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A case for late-Archaean continental emergence from thermal evolution models and hypsometry

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ABSTRACT

The secular cooling of the Earth's mantle and the growth of the continental crust together imply changes in the isostatic balance between continents and oceans, in the oceanic bathymetry and in the area of emerged continental crust. The evolution of these variables is of fundamental importance to the geochemical coupling of mantle, continental crust, atmosphere and ocean. To explore this further, we developed a model that evaluates the area of emerged continental crust as a function of mantle temperature, continental area and hypsometry.

In this paper, we investigate the continental freeboard predicted using different models for the cooling of the Earth. We show that constancy of the continental freeboard (±200 m) is possible throughout the history of the planet as long as the potential temperature of the upper mantle was never more than 110–210 °C hotter than present. Such numbers imply either a very limited cooling of the planet or, most likely, a change in continental freeboard since the Archaean. During the Archaean a greater radiogenic crustal heat production and a greater mantle heat flow would have reduced the strength of the continental lithosphere, thus limiting crustal thickening due to mountain building processes and the maximum elevation in the Earth's topography [Rey, P. F., Coltice, N., Neoarchean strengthening of the lithosphere and the coupling of the Earth's geochemical reservoirs, Geology 36, 635–638 (2008)]. Taking this into account, we show that the continents were mostly flooded until the end of the Archaean and that only 2–3% of the Earth's area consisted of emerged continental crust by around 2.5 Ga. These results are consistent with widespread Archaean submarine continental flood basalts, and with the appearance and strengthening of the geochemical fingerprint of felsic sources in the sedimentary record from ~2.5 Ga. The progressive emergence of the continents as shown by our models from the late-Archaean onward had major implications for the Earth's environment, particularly by contributing to the rise of atmospheric oxygen and to the geochemical coupling between the Earth's deep and surface reservoirs.

1. Introduction

The analysis of eustatic sea level (e.g. Vail et al., 1977; Haq et al., 1987; Miller et al., 2005) has led to the conclusion that continental freeboard (the average elevation of emerged continental crust above sea level) has remained constant throughout the Phanerozoic. This hypothesis has been extended to the entire Proterozoic on the premise that tectonic processes have been the same since the end of the Archaean (2.5 Ga) (Wise, 1974). This constancy in continental freeboard has also been used to constrain the evolution of other interdependent parameters such as crustal growth rates (Reymer and Schubert, 1984; Schubert and Reymer, 1985), sedimentation rates (McLennan and Taylor, 1983), ocean volume (Kasting and Holm, 1992; Harrison, 1999), crustal thickness (Hynes, 2001), depth of mid-oceanic ridges (Galer, 1991; Kasting and Holm, 1992) and mantle temperature (Galer, 1991; Galer and Mezger, 1998).

However, the widespread occurrence of continental flood basalts emplaced on top of immersed continents (Arndt, 1999; Kump and Barley, 2007) suggests higher sea levels in the Archaean. Furthermore, the geochemical fingerprints of felsic sources do not appear in the geological record before the end of the Archaean (Veizer and Compston, 1976; McLennan and Taylor, 1980; Taylor and McLennan, 1985; Shields and Veizer, 2002; Valley et al., 2005). These observations could indicate that the differentiated continental crust was mostly isolated throughout the Archaean (Rey and Coltice, 2008) because it was widely covered by mafic volcanism, and because the continents were largely immersed during this eon. The timing of continental emergence has long been questioned. Despite a lack of data and physical models, Hargraves (1976) proposed a late emergence during

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the Proterozoic, a hypothesis disputed by Windley (1977) and by Vlaar (2000) who both argued in favour of a continental emergence near the Archaean/Proterozoic boundary.

In this paper, we propose a model to quantify emerged crustal area as a function of mantle temperature, hypsometry (the global areal distribution of the Earth's surface elevations) and continental area. Because there is no consensual thermal evolution model for the Earth as yet, we present the continental freeboard and area of emerged continental crust as calculated using three different thermal evolution models: two models based on marginal stability analysis of the oceanic lithosphere (Korenaga, 2006) and one model based on the observation of the present-day distribution of seafloor ages (Labrosse and Jaupart, 2007). The choice of thermal evolution model is critical to the calculation of continental freeboard. For realistic thermal evolution models (Korenaga, 2006; Labrosse and Jaupart, 2007), constancy of continental freeboard (±200 m) can be achieved if the potential temperature of the mantle was never more than 110–210 °C hotter than present. Previous results from continental freeboard models (Galer, 1991; Galer and Mezger, 1998) fall within this range.

For a late-Archaean balance of parameters (mantle potential temperature 150 °C hotter than present and 80% of present continental area) and for constant hypsometry, we calculate that the area of emerged continental crust was less than 12% of the Earth's area. Physical models suggest that mountain building processes were less efficient during the Archaean, due to a reduced strength of the continental lithosphere (Rey and Houseman, 2006; Rey and Coltice, 2008). Including this effect in the evolution of the hypsometry has a major impact on the calculated emerged area of continental crust. For the late-Archaean, we calculate that the emerged area of land was only 2–3% of the Earth's area and therefore predict that the continents were largely flooded. These results are of fundamental importance to the understanding of the coupling of the Earth's geochemical reservoirs. We show that they are consistent with a large number of global geochemical trends in the late-Archaean record.

2. Model formulation

2.1. Isostasy

Because of isostatic equilibrium between continents and oceans, a change in sea level h_f with respect to present-day can be written as

$$\Delta h_f = \Delta d_r \left(1 - \frac{\rho_w}{\rho_m}\right) + \Delta d_{oc} \left(1 - \frac{\rho_{oc}}{\rho_m}\right) - \Delta d_{cc} \left(1 - \frac{\rho_{cc}}{\rho_m}\right), \quad (1)$$

where Δd_r, Δd_{oc} and Δd_{cc} are changes in the depth of ridge crests (Reymer and Schubert, 1984), the thickness of the oceanic crust (Galer, 1991) and the thickness of the continental crust (Galer and Mezger, 1998) respectively (all parameters are listed in Table 1). The densities are ρ_m for the mantle, ρ_{oc} for the oceanic crust, and ρ_w for seawater.

The thickness of continents may have changed over the Earth's history, although seismic studies (e.g. Christensen and Mooney, 1995) show that the present mean thickness of Precambrian shields and platforms (41.5±5.8 km) is similar to the average thickness of the continental crust (41.1±6.2 km). We therefore assume a constant thickness of the continental crust (Δd_{cc}=0 in Eq. (1)) in this study.

Change in Δd_{oc} is a function of the mantle temperature, and change in Δd_r depends upon both mantle temperature and ocean volumetric budget.

