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Global warming of the mantle beneath continents back to the Archaean

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ABSTRACT

Throughout its history, the Earth has experienced global magmatic events that correlate with the formation of supercontinents. This suggests that the distribution of continents at the Earth's surface is fundamental in regulating mantle temperature. Nevertheless, most large igneous provinces (LIPs) are explained in terms of the interaction of a hot plume with the lithosphere, even though some do not show evidence for such a mechanism. The aggregation of continents impacts on the temperature and flow of the underlying mantle through thermal insulation and enlargement of the convection wavelength. Both processes tend to increase the temperature below the continental lithosphere, eventually triggering melting events without the involvement of hot plumes. This model, called mantle global warming, has been tested using 3D numerical simulations of mantle convection [Coltice, N., Phillips, B.R., Bertrand, H., Ricard, Y., Rey, P. (2007) Global warming of the mantle at the origin of flood basalts over supercontinents. Geology 35, 391–394.]. Here, we apply this model to several continental flood basalts (CFBs) ranging in age from the Mesozoic to the Archaean. Our numerical simulations show that the mantle global warming model could account for the peculiarities of magmatic provinces that developed during the formation of Pangea and Rodinia, as well as putative Archaean supercontinents such as Kenorland and Zimvaalbara.

1. Introduction

The growth of the continental crust is an episodic process (Moorbath, 1978). Peak production of juvenile magmatic zircons is often used to date pulses of continent formation. Major growth events are dated at 2.7, 2.5, 1.9, 1.1, 0.48, 0.28 and 0.1 Ga and are associated with mantle magmatism and orogenic activity (Condie, 2004; Condie et al., 2009-this issue). Although no continental rock has been preserved from the Hadean, 4.1 Gy old zircons of Jack Hills in the Yilgarn cratons suggest the presence of early continents (Harrison et al., 2005; see also Ernst, 2009-this issue; Eriksson et al., 2009-this issue).

Some of the major events, correlating with mantle volcanism, occurred during the lifetime of supercontinents (Yale and Carpenter, 1998; Condie, 2004). Rodinia was breaking up at 0.75 Ga, Gondwana at 0.48 Ga and Pangea was stable between 0.28 and 0.2 Ga before disrupting into the continents we see today. There are suggestions of Archaean supercontinents especially between 2750 and 2650 Ga (Aspler and Chiarenzelli, 1998) in a period of rapid and worldwide formation of juvenile continental crust. The temporal coincidence between supercontinents and widespread magmatism suggests the role of the distribution of continents in triggering large-scale mantle melting. Mantle convection studies have shown that continents can impose their wavelength on the flow (Guillou and Jaupart, 1995; Phillips and Bunge, 2005), which can lead to an increase in mantle subcontinental temperature (Grigné et al., 2005). 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 12 mW m^{-2} beneath 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 warming as high as 100 °C, triggering melting in the asthenosphere and eventually the lithosphere over a very widespread area, conjuring the name mantle global warming model (Coltice et al., 2007). However, when plumes do form they tend to start and focus below continents because of the naturally hot, dynamically favorable conditions there (Gurnis, 1988; Zhong and Gurnis, 1993; Guillou and Jaupart, 1995; Phillips and Bunge, 2005, 2007). Until now, only the (super)plume model has been invoked to explain the magmatic activity during continental growth pulses. In most cases, plume evidence is very limited and the

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geological observations are ambiguous (Courtillot et al., 1999). 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, and temperature excess (Sleep, 1990). However, 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 and Samuel, 2005).

Some of the largest CFBs on Earth do not display evidence for the plume model, even taking into account some of the complexities of thermo-chemical plumes. Off course, for most of the geological record the only traces of CFBs left are large dykes, giant dyke swarms and/or large-scale basaltic eruptions from which the recognition of a plume pattern is very difficult (Ernst and Buchan, 2002). It is only for rocks younger than 200 Myr 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 mantle global warming model fits the observations well (Coltice et al., 2007) and raises the possibility that other CFBs in the past could also be explained by this mechanism.

In this paper, we investigate the role of the mantle global warming model in generating CFBs on supercontinents over the Earth's history. We first describe the model and show 3D spherical simulations that depict 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 that some CFBs over Earth's history could be explained by this model instead of by plume models.

2. The model

2.1. Physics and phenomenology

Heat from the core probably does not exceed 25–30% of the global heat budget (Bunge, 2005; Hernlund and Labrosse, 2007). Hence, most of the flow in the mantle is driven by the top boundary layer in which continental rafts constitute mechanical and thermal heterogeneities. There is a strong feedback between mantle convection and the distribution of continental rafts at the surface of the Earth. Cold downwellings pull continents together by focusing convergence while hot upwellings push them away. At the same time, the strength and insulating power of the aggregating continents act against sub-continental downwellings, which progressively disappear, replaced by increasingly hotter mantle. This led Anderson (1982) to suggest that the geoid high within the Atlantic is a remnant of hot mantle generated below Pangea.

Aggregation and dispersal of continents profoundly impact the mantle temperature field. The pronounced insulating power of supercontinents forces the flow towards longer wavelengths. Consequently, aggregated continents have hotter sublithospheric temperature than dispersed ones, even when there is no heat coming from the core in the form of hot plumes. Coltice et al. (2007) used 2D and 3D mantle convection models with continental rafts to illustrate and quantify this mechanism. They showed that continental aggregation can generate a temperature increase larger than 70 °C in about 100 Myr. However, the simulations contain simplifications including no heat input from the core, undeformable continents, linear rheology and, in 3D experiments, fixed continents. It is extremely challenging at present to integrate realistic rheologies (temperature, depth and stress dependent) in spherical models that generate self-consistent plate-like behaviour (Tackley, 2000; Richards et al., 2001). These complexities could be crucial because of the feedback between temperature, stress and the flow through viscosity. In the following, we use simple spherical models to explore how the mantle global warming model could have worked in the past.

