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# Contribution of mantle plumes, crustal thickening and greenstone blanketing to the 2.75–2.65 Ga global crisis

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#### Abstract

Assuming that the period 2.75–2.65 Ga corresponds to a single, but global, geodynamic event, we investigate—through numerical experiments—the mechanisms that could have led to the profound continental reworking that occurred at that time. Although the extent of the crisis at the Earth's surface pledges in favour of the involvement of mantle plumes, our numerical experiments suggest that the thermal impact of mantle plumes is unlikely to explain both the amplitude and timing of the thermal anomaly, as observed in the Superior Province (Canada) and the Yilgarn Craton (Australia). Similarly, moderate crustal thickening can not lead to significant reworking of the continental crust within the observed time constraint. Crustal thickening with a factor  $\geq 1.5$  is also unlikely because it is not consistent with the moderate metamorphic grade observed at the surface of many Archaean cratons. Burial of a radiogenic crust under a 10 km thick greenstone cover also falls short of explaining, not so much the amplitude and the extent, but the timing of the thermal anomaly. In contrast, the combination of the thermal anomaly related to the greenstone blanketing effect with the heat transfer from a plume head spreading at the top of the thermal boundary layer can adequately explain the amplitude, the timing, and the extent of the 2.75–2.65 Ga crisis.

Our favoured model involves a global rearrangement of convection cells in the deep mantle and formation of multiple mantle plumes. The greenstones emplaced at the surface and the plumes that spread in the thermal boundary layer contributed to heat the crust from both above and below. This produced massive crustal partial melting that reached its climax ca. 40 Myr after the emplacement of the plumes and associated greenstone cover rocks. This led to gravitational instabilities in the crust, as dense greenstone cover rocks began to sink into the thermally softened crust and granite domes rose in response. The extraction of heat-producing elements toward the upper part of the crust has contributed to the cooling and stabilisation of the cratons. This succession of events, which is not incompatible with plate-tectonic processes, may have profoundly changed the nature of the crust exposed at the surface and could explain the contrasting geochemical signatures of Archaean and post-Archaean shales. © 2003 Elsevier B.V. All rights reserved.

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# 1. Introduction

The thermo-tectonic evolution of Archaean continental lithosphere is a controversial issue, and more specifically, so is the origin of the thermal anomalies that generated large volumes of greenstones and granites during the Late Archaean. Greenstone blanketing (West and Mareschal, 1979), crustal thickening (England and Bickle, 1984; Barley et al., 1989), subduction of hot oceanic lithosphere (Martin, 1986), mantle plume emplacement (Campbell and Hill,

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1988), and plume-arc interaction (Wyman et al., 1999) have all been proposed to explain the creation of juvenile crust and the massive reworking of pre-existing continental crust between 2.75 and 2.65 Ga in various Archaean cratons. That period of the Earth's history corresponds to the largest anomaly (Fig. 1) in the isotopic age distribution of juvenile continental crust (Condie, 1995). It corresponds also to a dramatic change in the composition of the crust exposed at the surface of the Earth (Taylor and McLennan, 1985). In particular, the deposition at the Archaean/Proterozoic boundary of fine-grained sediments with negative Eu-anomalies and radiogenic element enrichment points toward a major period of intracrustal partial melting and crustal differentiation during the Late Archaean (Taylor and McLennan, 1985, 1995). This period, which represents the largest global thermo-tectonic crisis in the history of the Earth, could be the consequence of the development of large-scale instabilities in the mantle (Stein and Hofmann, 1994; Breuer and Spohn, 1995; Condie, 1995, 1998, 2000). If this is true, one single set of related events, directly linked to mantle plumes or not, could have led to the production of juvenile crust and the reworking of pre-existing crust (see Machetel and Thomassot, 2002, for a modern analogue). Based on a comparison of the ages of Late Archaean granites and greenstones in the eastern part of the Yilgarn Craton and the Abitibi subprovince of the Superior Province, Nelson (1998) agreed that Late Archaean greenstones were related to global-scale catastrophic convective overturn of the Earth's mantle, whereas Late Archaean



Fig. 1. Evidence for global reworking of the continental crust during the Late Archaean. (a) Distribution of U/Pb zircon ages in juvenile continental crust between 3 and 1 Ga. Abundance is proportional to surface distribution of juvenile age provinces. This graph (redrawn from Condie, 2000), supports the hypothesis that the 2.75–2.65 Ga event was the most prominent thermo-tectonic crisis enjoyed by the Earth... so far. (b) Secular variation of Eu/Eu\* ratio in fine-grained sedimentary rocks. (c and d) Secular variation of Th and U concentration in continental shales. Graphs b, c and d redrawn from Taylor and McLennan (1985). Together these graphs show that the average composition of the crust exposed at the surface of the Earth dramatically changed during the Late Archaean. Granitoids, derived from partial melting in the stability field of plagioclase, and emplaced near the Earth surface could explain the sudden enrichment of the upper crust in radiogenic elements, and the decrease of the Eu anomaly. We argue that the 2.75–2.65 Ga event was responsible for this dramatic change.

granites were derived from local plate-tectonic processes similar to present-day, tectonic plate interaction. This paper discusses some possible links between the global reorganisation of mantle convection and the events that affected the continental crust at that time.