2.2. Thickness of the oceanic crust

Sleep and Windley (1982) suggested that a hotter mantle in the Archaean would imply a thicker oceanic crust. Models of isentropic decompression melting have been developed to predict the thickness of the oceanic crust as a function of mantle potential temperature T_p (McKenzie, 1984), eventually taking into account the limiting effect induced by the pressure of the crustal over- burden (Vlaar and van den Berg, 1991). However, this effect is significant only when considering high mantle temperatures (T_p>1560 °C). When using isentropic decompression melting models, the thickness of the oceanic crust is generally assumed to be equal to the calculated thickness of melt, which completely ignores the role of dynamic extraction that could significantly affect the results (Šramek et al., 2007). Therefore, changes in oceanic crustal thickness should be assigned ±2.5 km uncertainties for a given temperature (Galer and Mezger, 1998). We compared the oceanic crustal thickness calculated using the model of McKenzie (1984) and the parametrisation of the peridotite solidus and liquidus given by McKenzie and Bickle (1988) with that calculated using the model of Korenaga (2006). The difference between both parametrisations for a mantle potential temperature 200 °C hotter than present is very close to the uncertainty range: the oceanic crustal thickness calculated by McKenzie and Bickle (1988) is only 2.7 km thicker than the one calculated by Korenaga (2006). The melt height produced for present-day upper mantle potential temperature (T_p=1280 °C) fits the observed 7 km-thick oceanic crust for both parametrisations. Due to the lack of a two-phase model taking dynamic extraction into account, the model of McKenzie (1984) is used in this study.

Table 1
Variables used in the models. Asterisks indicate present-day values

Symbol	Meaning	Value	Standard deviation	Unit
α_R	Coefficient of thermal expansion (rectangular system)	2.3×10^{-5}		K^{-1}
α_L	Coefficient of thermal expansion (triangular system)	2×10^{-5}		K^{-1}
κ	Thermal diffusivity	8×10^{-7}		$m^2\ s^{-1}$
ρ_{cc}	Mean density of the continental crust	2800	±100[a]	$kg\ m^{-3}$
ρ_{oc}	Mean density of the oceanic crust	3000	±100[a]	$kg\ m^{-3}$
ρ_m	Mean density of the upper mantle	3300	±100[a]	$kg\ m^{-3}$
ρ_w	Mean density of sea water	1030	±10[a]	$kg\ m^{-3}$
A^*_{cc}	Total area of continents	2.17×10^{14}[b]		m^2
A_E	Area of the Earth	5.1×10^{14}		m^2
A^*_f	Area of emerged continents	1.4×10^{14}		m^2
A^*_{oc}	Total area of oceans	2.93×10^{14}[b]		m^2
A^*_{sh}	Area of the continents, shelf included	1.75×10^{14}		m^2
a	Constant in the hypsometric function	0.0388		
b	Constant in the hypsometric function	0.608		
d^*_b	Average depth below ridges	1859		m
d^*_{cc}	Continental crustal thickness	41.1[c]	±6.2[c]	km
d_{cc}	Archaean continental crustal thickness	41.1	±7[d]	km
d_{max}	Maximum depth of the continental slope	2446		m
d^*_{oc}	Oceanic crustal thickness	7075	±700[e]	m
d_{oc}	Oceanic crustal thickness for T^*_p=1430 °C	20870	±2500[d]	m
d^*_r	Ridge depth	2446		m
E	Activation energy in the viscosity law	300		kJ
f^*	continental freeboard	8850		m
h^*_f	Sea level above the edge of the continental shelf	200		m
h^*_{max}	Maximum elevation above continental shelf	9050		m
R	Universal gas constant	8.314		$J\ mol^{-1}K^{-1}$
t^*_s	Mean subduction age of the ocean floor	90		Ma
t^*_{max}	Maximum age of the oceanic lithosphere	180		Ma
T^*_p	Potential temperature of the upper mantle	1280		°C
u^*	Oceanic half-spreading rate	2.59[f]		$cm\ a^{-1}$
V_o	Total volume of oceans	1.36×10^{18}	$\pm2\times10^{17}$[g]	m^3
z	Exponent in the hypsometric function	0.706		

[a] Galer (1991).
[b] Schubert and Reymer (1985).
[c] Christensen and Mooney (1995).
[d] Galer and Mezger (1998).
[e] White et al. (1992).
[f] Cogné and Humler (2004).
[g] Harrison (1999).

2.3. Ocean volumetric budget

As depicted in Fig. 1, the volume of the ocean V_o is divided into V_a the reservoir above the mid-oceanic ridge, V_b the reservoir below the mid-oceanic ridge, V_s the reservoir above the continental slope, and V_l the volume due to change in the hypsometric curve after water loading or unloading:

$$V_o = V_a + V_b + V_s + V_l. \tag{2}$$

The volume V_a depends on the total oceanic area A_{oc} and depth of the ridges d_r and is written

$$V_a = A_{oc} d_r. \tag{3}$$

2.3.1. Bathymetry

The volume of the reservoir V_b depends on the seafloor bathymetry, caused by cooling of the ocean lithosphere, which is modelled here with the theory of semi-infinite half-space cooling combined with isostasy (e.g. Turcotte and Schubert, 1982):

$$d_b(t) = \frac{2\rho_m \alpha T_p}{\rho_m - \rho_w}\sqrt{\frac{\kappa t}{\pi}},$$

where d_b is the depth below ridge axis, α is the thermal expansivity, κ is the thermal diffusivity and t is the age of the oceanic floor. For a given distribution of area of the ocean floor with age dA/dt, the volume of the reservoir V_b can therefore be written as

$$V_b = \int_0^{t_{max}} d_b(t)\frac{dA}{dt}dt,$$

where t_{max} is the maximum age of the ocean floor. Different age distribution models can be derived, depending on the mode of mantle cooling (Labrosse and Jaupart, 2007). In the following, we explore changes in bathymetry as a function of the potential temperature of the mantle T_p and of the area of continental crust A_{cc} for different distribution models.

Two different approaches can be used to calculate the distribution of seafloor ages, but they are mutually exclusive. The first is based on boundary layer stability analysis, stating that subduction starts when a plate reaches a threshold age at which the thermal boundary layer becomes gravitationally unstable. As a consequence, such models imply a rectangular distribution of seafloor ages:

$$\frac{dA}{dt} = C_0 \text{ for } t < t_s,$$

where C_0 is the global rate of seafloor spreading and t_s is the age of subduction of the oceanic floor. For this distribution, the total area of the ocean floor is:

$$A_{oc} = C_0 t_s,$$

and the present-day subduction age is $t_s^* = 90$ Ma. The boundary layer theory implies that the age of subduction depends on mantle convective vigour through convection velocity:

$$t_s = t_s^* u^* / u,$$

where u is the half spreading rate at mid-oceanic ridges. For Rayleigh–Bénard convection, a given potential temperature T_p gives u as

$$u(T_p) = u^* \left(\frac{T_p}{T_p^*}\frac{\eta(T_p^*)}{\eta(T_p)}\right)^{2/3},$$

where the viscosity itself depends on the potential temperature of the mantle T as (Davies, 1980)

$$\eta = \eta_0 \exp\left(\frac{E}{RT}\right),$$

where E is the activation energy. This model is referred to hereafter as the Boundary Layer Theory (BLT). It implies a strong feedback between T_p and heat flow, suggesting larger radiogenic heat production or core

