 Ga (e.g. Heaman and Grotzinger, 1992; Ernst and Buchan, 2001 and reference therein) for the latter. The Keweenawan province is the best studied of the two Laurantian provinces and is a bimodal

province including mafic (typically olivine tholeiites; Paces and Bell, 1989) and minor silicic (Vervoort and Green, 1997; Vervoort et al., 2007) rocks. Element and Sr-Nd-Pb isotope geochemistry studies done so far indicate that the magma is derived from enriched mantle reservoirs (Shirey et al., 1994) and share common features with LIPs like Umkondo and Karoo. For example, both Karoo and Keweenawan eruptive sequence have an early activity represented by picrites with a strong lithospheric signature (Shirey et al., 1994; Jourdan et al., 2007). The mantle sources proposed to explain the origin of this problem comprise a variety of scenario ranging from astenospheric/subasthenospheric mantle plume to enriched lithospheric mantle (or a combination of both), altogether with variable degree of crustal contamination (e.g. Paces and Bell, 1989; Nicholson and Shirey, 1990; Vervoort et al., 2007 and references therein). Crustal contamination affects mostly the silicic rocks that derive partially from the fusion of the Archean crust (Vervoort et al., 2007). The South Western U.S. diabase are much less studied and mostly geochronology and paleomagnetic data are available (e.g. Shastri et al., 1991; Heaman and Grotzinger, 1992; Ernst & Buchan, 2001; Ernst et al., 2008). This province consists mostly of up to 450m-thick doleritic sills and no chemical data has been published to our knowledge. This currently makes it difficult to correlate with other LIPs.

3.2.3 The Warakurna large igneous province (1.05Ga)

This province regroups coeval magmatism distributed over ~$2.5 \cdot 10^6$ km² and distributed mostly over Western and northern Australia (Wingate et al., 2004). It is represented mostly by silicic and basaltic lava flows and mafic and ultramafic intrusions. The age of the province has been determined using zircon SHRIMP U/Pb technique and yielded a mean age of 1076 ± 3 Ma although minor magmatism intrusion continue until ~1050 Ma. Investigation using the more robust CA-TIMS analysis would be desirable to obtain a more accurate estimate of the age along with a better estimation of the magmatism duration. Nevertheless, an age of ~1080 Ma makes the Warakurna province synchronous with the Pinjarra orogen, Western Australia (Wingate et al., 2004; Bruguier et al., 1999; Cobb et al., 2001) and may indicate a causal relationship between the Pinjarra subduction and emplacement of the Warakurna magmatic province.

The rocks consist of dolerite with sericitized plagioclase and amphibolitized pyroxene and show an enriched tholeiitic composition with SiO_2 ranging from 48 to 55 wt%. They are enriched in LILE and show strong Nb anomaly that Zhao and McCulloch (1993) and Glikson et al. (1996) attributed to the melting of subduction-modified lithospheric mantle. Zhao and McCulloch

(1993) invoked a rising mantle plume to explain the thermal anomaly required to melt the SCLM. No mantle plume track has been found suggesting that the role of a mantle plume, if any, was confined as heat purveyor.

2.5. Archean provinces

Looking for CFBs caused by global warming beneath supercontinents is a difficult task because of the intrinsic difficulties of working on such old rocks. Although, the magmatic evolution of the mantle within the first couple of billion years involved sudden and violent crises, possibly mantle overturns and/or superplume events (Stein & Hofmann, 1994; Breuer & Spohn, 1995; Barley et al., 1998; Condie, 2001), that punctuated periods of relative quietness. In this context, the period between 2.75 and 2.65 Ga is one of the most dramatic in the Earth's history. In that period, most Archaean cratons were covered by 5 to 15 km thick, komatiite-bearing, basalt dominated, greenstone covers (Nelson, 1998), while isotopic age distribution of detrital zircons point to a major peak production of juvenile continental crust (Gastil, 1960; Condie, 2001; Rino, 2004). In many cratons, this event preceded and overlapped with a profound episode of crustal anatexis and differentiation (Rey et al., 2003). Although partial of melting of volatile-rich mantle above a subduction zone can generate komattites, the geochemistry of most late Archaean komatiites demand a deep source origin and request either a large melt fraction (50%) or the partial melt of a depleted mantle source (Arndt, 2003). Considering the global character of the late Archaean crisis, a superplume event is indeed very appealing (Isley and Abbott, 1999; Condie 2004; Barley et al., 1998), however, it is not without problems. In this section, we review the extent of the late Archaean volcanism and the problems linking this volcanism to superplumes.