In the past two decades, the geological and geochronological database on some Late Archaean cratons has progressed to the point that it is now possible to test a number of scenarios that could explain the extent and magnitude of the Late Archaean global event, as well as the timing of its development. As well, realistic constraints have been placed on the average composition of the Archaean crust (Taylor and McLennan, 1985, 1995), its average thickness now and in the Late Archaean (Galer and Mezger, 1998), the amount of erosion since the stabilisation of the crust (Galer and Mezger, 1998), and the magnitude of the mantle basal heat flow under Archaean terranes (e.g. Boyd and Gurney, 1986; Jaupart et al., 1998; Lenardic and Moresi, 2000; Russell et al., 2001). We use these constraints to define a reasonable steady-state thermal structure for an average 2.75 Ga continental lithosphere, and to investigate the thermal impact from (1) the heat transfer related to the emplacement of a mantle plume, (2) homogeneous crustal thickening, and (3) the emplacement of a greenstone cover. Our study suggests that the combination between heat transfer from multiple mantle plumes and thermal insulation from a thick greenstone cover can lead to massive crustal reworking within a few tens of million years after the emplacement of the plume and greenstones.

# 2. Geological constraints

Many Late Archaean cratons share a number of first order features. Rather than giving an exhaustive description of the geology of each of them, we will focus on key observations made in the Yilgarn Craton (Australia) and the Superior Province (Canada), two of the largest Archaean cratons. Both cratons have a Late Archaean history (Fig. 2), starting with the emplacement of a greenstone cover at ca. 2.75–2.68 Ga that precedes, and overlaps with, the emplacement of large volumes of granitic melts and felsic volcanism at 2.72–2.65 Ga (e.g. Nelson, 1998). Therefore, the time lag between greenstone emplacement and granite formation is of about 20–40 Myr (Fig. 2), with the peak production of granite occurring ca. 40 Myr after the emplacement of the oldest greenstones. The main regional phase of deformation corresponds, for both cratons, to a number of short-lived contractional events that are broadly contemporaneous with granite emplacement.

In both the Superior Province and Yilgarn Craton, little of the pre-2.75 Ga history was preserved. In the eastern part of the Yilgarn Craton, both the granites and felsic volcanics preserve zircon xenocrysts with ages ranging from >2.90 to 2.69 Ga (Hill et al., 1989, 1992; Hill and Campbell, 1993; Nelson, 1997), while rare remnants of ca. 2.96–2.93 Ga greenstones are preserved near Norseman (Hill et al., 1992; Nelson, 1997). The western part of the Yilgarn Craton contains older gneisses and greenstones in the range of 3.0–2.8 Ga (Pidgeon and Wilde, 1990; Nelson, 1995; Savage et al., 1995; Schiøtte and Campbell, 1996; Wang et al., 1996, 1998).

Although the Abitibi subprovince developed in a back-arc setting starting at 2.73 Ga, events at 3.3–3.2 Ga, 3.0–2.90 Ga and 2.87–2.85 Ga are locally preserved in some of the oldest blocks of the Superior Province (see Easton, 2000, and references therein). As for the Yilgarn Craton, available geochemical data points toward an older basement (Barley, 1986; Champion and Sheraton, 1997; Henry et al., 2000).

The above data implies for both cratons that the ca. 2.75-2.68 Ga greenstone cover was largely emplaced on an older felsic crust. However, the near absence of granites and felsic gneisses older than the ca. 2.75-2.68 Ga greenstone sequences and the large production of 2.7-2.65 Ga granitoids imply a complete reworking of the older crust for both cratons, and in the case of the Abitibi subprovince, a profound reworking of the juvenile volcanic arc. In both cratons, the reworking event has occurred after the emplacement of the greenstones, and since no stratigraphic contact with an older basement is preserved at the base of the greenstones, the reworking must have affected the crust up to the base of the greenstone cover. Indeed, investigations of metamorphic rocks in both cratons show that temperatures of up to 750 °C were reached at a depth of 15-20 km (e.g. Binns et al., 1976; Gunn, 1976; Lindsley, 1983; Easton, 2000).