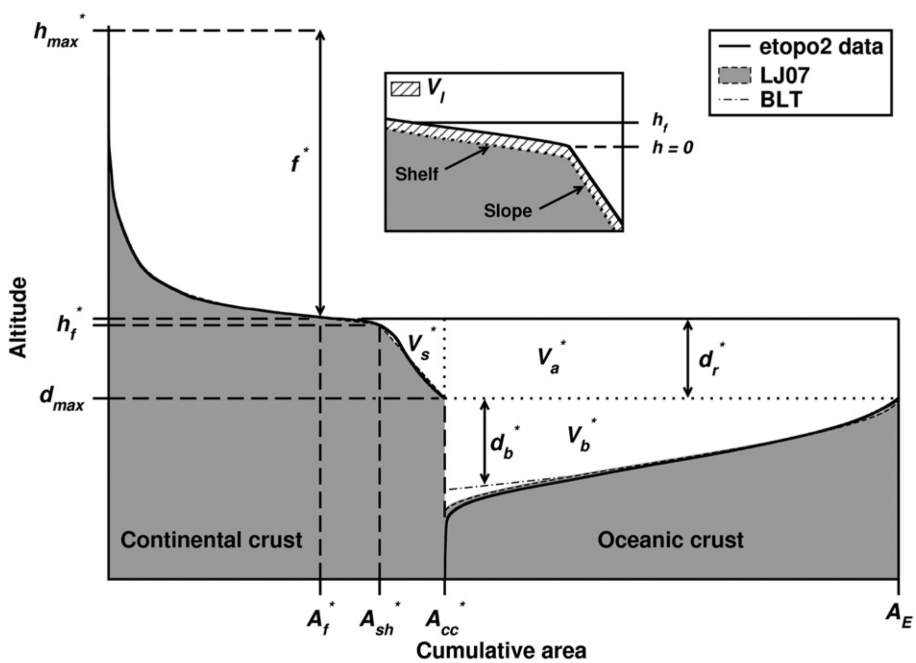

Fig. 1. Schematic diagram illustrating the parameters relating to the area of emerged continents (see Table 1 for definitions). Changes in continental freeboard f are equal and opposite to changes in sea level h_f. The oceanic portion of the observed hypsometry is flipped horizontally to better illustrate tectonic processes. Both rectangular (BLT) and linear (LJ07) seafloor age distributions are shown. The inset diagram shows the relationships between reference level $h = 0$, sea level h_f, continental shelf and continental slope (not to scale). The volume caused by isostatic adjustment of the continent V_l is also illustrated in the inset diagram for a hypothetical raise in sea level. Note that $V_l^* = 0$.

heat flow than what is usually proposed from cosmochemical arguments (see Jaupart et al., 2007, for a recent review).

For this reason, it has been proposed (Arndt, 1983; Korenaga, 2006) that the positive buoyancy of the oceanic lithosphere could delay the onset of subduction if the oceanic crust is thick enough. The positive buoyancy of the oceanic lithosphere is due to the change of partitioning intrinsic density between the oceanic crust and the underlying depleted mantle with change in mantle temperature. Including this chemical effect and small scale convection in the classical boundary layer theory reduces the dependence of u on T_p. This model, hereafter referred to as K06, proposes a moderate thermal evolution of the mantle which is more consistent with petrological constraints on the mantle temperature in the Archaean (Abbott et al., 1994). The advantage of the boundary layer approaches BLT and K06 is that the distribution of seafloor ages can readily be calculated for any given mantle temperature.

However, these models could be in contradiction with the observed present-day distribution of ages of the seafloor. Indeed, different oceanic lithospheres have different ages of subduction and the probability of subduction is approximately the same for any age (Sclater et al., 1981; Parsons, 1982; Rowley, 2002; Cogné and Humler, 2004; Labrosse and Jaupart, 2007). Hence the distribution is triangular:

$$\frac{dA}{dt}(x) = C_0\left(1 - x\frac{t}{t_{\max}}\right),$$

where $t_{\max}$ is the maximum age of the lithosphere and $x = A_{cc}/A_{cc}^*$ is the areal fraction of present continental crust. For this distribution, the area of oceans can be derived as

$$A_{oc} = \frac{(2-x)C_0 t_{\max}}{2}.$$

One problem of this approach is that there is no physical model as yet for a relationship between the maximum age of the lithosphere $t_{\max}$ and T_p. Labrosse and Jaupart (2007) inferred from their parametrisation of the thermal evolution of the Earth that $t_{\max}$ must be larger than 104 Ma at all ages. The authors also proposed that the deviation from the rectangular distribution of ages is caused by the presence of continents. Hence, as in their models, our model for the seafloor age distribution evolves from rectangular to triangular as a linear function of x (Labrosse and Jaupart, 2007). This model is referred to hereafter as LJ07.

For each model, the coefficient of thermal expansion α is adjusted so that the calculated ridge depth is equal to the present-day ridge depth (see Table 1). The area of continents at a depth greater than $d_{\max}$ is neglected in all models. The depth of mid-oceanic ridges is allowed to change through time in Eq. (1). The plots of present-day bathymetry for both distribution systems are shown in Fig. 1. The triangular distribution gives a better fit to the data as expected since it is an empirical model.

2.3.2. Hypsometry

The volume V_s of the water reservoir over the continent is a function of the hypsometric curve $f(A)$:

$$V_s = (A_{cc} - A_f)(h_f + d_{\max}) - \int_{A_f}^{A_{cc}} f(A)dA, \tag{4}$$

where A_f is the area of emerged continents. The hypsometric curve is described by a function modified from Strahler (1952) and Harrison et al. (1981) until the edge of the continental shelf, and the continental slope is taken as a linear function. This can be written as

$$f(A) = \begin{cases} b \cdot h_{\max}\left(\dfrac{1 - A/A_{sh}}{a + A/A_{sh}} \cdot a\right)^z & \text{if } A \le A_{sh} \\ \dfrac{A - A_{sh}}{A_{cc} - A_{sh}} d_{\max} & \text{if } A > A_{sh}, \end{cases} \tag{5}$$

where $h_{\max}$ is the maximum elevation of the continents, and a, b and z are constants determined to fit the eTOPO2 data (NOAA, 2006). Values are given in Table 1 for a correlation coefficient $R^2 = 0.999$. Sea level is always above the reference level ($h = 0$), defined at the edge of the continental shelf in $A = A_{sh}$.