3.3.1. *Surface extent of the 2.75-2.65 Ga magmatic crisis*

In the Kaapvaal Craton, the up to 8 km thick Ventersdorp extents over 3e5 km^2 (van der Westhuizen et al. 1991, Eriksson et al., 2002) and includes 2.72 Ga to 2.69 Ga old subaerial continental komatiitic basalts, tholeiitic basalts and sedimentary rocks (Armstrong et al, 1991), that precedes and overlaps with crustal melting, plutonism and coeval sedimentation (Schmidz and Bowring, 2003).

In the Zimbabwe craton, the base of the 2.7 Ga old craton wide Ngezi Group includes the Zeederberg continental flood basalt, which covers an area over 2.5e5 km^2 (Prendergast, 2001).

Its base is made of the submarine Reliance formation, which includes up to 2km thick interlinked komatiite sills extending over 100 km intercalated within sandstones and pillowed basalts (Prendergast, 2001). This event precedes and overlaps with Sesombi-Wedza granitoids emplaced between 2.7 and 2.65 Ga.

The evolution of the Superior Province culminated between 2.75 and 2.7 Ga with the emplacement of submarine magmatism involving komatiites before craton-scale crustal anatexis and plutonism, the bulk of which occurred between 2.71 and 2.66 Ga.

In the Slave Province, the 2.74 to 2.69 Ga Yellowknife greenstone belt includes up to 6 km thick pillowed and massive flow of tholeiitic basalts, minor komatiites and rhyolitic tuff intercalations (Bleeker, 2005). This volcanic event, which covers > 10^5 km^2 (Bleeker, 2005), was followed from 2.69 to 2.66 Ga by calc-alkaline volcanism and Tonalite-Trondhjemite-Granodiorite plutonism (Bleeker, 2005).

In the East goldfield province of the Yilgarn Craton, a 12±2km thick package (>10^5 km^2) of 2.72 to 2.70 Ga marine tholeiite–komatiite and tholeiite–calc-alkaline associations (Barley, et al., 1998) is followed by episodic deep-water volcaniclastic sedimentation from to 2.70 to 2.66 Ga coeval with craton-scale crustal anatexis.

In the Pilbara craton, all the older units are unconformably overlain by the mainly subaerial Fortescue Group (2.765-2.687 Ga) (Nelson et al., 1992; Blake, 1993; Arndt et al., 2001; Blake, 2001). The Fortescue, which covers at least 1.8e5 km^2 (Eriksson et al., 2002), is made of ca. 7 km thick volcanics which includes a series of tholeiitic basalts, minor komatiites, felsic volcanics and clastics rocks (Hickman, 1983; Thorne and Tyler, 1997; Thorne and Hickman, 1998; Blake, 2001).

Greenstones emplaced at 2.7 Ga have also been described in the Wyoming craton (e.g. Stillwater intrusion), Dharwar craton (e.g. Ramagiri-Hungund composite greenstone belt, Gadwal greenstone belt)

3.3.2. *Models for the 2.75-2.65 Ga magmatic crisis*

In all the above cratons, 1/ the synchronicity of bimodal volcanism, felsic plutonism and sedimentation, 2/ the calc-alkaline composition of some basalts, despite minor komattites, and 3/ the syn- to late-contractional deformation, have lead to interpretative models involving multiple subduction zones and discrete back-arc basins between micro-plates. The collisional aggregation these micro-plates finally led to the formation of late Archaean cratons. This model

has been proposed for the late Archaean of the Yilgarn craton (Myers, 1993; Swager et al., 1997; Barley et al., 1989, 1998); the Superior Province (Corfu, 1987; Percival and Williams, 1989; Card, 1990; Ludden et al., 1993; Wyman et al., 2002); the Slave Province (Kusky, 1989; 1990); the Zimbabwe craton (Dirks and Jelsma, 1998; Jelsma and Dirks, 2002; Horstwood et al., 1999); and the Kaapvaal craton (REF). Yet, in all the above cratons, detailed field studies have noted the strong coherence of the greenstone stratigraphy across so-called terrane boundaries as well as the lack of crustal scale features validating collisional tectonics. This prompted the proposition of an alternative model based on the emplacement of craton-scale continental flood basalts, a model which has been proposed in the Slave province (Bleeker et al., 1999a, 1999b), Zimbabwe (Prendergast, 2001; Wilson, 1979), in the Superior Provice (Heather et al., 1995; Heather, 1998; Ayer et al., 1999, 2002; Thurston, 2002; Benn, 2006), and the Yilgarn (Rey et al., 2003). The profound phase of crustal anatexis that overlapped and post-dated continental flood basalts in many cratons could be linked to the thermal insulation effect associated with the emplacement the thick greenstones cover on a radiogenic crust (Rey et al., 2003). It is worth noting that recent Hafnium isotopes data on zircon show that, contrarily to the 3.3 Ga and 1.9 Ga events, the large anomaly in the isotopic age distribution of zircons is not accompanied with the formation of juvenile crust (Kemp et al., 2006). Intra-crustal anatexie alone, rather than the formation of juvenile crust in a subduction zones, could explain the isotopic age distribution of detrital zircons as well as the change of the average composition of emerged landmasses recorded in black shales (Taylor and McLennan, 1985; Rey and Coltice, 2008).