A number of hypotheses can be made on the origin of the thermal anomaly that produced the partial melting of the crust. One model involves relaxation



Fig. 2. Summary of Late Archaean geological events in (a) the Yilgarn Craton and (b) the Superior Province. Data sources for the Yilgarn: Nelson (1995, 1997, 1998), Hill et al. (1989, 1992), Hill and Campbell (1993), Nemchin et al. (1994) and Kent et al. (1996). Data sources for the Superior Province: Corfu (1993), Corfu and Davis (1991), Davis et al. (2000), Davis (2002) and references therein.

through time of a temperature anomaly produced by crustal thickening (e.g. England and Thompson, 1984). This event, if it existed, must have occurred after the deposition of the greenstone cover since deformation post-dates the greenstones. In the past decade, the hypothesis of crustal thickening related to the amalgamation of a number of terranes has been widely accepted in both the Superior Province (e.g. Corfu, 1987; Percival and Williams, 1989; Card, 1990; Ludden et al., 1993; Kimura et al., 1993) and the Yilgarn Craton (Myers, 1993, 1997; Barley et al., 1998). Although well documented in the Abitibi-Wawa subprovince, the basis for this model seems to be increasingly less convincing. Indeed, geochronological data no longer supports the view that deformation and metamorphism migrate across the province (Davis et al., 2000), and greenstones investigated across terrane boundaries display remarkably similar characteristics and ages (Heather et al., 1995; Heather, 1998; Ayer et al., 1999, 2002). This may suggest that either the greenstones post-date crustal accretion, or deformation was intracratonic and did not involve accretion. In addition, the presence of unconformities in many greenstone sequences of the Superior Province, including in the Abitibi-Wawa subprovince (Thurston, 2002), suggests an autochthonous or parautochthonous origin for the greenstones, at least locally. This is further supported by the presence in younger strata of inherited zircons with U-Pb ages equivalent to that of zircons from underlying strata (Heather, 1998; Ayer et al., 1999, 2002). This may suggest that accretion of crustal blocks and volcanic arcs only reflects part of the craton's development, or that it only applies to the Abitibi-Wawa subprovince.

In the Yilgarn Craton, the emplacement of a mantle plume was proposed as an alternative to crustal thickening for the generation of Late Archaean granites (Campbell and Hill, 1988). In this model, partial melting in the crust to generate granitoids was interpreted as the consequence of the thermal impact of the mantle plume that first produced the greenstones (Campbell and Hill, 1988). The genetic link between a mantle plume and greenstones is strongly supported by geochemical evidence, which shows that komatiite lavas and other high-Mg basalts in the greenstone cover were derived from melts extracted from a mantle plume (e.g. Campbell and Hill, 1988; Bateman et al., 2001). However, crustal melting as a direct consequence of the emplacement of a plume is more contentious. In particular, the amplitude of the thermal impact in the crust depends on the thickness of the lithosphere, or the ability of the plume to move through the lithospheric mantle. Studies of modern plumes show that, although not impossible, mantle plumes are unlikely to travel through the lithosphere, in particular if the lithosphere is moving relative to the plumes (Olson et al., 1988; Monnereau et al., 1993; Davies, 1994).

The third model invokes the thermal blanketing effect of the greenstone cover (West and Mareschal, 1979). According to this hypothesis, the greenstone cover insulates and buries the heat-producing continental crust, and the geotherm subsequently raises. The insulation effect works better when the conductivity of the cover is smaller than that of the underlying radiogenic layer. This effect was most likely significant in the Archaean because of larger radiogenic-heat production in the crust. Indeed, a thick greenstone cover deposited on a radiogenic crust could raise the temperature in the crust and the lithospheric mantle by a few hundreds of degrees (West and Mareschal, 1979), which could be enough to produce large volumes of granitoids.

To account for the production of large volumes of granitoids in Late Archaean terranes, we assume that any valid model should be able to raise temperatures well above the wet tholeiite solidus to produce TTG, and well above the tonalite solidus to produce granites. Melting experiments at 500–1000 MPa show that a temperature of ca. 900 °C is required to produce 20–40% volume liquid from amphibolite and tonalite dehydration melting (e.g. Wyllie et al., 1997; Rutter and Wyllie, 1988). In addition, any valid model should also account for the ca. 40 Myr delay observed between the beginning of greenstone emplacement and the peak of granite production. Finally, the proposed model should be able to account for the global-scale of the 2.75–2.65 Ga event.

# 3. Modelling approach, main assumptions and simplifications

Numerical experiments on simplified lithosphere systems provide quantitative insights into geodynamic processes and allow us to test tectonic models. The numerical experiment involves the building of a simplified, but realistic, average Archaean lithosphere. In addition, perturbation of the steady-state thermal structure of the lithosphere is made in a way that simulates the following three processes: crustal thickening, greenstone blanketing, and heat transfer from a mantle plume emplacement. Since mantle plume and greenstone blanketing are expected to be related, we have also investigated the combination of both processes.

Following these perturbations, transient geotherms are derived from the one-dimensional diffusion–advection equation:

$$\frac{\partial T}{\partial t} = \frac{\kappa \partial^2 T}{\partial z^2} + \frac{A(z)}{\rho C_{\rm p}} - v \frac{\partial T}{\partial z} \tag{1}$$

with  $\kappa$  the thermal diffusivity, A(z) the radiogenic-heat production,  $\rho$  the density,  $C_p$  the heat capacity and v the velocity of the medium relative to the surface. Boundary conditions involve a constant heat flow entering the base of the lithosphere and a constant temperature at the surface. All parameters are listed in Table 1.