The volume V_l, caused by isostatic adjustment after a sea level change over the continent (see Fig. 1), is given by

$$V_l = \begin{cases} \dfrac{\rho_w}{\rho_m}\left[\displaystyle\int_{A_f}^{\hat{A}_f}(h_f - h)dA + \int_{\hat{A}_f}^{A_{cc}}\left(h_f - \hat{h}_f\right)dA\right] & \text{if } h_f > \hat{h}_f \\ \dfrac{\rho_w}{\rho_m}\left[\displaystyle\int_{\hat{A}_f}^{A_f}(h - h_f)dA + \int_{A_f}^{A_{cc}}\left(h_f - \hat{h}_f\right)dA\right] & \text{if } h_f < \hat{h}_f, \end{cases}$$

with $\hat{A}_f$ and $\hat{h}_f$ being respectively the area of emerged continents and sea level before water loading or unloading. h is defined using the hypsometric curve $f(A)$ described in Eq. (5). For a transgression, $A_f < \hat{A}_f < A_{cc}$ and V_l is positive. As for a regression, $\hat{A}_f < A_f < A_{cc}$, and V_l is negative, therefore the continent is uplifted. V_l is zero today since the observed hypsometric curve is the reference.

Eustatic sea level is determined by iterative computing of V_o from Eq. (2) until constancy of ocean volume is reached. In this process, V_a is calculated from Eq. (3) with d_r computed from Eq. (1) and the integral of $f(A)$ in Eq. (4) is calculated numerically.

3. Influence of thermal evolution models on continental freeboard

In this section, we explore the dependencies of continental freeboard and area of emerged continents on mantle potential temperature and continental area, assuming constant hypsometry and ocean volume.

3.1. Continental freeboard

For a given continental area, sea level increases as temperature increases, due to the combination of two processes: thickening of the buoyant oceanic crust and flattening of the seafloor. The former process has an identical effect in the BLT, K06 and LJ07 models. However, models differ in the parametrisation of the flattening of the bathymetry. Indeed, in the BLT model, spreading rates are very sensitive to mantle temperature. Hence, for mantle temperatures more than 100 °C hotter than present-day, the bathymetry is almost flat, the volume of the reservoir below ridges V_b is reduced and sea level is high. The K06 model, however, proposes a reduced sensitivity of spreading rate to temperature and the LJ07 model proposes a modest change in the maximum age of the ocean floor. Flattening of the bathymetry is thus buffered in these two models. As a consequence, K06 and LJ07 models have a lower sensitivity to T_p than the BLT model.

Continental freeboard is also sensitive to the fraction of the Earth's area covered by continental plates. As new continental crust is produced, the total area of the continents A_{cc} increases while the total area of the oceans decreases. Removing the continental crust entirely would result in a decrease in sea level of ~1000 m. The dependency of continental freeboard on continental area is the same in the BLT and K06 models but is slightly different in the LJ07 model, in which the seafloor age distribution changes with continental area.

Fig. 2 shows the contours of constant continental freeboard as a function of continental area and mantle temperature for each model. The constant freeboard hypothesis allows for sea level variations of ±200 m, which is the amplitude of Phanerozoic sea level change (Haq et al., 1987; Miller et al., 2005). The slope of the contours shows that continental freeboard calculated using the BLT model is more sensitive to mantle temperature than other thermal evolution models. Constant freeboard (±200 m) can be achieved if mantle temperature did not exceed its present-day value by more than 90 °C (75–110 °C) for the

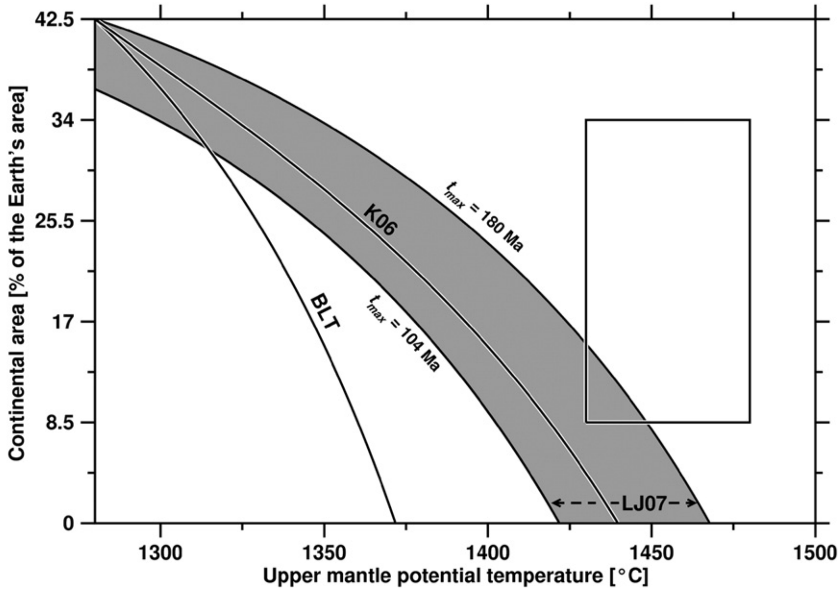

Fig. 2. Contours of constant continental freeboard as a function of continental area and mantle temperature, for present hypsometry. Results are presented for each model. The shaded area shows the variability of the continental freeboard with t_{max} for the linear seafloor age distribution model (LJ07). The box shows the possible range of parameters in the Archaean: the mantle was probably 150–200 °C hotter than present and the continental area increased from 20% to 80% of present.

BLT model, 160 °C (130–190 °C) for the K06 model, 190 °C (165–210 °C) for the LJ07 model with t_{max} = 180 Ma, and 140 °C (120–165 °C) for the LJ07 model with t_{max} = 104 Ma. The temperature calculated by the K06 model is close to the temperature suggested by Galer (1991). Both the K06 and LJ07 models can accommodate a greater mantle temperature than the BLT model, keeping continental freeboard constant.

To achieve constancy of the freeboard, a change in mantle temperature must be counter-balanced by a change in continental area (Schubert and Reymer, 1985). For the BLT model, a cooling of 30 °C is necessary to balance 30% of crustal growth, while a cooling of ~60 °C is required for other models. If most of the continents were produced very early in the Earth's history, the constancy of continental freeboard within the Archaean and Proterozoic could not be achieved because cooling is supposedly between 50 and 100 °C Ga^{-1} (Jaupart et al., 2007). For all thermal evolution models, keeping continental freeboard constant throughout most of the Earth's history could be very difficult if the episodicity of crustal production (Rudnick, 1995; Condie, 2000) is not matched by an episodicity of mantle cooling.

3.2. Area of emerged continental crust

One novel aspect of our approach is the calculation of the area of emerged continental crust. Today, continents represent 42.5% of the Earth's area (Schubert and Reymer, 1985), and emerged crustal area represents only 27.5%. At constant freeboard, the change in the fraction of emerged crust is proportional to the change in continental area (see Fig. 3). However, as expected from the freeboard equations, the greater the mantle temperature, the higher the sea level and the smaller the emerged crustal area. For present-day continental area and hypsometry, an increase in mantle temperature of 100 °C would reduce the area of emerged continental crust to ~9–16% depending on the thermal evolution model.

