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Abstract

Large basaltic provinces as much as 15 km thick are common in Archean cratons. Many of these flood basalts erupted through continental crust but remained at sea level. Although common in the Archean record, subaqueous continental flood basalts (CFBs) are rare to absent in the post-Archean. Here we show that gravity-driven lower crustal flow may have contributed to maintaining Archean CFBs close to sea level. Our numerical experiments reveal that the characteristic time to remove the thickness anomaly associated with a CFB decreases with increasing Moho temperature (T_M), from 500 m.y. for $T_M \approx 320^\circ\text{C}$ to 1 m.y. for $T_M \approx 900^\circ\text{C}$. This strong dependency offers the opportunity to assess, from the subsidence history of CFBs, whether continental geotherms were significantly hotter in the Archean. In particular, we show that the subsidence history of the ca. 2.7 Ga upper Fortescue Group in the East Pilbara Craton, Western Australia, requires Moho temperatures $>>700^\circ\text{C}$. Applied to eight other unambiguous subaqueous Archean CFBs, our results indicate Moho temperatures $>>650^\circ\text{C}$ at the time of eruption. We suggest that the decrease in the relative abundance of subaqueous CFBs over Earth's history could reflect the secular cooling of the continental lithosphere due to the decrease in radiogenic heat production.

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Lower crustal flow kept Archean continental flood basalts at sea level

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ABSTRACT

Large basaltic provinces up to 15 km thick are common in Archean cratons. Many of these flood basalts erupted through continental crust but remained at sea level. While common in the Archean, subaqueous Continental Flood Basalts (CFBs) are rare to absent in the post-Archean. Here we show that gravity-driven lower crustal flow may have contributed to maintaining Archean CFBs close to sea level. Our numerical experiments reveal that the characteristic time to remove the thickness anomaly associated with a CFB decreases with increasing Moho temperature (T_M) from 500 million years (Myr) for $T_M \approx 320$ °C to 1 Myr for $T_M \approx 900$ °C. This strong dependency offers the opportunity to assess, from the subsidence history of CFBs, whether continental geotherms were significantly hotter in the Archean. In particular, we show that the subsidence history of the ~2.7 billion-year-old upper Fortescue group in the East Pilbara Craton, Western Australia, requires Moho temperatures much greater than 700 °C. Applied to eight other unambiguous subaqueous Archean CFBs, our results indicate Moho temperatures well in excess of 650 °C at the time of eruption. We suggest that the decrease

in the relative abundance of subaqueous CFBs over Earth's history could reflect the secular cooling of the continental lithosphere due to the decrease in radiogenic heat production.

INTRODUCTION

A salient feature of the Archean geological record is the abundance of subaqueous greenstone belts 5–15 km thick, dominated by mafic volcanics and covering areas $> 100,000 \text{ km}^2$ (de Wit and Ashwal, 1997; Arndt, 1999; Kump and Barley, 2007). Archean greenstone belts differ in composition, structure and geological setting, including arc-like and mantle plume settings. We aim to explain why thick CFBs failed to emerge and thus why subaqueous CFBs were relatively common in the Archean (Table 1; Arndt, 1999; Kump and Barley, 2007) but became rarer through geological time. The only two unambiguous post-Archean occurrences of subaqueous CFBs are the early Proterozoic komatiites of the Gilmour Islands, Hudson Bay (Arndt, 1982) and the *c.* 250-Ma Siberian Flood Basalts (Czamanske et al., 1998). We note that maintaining Archean CFBs up to 10 km thick and emplaced in $< 70 \text{ Myr}$ (Table 1) below sea level would require a change in sea level of $> 1 \text{ km}$ over a few tens of million years. Changes in eustatic sea level do not suffice to explain this observation since they have an amplitude of $\sim 200 \text{ m}$ in the Phanerozoic (Müller et al., 2008) and of $< 1 \text{ km}$ over Earth's history (Flament et al., 2008). In this paper, we focus on intraplate subaqueous Archean flood basalts that erupted away from continental rifting (Arndt, 1999; Tomlinson and Condie, 2001; Table 1), which allows us to characterize gravity-driven subsidence. Our generic thermo-mechanical numerical experiments suggest that syn- to post-emplacement lateral ductile flow of hot lower crust could have maintained Archean CFBs close to sea level. Comparing our modeling results to the eruption time of subaqueous Archean CFBs, we place broad constraints on the continental geotherm at the time of their eruption.