2.2. Numerical models

Modeling the thermal evolution of the convective mantle with continental rafts requires several ingredients. First, a physical and numerical model of mantle convection has to be used. Then, a physically consistent formulation of continents has to be inserted within the convective framework. Interaction between mantle flow and continents has to be implemented without violating the conditions of free convection in which self-organization is crucial. We use the code TERRA (Bunge and Baumgardner, 1995), which is a coupled thermal-mechanical code that satisfies the above requirements by solving the incompressible Navier–Stokes equations with continental caps in 3D spherical geometry (Bunge et al., 1997; Phillips et al., this volume). The viscosity is laterally constant but increases by a factor of 30 at the 660 km 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 current numerical models for highly vigorous convection in spherical geometry. Additionally, temperature-dependent viscosity would freeze the cold surface, generating a stagnant lid that is unrealistic for the Earth (Christensen, 1984). Our model mantle receives 85% of its heat internally via radioactive decay and the balance of 15% from the core. The Rayleigh number based on the upper mantle viscosity is 10^7, close to the assumed present-day convective vigor (Turcotte and Schubert, 2002).

Due to numerical difficulties (Lenardic et al., 2003), no 3D spherical model of mantle convection involving deformable continents has been published so far. In our models, continents are simulated by 220 km thick rigid lids. This thickness is consistent with that of stable continental lithosphere determined by geophysical observations (Artemieva and Mooney, 2001). These lids are fixed (zero velocity) and act as undeformable lithospheres. Although crude, this simplification allows for the exploration of first order thermal and mechanical impacts of continents on the convective mantle (Gurnis, 1988; Guillou and Jaupart, 1995; Lowman and Jarvis, 1999).

2.3. Simulations

Our goal is to explore the potential for mantle global warming in a younger Earth involving smaller continents. Indeed, Archaean and Proterozoic continents could have been smaller than today, making mantle melting caused by subcontinental warming after aggregation more questionable. Models are run for several convective overturns in order to obtain statistically steady-state values for the average subcontinental temperature. The total continental cover varies between 10% and 30% of the Earth's surface and involves dispersed or aggregated lids.

The temperature beneath supercontinents is always hotter than beneath dispersed continents regardless of the total continental surface area. Upon aggregation, a widespread zone of hotter mantle develops below the supercontinent because of both thermal insulation and the longer wavelength of the flow (Fig. 1). In contrast, the average temperature beneath oceanic areas (areas without continents) is similar in the aggregated and dispersed states. We compare models by taking the difference between the averaged subcontinental temperature of the aggregated and dispersed states, normalized by the average temperature beneath the oceanic area. Fig. 2 shows that increasing the size of continents increases both insulation and flow wavelength, which consequently increase the temperature. Indeed, the average temperature 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 present-day continental cover. Therefore, the larger the continental cover the larger the temperature difference between aggregated and dispersed states. These numbers are not very sensitive to the Rayleigh number. While seemingly small, a 5% temperature increase would equate to a warming of 75 K in the Archaean, assuming a conservative 1500 K temperature difference over the boundary layer at that time (Jaupart et al., 2007). Such warming is significant enough for the geotherm to cross the mantle solidus. Considering the secular