All other parameters being constant, the area of emerged continental crust varies with the total area of continental crust. However, there is a trade-off between continental area and emerged area: a larger continental area generally leads to a greater emerged area, but it also triggers an increase in sea level. As expected, the BLT model presents the highest sensitivity to mantle temperature. For a hotter mantle, this model thus proposes an emerged continental area much lower than that proposed by the more conservative LJ07 and K06 models (Fig. 3).

For low temperature and large continental area, the area of emerged continental crust is not sensitive to continental growth. The emerged area can only increase with continental growth for a hot mantle, where continental freeboard is highly sensitive to mantle temperature. Hence, a pulse of crustal growth during the Archaean would have resulted in a net increase of emerged area. The area of emerged continental crust depends mainly upon mantle temperature whereas continental freeboard depends on the ratio of crustal growth to mantle cooling.

4. Area of emerged continental crust in the Archaean

4.1. Mantle temperature and continental area in the Archaean

It is generally agreed that the temperature of the mantle has decreased over time while the continental area has increased. The main argument for a hotter mantle in the Archaean is based on the petrography and geochemistry of komatiites (Green et al., 1975; Nisbet et al., 1993). Temperature estimates depend on the water content of komatiites. Arndt et al. (1998) argued that while some komatiites formed from hydrous melt, most were dry. Dry komatiites, erupted in a plume environment, would suggest a mantle temperature 200–300 °C hotter at 3.5 Ga (Nisbet et al., 1993) compared to a 100–200 °C hotter mantle for hydrated komatiites erupted in a subduction environment (Grove and Parman, 2004). Abbott et al. (1994) concluded from their petrological and geochemical study on mid-ocean ridge basalts that the potential temperature of the mantle at 2.8 Ga was 137–187 °C greater than present. Using these constraints, we assume that the mantle was 150–200 °C hotter during the Archaean.

Most models of continental growth show that the volume of continental crust has increased over time (e.g. Campbell (2003) and Rino et al. (2004) for recent reviews), with many pointing toward a major growth pulse during the Neoarchaean (2.8–2.5 Ga). Following the work of Campbell (2003) based on Nb/U ratios in Archaean basalt–komatiite suites, we assume that the continental crust grew from ~20% of the present continental area at around 4 Ga to 80% by 2.5 Ga.

Putting these considerations of mantle temperature and crustal growth together, and assuming constant hypsometry, the calculated area of emerged continents in the Archaean is between 1 and 12% of the Earth's area (Fig. 3).

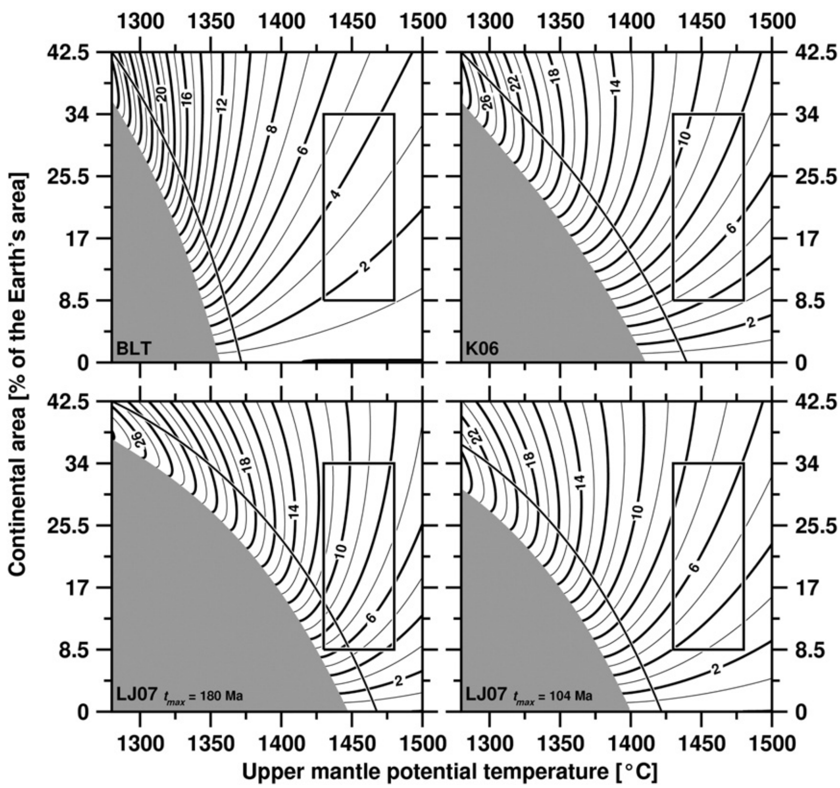

Fig. 3. Contours of the area of emerged continental crust (in % of the Earth's area) as a function of continental area and mantle temperature for present hypsometry. Results are presented for each model. The shaded area in each diagram shows the conditions within which the continental slope would emerge. The thick black lines are the contours of constant continental freeboard. The boxes are as defined in Fig. 2.

4.2. Hypsometry in the Archaean

The hypsometric curve is an expression of the balance between mountain building processes and erosion processes. Orogenies are expected to produce high plateaus and high peak elevations that erosion processes can flatten down in a few hundred million years (Harrison, 1994). Considering the present-day variation in hypsometry from one continent to another (e.g. Harrison et al., 1981) we propose that the global hypsometry could have fluctuated during the Phanerozoic between an orogenic end-member and an anorogenic end-member (see Fig. 4). The orogenic end-member for the Phanerozoic is illustrated by the present-day hypsometry of Asia,

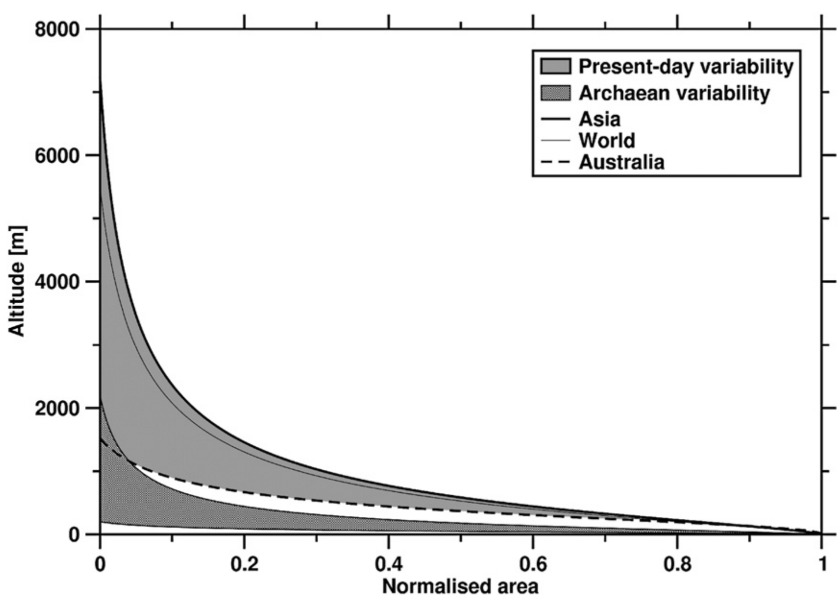

Fig. 4. Variability of continental hypsometric curves for the present-day and for the Neoarchaean. The area is normalised. The datum of the ordinate axis is the altitude at the edge of the continental shelf (presently −200 m). The hypsometric curves for Asia and Australia are built using parameters from Harrison et al. (1981). The world hypsometric curve is that compiled from the eTOPO2 data.