Both the high radiogenic-heat production (to account for radioactive decay since the Archaean) and the large value of mantle heat flow (to account for hotter mantle temperature in the Archaean) point toward a warmer continental geotherm, and therefore, a thinner continental lithosphere. This is in apparent contradiction with the discovery of Archaean diamonds in xenoliths from the Archaean lithospheric mantle, which suggests that the base of the Archaean continental lithosphere reached a depth greater that 150 km at least locally (Richardson et al., 1984; Boyd et al., 1985). However, recent mantle convection models suggest that the mantle heat flow could have been

Table 1

Parameters used in the definition of the 2.75 Ga continental lithosphere

Thickness of the continental crust $(z_c, m)$	42000
Thickness of the whole lithosphere $(z_1, m)$	112000
Depth-independent crustal heat production	0.98
$(A, 10^{-6} \mathrm{W}\mathrm{m}^{-3})$	
Density of the crust ( $\rho$ , kg m <sup>-3</sup> )	2750
Basal heat flow $(Q_0, 10^{-3} \text{ W m}^{-2})$	25
Thermal diffusivity ( $\kappa$ , $10^{-6} \text{ m}^2 \text{ s}^{-1}$ )	1
Heat capacity ( $C_p$ , J kg <sup>-1</sup> K <sup>-1</sup> )	1000
Thermal conductivity (K, W m <sup>-1</sup> K <sup>-1</sup> )	2.75

strongly partitioned into the oceanic lithosphere during the Archaean (Lenardic and Moresi, 2000). Therefore, this may imply that the heat flow at the base of the continental lithosphere was highly variable and only marginally higher than present-day values, at least locally, as supported by studies of Archaean continental metamorphic gradients. We can therefore safely assume that the basal heat flow in the Late Archaean was  $\geq 15 \text{ mW m}^{-2}$ , the average value for many Archaean cratons (Jaupart et al., 1998; Jaupart and Mareschal, 1999; Russell et al., 2001).

Other parameters, including the thickness of the crust, the concentration in radiogenic elements, and their vertical distribution, control the geotherm in addition to the mantle heat flow. For K, Th and U concentrations in the continental crust, we used the average composition of present-day Archaean crust as determined by Taylor and McLennan (1985, 1995) (K<sub>0</sub>: 7500 ppm, Th<sub>0</sub>: 2.9 ppm, U<sub>0</sub>: 0.75 ppm). We calculated the concentrations of radiogenic elements at times *t* (Table 2) in the past using the following equations:

$$K_t (ppm) = K_0 \exp(\lambda_K t),$$
  

$$Th_t (ppm) = Th_0 \exp(\lambda_{Th} t),$$
  

$$U_t (ppm) = U_0 (0.992849 \exp(\lambda_{238} t))$$
  

$$+0.00725 \exp(\lambda_{235} t))$$

We assumed that the Late Archaean thermal event affected a largely undifferentiated continental crust. Consequently, we used for the crust a depth-independent distribution of radiogenic-heat production of  $0.98 \text{ mW m}^{-3}$ , and we disregarded the radiogenic-heat production in the mantle. Based on the burial pressures at the present exposure levels of 10 undisturbed Archaean terranes, including the Superior Province and the Yilgarn Craton, Galer and

Гa	ble 2	2								
K.	Th	and	U	concentrations	in	the	crust	at	different	times

	Decay constant $(\lambda)$	Crustal co (ppm) @	1	
		2.75 Ga	2.60 Ga	Present
K	$5.543 \times 10^{-10}$	35412	32981	7500
Th	$4.9475 \times 10^{-11}$	3.32	3.298	2.9
U	$^{238}$ U: 1.55125 × 10 <sup>-10</sup> ; $^{235}$ U: 9.8485 × 10 <sup>-10</sup>	1.22	1.185	0.75

Mezger (1998) inferred a mean continental crust thickness of 46 km at the time of crustal stabilisation (i.e. 2.6 Ga). To estimate the mean continental crust thickness before the 2.75-2.65 Ga event, we assumed that ultimately the average steady-state thickness of the crust was limited by its rheology, therefore by its geotherm. Indeed, a thermally softened thick crust will flow toward adjacent ocean basins, to reduce gradient in gravitational potential energy, until the crust is thin, cold and strong enough to sustain gravitational stresses (Bailey, 1999). In modern continental crust, efficient gravity-driven flow is expected when the temperature at the Moho is higher than 700 °C (Sonder et al., 1987). Because of a more mafic composition in the Archaean, we assumed that the steady-state crustal thickness was limited by a temperature of 800 °C at the Moho.