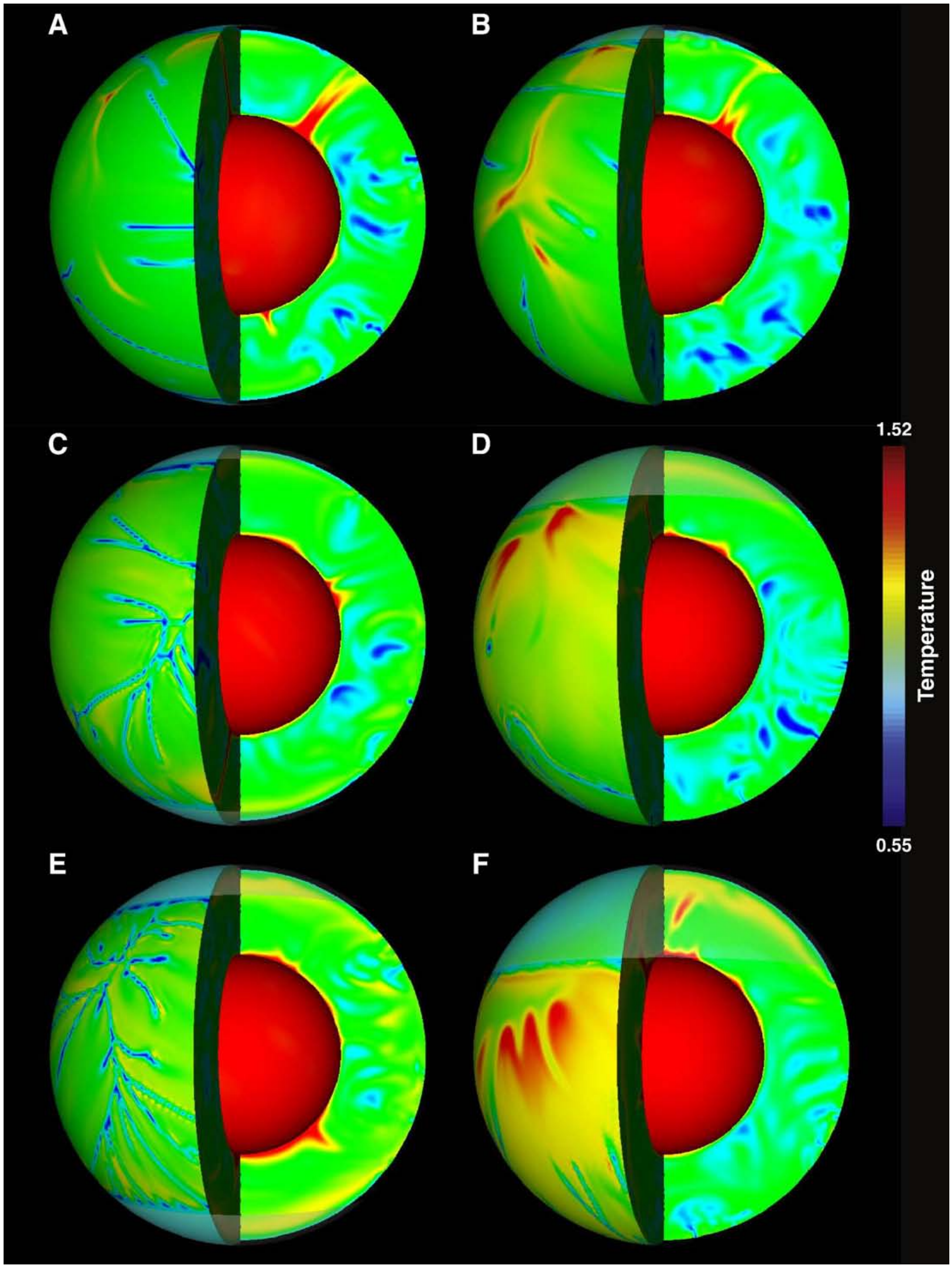

Fig. 1. Snapshots of temperature for six spherical convection models. The temperature range spans non-dimensional values of 0.55 (blue) to 1.52 (red). Gray caps represent fixed continents. The models contain A) two antipodal continents, each of which covers 5% of the surface, B) one continent covering 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. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

growth of the continents and cooling of the mantle, mantle global warming upon aggregation should have increased slightly with time, and probably remained in the range of 50 to 150 K since the Archaean according to our calculations.

3. Global warming provinces: case studies

We can compare the results of our models to observations of various large igneous provinces, for which strong evidence of plume involvement is sometimes lacking.

3.1. Pangea provinces

Pangea is the last supercontinent, the assembly and break-up of which are relatively well constrained. Therefore, it is one of the best examples for testing the mantle global warming versus hot plume models. Pangea's final stages of assembly occurred between 320 and 250 Ma (Cawood and Buchan, 2007), while its dispersal progressively unfolded 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 well constrained,

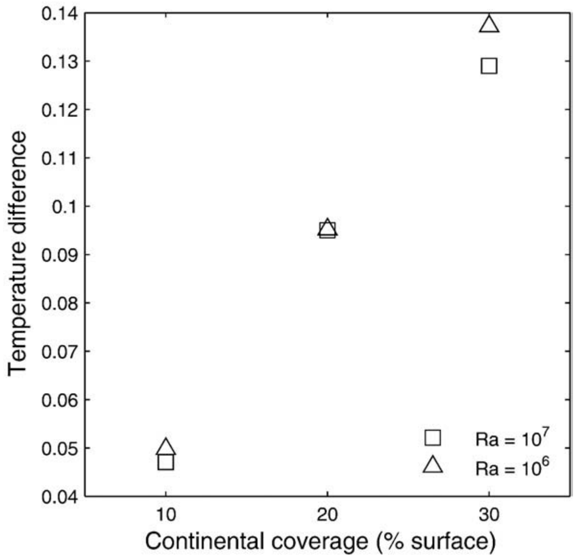

Fig. 2. Subcontinental nondimensional temperature excess of the aggregated state relative to the two continent dispersed state as a function of total continental coverage and Rayleigh number. The $Ra = 10^7$ values correspond to the models shown in Fig. 1.

allowing for comparison with CFBs linked to plumes and previous supercontinents.

Recording a 170 Myr-long story of continental dispersal, six CFBs were emplaced after the complete aggregation of Pangea: the 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 and Renne, 2003). Much debate occurs to explain the origin of Pangean CFBs, with models ranging from mantle plumes to decompression melting of the lithospheric mantle. The four youngest CFBs (Parana-Etendeka, Deccan, North Atlantic, Ethiopia–Yemen) were produced at a time when Pangea was already largely dismembered and will not be discussed here. The plume model has been questioned for the two oldest CFBs (CAMP and Karoo), emplaced while Pangea was still a supercontinent (McHone, 2000; DeMin et al., 2003; Jourdan et al., 2004; Beutel et al., 2005; Deckart et al., 2005; McHone et al., 2005; Verati et al., 2005; Jourdan et al., 2006, 2007c).

In what follows, we examine critical features of the CAMP and Karoo CFBs in order to assess the pertinence of the plume versus global warming models. These features 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 and the chemical composition of the magmas.

3.1.1. The CAMP case (200 Ma)

The CAMP is the largest CFB on Earth, extending over four continents (Europe, Africa, North America and South America) and covering ca. 10^7 km^2 (Fig. 3). It consists of a) huge sills (up to 10^6 km^2) 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 the Eastern North America margin and in West Africa and Guyana (Dupuy et al., 1988; Deckart et al., 1997; Verati et al., 2005; McHone et al., 2005) and c) some lava flow 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 m thick. The CAMP extends mainly over the peri-cratonic Pan-African to Variscan orogens and intra-cratonic basins (Brazilian and West-African cratons).

Basalts, gabbros and ultramafic cumulates of the CAMP are dominantly low-Ti tholeiites (Bertrand, 1991), with minor high-Ti tholeiites restricted to limited areas in Liberia, Guyana and Brazil (Dupuy et al., 1988; DeMin et al., 2003; Deckart et al., 2005).