3.3.3. *The cause of the 2.75-2.65 Ga volcanic crisis: Plume vs mantle golbal warming*

Hitherto, only mantle plumes were considered a viable setting of deep mantle partial melting. The application of the superplume model in the late Archaean is not without problem.

1/ Most plume-related Phanaerozoic continental flood basalts were emplaced over a very short periods typically less than a few Myr. In contrast, the late Archaean continental flood basalts were emplaced over a few tens of Myr.

2/ Subduction-related calc-alkaline volcanism is often spatially and temporarily associated with komatiites. Although, models have been proposed to explain the production of komatiitic magma in a subduction setting, most late Archaean komatiites did not form in a subduction zone (Arndt, 2003). This leads to very complex tectonic models involving coeval upwelling flow (plume) and downwelling mantle flow (subduction), both in the same location.

3/ Komatiite in Archaean CFB are generally less than 5% (Viljoen & Viljoen 1969 de Wit & Ashwal 1997) and the bulk of continental flood basalt are made of basalt involving lower degrees of melting. On that argument, large partial melt of a deep fertile plume could be ruled out in favor of a smaller melt fraction in a refractory upper mantle source.

4/ In the Yilgarn, the geochemistry of 2.72-2.68 Ga basalts indicates extraction from a depleted mantle with evolution from residual-plagioclase to residual-garnet. This suggests that komatiites and basalts in the Eastern goldfield province derived from melting that took place at progressively greater depths (Bateman et al., 2001). Interestingly, the same conclusion was proposed in the Pilbara to explain the increase in FeO and incompatible elements in progressively younger terms of the Fortescue Group (Arndt, 2001). In the context of a rising plume, the downward migration of the melting source is not clear.

Mantle global warming above aggregating continents could provide a viable alternative to surperplume events. Williams et al. (1991) proposed that a supercontinent, Kenorland, formed at around the 2.7 Ga via the aggregation of the North American shields (the Slave Province, Superior Province, Wyoming craton), joined perhaps by the Baltic and Siberian shields (Bleeker, 2005). A possible second supercontinent, the Zimvaalbara, may have included the Yilgarn craton, the Pilbara craton, The Gawler craton, the Zimbabwe craton, the Kaapvaal craton, the Congo craton, the São Francisco craton, and possibly the Dharwar craton (Aspler and Chiarenzelli, 1998). It is not clear how mantle plumes could have dominated the mantle flow beneath supercontinents at the stage of their aggregation. In contrast, we showed that the formation of such supercontinents could trigger the progressive warming of the sub-continental mantle. The mantle global warming model involves a mantle scale upwelling to explain the occurrence of komatiite. Therefore, there is no need to invoke complicated models involving mantle plumes impinges on a subducting slabs to explain the intercalation of komatiites with basalts.

Although most late Archaean continental flood basalts emplaced onto flooded continents, the Pilbara and the Kaapvaal cratons were largely emerged at the time of volcanism. Therefore, although continental emergence was not yet the rule, global high sea level is not expected in the mantle warming model as mantle warming doesn't affect oceanic areas.

Due to the progressive warming of the sub-supercontinental mantle, the mantle geotherm would cross the mantle solidus at a progressively increasing depth. This could explain the progressive

deepening of the melting source as recorded by the Komatiites from the Yilgarn cratons (Bateman et al., 2001) and the Pilbara craton (Arndt, 1999).

Finally, the process of mantle warming and melting underneath insulating supercontinent is similar to crustal warming and anatexie due to thermal insulation of the craton under greenstone cover proposed in the Yilgarn (see Fig.6) by Rey et al. (2003). Therefore, the late Archaean volcanic crisis and plutonic crisis can be seen in a broad context of nested insulation processes.