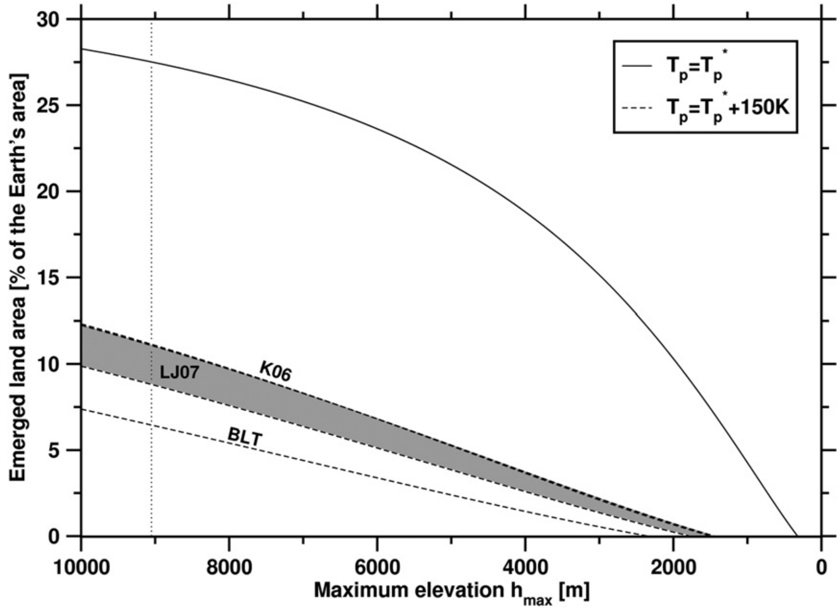

Fig. 5. Changes in the area of emerged continental crust as a function of the elevation of the highest peak on Earth, for present-day continental area ($x=1$) and hypsometry. The plain curve is calculated for present mantle temperature and is the same for all models. The dashed curves are illustrated for each model and for a mantle temperature 150 °C greater than today. The shaded area shows the variability of emerged area with t_{max} for the linear seafloor age distribution model (LJ07). In this calculation the results for the K06 model are very close to those for the LJ07 model with $t_{max}=180$ Ma. The vertical dotted line shows the present-day maximum elevation.

which supports the highest plateau and highest peak on Earth, built over the last 50 Ma. The Earth's present-day hypsometry is actually very close to this end-member. In contrast, the global hypsometry during the upper-Jurassic/lower-Cretaceous, a time of continental flooding, is likely to have been closer to that of the anorogenic end-member, present-day Australia.

Assuming that the overall shape of the end-member hypsometric curves (described by Eq. (5)) has remained constant over time, the secular evolution of the global hypsometry depends only upon the maximum elevation. Harrison (1994) proposed that early in the Earth's history, a greater generation of radioactive heat would imply greater rates of mountain building and thus higher continental elevations. In contrast, Rey and Houseman (2006) argued that a greater production of radiogenic crustal heat and a greater mantle heat flow would reduce the strength of the continental lithosphere and thus its ability to sustain high mountain belts (Rey and Houseman, 2006). Assuming conservative present-day strain rates (10^{-15} to 10^{-14} s^{-1}), the maximum elevation of orogenic plateaus during the Neoarchaean would have been 1800–2200 m (Rey and Coltice, 2008), allowing for a dynamically supported maximum elevation up to ~3600 m. Fig. 4 shows possible orogenic and anorogenic hypsometry end-members for the Neoarchaean. The proposed Neoarchaean orogenic end-member is very close to the present-day anorogenic end-member.

This reduced maximum elevation is at odds with some previous works, which have argued that Archaean orogenic processes and mountain belts were similar to their modern counterparts. For instance,

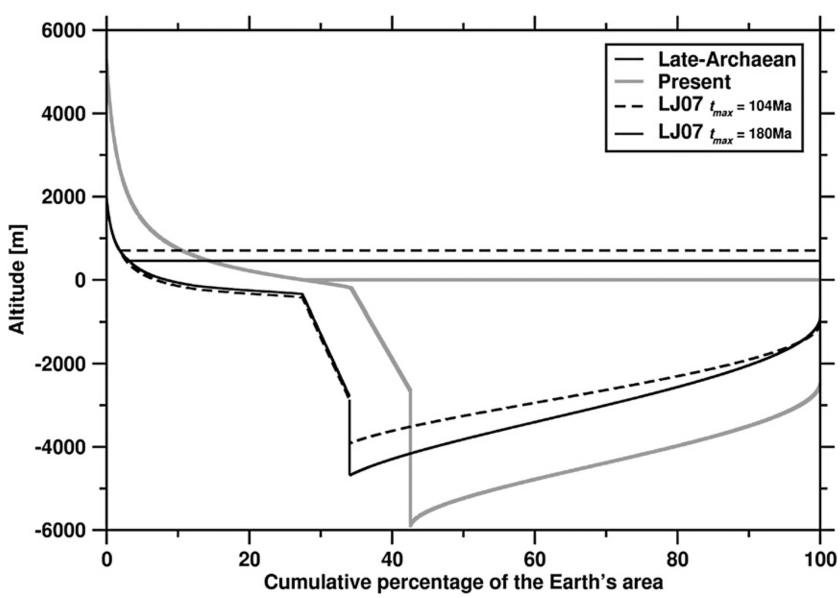

Fig. 6. Calculated hypsometry, bathymetry and sea level for a late-Archaean configuration (~2.5 Ga, $x=0.8$, $T_p=1430$ °C, $h_{max}=3615$ m) for the LJ07 model. The plain dark lines are for a maximum oceanic floor age of 180 Ma and the dashed lines for a maximum oceanic floor age of 104 Ma. The lighter lines represent present-day hypsometry, bathymetry and sea level for the LJ07 model.

England and Bickle (1984) argued that orogenic crustal thicknesses could have been up to 70 km in the Archaean "by noting that pressures in high-grade gneiss belts of 8–10 kb imply overburdens of 28 to 35 km [...] with 35 km of crust preserved beneath the sampled locality". Their paper is one of the cornerstones of uniformitarianism as it concludes that Archaean geothermal gradients were similar to present-day ones. A quarter of a century of research has shown that erosion following instantaneous thickening is not the only process able to exhume high-grade metamorphic terranes. Firstly, it should be noted that exhumation of deep crustal levels may occur with little finite crustal thickening when erosion rates are similar to crustal thickening rates. Secondly, exhumation of deep crustal levels in metamorphic core complexes occurs in association with horizontal extension and thinning of the crust. Thirdly, the coeval sagduction of dense greenstone covers and exhumation of their felsic basement, a process described in many Archaean cratons (McGregor, 1951; Dixon and Summers, 1983; Bouhallier et al., 1995), is known to have exhumed kyanite bearing assemblages (Delor et al., 1991; Collins et al., 1998). Therefore, one cannot assume that 8–10 kb high-grade gneiss terranes are proof of a 70 km-thick orogenic crust in the Archaean, nor that Archaean continental geotherms were similar to modern ones.