HYPOTHESIS AND MODEL

We make the hypothesis that the ductile lower continental crust can flow under the weight of a CFB to minimize the lateral pressure contrast in the lithosphere. To evaluate the time needed for the crust to reach mechanical equilibrium, we design a 2D-cartesian model of a vertically stratified continental lithosphere on which a CFB has been emplaced and is in isostatic equilibrium (Fig. 1). The models are 800×100 km and they consist, from top to bottom, of a 20 km thick compressible air-like layer, a volcanic plateau, a 40 km thick continental crust and the uppermost 40 km of the mantle (Fig. 1). We consider CFBs of width 300 km, of thickness 3, 6, 9 and 12 km and we use a conservative density of 2840 kg m^{-3} since CFBs consist of up to 50% sediments. We adopt a visco-plastic rheological model, combining frictional flow (supplementary methods) and the temperature-dependent viscosity law (modified from Solomatov and Moresi, 1997)

$$\eta = \eta_0 \exp(-\gamma T), (1)$$

where T is the temperature, η_0 is the reference viscosity and $\gamma = Q/RT_p^2$, with Q being the activation enthalpy, R the gas constant and T_p the temperature at the base of the lithosphere (parameter values are given in Table DR1). We limit the range of viscosities to $5 \times 10^{17} - 5 \times 10^{22} \text{ Pa s}$ to optimize calculation time.

We use a thermal model in which the depth-independent crustal heat production is calculated back in time (supplementary methods) from the present-day concentration of the bulk Archean continental crust (Taylor and McLennan, 1995; Tables DR1 and DR2). We allow the mantle heat flow to vary from a conservative present-day cratonic value of 12 mW m^{-2} (Jaupart and Mareschal, 1999) to 33.5 mW m^{-2} (Table DR2). Using the composition of present-day

cratons is conservative since the most radiogenic uppermost 3–7 km of Archean continental crust have been eroded over time (Galer and Mezger, 1998). We compute an initial series of six geotherms with $400\text{ }^{\circ}\text{C} < T_M < 900\text{ }^{\circ}\text{C}$ (Fig. 1 and Table DR2), that result in lithospheric rheological profiles in which the viscosity of the lower continental crust decreases from $5 \times 10^{21}\text{ Pa s}$ for $T_M \approx 400\text{ }^{\circ}\text{C}$ to $5 \times 10^{18}\text{ Pa s}$ for $T_M \approx 680\text{ }^{\circ}\text{C}$ (Fig. 1), in agreement with laboratory estimates of dislocation creep laws of quartzite (Gleason and Tullis, 1995).

RESULTS

For $T_M > \sim 600\text{ }^{\circ}\text{C}$, we observe that the removal of the contrast in thickness of the continental crust due to the CFB occurs through lateral flow in a weak lower crustal channel (Fig. 2). The horizontal pressure gradient due to the final width of the CFB drives the flow. The channel is characterized by a Poiseuille-like velocity profile, and its thickness h increases from 10 to 15 km for increasingly hot geotherms. The removal of thickness contrast, determined from two Lagrangian position trackers (Fig. 1), follows an exponential evolution (Fig. DR1)

$$w(t) = w_0 \exp(-t/\tau_f), \quad (2)$$

where $w(t)$ is the cumulative thickness of the crust and CFB after a time t , w_0 is the initial cumulative thickness of the crust and CFB and τ_f is the characteristic relaxation time of thickness removal by lower crustal flow. We obtain τ_f for each model by fitting the calculated evolution of the thickness of the continental crust (Fig. DR1).

Two series of models with varying CFB density and increasing CFB half-width λ allows us to verify that τ_f is inversely proportional to CFB density and that τ_f increases with λ until it reaches a near-constant value for $\lambda/h \geq 10$ (McKenzie and Jackson, 2002). In addition, τ_f is independent of the thickness of the CFB (Kruse et al., 1991; Fig. 3). Our key result is the exponential decrease of τ_f with increasing T_M , from $\sim 500\text{ Myr}$ for $T_M \approx 320\text{ }^{\circ}\text{C}$ to 1 Myr for

$T_M \approx 900$ °C (Fig. 3). This relationship is due to the dependency of τ_f on viscosity (McKenzie and Jackson, 2002) that itself depends exponentially on temperature (Equation 1). Fitting the results for the series of six geotherms with various temperature dependence of the rheology gives a scaling law with an exponential constant of the order of $\gamma_{cc}/3$ (Fig. 3). This suggests that the effective viscosity of the modeled lithosphere during the relaxation of the topographic load is an average between the viscosity of the fast-flowing weak lower crust and the larger effective viscosity of the slow-deforming upper crust and upper mantle. In addition, the consistency of the fitted exponential constant for different rheological laws and continental geotherms suggests that the scaling laws we derived are robust.