2.5.1.Conclusions

The presence of a supercontinent at the surface of the Earth has a major impact on the convective flow within the mantle, as shown by our 3D spherical convection simulations. As a consequence, the subcontinental temperature is 5-15% higher than for cases with dispersed continents. The smaller the continental cover, the smaller is the temperature increase with aggregation. For a continental cover larger than 10% of the Earth's surface, it is expected that the subcontinental mantle could warm by more than 70K over an area comparable to the size of the supercontinent. The mechanism of mantle global warming, alternative to mantle plumes, is then a viable hypothesis for the origin of some continental flood basalts since the Archean.

The numerical models are used to explore realistic physical mechanism from which quantitative predictions can be made and tested against data. In the framework of the mantle global warming model, the models predict that

- the temperature increase causing melting does not exceed more than 100K
- heating occurs over an area comparable to the supercontinent
- magmatic sources are mostly asthenosphere and continental lithosphere
- melt extraction is controlled by tectonics.

The mantle global warming model is consistent with the observations made on some CFBs in the geological record. The best example is the CAMP (200Ma), the largest magmatic province on Earth in area, but more and more evidence point also to the Karoo (180Ma). On the basis of geochronology, geochemistry and tectonics, we suggest that during the Proterozoic, the

Umkondo, Laurentian and Warakurna CFBs are derived from mantle global warming beneath the Rodinia supercontinent around 1.1Ga. The global magmatic crisis around 2.7Ga in the Archean could also be explained by heating below the formation of Kenorland and Zimvaalbara supercontinents, for which plumes are difficult to invoke.

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Fig.1: Snapshots of temperature for six spherical convection models. The temperature range spans nondimensional values of 0.013 (blue) to 0.036 (red). Gray caps indicate the surface location of fixed continents. The models contain A) two antipodal continents, each of which covers 5% of the surface, B) one continent covering 10% of the surface, C) two continents with 10% coverage, D) one continent with 20% coverage, E) two continents with 15% coverage, and F) one continent with 30% coverage. Respective pairs from left to right give a sense for the impact on the mantle due to the aggregation of two smaller continents into a single large continent. An apparent impact on subcontinental temperature occurs only once the size of an individual continent reaches 20% coverage. The full supercontinent of F) is unique in promoting a strong hemispheric scale hot anomaly shown by the far reaching extent of warm colors, indicative of global mantle warming.

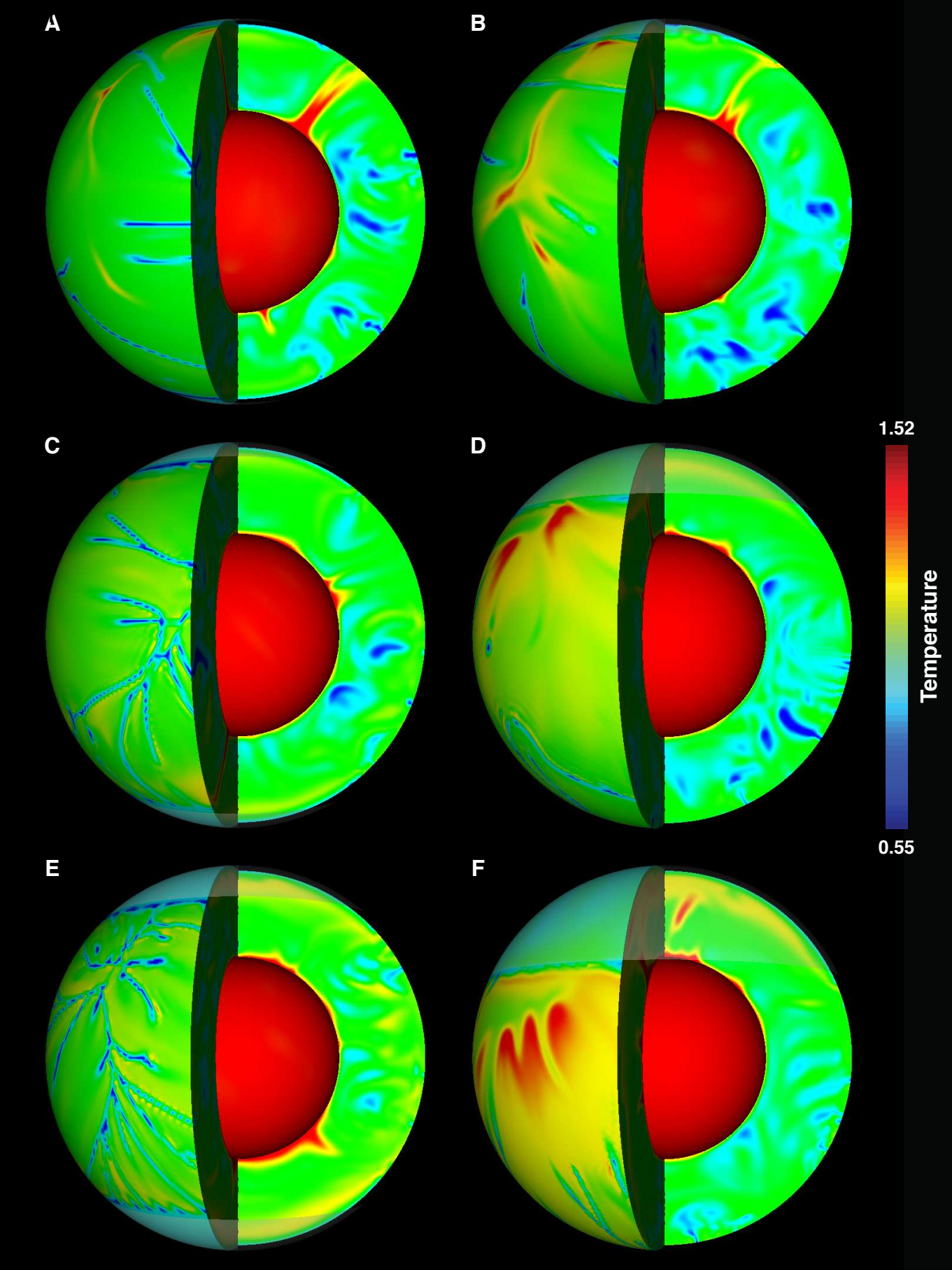
Fig.2: Subcontinental nondimensional temperature excess of the aggregated state relative to the two continent dispersed state as a function of total continental coverage and Rayleigh number. The Ra = 10^7 values correspond to the models shown in Fig. 1.