The CAMP is the oldest CFB, postdating by several tens of millions of years the final stages of Pangea's aggregation. It predates by ca. 10 Myr the earliest disruption of the supercontinent and the initial opening stages of the central Atlantic Ocean ~190 Ma ago (Deckart et al., 1997; Marzoli et al., 1999; Sahabi et al., 2004; Nomade et al., 2007). 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; Hames et al., 2000; Marzoli et al., 2004; Knight et al., 2004; Verati et al., 2007), taking into account a ~1% bias between $^{40}Ar/^{39}Ar$ and U/Pb chronometers (Schaltegger et al., 2008).

The origin of the CAMP is often attributed to the impingement of a mantle plume head 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). More recently, however, this model has been refuted in favor of a shallow mantle origin (McHone, 2000; DeMin et al., 2003; Beutel et al., 2005; Deckart et al., 2005; McHone et al., 2005; Verati et al., 2005) on the basis of several features including the following:

- No hotspot track, synchronous with the opening of the Atlantic ocean, is visible on the ocean floor (McHone, 2000; McHone et al., 2005).
- The geometry of the CAMP is not compatible with the plume model, which predicts a radial spreading over an area 2000–2500 km across (Campbell and Griffiths, 1990). In contrast, the CAMP covers a region elongated (Fig. 3) over ~8000 km 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). This scenario is however invalidated by the absence of a northward age propagation. On the contrary, if any migration of the magmatic activity existed it propagated southward (Baksi, 2003; Nomade et al., 2007).
- Contrasting with its large surface area, 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 (~2–3 km) observed in plume-related CFBs such as the Deccan or Ethiopian traps. The preservation of thin lava units interstratified into sedimentary horizons 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 upon 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 dykes (see 1:1,000,000 geological map of Mauritania, 1975). In addition, the subsurface dyke swarm hypothesized in Senegal based on a magnetic survey is of unknown age and origin. Therefore, this ENE–WSW to E–W branch of the radial dyke pattern is not constrained.
b) Dykes that would fit the radial pattern (e.g. NE–SW dykes in northern CAMP) display inherited trends reactivated during the CAMP event. Some dyke swarms (e.g. in Guyana–Surinam) are composite and include both Proterozoic and CAMP dykes (Deckart et al., 1997, 2005), whereas the dyke swarm intruding the Reguibat

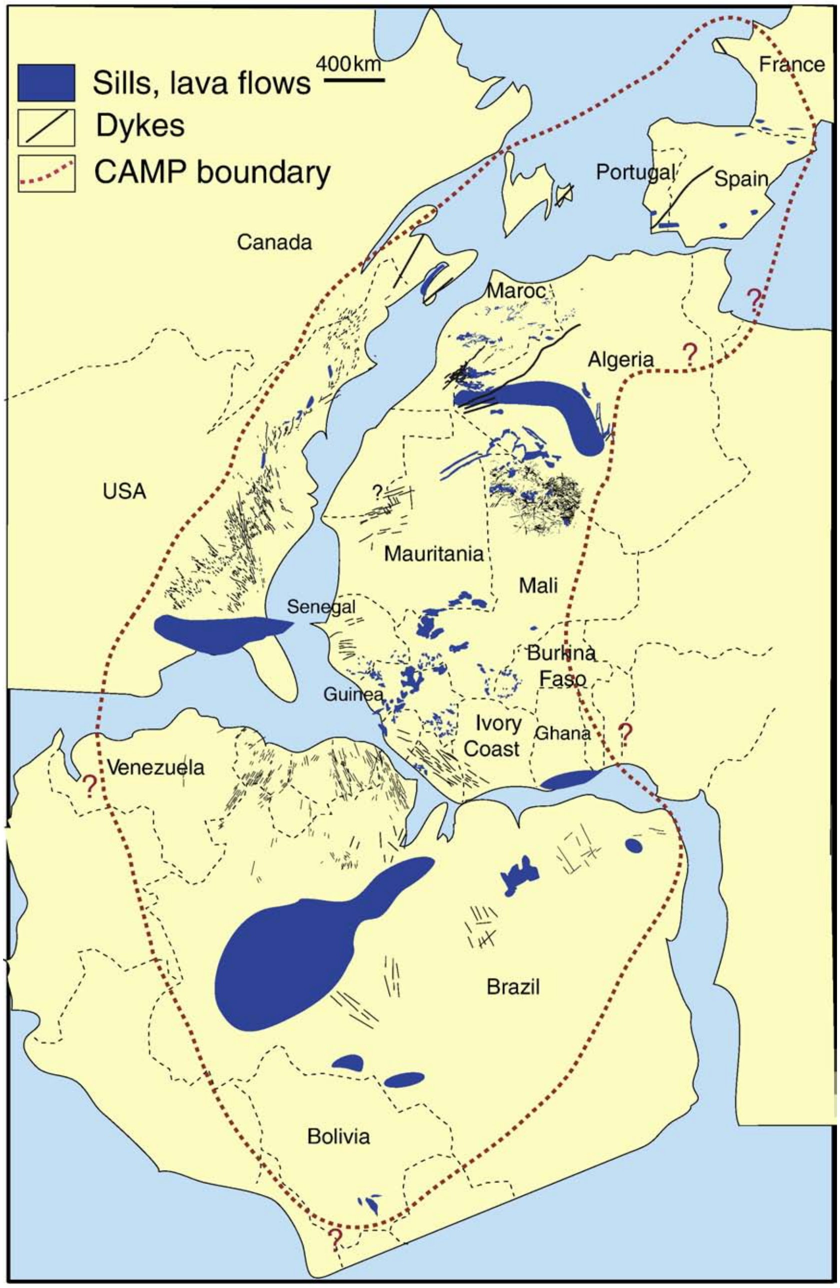

Fig. 3. Extension of the Central Atlantic Magmatic Province at 200 Ma. Compiled from Bertrand (1991), Deckart et al. (1997), Marzoli et al. (1999), McHone (2000), Bertrand et al. (2005), and Chabou et al. (2007).

shield in NW Mauritania is poorly dated providing so far only Proterozoic ages (Dosso, 1975).

c) Some dyke swarms display a more complex distribution than that expected in a "radial pattern". For example, 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 previously enriched by ancient subduction processes (Bertrand et al., 1982; Bertrand, 1991; Pegram, 1990; Puffer, 2001; Cebria et al., 2003; DeMin et al., 2003; Deckart et al., 2005; Verati et al., 2005). The subordinate chemically and isotopically distinct high-Ti group indicates the contribution of the asthenospheric mantle (Dupuy et al., 1988; Deckart et al., 2005).