4.3. Flooded continents in the late-Archaean

Assuming that the hypothesis of warmer and weaker continental lithospheres in the Archaean is valid, the effect of decreasing the maximum elevation in the present-day global hypsometric curve is a decrease in the area of emerged continental crust as shown in Fig. 5. Note that, however, the relationship between maximum elevation and area of emerged continental crust can differ depending on mantle temperature. We represent a late-Archaean configuration in Fig. 6, based on the hypsometry of our proposed Neoarchaean orogenic end-member (see Fig. 4). We use the LJ07 model, allowing the maximum age of the ocean floor to vary between 104 and 180 Ma, since the results for the K06 model fall within these limits. For a maximum oceanic floor age t_{max} of 180 Ma, the calculated emerged crustal area is 3.1% of the Earth's area. For t_{max} = 104 Ma, it is 1.8%. It is worth noting that a small increase in total continental area from 34% to 42.5% since the end of the Archaean has resulted in a large increase in emerged area, of approximately 25% of the Earth's area.

Our models show that large submerged continental areas existed in the late-Archaean. For a maximum oceanic floor age of 180 Ma (104 Ma) we calculate that ~24% (~25.5%) of the Earth's area is covered by water ~800 m (~1100 m) deep. Note that these are maximum depths, as sedimentation processes are not taken into account here.

4.4. Uncertainty in the calculation of the area of emerged land

We have shown that the calculation of the area of emerged land depends upon the chosen thermal evolution model for the Earth. The results of the calculation are also affected by the natural variability associated with each of the parameters used for the computation, especially the thickness of the continental crust and the volume of the ocean. To illustrate the uncertainty in the calculation of the area of emerged land due to this natural variability, we performed a Monte Carlo analysis for the model presented in Fig. 6 assuming a maximum oceanic floor age of 104 Ma. The input parameters are described by Gaussian distributions using the mean values and standard deviations listed in Table 1. From 10^5 realisations, we predict a distribution of the area of emerged land and of sea level.

We analyse both the predicted distributions for the effect of density, crustal thicknesses and ocean volume variabilities. Three types of configurations can be obtained in which continents are (a) entirely emerged, (b) partially flooded or (c) entirely flooded. The input and output of these analysis are summarized in Table 2. The

Table 2
Effect of uncertainties on densities and natural variability in continental and oceanic crustal thicknesses and oceanic volume for the calculation of the area of emerged land for the example in Fig. 6 with t_{max} = 104 Ma

Variability incorporated	Result	N	Input error and uncertainties						Monte Carlo output	
			ρ_{oc}/ρ_m	ρ_{cc}/ρ_m	ρ_w/ρ_m	Δd_{oc}[km]	Δd_{cc}[km]	V_o[×10^{18}m^3]	h_f[m]	A_f [%A_E]
None	$1<A_f<A_{sh}$	–	0.91(–)	0.85(–)	0.312(–)	13.8(–)	0.0(–)	1.36(–)	855.9(–)	2.04(–)
Densities	$0<A_f<A_{sh}$	95,900	0.91(3.7×10^{-2})	0.85(3.9×10^{-2})	0.312(9.5×10^{-3})	–	–	–	727.9/718.3(344.2)	4.28/2.75(4.39)
	Entirely emerged	4099	1.00(1.6×10^{-2})	0.89(3.9×10^{-2})	0.327(8.5×10^{-3})	–	–	–	–	–
	Entirely flooded	1	0.74 (–)	0.80(–)	0.266(–)	–	–	–	––	
Thicknesses	$0<A_f<A_{sh}$	69,560	–	–	–	13.9(2.6)	−2.5/−2.1(5.6)	–	965.2/919.2(571.6)	4.05/1.79(5.49)
	Entirely emerged	24,015	–	–	–	13.2(2.6)	12.0/11.0(4.8)	–	–	–
	Entirely flooded	6425	–	–	–	14.7(2.6)	−18.1/−17.4(3.9)	–	–	–
Ocean volume	$0<A_f<A_{sh}$	99,284	–	–	–	–	–	1.36(0.20)	707.2/703.2(276.5)	3.84/2.84(3.32)
	Entirely emerged	716	–	–	–	–	–	0.8(0.06)	–	–
All	$0<A_f<A_{sh}$	63,995	0.91(3.8×10^{-2})	0.85(3.9×10^{-2})	0.312(9.7×10^{-3})	13.9(2.6)	−2.1/−1.9(6.3)	1.38(0.20)	942.8/883.6(577.6)	4.29/1.93(5.66)
	Entirely emerged	26,683	0.93(4.2×10^{-2})	0.85(4.1×10^{-2})	0.314(1.0×10^{-2})	13.4(2.6)	10.1/9.8(6.1)	1.29(0.20)	–	–
	Entirely flooded	9322	0.88(3.5×10^{-2})	0.82(3.3×10^{-2})	0.307(8.6×10^{-3})	14.6(2.6)	−14.1/−13.8(5.7)	1.43(0.22)	–	–

The information is given in the form mean (standard deviation) or mean/median(standard deviation) when three numbers are present.

predicted distributions have large standard deviations, up to 100% of the mean value for A_f and 50% for h_f. This is caused by the large variability we have considered for densities, thicknesses and ocean volume.

The predicted distributions of the area of emerged land present a positive skew which explains the observed differences between mean and median. The medians for A_f range between 1.79 and 2.84% of the Earth's surface. Hence, the model presented before in Fig. 6 is one of the most probable.

Taking all variabilities into account, the input parameters of models predicting partial flooding (64% of all the realisations) are consistent with that obtained in the determinist approach for a change in the thickness of the continents of −2.1 ± 6.1 km. For complete flooding to be obtained (9% of the trials) the continental crust has to be about 10 km thinner than present. For a crust much thicker than present (10.2 ± 6.1 km), and a reduced ocean volume (about 10% lower than present), continents can be entirely emerged. This situation is obtained for 27% of the trials. The calculated thickening of the crust necessary to entirely emerge the existing continents in the Archaean is twice that proposed by Galer and Mezger (1998) and abundant geological and geochemical evidence suggesting flooded continents in the Archaean (see next section) would advocate for Archaean continents slightly thinner or as thick as present-day ones.