Fig. 3 shows the relationship between crustal thickness, lithospheric thickness, basal heat flow and Moho temperature at 2.75 Ga (i.e. before the Late Archaean thermo-tectonic crisis), at 2.6 Ga (i.e. after the crisis), and for present-day Archaean cratons. For a given conductivity, concentration, and distribution of heat-producing elements, knowledge of two of the four variables mentioned above fixes the two others. Present-day Archaean cratons have a crustal thickness in the range of 35-50 km, a basal heat-flow of  $10-20 \,\mathrm{mW}\,\mathrm{m}^{-2}$ , a lithospheric thickness in the range of 150-250 km, and a temperature at the Moho as low as 300 °C (e.g. Jaupart and Mareschal, 1999; Poudjom Djomani et al., 2001; Artemieva and Mooney, 2001). Measurement of the surface heat flow, as well as the concentration of heat-producing elements at the surface suggests that the upper crust of Archaean cratons is enriched in radiogenic elements compared to the lower crust (e.g. Taylor and McLennan, 1985, 1995). For many Archaean cratons, this differentiation occurred during the Late Archaean. We assume that this resulted in the concentration of radiogenic elements toward the surface, with the radiogenic-heat production decreasing exponentially with depth with a length scale, D, equal to one third of the crustal thickness (i.e.  $A(z) = A_0 \exp(-z/D)$ ). The striped domain in Fig. 3c shows the range of present-day Archaean lithosphere based on the constraints defined above. The graph of Fig. 3b shows the same differentiated Archaean cratons at 2.6 Ga. The difference is solely related to a higher radiogenic-heat production. The possible

range of steady-state Archaean lithosphere at 2.6 Ga is defined assuming 5 km of erosion since 2.6 Ga, a marginally higher basal heat flow, and a temperature at the Moho necessarily lower than  $800 \,^{\circ}$ C. The graph of Fig. 3a shows Archaean cratons at 2.75 Ga, before differentiation. The total radiogenic-heat production is marginally higher than that at 2.6 Ga, and the main difference is that the radiogenic-heat production is assumed to be homogeneous throughout the crust. The same criteria used for 2.6 Ga cratons are used to define the possible range of Archaean lithosphere.

In our numerical experiments we used an 'average' 2.75 Ga continental lithosphere with a mantle heat flow of  $25 \text{ mW m}^{-2}$  entering a 42 km thick continental crust. Such a lithosphere has a temperature at the Moho slightly lower than 700 °C and a total thickness of 112 km (i.e. star on Fig. 3a). In our numerical experiments, the thermal structure of this reference lithosphere is impacted upon by four processes: (1) instantaneous homogeneous thickening by a factor of 1.25-1.5, (2) emplacement of a 6-12 km thick greenstone cover, (3) instantaneous emplacement at various depth of a 50 km thick sill of asthenosphere representing the spreading of a plume head 400 km in diameter, underneath a 920 km long craton, and (4) a combination of (2) and (3). Partial melting and crustal reworking of the crust could, in principle, occur in response to underplating from magmas extracted from a deeper plume. This scenario is not investigated here because, contrarily to what is observed, partial melting of the crust would be coeval with the emplacement of the plume.

# 3.1. Model 1: crustal thickening

In this model (Fig. 4), instantaneous thickening is followed by relaxation of the thermal anomaly. For simplicity, there is no erosion and lateral heat flow effects are neglected, which implies that the conclusions drawn from this model will be valid for large orogens (Gaudemer et al., 1988). Finite thickening over a period of 10-20 Myr and erosion can only delay the temperature rise and increase the time necessary for partial melting to occur at any depth in the crust. The modelling presented here is therefore conservative. It is unlikely that our reference lithosphere could have sustained a large amount of thickening. Firstly, because with a temperature at the Moho close to 700 °C,



Fig. 3.  $z_c-z_l$  planes contoured for both basal heat flow and temperature at the Moho for Archaean lithospheres at 2.75 Ga (a), 2.6 Ga (b) and present time (c). The Archaean crust at 2.75 Ga is not differentiated, i.e. the volumetric radiogenic-heat production is constant with depth (0.98  $\mu$ W m<sup>-3</sup>). For Archaean crusts at 2.6 Ga and present day, we assume that differentiation resulted in a volumetric radiogenic-heat production decreasing exponentially with depth with a length scale of  $z_c/3$ . The crossed fields represent the continental crust sitting above the mantle. The striped area in each plane covers a possible range of realistic steady-state Archaean lithospheres. The grey area covers all lithospheres with a Moho temperature over 800 °C. These lithospheres are too weak to sustain large crustal thickness and are therefore excluded from the possible domain of steady-state lithospheres. The star in plane (a) represents the reference lithosphere used in our modelling.



Fig. 4. Transient geotherms following instantaneous homogeneous thickening by a factor of 1.25 (a) and 1.5 (b), wts: wet tholeiite solidus, i.e. minimum temperature for partial melting; tnls: biotite dehydration (tonalitic) solidus (Wyllie, 1977).

the crust would have been too soft to sustain large contrasts in gravitational potential energy. Secondly, the more buoyant Archaean lithospheric mantle would have contributed to limit crustal thickening by promoting larger (horizontal) extensional gravitational forces. With this in mind, we consider here a thickening factor ranging between 1.25 and 1.5 which brings the crustal thickness to a range of 52-63 km. The chosen thickening factor must be seen as the thickening factor averaged over a large portion of the Earth's continental lithosphere. Finally, as the lithospheric mantle was more buoyant in the Archaean (Jordan, 1975; Griffin et al., 1998; Poudjom Djomani et al., 2001), convective thinning (Houseman et al., 1981; Houseman and Molnar, 1997) can not be invoked as a cause for further heating and was therefore disregarded.