Finally, none of the CAMP areas on which Nd–Sr–Pb isotope data are available (Pegram, 1990; Jourdan et al., 2003; Cebria et al., 2003; Deckart et al., 2005) bears the HIMU signature diagnostic of the present-day Atlantic mantle plumes (Cape Verde, Fernando do Noronha, Asuncion), which would have been in a favorable position in late Triassic times to trigger the CAMP magmatism (Morgan, 1983).

3.1.2. The Karoo case (180 Ma)

The Karoo CFB extends over more than 3×10^6 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 with 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 Archaean Kaapvaal–Zimbabwe cratons, the Proterozoic Limpopo belt and the Permian–Jurassic Karoo sedimentary basins.

Karoo lava remnants consist of: a) tholeiitic basalt lava piles in Lesotho (Marsh et al., 1997), Botswana and adjacent Zambia–Zimbabwe (Wigley, 1995; Jones et al., 2001; Jourdan et al., 2005) and 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 include 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. 4), 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 My 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,b). The increase of magmatic activity early in the history of the province and the emplacement of abundant sills in C-rich layers from the Karoo sedimentary basins seem 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 volcanism (e.g. Campbell and Griffiths, 1990; Courtillot et al., 1999; Ernst and Buchan, 2002). One of the main arguments supporting this plume model was the presence of the so-called triple junction supposedly 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 with geochemical analyses, reveal that these dyke swarms unambiguously include Proterozoic dykes (Jourdan et al., 2004, 2006). For instance, 12% of the N110° Okavango dyke swarm is composed 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 plume head but was inherited from older structures in 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.

The lack of a volcanic track linking the Karoo CFB to a present day hot spot also questions the validity of the hypothesized Karoo mantle plume.

The involvement of a deep mantle plume in the genesis of the Karoo CFB is also questioned on the basis of geochemical arguments. Only a few geochemical studies conclude that a mantle plume contributes to the chemical composition of the Karoo magmas (Ellam and Cox, 1991). Based on isotopes and trace element 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

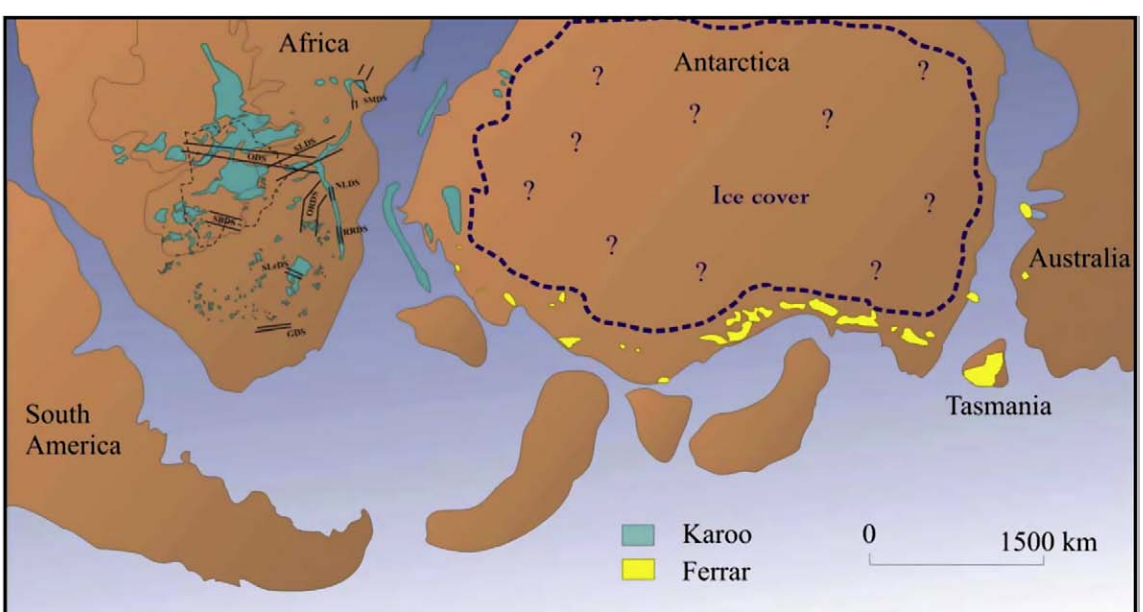

Fig. 4. Extension of the Karoo and Ferrar provinces at 180 Ma. Modified from Jones et al. (2001) and Hergt et al. (1991).

distribution is pre-requisite of any petrogenetic model applied to the Karoo CFB (Sweeney and Watkeys, 1990; Jourdan et al., 2007c).

3.2. Rodinia provinces

As for Pangaea, the assembly and breakup of the Proterozoic mega continent Rodinia was accompanied by the emplacement of major CFBs and dyke swarms (e.g. Ernst and Buchan, 2001; Ernst et al. 2008). Since most lava flow components have been eroded, most of these CFBs are scarce remnants occurring as altered and metamorphosed sills and dyke swarms.