CASE STUDY: SUBSIDENCE HISTORY OF THE UPPER FORTESCUE GROUP

The strong dependency of the relaxation time on Moho temperature offers the opportunity to constrain the geotherm at the time of eruption of a CFB from its subsidence history. In the following, we apply this concept to the Neoproterozoic upper Fortescue Group, Pilbara Craton, Western Australia. In the East Pilbara Craton, the 6.5-km-thick, 2775–2630-Ma Fortescue Group (Thorne and Trendall, 2001) is preferentially preserved in broad synclines and centroclines. While the *c.* 2772-Ma Mount Roe Basalt and the 2765–2752-Ma Hardey Formation were folded and faulted during their deposition, markers of syn-depositional extension are lacking in the upper (< 2752 Ma) Fortescue Group (Williams and Bagas, 2007). A paleostress analysis in the Meentheena Centrocline reveals that a radially symmetric stress regime prevailed during its formation (Fig. DR2). This suggests that continental extension, which would be characterized by a unidirectional minimum horizontal stress, exerted little to no control on the emplacement and subsidence of the $> \sim 2.5$ km thick upper Fortescue Group (Fig. 4). In addition, there is abundant evidence for subaqueous sedimentation and volcanism throughout the

Fortescue Group (Thorne and Trendall, 2001). Despite being mostly subaerial in the East Pilbara, the effusive volcanics of the Kylenea (2749–2735 Ma), Tumbiana (2729–2715 Ma) and Maddina (2718–2713 Ma) Formations alternate with stromatolite-bearing shallow-water sedimentary rocks of the Mopoke, Meentheena and Kuruna Members (Thorne and Trendall, 2001; Williams and Bagas, 2007; Fig. 4). This alternation indicates that the erupting basalts were episodically brought back to sea level. In the absence of tectonic extension and of significant erosional features during the emplacement of the upper Fortescue Group, we suggest that the subsidence of these basalts is the surface expression of lower crustal flow. The Moho temperature at the time of emplacement of the upper Fortescue Group can be estimated from its subsidence history and our modeling results. The ≤ 1400 -m-thick basalts of the Kylenea Formation erupted in ~ 14 Myr (Williams and Bagas, 2007; Fig. 4) and are stratigraphically bounded by the shallow-water sediments of the Mopoke and Mingah Members (Fig. 4). The lower ~ 600 -m-thick basalts of the Maddina Formation erupted in ≤ 11 Myr (Blake et al., 2004; Fig. 4) and are stratigraphically bounded by the shallow-water sediments of the Meentheena and Kuruna Members (Fig. 4). For these formations that were episodically below sea level, the thickness removal time is necessarily much shorter than the eruption duration. Thus, assuming constant sea level, the subsidence of these basalts requires Moho temperatures at the time of CFB emplacement much greater than 680°C and than 700°C , respectively (Fig. 3). Given the incertitude on the dependency between rheology and temperature, the uncertainty on these Moho temperatures is $\pm 100^\circ\text{C}$.

DISCUSSION

The example of the upper Fortescue Group discussed above illustrates that subaqueous Archean CFBs up to 10 km thick could have resulted from lower crustal flow during the repeated accumulation of kilometer-thick basaltic piles. Our model can be tested on at least eight other