Fig.3: Extension of the Central Atlantic Magmatic Province at 200Ma.

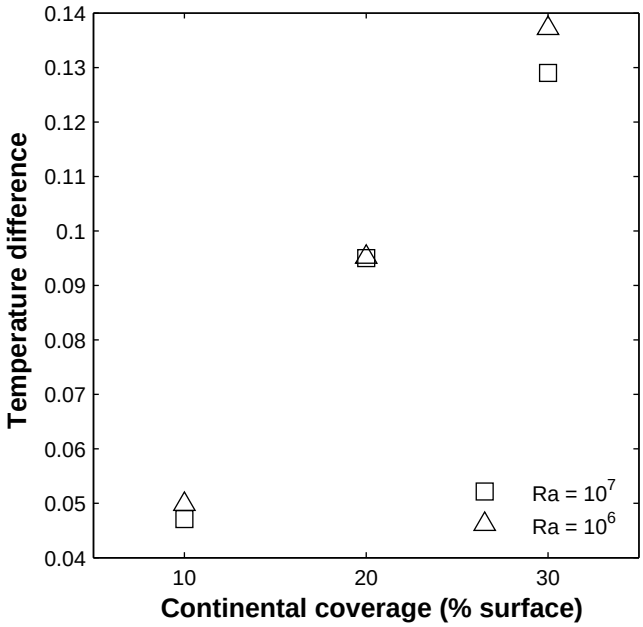
Fig.4: Extension of the Karoo Province at 180Ma.

Fig.5: Estimated extension of the Umkondo, Warakurna and Keweenawan provinces at the time of Rodinia aggregation.