From 1.3 Ga until its final assembly (~900 Ma; Li et al., 2008) and subsequent breakup, Rodinia recorded many CFB events (e.g. Ernst et al., 2008). Of particular interest is a cluster of CFBs at around 1.1 Ga (Fig. 5), each of them covering a surface of several 10^6 km^2 and thus, equivalent in size to the Pangean provinces mentioned previously.

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

3.2.1. The Umkondo case (1.1 Ga)

Paleomagnetic and geochronological (mainly zircon U/Pb analyses) investigations attribute a common origin to 1.1 Ga tholeiitic rocks occurring throughout northern South Africa, Botswana, Zimbabwe and possibly Antarctica (Hanson et al., 2004). This CFB is called Umkondo based on the name of the Umkondo Zimbabwe dolerite formation of the same age (Hanson et al., 1998). It consists of tholeiitic mafic intrusions (sills and dykes swarms, e.g. Jourdan et al., in press) and scarce remnants of eroded basaltic lava-flows emplaced over an estimated surface of $\sim 2.5 \times 10^6$ km^2 (Hanson et al., 2004). Robust ages, clustering around 1108 ± 3 Ma, have been obtained using zircon and baddeleyite U/Pb TIMS techniques. These ages reveal magmatism of relatively short duration synchronous with the formation of Rodinia (Hanson et al., 1998; Dalziel et al., 2000), and in particular with the collision between the Laurentia and Kalahari cratons (Kibaran Grenville-Llano and Namaqua-Natal Orogenies; 1.4–1.0 Ga, Jonhson and Oliver, 2000).

The Umkondo CFB mostly consists of low-Ti basaltic rocks. 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, and 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., in press). The compositional similarity with the spatially overlapping 180-Ma Karoo CFB suggests that both provinces originated from a very similar mantle source such as a common enriched stabilized SCLM attached to the African plate and is hard to reconcile with the melting of two distinct mantle plumes. A model such as the mantle global warming model presented in this paper is a viable scenario to explain the fusion of the African SCLM.

3.2.2. The Laurantian large igneous provinces (1.1–1.07 Ga)

The Laurantian provinces include the Keweenawan "mid-continental rift" ($\sim 2 \times 10^6$ km^2; Cannon, 1992) and the South Western U.S. diabase ($\sim 0.4 \times 10^6$ km^2; Ernst et al., 2008) provinces. Both provinces were emplaced sub-synchronously with 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, the best studied of the two Laurantian provinces, 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. Major elements and Sr–Nd–Pb isotope indicate that the magma is derived from enriched mantle reservoirs (Shirey et al., 1994) and share common features with CFBs such as Umkondo and Karoo. For example, both Karoo and Keweenawan eruptive sequences have an early activity represented by picrites with a strong lithospheric signature (Shirey et al., 1994; Jourdan et al., 2007a,b). Possible mantle sources range from astenospheric/subasthenospheric mantle plume to enriched lithospheric mantle (or a combination of both) with variable degrees 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 from the partial melting of Archaean crust (Vervoort et al., 2007). The South Western U.S. diabase is less known, with mainly geochronological and paleomagnetic data available (e.g. Shastri et al., 1991; Heaman and Grotzinger, 1992; Ernst and Buchan, 2001; Ernst et al., 2008). This province consists mostly of up to 450 m-thick doleritic sills for which no chemical data has been published to our knowledge. Therefore, no correlation with other CFBs can be made at this stage.

3.2.3. The Warakurna large igneous province (1.05 Ga)

This province regroups coeval magmatism distributed over $\sim 2.5 \times 10^6$ km^2, mostly across Western and northern Australia (Wingate et al., 2004). It is primarily composed of silicic and basaltic lava flows and

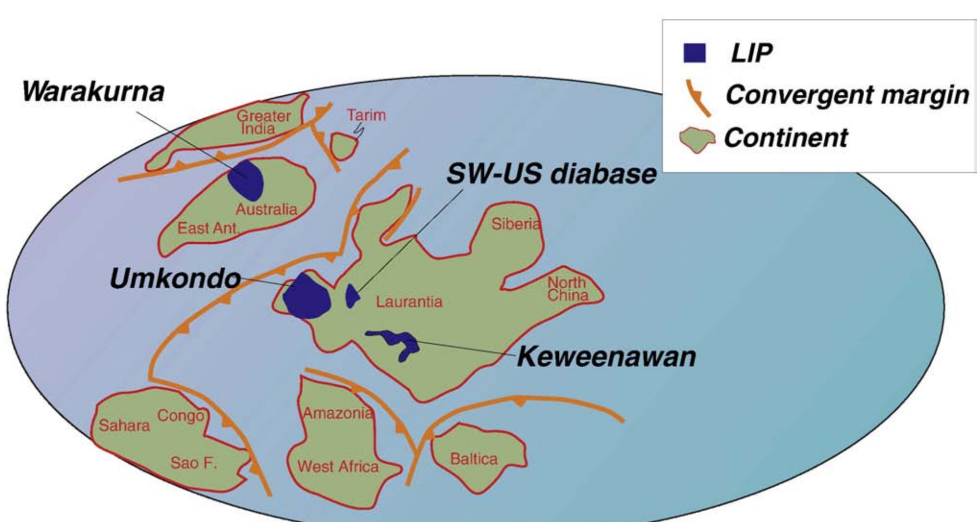

Fig. 5. Estimated extension of the Umkondo, Warakurna and Keweenawan provinces at the time of Rodinia aggregation (modified from Li et al., 2008 at 1050 Ma).

mafic and ultramafic intrusions. The age of the province, determined using zircon SHRIMP U/Pb techniques, yielded a mean age of 1076 ± 3 Ma, although minor intrusions continue until ~1050 Ma. Investigation using more robust CA-TIMS U/Pb zircon analysis would be desirable to obtain a more accurate estimate of the age along with a better control on the duration of the magmatism. 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 Warakurna CFB consists of dolerite with sericitized plagioclase and amphibolitized pyroxene and shows an enriched tholeiitic composition with SiO_2 ranging from 48 to 55 wt.%. They are enriched in LILE and show a 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. Yet, no mantle plume track has been found suggesting that the role of a mantle plume, if any, was confined to that of a heat purveyor. We propose here that the mantle global warming model might reconcile all of these observations and provide the temperature anomaly required to melt an easily fusible subduction-enriched SCLM.