subaqueous Archean CFBs for which available high-precision geochronological data indicate that the eruption duration of individual basaltic event was $< \sim 20$ Myr (Table 1). Because τ_f is necessarily much shorter than 20 Myr for these subaqueous CFBs, our modeling results indicate Moho temperatures much greater than ~ 650 °C (Fig. 3) and viscosities of the lower crust much lower than $\sim 1.4 \times 10^{19}$ Pa s (Equation 1). This temperature estimate is $> \sim 250$ °C hotter than the present-day Moho temperature of Archean cratons of 400 ± 100 °C (Artemieva, 2009). The question arises whether the Moho temperatures that we derived at the time of emplacement of a CFB are representative of a steady-state geotherm, or of a geotherm transiently hot due to thermal anomaly at the origin of the CFB. Rey et al. (2003) showed that it takes > 20 Myr for a thermal anomaly at the base of the mechanical lithosphere (isotherm 900 °C) to heat up the Moho. Therefore, the Moho temperatures derived from the subsidence history of each of the basaltic events that constitute subaqueous CFBs represent steady-state geotherms rather than transient geotherms. The fact that Phanerozoic CFBs erupted in < 20 Myr, including the Deccan, Paraná-Etendeka and Karoo basalts (Richards et al., 1989), are subaerial confirms that the lower crust remained cool during their eruption. Indeed, the greater abundance of subaqueous CFBs in the Archean indicates that Archean continental geotherms were significantly hotter than their present-day counter-parts. Our models suggest that the flow of hot, ductile lower crust was a key process that maintained CFBs at sea level. This is consistent with high-temperature shear fabrics described in Archean lower crust (Sandiford, 1989; Krogh, 1993). As the continental lithosphere became cooler over time, both CFBs and the underlying continental crust emerged, with major consequences for the composition of the atmosphere and of the oceans (Kump and Barley, 2007; Rey and Coltice, 2008; Flament et al. 2008). Other processes have been put forward to explain the anomalous subsidence observed in some

volcanic provinces. For instance, Elkins Tanton and Hager (2001) attributed the anomalous subsidence of the Phanerozoic Siberian Flood Basalts to the removal of the mantle lithosphere via Rayleigh-Taylor instability, caused by the combined decrease of the viscosity of the lithosphere heated by intruding melts, and increase in lithospheric density when dikes solidify as eclogites. This model, which requires a very thin lithosphere, predicts an initial uplift due to the rising plume head followed by subsidence due to the delamination of the lithosphere throughout a short eruption ($< \sim 5$ Myr). Alternatively, Leng and Zhong (2010) proposed that the ponding of mantle plume head at the 660-km boundary could result in protracted, small-amplitude subsidence followed by rapid and important uplift *prior* to the eruption of a CFB. Neither of these processes reproduces the alternation between subaerial and subaqueous facies that is observed over > 35 Myr in the Fortescue Group (Fig. 4).

CONCLUSIONS

Our numerical experiments and the field examples discussed above suggest that CFBs can remain at sea level if the lower crust can flow at a rate comparable to that of eruption. We show that the characteristic time to remove the thickness anomaly associated to a CFB by flow of the lower crust strongly depends on Moho temperature. This allows us to use the subsidence history of subaqueous Archean CFBs emplaced in an intraplate setting to place broad constraints on the continental geotherms at the time of their eruption. We show that the subsidence history of the upper Fortescue Group in the East Pilbara Craton and that of other subaqueous Archean CFBs are compatible with lower crustal flow. Together with our modeling results, this points to Archean Moho temperatures much greater than 650°C . For intraplate CFBs erupted over hot continental crust, our model predicts (a) alternating subaerial and subaqueous volcano-sedimentary facies during eruption, (b) shearing in the ductile lower continental crust upon

loading, and (c) limited crustal thickness gradient. We speculate that subaqueous CFBs became less common over Earth's history because lower crustal flow became less efficient as the continental lithosphere cooled down due to lesser radioactive heat production.

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FIGURE CAPTIONS

Figure 1. Setup to model crustal removal following the eruption of a CFB. Geometry, material densities, continental geotherms, viscosity profiles and Lagrangian position trackers used in the models. The thick, dark-gray lines in the continental crust are strain markers. The area within the dashed frame is shown in Figure 2.

Figure 2. Thickness removal and subsidence of the CFB by lower crustal flow. Strain markers after a) 7.5 Myr and b) 15 Myr for a numerical simulation with a Moho temperature of $\sim 680^\circ\text{C}$ and a CFB 6 km thick.

Figure 3. Scaling laws (black lines) for the characteristic time of thickness removal as a function of Moho temperature. Results are presented for three series of models: 1) reference model for a thickness of the CFB of 12 km (empty circles), 9 km (empty diamonds), 6 km (gray triangles), and 3 km (black squares); 2) a thickness of the CFB of 6 km and $\gamma = 2/3 \gamma_{cc}$ in Equation 1 (crosses); 3) a thickness of the CFB of 6 km, $\gamma = \gamma_{cc}$ in Equation 1, and a continental crust 30 km

thick (plus symbols). Thick gray lines show the maximum relaxation time and minimum Moho temperature for subaqueous Archean CFBs (Table 1).