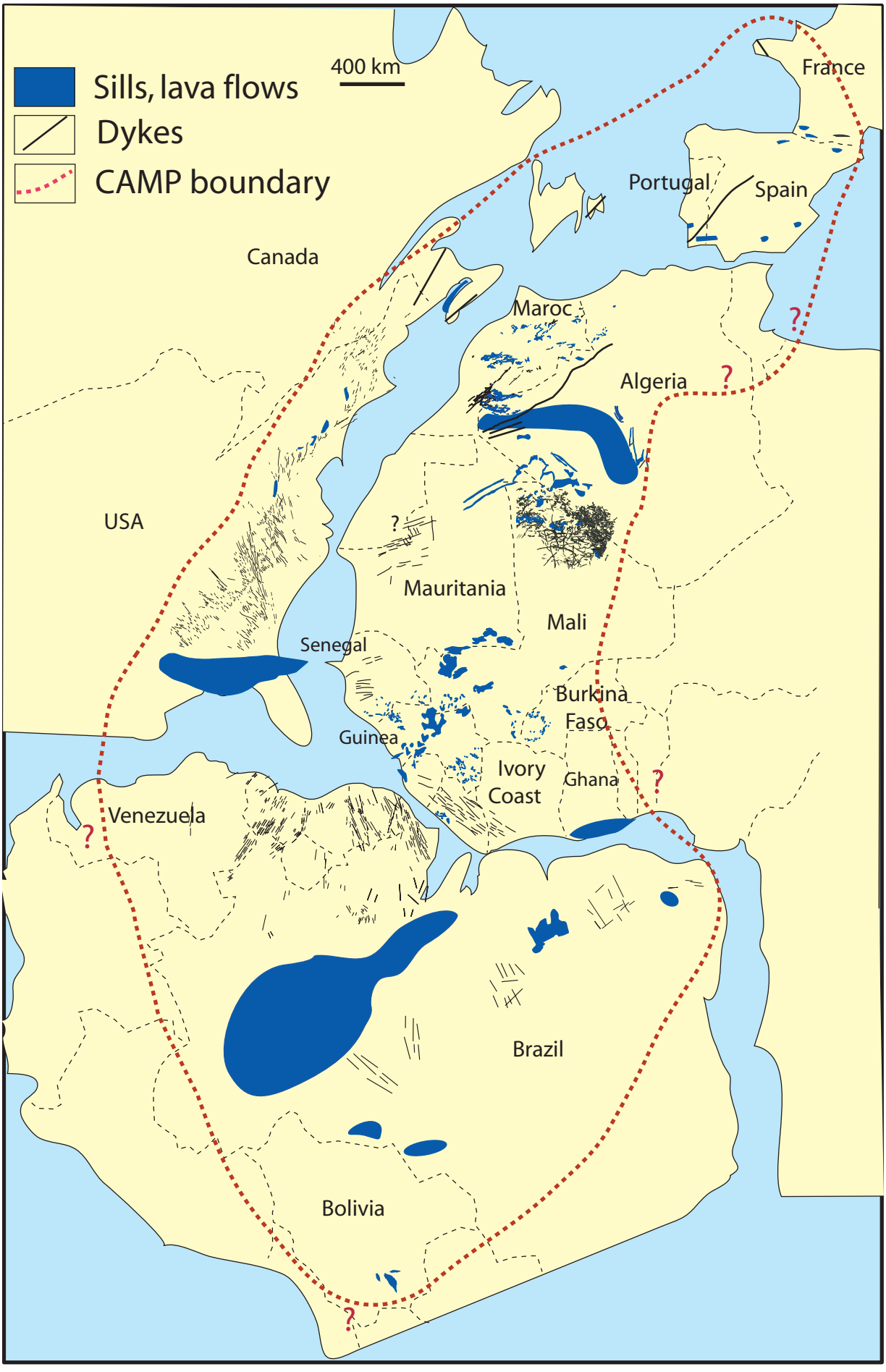
Fig.6: Continental warming following the emplacementof a 6 km thick CFB on a 2.75 Ga old continental crust (heat production of 10^{-6} W m^{-3}, mantle heat flow of 0.025 W m^{-2}, : 1000 J mol^{-1}, thermal diffusivity of 0.9 10^{-6} m^2 s-1, continental crust thickness of 40 km). a/ Geotherm at to, to+5my, to+25 my, to+50 my, to+100 my, to+250 my and b/ evolution of the Moho temperature after emplacement of a 6 and 12 km thick CFB. Continental warming and anatexy is inescapable and does not require the direct heat input from a mantle plume.