3.3. Archaean provinces

Looking for CFBs caused by global warming beneath putative Archaean supercontinents is a difficult task because of the intrinsic difficulties of working on such old rocks. The evolution of the Earth during its first couple of billion years certainly involved sudden and violent crises, possibly including large scale reorganizations of mantle convection or superplume events (Stein and Hofmann, 1994; Breuer and 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 the isotopic age distribution of detrital zircons points to a major peak production of juvenile continental crust (Gastil, 1960; Condie, 2001; Rino et al., 2004, 2008). 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 komatiites, the geochemistry of most late Archaean komatiites demands a deep source origin and requests 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 involving several giant plumes is indeed very appealing (Isley and Abbott, 1999; Condie, 2004; Barley et al., 1998), but not without problems. In this section, we review the extent of late Archaean volcanism and the problems with linking this volcanism to a superplume event.

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 Supergroup extents over 3×10^5 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 (Schmitz and Bowring, 2003).

In the Zimbabwe craton, the base of the 2.7 Ga old craton-wide Ngezi Group includes the Zeederberg CFB, which covers an area over 2.5×10^5 km^2 (Prendergast, 2004). Its base is made of the submarine Reliance formation, which includes up to 2 km thick interlinked komatiite sills extending over 100 km across and intercalated within sandstones and pillowed basalts (Prendergast, 2004). 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, 2002). This volcanic event, which covers a surface over 10^5 km^2 (Bleeker, 2002) was followed from 2.69 to 2.66 Ga by calc-alkaline volcanism and Tonalite–Trondhjemite–Granodiorite plutonism (Bleeker, 2002).

In the East goldfield province of the Yilgarn Craton, a 12 ± 2 km 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, 1997; Blake and Barley, 1993; Arndt et al., 2001; Blake, 2001). The Fortescue, which covers a surface of at least 1.8×10^5 km^2 (Eriksson et al., 2002), is made of ca. 7 km thick volcano-sedimentary rocks including a series of tholeiitic basalts, minor komatiites, felsic volcanics and clastics rocks (Hickman, 1983; Thorne and Tyler, 1996; Thorne and Hickman, 1998; Blake, 2001).

Other greenstones emplaced at 2.7 Ga include the Stillwater intrusion in the Wyoming craton and the Ramagiri–Hungund composite greenstone belt and Gadwal greenstone belt in Dharwar craton.

Despite the preferential preservation of their stronger nucleus and their overall limited extent, most if not all Archaean cratons recorded a major magmatic event at 2.7 ± 0.05 Ga. This event covered a cumulative surface area $>10^7$ km^2 to thicknesses up to 15 km.

3.3.2. Models for the 2.75–2.65 Ga magmatic crisis

In all the above cratons, 1) the synchronism of bimodal volcanism, felsic plutonism and sedimentation, 2) the calc-alkaline composition of some basalts, despite minor komatiites, and 3) the syn- to late-contractional deformation, have led to interpretive models involving multiple subduction zones and discrete back-arc basins between micro-plates. The collisional aggregation of these micro-plates eventually led to the formation of late Archaean cratons. This model has been proposed for 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 and the Kaapvaal craton (Dirks and Jelsma, 1998; Jelsma and Dirks, 2002; Horstwood et al., 1999). 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 CFBs. Such a model has been proposed for the Slave province (Bleeker et al., 1999a,b), the Zimbabwe craton (Prendergast, 2001; Wilson, 1979), the Superior Province (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 CFBs in many cratons can be linked to the thermal insulation effect associated with the emplacement the thick greenstone covers on a continental crust enriched in radiogenic elements (Rey et al., 2003). It is worth noting that recent Hafnium isotopic data on zircons show that, contrary 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 anatexis alone, rather than the formation of juvenile crust in 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 (McLennan and Taylor, 1980; Rey and Coltice, 2008).

3.3.3. The cause of the 2.75–2.65 Ga volcanic crisis: plume versus mantle global warming

Hitherto, only mantle plumes were considered a viable setting for deep mantle partial melting. However, the application of the hot plume model in the late Archaean is not without problems.

- The late Archaean CFBs were emplaced over a few tens of Myr, which is hardly compatible with the mantle plume model that requires a short 1–2 Ma duration (although we note that a short duration, e.g. for CAMP, is on the other hand not proof for a mantle plume source). However, in a plume framework, such a long duration could be at best explained by multiple and diachronous plume heads impinging in succession on the base of the lithosphere.
- Subduction-related calc-alkaline volcanism is often spatially and temporarily associated with komatiites (eg. Wyman et al., 2002). While 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) in the same location.
- Komatiites generally account for less than 5% of Archaean CFBs (Viljoen and Viljoen, 1969; de Wit and Ashwal, 1997), the bulk of which are made of basalt involving lower degrees of melting. Therefore, large partial melting of a deep fertile plume can be ruled out in favor of a smaller melt fraction in a refractory upper mantle source.
- 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). The downward migration of the melting source is difficult to explain in the context of a rising plume.
- The deepening of the melting source is also proposed in the Pilbara to explain the increase in FeO and incompatible elements in progressively younger terms of the Fortescue Group (Arndt et al., 2001).