Figure 4. Sedimentology and stratigraphy of the upper Fortescue Group in the Meentheena Centrocline. The three carbonate members display shallow-water facies. a. Wavy bedding (mudstone-sandstone interlayers) with laminated, rippled sand sheets indicates an overbank and floodplain depositional environment (Kuruna Member). Diameter of coin: 25 mm; b. Laterally-linked stromatolites (St) and their lateral sediments with climbing ripples (CR) overlying imbricate flat pebble conglomerate (FPC) indicate a tidal-flat environment (Meentheena Member). Length of pen: 14 cm; c. Silicified carbonate precipitates (CP), stromatolites (St), and fenestrae (F) in restricted tidal-flat environments (Mopoke Member). Diameter of lens cap: 5 cm; d. Simplified stratigraphic column. Geochronological data are from Blake et al. (2004); ages for the Kylena Formation and all thicknesses are from Williams and Bagas (2007). The dashed section of the Maddina Formation does not crop out in the Meentheena Centrocline.

¹GSA Data Repository item 2011xxx, extended methods, supplementary references, Figures DR1 and DR2, and Tables DR1 and DR2, is available online at www.geosociety.org/pubs/ft2011.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

TABLE 1. CHARACTERISTICS OF SELECTED SUBAQUEOUS ARCHEAN CFBs

Name	Craton	Events*	Thickness (km)	Age (Ma)	Duration (Myr)	References
Lower Maddina Fm., Fortescue Gp.	Pilbara	1	~ 0.6	2718 ± 3 to 2713 ± 3	≤11	Blake et al. (2004)
Kylena Fm., Fortescue Gp.	Pilbara	1	≤1.4	2749-2735	14	Thorne and Trendall (2001)
Honman Fm., Lake Johnston gr. belt	Yilgarn	1	≤1.2	2921 ± 4 to 2903 ± 5	≤28	Wang et al. (1996)
Ngezi Gp., Belingwe gr. belt	Zimbabwe	2	6.5	2692 ± 9	18?	Tomlinson and Condie (2001); Chauvel et al. (1993)
Upper Kam Gp., Yellowknife belt	Slave	3	~ 6	2722 ± 2 to 2701 ± 1	≤24	Isachsen and Bowring (1994)
Kambalda gr. belt	Yilgarn	4	~ 4	2726 ± 30 to 2690 ± 5	≤71	Tomlinson and Condie (2001)
Lower Warrawoona Gp., Marble Bar gr. belt	Pilbara	sequence	~ 3	3490-3469	>21	Van Kranendonk et al. (2004)
Balmer assemblage	Superior	sequence	<10	2992-2964	>28	Tomlinson and Condie (2001)
Lumby Lake gr. belt	Superior	sequence	≤7	<2963-2898	<70	Tomlinson and Condie (2001)
Kolar schist belt	Dharwar	sequence	~ 4	~ 2700	N.D.	Krogstad et al. (1989)

*Number of distinct formations representing distinct magmatic events.
N.D.—no data.

a.

$t = 7.5 \text{ Myr}$

50 km

b.

$t = 15 \text{ Myr}$

SUPPLEMENTARY METHODS

Because of the symmetry of the problem, only one half of the CFB is modeled. Grid cells are 1560×1560 m and each cell includes 36 particles, so that the effective resolution is 260×260 m. All boundaries in the models are rigid, undeformable and free-slip, and neither forces nor velocities are applied at any of the boundaries.

Rheological model

Frictional flow is described using the Coulomb criteria (Moresi and Solomatov, 1998) combined with a strain softening function (Wijns et al., 2005)

$$\sigma_{yield} = (C_0 + \tan(\Phi) \sigma_n) f(\varepsilon), \quad (S1)$$

where σ_{yield} is the shear stress, C_0 is the cohesion of the material at atmospheric pressure, Φ is the angle of internal friction, σ_n is the stress normal to the failure plane approximated by the lithostatic pressure (Moresi and Solomatov, 1998) and

$$f(\varepsilon) = \begin{cases} 1 - \beta(\varepsilon / \varepsilon_0) & \varepsilon < \varepsilon_0 \\ 1 - \beta & \varepsilon \geq \varepsilon_0 \end{cases}, \quad (S2)$$

where ε is the accumulated strain, taken as the second invariant of the deviatoric plastic strain tensor, and ε_0 is the “saturation” strain from which the yield stress is reduced by a proportion β (Wijns et al., 2005). Values for these parameters are given in Table DR1.