A
B
C
D
E
F
1.52
Temperature
0.55

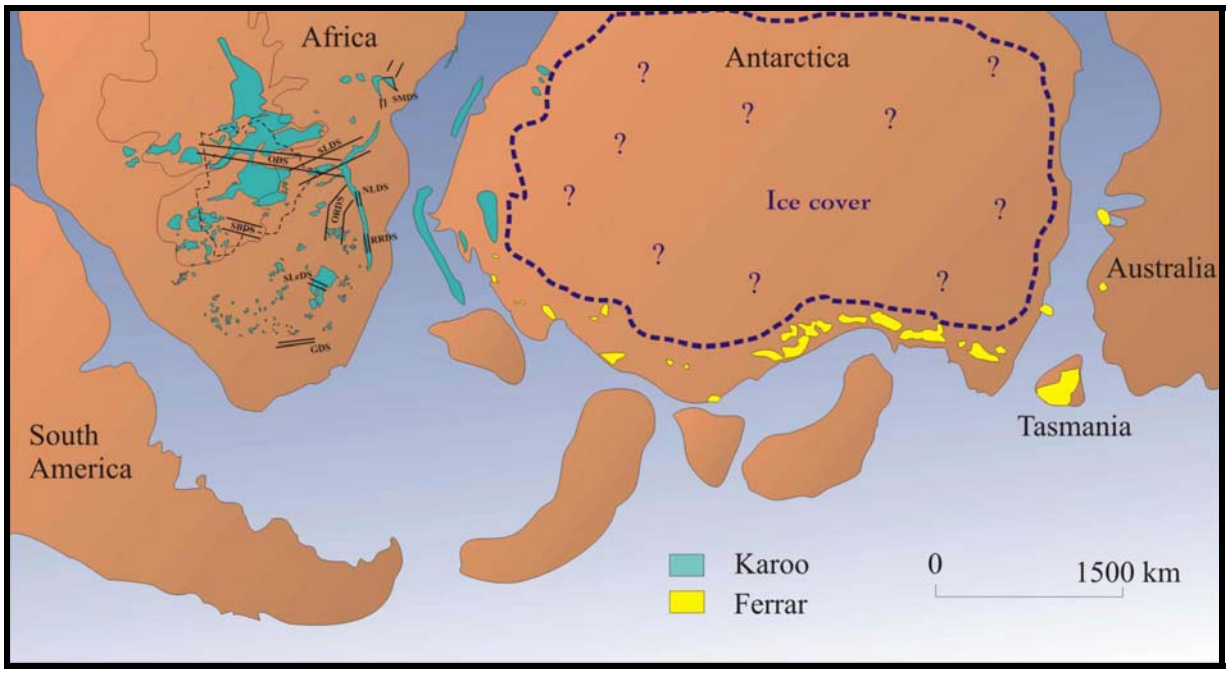
Figure


0.14
0.13
0.12
0.11
0.1
0.09
0.08
0.07
0.06
0.05
0.04
Temperature difference
10
20
30
Continental coverage (% surface)
Ra = 10^7
Ra = 10^6

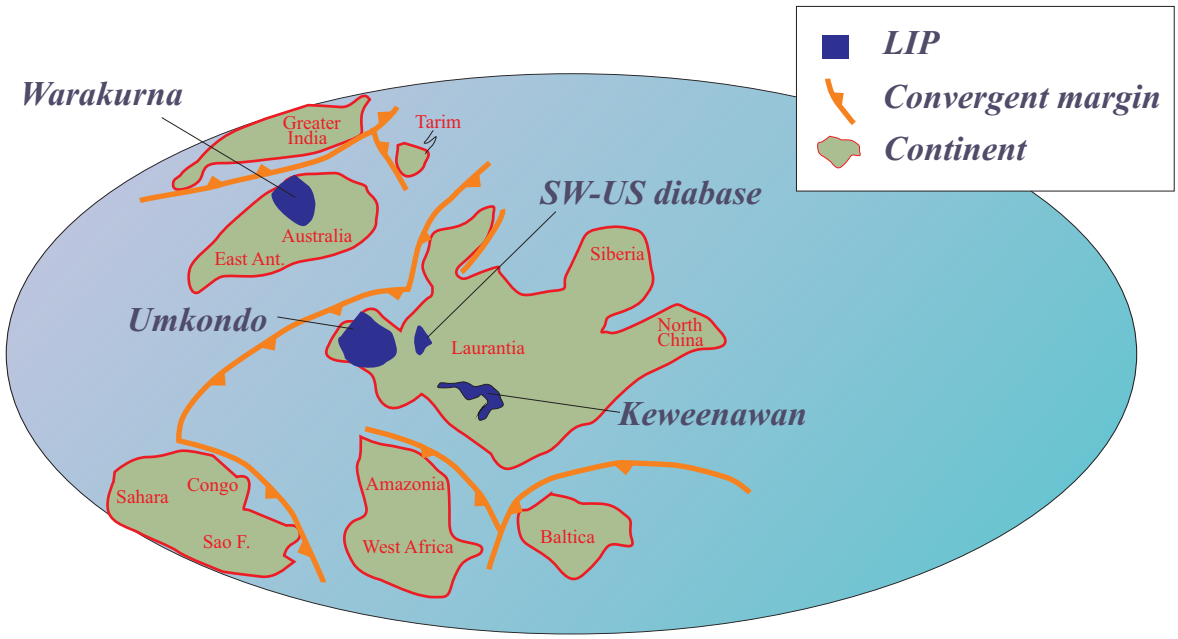
Figure


Sills, lava flows
Dykes
CAMP boundary
400 km
France
Portugal
Spain
Canada
Maroc
Algeria
USA
Mauritania
Mali
Senegal
Burkina
Faso
Guinea
Ivory
Coast
Ghana
Venezuela
Brazil
Bolivia