Fig. 4 shows the evolution of the continental geotherm through time. Following homogeneous deformation, melting of wet tholeiites can occur during, or shortly after, thickening to produce TTG. However, the formation of granite by melting of tonalite is not to be expected before 40–70 Myr after thickening. With the model assumptions, thermal conduction following crustal thickening can explain the timing of

the thermal anomaly observed in the Yilgarn Craton and Superior Province, but only if a thickening factor larger than 1.5 is considered. The amplitude of the anomaly is, however, too weak to produce large volumes of granitic magma in the time scale considered here (20–40 Myr). Therefore, our modelling suggests that crustal thickening, if moderate, can not produce the temperature increase necessary for a near complete reworking of the continental crust within the observed time constraint.

# 3.2. Model 2: mantle plume

In this model, the head of a mantle plume spreads under the continental lithosphere into a 50 km thick and 1700 °C hot layer (Campbell and Hill, 1988). This would be equivalent to a 400 km large plume head spreading into a 50 km thick and 920 km diametre disk, which is consistent with the size of the Yilgarn Craton. Investigations of the behaviour of modern mantle plumes at viscosity interfaces show that the hot and weak material of mantle plumes is unlikely to penetrate far into the lithospheric mantle and affect the temperature distribution above the 800 °C isotherm. The



Fig. 5. Transient geotherms following emplacement and spreading of a mantle plume. Three cases are presented that differ by the depth at which the plume head spreads: (a) 112 km; (b) 64 km (top of the thermal boundary layer) and (c) 42 km. In between  $t_{20}$  and  $t_{100}$  geotherms are 20 Myr apart.

potential for thermal erosion is, however, enhanced if the plume spreads under a thin lithosphere, or, alternatively, if small-scale convection cells advect heat in the plume head (Olson et al., 1988; Monnereau et al., 1993; Davies, 1994). We have investigated three situations that only differ by the spreading depth of the plume head.

In the first case (Fig. 5a), the plume head spreads at 112 km depth (i.e. under the base of the lithosphere). Disregarding magmatic underplating and considering only heat transfer by conduction, the thermal anomaly is far too weak to perturb the isotherm in the upper part of the lithosphere. In the second case (Fig. 5b), the plume head penetrates the thermal boundary layer (i.e. the weakest portion of the lithospheric mantle) and spreads at 64 km depth, which corresponds to the 900 °C isotherm marking the limit between the mechanical and thermal boundary layers (e.g. Houseman and Molnar, 1997). In that case the peak of the thermal perturbation reaches the crust 20 Myr after the emplacement of the plume. The thermal anomaly can explain the production of TTG magmas by melting of the lower crust if it is made of wet tholeiites. However, the amplitude of the perturbation cannot explain the formation of large volumes of granite by melting tonalites.

In the third case (Fig. 5c), the plume head spreads under the crust at 42 km depth. Both the wet tholeiites and tonalites solidus are reached. Partial melting affects the lower half of the crust. In this scenario, however, the modelling shows that peak granite production is expected to be coeval with plume and greenstone emplacement. Although there are felsic volcanics interbedded within the greenstone sequences, and despite the oldest granites overlapping in age with youngest greenstones, the peak granite production is always significantly younger than the komatiites that mark the earliest stage of plume emplacement (Nelson, 1997; Davis et al., 2000). This makes the thermal impact of a mantle plume an unlikely primary cause for reworking the crust in both the Yilgarn Craton and the Superior Province.

#### 3.3. Model 3: thermal blanketing effect

In this model (West and Mareschal, 1979), a 6–12 km thick greenstone cover, a common thickness range for greenstone sequences (Anheusser et al.,



Fig. 6. Transient geotherms following deposition of a 6–12 km thick greenstone cover. The cover has the same conductivity and K, U and Th concentrations than the basement. Its internal temperature is that of the surface, which implies that the cover was emplaced in successive volcanic events over a short period of time.

1969; de Wit and Ashwal, 1997; Van Kranendonk et al., 2002), is instantaneously emplaced at the surface of the continental crust. We assume that the cover has the same conductivity and radiogenic-heat production as the crust. Fig. 6 illustrates the evolution of transient geotherms. For a 10 km thick cover, the blanketing effect is equivalent to crustal thickening by a factor of 1.25, and therefore, the amplitude and timing of the anomaly are similar to the that observed in Model 1. For a cover in the upper thickness range (10–12 km), TTG magmatism would be expected to occur within the first few million years after the



Fig. 7. Transient geotherms following deposition of a 6-12 km thick greenstone cover and the spreading of a plume head at the top of the thermal boundary layer. Note that the crust gets heated from both its top and bottom ends.

emplacement of the cover, whereas melting of lower crustal tonalites would not start before 50-60 Myr after the emplacement of the greenstones. It seems that, unless the cover is thicker than 12 km, the effect of thermal blanketing is inadequate to produce the massive reworking of the continental crust.