Mantle global warming above aggregating continents could provide a viable alternative to plume events. Williams et al. (1991) proposed that a supercontinent, Kenorland, formed at around 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, 2002). 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). Since supercontinents aggregate preferentially above mantle downwellings, it is not clear how mantle plumes could have dominated the flow field beneath supercontinents during their aggregation. In contrast, our numerical experiments show that a supercontinent covering a mere 10% of the Earth's surface could have triggered the progressive warming of the sub-continental mantle by up to 75 °C. In the context of an Archaean mantle 100 to 150 °C hotter than the present-day mantle, the mantle global warming model could elegantly explain not only the occurrence of komatiites (deep partial melting) but also the progressive deepening of the melting source as recorded by the chemistry of basalts. Indeed, upon the progressive warming of the sub-supercontinental mantle, the partially melted region would involve progressively deeper mantle. Therefore, there is little rationale to invoke complicated models involving mantle plumes impinging on subducting slabs to explain the intercalation of komatiites with basalts.

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

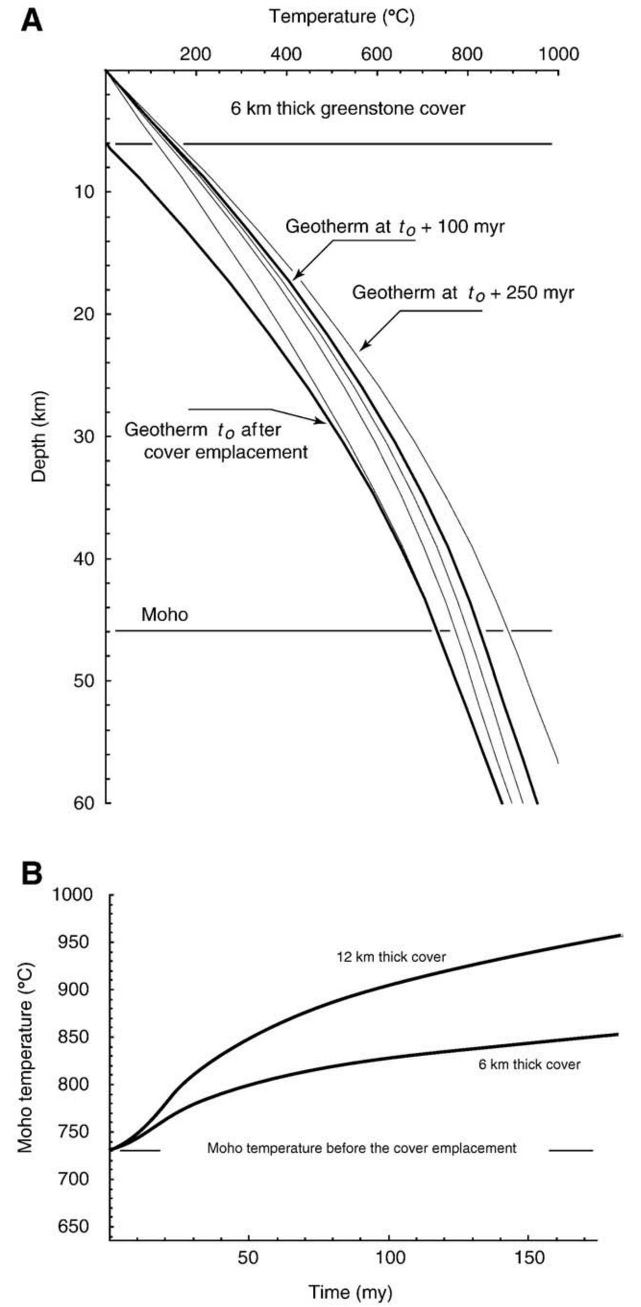

Fig. 6. Continental warming following the emplacement of a 6 km thick CFB on a 2.75 Ga old continental crust (heat production of 10^{-6} W m^{-3}, mantle heat flow of 0.025 W m^{-2}, 1000 J mol^{-1}, thermal diffusivity of 0.9 10^{-6} m^2 s^{-1}, continental crust thickness of 40 km). A) Geotherm at to, to + 5 Myr, to + 25 Myr, to + 50 Myr, to + 100 Myr, to + 250 Myr 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.

4. Conclusions

The numerical models of 3D mantle convection presented here explore a realistic physical mechanism from which quantitative predictions can be made and tested against data. In the framework of the mantle global warming model associated with supercontinents, the models predict that:

- the temperature increase causing melting does not exceed 100 °C
- heating occurs over an area comparable to the supercontinent
- magma is sourced mostly from the asthenosphere and continental lithosphere
- melt extraction is controlled by tectonics.

As shown by our 3D spherical convection simulations, the presence of a supercontinent at the surface of the Earth has a major impact on the convective flow within the mantle. As a consequence, the subcontinental temperature is 5–15% higher than for cases with dispersed continents. The smaller the continental cover, the smaller 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 75 °C over an area comparable to the size of the supercontinent. The mechanism of mantle global warming, as an alternative to mantle plumes, is then a viable hypothesis for the origin of some CFBs since the Archaean. However, the plume model remains valid for many flood basalts and particularly for those that are not related to supercontinents.

The mantle global warming model is consistent with the observations made on some CFBs in the geological record. The best example is the CAMP (200 Ma), the largest magmatic province on Earth, but a growing number of observations point also to the Karoo CFB (180 Ma) as a candidate for a global mantle warming origin. 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 1.1 Ga. The global magmatic crisis around 2.7 Ga in the Archaean could also be explained by heating beneath the forming Kenorland and Zimvaalbara supercontinents, for which plumes are difficult to invoke.

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