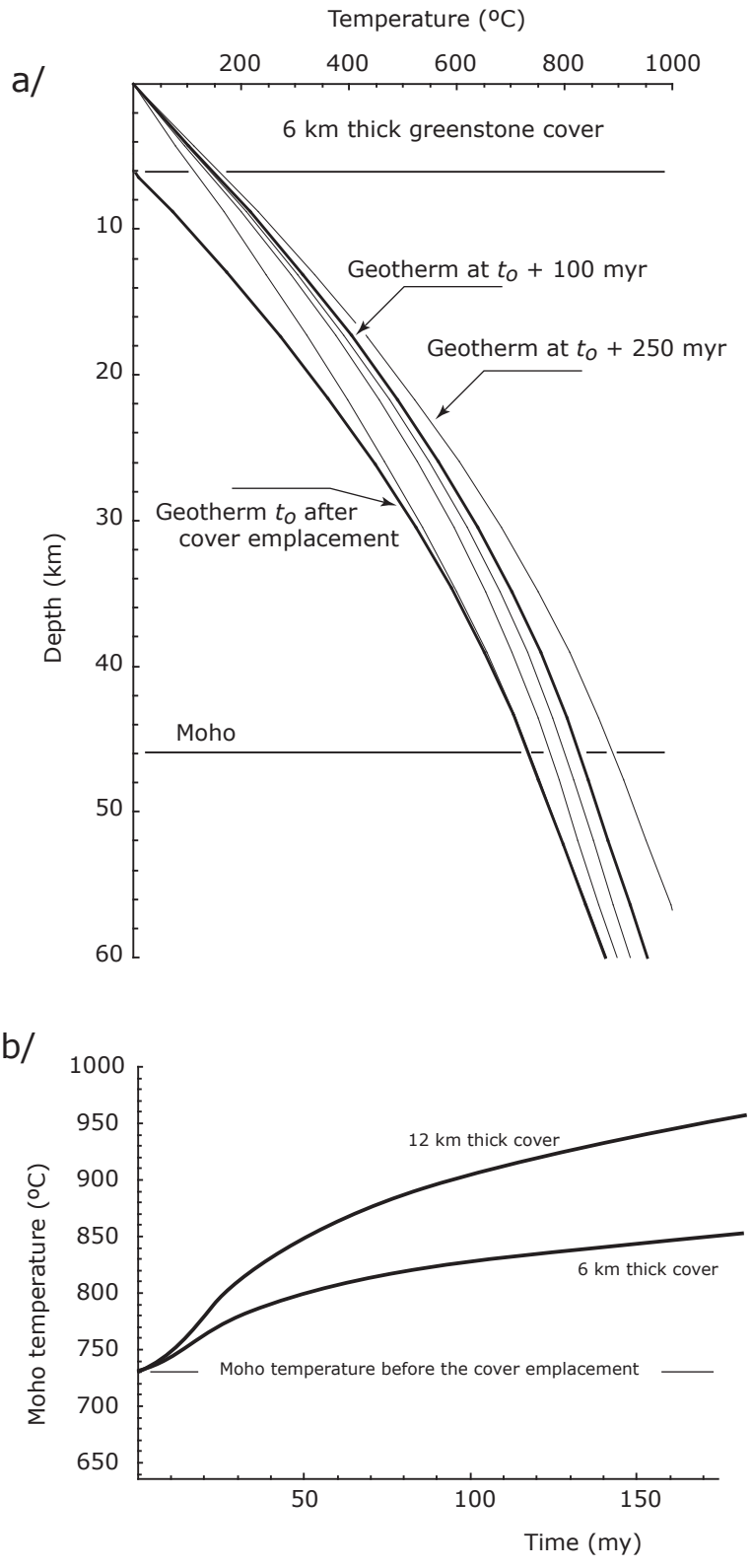
Figure


Africa
Antarctica
Ice cover
Australia
Tasmania
South
America
SMDS
SLDS
ODS
NLDS
ORDS
NRDS
IRRDS
SLeDS
GDS
Karoo
Ferrar
0
1500 km

Figure


LIP
Convergent margin
Continent
Warakurna
Greater India
Tarim
SW-US diabase
Australia
East Ant.
Siberia
Umkondo
Laurantia
North China
Keweenawan
Sahara
Congo
Sao F.
Amazonia
West Africa
Baltica

Figure


a/
Temperature (ºC)
200
400
600
800
1000
6 km thick greenstone cover
Geotherm at t_o + 100 myr
Geotherm at t_o + 250 myr
Geotherm t_o after cover emplacement
Moho
Depth (km)
10
20
30
40
50
60
b/
Moho temperature (ºC)
1000
950
900
850
800
750
700
650
12 km thick cover
6 km thick cover
Moho temperature before the cover emplacement
50
100
150
Time (my)