# 3.4. Model 4: greenstone plus plume

Since it is well established that greenstones are derived from mantle plumes (e.g. Campbell and Hill, 1988; Wyman et al., 1999; Bateman et al., 2001), it is probable that the Late Archaean crustal reworking event could have been caused by a combination of a thermal blanketing effect and heat transfer from a plume. Fig. 7 shows the transient geotherms following the spreading of the plume head at the base of the thermal boundary layer and associated deposition of a 6–12 km thick greenstone cover. In the lower crust, the temperatures are well above the solidus for both wet tholeiites and tonalites within 20-40 Myr following emplacement of the plume and coeval deposition of the greenstone cover. Moreover, the peak temperature at any depth in the crust is reached 40 Myr after emplacement. It seems that this model has the ability to explain both the amplitudes and timing of the thermal anomaly. Furthermore, because the plume is the ultimate cause for the thermal perturbation at both ends of the crust, the broad extent of the anomaly is explained by its origin as a large-scale mantle feature.

# 4. Discussion

The production of large volumes of granitoids and the intense reworking of pre-existing crust in the Late Archaean terranes have been identified in many Archaean cratons (e.g. Zimbabwe Craton, Slave Province, Yilgarn Craton, Superior Province). The volume of granitoid produced and the intensity of the reworking suggest that at least the lower third of the continental crust reached its solidus. This plutonic event overlapped with, and followed the emplacement of, komatiite-bearing greenstones with a ca. 40 Myr delay relative to the peak granite production. It is worth noting that the volcanic event is more widespread and global than the subsequent plutonic event. Indeed, the Fortescue Group in the Pilbara Craton (Australia) and the Vendersdorp Group in the Kaapvaal Craton (South Africa) were emplaced ca. 2.7 Ga ago on older, 3.5 Ga Archaean basement and were not followed by significant crustal reworking. Taken together, the 2.75–2.65 Ga magmatic and plutonic events point toward a global-scale event, which could explain the anomaly in the isotopic age distribution of juvenile continental crust (Condie, 1995) and the major change in the composition of the crust exposed at the surface of the Earth (Taylor and McLennan, 1985) (Fig. 1).

Crustal thickening related to accretion processes has been advocated to explain the reworking of pre-existing crust and regional granitoid emplacement in both the Yilgarn Craton and Superior Province (e.g. Card, 1990; Myers, 1997). In both cases, thickening was inferred to post-date the emplacement of the greenstone cover and to be broadly contemporaneous with granitoid emplacement. Finally, because deformation, magmatism and metamorphism took place without significant spatial and temporal variations across the entire Yilgarn Craton and Superior Province, crustal thickening should have been at the craton-scale and of short duration. Our modelling suggests that significant crustal reworking can be expected over such a time scale if thickening resulted in a >65 km thick crust (Fig. 4b). Crustal thickening can certainly explain local features in many Late Archaean terranes, except that none record high-pressure metamorphic assemblages indicative of significant overthrusting. However, significant crustal thickening as the cause for major global-scale crustal reworking in the Late Archaean is unlikely because the crust would have been far too weak and buoyant to sustain a significant lithospheric thickness gradient (Rey et al., 2001). In addition, the average crustal thickness of 41 km for present-day Archaean crust would imply 20-25 km of erosion since the Late Archaean. This contrasts with the moderate long term uplift and erosion of  $5 \pm 2 \,\text{km}$  (Galer and Mezger, 1998) as shown by the general low metamorphic grade preserved at the surface of Archaean terranes.

Having assumed that the 2.75–2.65 Ga event was profound and global, it is most likely that it had its origin deep in the mantle. Heat transfer from mantle plumes has been proposed as the cause for crustal reworking in the Yilgarn Craton (Campbell and Hill, 1988). Mantle plumes are also invoked as part of a composite scenario involving simultaneous generation of magmas from mantle plume, mantle wedge and subducted slab in the Abitibi-Wawa greenstone province (Wyman et al., 1999, 2002; Wyman and Kerrich, 2002). These composite magmatic events would have occurred shortly before the Abitibi-Pontiac collision, which was completed by 2.68 Ga (Corfu et al., 1991), with the majority of granitoids emplaced between 2.69 and 2.67 Ga. Assuming that mantle plumes can travel through the lithospheric mantle, our numerical experiments show that a plume head spreading closer to the Moho will cause the expected partial melting of the crust, but at a too early stage. On the other hand, when the plume head spreads too far away from the base of the crust, the thermal anomaly is far too weak to perturb the crustal geotherm. When the plume spreads at the top of the thermal boundary laver (isotherm 900 °C), the thermal anomaly produced reaches its peak in the crust 20-40 Myr after emplacement (Fig. 8). Although this anomaly may produce TTG if the lower crust is made of wet tholeiites, it may be too weak to produce granitic magmas from the melting of tonalites. In conclusion, our modelling suggests that the thermal anomaly related to the emplacement of mantle plumes cannot, on its own, explain the massive crustal reworking that did occur in the Late Archaean.