Thermal model

The crustal heat production is calculated back in time using

$$H = \sum_i \alpha_i C_0^i H^i \exp(a \ln(2) / \tau_{1/2}^i), \quad (S3)$$

where i is a radioactive element amongst U^{238} , U^{235} , Th^{232} and K^{40} , α^i is the natural

proportion of the radioactive element, C_0^i is the concentration of the element (Table DR1), H^i is rate of heat release of the element, a is the age for which the radiogenic heat production is calculated and $\tau_{1/2}^i$ is the half-life of the element.

SUPPLEMENTARY REFERENCES

Moresi, L., and Solomatov, V., 1998, Mantle convection with a brittle lithosphere: thoughts on the global tectonic styles of the Earth and Venus: *Geophysical Journal International* v. 133, p. 669-682.

Wijns, C., Weinberg, R., Gessner, K., and Moresi, L., 2005, Mode of crustal extension determined by rheological layering. *Earth and Planetary Science Letters* v. 236, p. 120-134.

SUPPLEMENTARY FIGURES

Figure DR1. Relaxation of the crustal thickness anomaly. Evolution of the elevation difference between the two Lagrangian position trackers shown in Fig. 1. A logarithmic scale is used for the vertical axis. Results are presented for a CFB 9 km thick and for a continental crust 40 km thick.

Figure DR2. Directions of the maximum horizontal compressive paleostress σ_H in the Meentheena Centrocline. These directions are derived from the multiple measurements of conjugate fractures and bedding plane at 33 sites, assuming that σ_H is aligned with the acute bisector of each pair of fractures. Insets *a* and *b* illustrate local variability. The simplified geology is from Williams and Bagas (2007).

SUPPLEMENTARY TABLES

TABLE DR1. LIST OF PARAMETERS USED IN THE MODELS

Parameter*	Meaning	Value(s)	Unit
<i>a) Mechanical parameters</i>			
β	Strain weakening factor	0.8	
ε_0	Strain from which weakening is maximum	0.5	
ρ_{atm}	Atmospheric density	2.5	kg m ⁻³
ρ_{cc}	Crustal density	2720	kg m ⁻³
ρ_{CFB}	CFB density	2840	kg m ⁻³
ρ_m	Mantle density	3370	kg m ⁻³
σ_{cc}	Crustal maximum yield stress	250	MPa
σ_m	Mantle maximum yield stress	500	MPa
ϕ_{cc}	Crustal internal angle of friction	15	°
ϕ_m	Mantle internal angle of friction	25	°
b_{atm}	Atmospheric pre-exponential constant	$7.5 \cdot 10^{18}$	Pa s
η_{0cc}	Crustal reference viscosity	$1.48 \cdot 10^{29}$	Pa s
η_{0m}	Mantle reference viscosity	$1.19 \cdot 10^{28}$	Pa s
C_{cc}	Crustal cohesion	10	MPa
C_m	Mantle cohesion	200	MPa
g	Acceleration of gravity field	9.81	m s ⁻²
Q_{atm}	Atmospheric activation enthalpy	0	kJ mol ⁻¹
Q_{cc}	Crustal activation enthalpy	540	kJ mol ⁻¹
Q_m	Mantle activation enthalpy	280	kJ mol ⁻¹
R	Gas constant	8.31	J mol ⁻¹ K ⁻¹
T_p	Upper mantle potential temperature	1330	°C
<i>b) Thermal parameters</i>			
κ	Thermal diffusivity	$0.9 \cdot 10^{-6}$	m ² s ⁻¹
C_p	Heat capacity	1000	J kg ⁻¹ K ⁻¹
[U]	Uranium concentration [†]	0.75	ppm
[Th]	Thorium concentration [†]	2.9	ppm
[K]	Potassium concentration [†]	0.75	wt. %
H_{cc}	Crustal heat production	Table DR2	$\mu W m^{-3}$
q_m	Basal mantle heat flux	Table DR2	mW m ⁻²

*Non-listed parameters for the CFB are taken equal to crustal ones.

[†]From Taylor and McLennan (1995)

TABLE DR2. THERMAL VALUES

a^*	H_{cc}	q_m	d_{cc}	T_{Moho}
Ga	$\mu W m^{-3}$	mW m ⁻²	km	°C
1.65	0.69	12.0	40	404
2.75	0.98	12.0	40	496
2.30	0.84	23.0	30	322
			40	598
2.75	0.98	26.0	40	406
			40	682
3.25	1.19	29.5	30	460
			40	797
3.60	1.39	32.0	30	535
			40	896
			30	598

*To compute radiogenic heat production