As demonstrated by our numerical experiments, the thermal blanketing effect is very efficient to increase the temperature in the crust, even if the cover is devoid of heat-producing elements (West and Mareschal, 1979). This process can be seen as another mode of crustal thickening with similar thermal implications. Therefore, and because of the ubiquity of greenstone cover in all Archaean cratons, thermal blanketing can not be disregarded when considering the evolution of Archaean cratons. On its own, however, this process cannot easily explain both the amplitude and the timing of the thermal anomaly, unless the cover is >12 km thick. Such thickness has been observed in some cratons (Superior, Pilbara, Kaapvaal), but cannot be regarded as an average value for all.

Our numerical experiments show that, when the thermal anomaly is related to a ca. 10 km thick greenstone cover combined with a plume head spreading at ca. 60 km depth (the top of the thermal boundary layer), both the timing and amplitude of the anomaly match the geological record. In particular, the peak of the thermal anomaly (>900 °C) is compatible with the production of 20–40% volume liquid, and it is reached 40 Myr after the emplacement of the plume and the deposition of the greenstone sequence (Fig. 8). This scenario is indeed extremely efficient because the temperature increases simultaneously at the top and at the bottom of the crust. The amplitude of the anomaly is also sufficient to produce firstly TTG and, later on,



Fig. 8. Evolution of the temperature at the Moho for different models. The evolution of temperature at the Moho that follows the emplacement of a 10 km thick greenstone cover (thin curve), is similar to that observed following the homogeneous thickening of the lithosphere by a factor of 1.25. Note that the combination of plume and greenstone leads to a thermal anomaly that peaks at >900 °C ca. 40 Myr after their emplacement.

granitic magmas. This model is, however, dependent on the ability of a plume to travel through the thermal boundary level of the lithosphere.

As shown by Mareschal and West (1980), the emplacement of a greenstone cover on a felsic basement introduces a density inversion that becomes increasingly unstable as the viscosity of the crust decreases in relation to the temperature increase. As thermal relaxation proceeds, the heavy greenstones progressively sink into its partially molten, rheologically weak, reactivated basement, which rises into granitoid domes through the transfer of partial melts (see Pawley et al., 2003). This gravitational instability will result in partial convective overturn of the crust (Dixon and Summers, 1983; Collins et al., 1998) with the characteristic features of diapirs (e.g. Bouhallier et al., 1993, 1995; Jelsma et al., 1993; Collins et al., 1998; Chardon et al., 1996, 1998; Van Kranendonk et al., 2003). This model is consistent with the broad overlap between granite emplacement and regional contractional deformation in the two cratons considered here. It is also consistent with the relative homogeneity of the erosional level now exposed at the surface, the craton-scale distribution of deformation, and the absence of significant age gradients in the tectonic, metamorphic and magmatic activities across entire cratons.

#### 5. Conclusion

Among the models investigated here, the combined effect of greenstone cover and mantle plume best explains the amplitude and the timing of the thermal



Fig. 9. Two-stage model to explain reworking of the crust during the Late Archaean. (a) Plume heads spread under Mechanical Boundary Layer (MBL), following major reorganisation of convection in the deep mantle. Locally, plumes interfere with plate-tectonic processes. Melt extracted from the plumes forms either thick greenstone covers that bury pre-existing continental crusts or plateaux in oceanic basins. (b) The insulation of heat-producing crust (felsic or mafic) combined with heat transfer from the plume lead to the partial melting of the crust 20–40 Myr after the emplacement of the plume. Because of the density inversion represented by a greenstone cover over a felsic basement and following the thermal softening of the felsic basement, gravitational instabilities develop during which greenstone covers sink into the partially molten crust, thus providing space for granitoid domes. Again gravity-driven processes can interfere with plate-boundary processes (Choukroune et al., 1995, 1997). As radiogenic elements concentrate in granitoids, they accumulate in the upper part of the crust inducing cooling, strengthening and finally stabilisation of the craton.

anomaly that profoundly reworked the continental crust in the Late Archaean. We envision a succession of events that started with a "mantle overturn" (Stein and Hofmann, 1994) and the formation of plumes at the core-mantle boundary. As the plumes impinge on the lithosphere, the extracted melts produced the greenstone covers at the surface of continents and plateaux in oceanic basins (Fig. 9a). Greenstones that were erupted onto undifferentiated radiogenic crust (either mafic or felsic) triggered a thermal anomaly that, when combined with the anomaly related to the heat transferred from the plume head, resulted in the massive partial melting and reworking of the crust (Fig. 9b). Finally, greenstones became unstable and sank into their basement, thus producing dome and basin patterns as a result of partial convective overturn of the upper and middle crust.

It is important to emphasize that there is no incompatibility between the processes envisaged in this model and plate-tectonic processes. The proposed model can take place together with whatever (other) tectonic processes were operational at the surface of the Earth at that time. The evolution of many Archaean cratons could therefore be better explained by a combination of plume-related and plate-tectonic processes.

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