4.5. Geological and geochemical footprints for a late-Archaean emergence of continents

Largely flooded continents at the end of the Archaean as shown by our models are consistent with the observation of widespread volcanism on submerged continental platforms (Arndt, 1999; Kump and Barley, 2007). The histogram in Fig. 7a shows that more than 80% of the Archaean Large Igneous Provinces (LIPs) were emplaced in subaqueous conditions, compared to less than 30% after 2.5 Ga (Kump and Barley, 2007). Assuming no sampling bias, this fraction of emerged LIPs is a direct measure of the fraction of emerged continents. Note that the proportion of emerged continental crust at 2.5 Ga calculated using our model (10%, Fig. 6) is consistent with the observation that ~15% of the LIPs emplaced between 2.75 and 2.5 Ga are subaerial (Fig. 7a). Further research regarding the subaerial or submarine nature of LIPs of different ages would better constrain the fraction of emerged continental crust over time. Moreover, this increase in the amount of stable-shelf sediments throughout the Archaean (Eriksson and Fedo, 1994) leading up to the formation of giant carbonate platforms at 2.9 Ga (Wilks and Nisbet, 1985) and between 2.6–2.4 Ga (Eriksson et al., 2005) confirms that continental crust mainly evolved below sea level.

The reduced area of emerged continents (up to 3% of the Earth's area, which is roughly the size of South America) and maximum elevations (approximately 3600 m) calculated using our models would still account for the common occurrence of clastic sediments and basins in the Archaean (Nisbet, 1987), including the oldest known unconformity dated at 3.46 Ga (Buick et al., 1995). Nevertheless, Archaean clastic sediments are known to be not as well mixed as present-day ones (Sircombe et al., 2001), and some authors have noted that limited sediments were supplied to continental margins in the early Archaean (Lowe, 1994). This is compatible with a reduced emerged continental area and smaller drainage basins (Bleeker, 2002).

A reduced area of emerged continental crust is also consistent with the lack of felsic fingerprints in the sedimentary record before ~2.5 Ga, which was explored by: a) Taylor and McLennan (1985) who showed that the geochemical composition of black shales, a proxy for the composition of emerged continental crust, changed from mafic-dominated to felsic-dominated at ~2.5 Ga; b) Veizer and Compston (1976) who showed that the isotopic ratio $^{87}Sr/^{86}Sr$ of marine carbonates separated from the calculated mantle evolution before ~2.5 Ga towards more radiogenic values after ~2.5 Ga (Fig. 7b) –

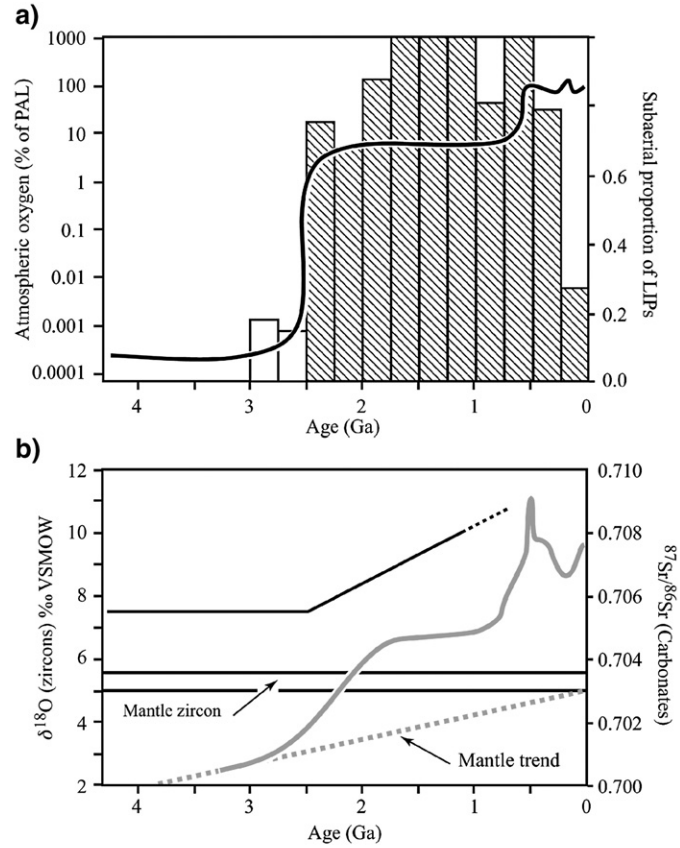

Fig. 7. Possible geological and geochemical lines of evidence for the proposed late-Archaean emergence of the continents. a) Histogram: proportion of subaerial LIPs through time (from Kump and Barley, 2007); Curve: evolution of atmospheric oxygen in percentage of Present Atmospheric Level (from Kump, 2008). b) Black lines: evolution of the $\delta^{18}O$ of detrital and inherited zircons (from Valley et al., 2005). The mantle value is 5.3 ± 0.3‰; Grey lines: evolution of the isotopic ratio $^{87}Sr/^{86}Sr$ of the mantle (dotted line) and of marine carbonates (plain line) (from Shields and Veizer, 2002). These signals all indicate a major change at the end of the Archaean (2.5 Ga).

this shift in composition implies a change from a "mantle"-buffered to a "river"-buffered ocean because the felsic continental crust is a major reservoir for radiogenic strontium (Shields and Veizer, 2002) –; and c) Valley et al. (2005) who showed that the $\delta^{18}O$ signature of igneous zircons changed from mildly-evolved magmatic values ($5 < \delta^{18}O < 7.5$) before 2.5 Ga to values increasingly greater than 7.5 after 2.5 Ga (Fig. 7b). They interpreted this change in $\delta^{18}O$ as an increase in the intra-crustal recycling of high $\delta^{18}O$ felsic sediments.

The appearance of the felsic signature in the sedimentary record is generally interpreted as a major pulse in crustal growth at ~2.7 Ga (Taylor and McLennan, 1985; Shields and Veizer, 2002; Valley et al., 2005). However, the discrepancy between the U–Pb crystallisation age and the Hf model age of low $\delta^{18}O$ detrital and inherited zircons of the Gondwana supercontinent (Kemp et al., 2006) has lead to the conclusion that it could take up to one billion years for newly formed continental crust to dominate the sedimentary record (Hawkesworth and Kemp, 2006). Hawkesworth and Kemp (2006) attributed this time lag to the emplacement of new crust by under- or intra-plating within older crust, delaying its erosion and contribution to the sedimentary record. However, we attribute this delay to the fact that the felsic continental crust was both largely covered under LIPs and mostly flooded during the Archaean. Hence, the anomalies in the sedimentary record after 2.5 Ga do not necessarily represent a pulse in crustal growth at ~2.7 Ga, but rather the emergence of the continents allowing for the erosion of the mafic cover and the exhumation of the felsic reservoir. Finally, this emergence of the continents would most likely contribute to the rise of the atmospheric oxygen at ~2.5 Ga (see Kump, 2008, for a recent review; Fig. 7a) as it would result in the reduction of the proportion of submarine LIPs, a major sink of oxygen (Kump and Barley, 2007), and in an increase in the weathering of emerged felsic material, a major sink of carbon dioxide (e.g. Kramers, 2002; Lowe and Tice, 2004).

5. Conclusions

The models presented in this study allow the calculation of emerged crustal area as a function of mantle temperature, hypsometry and continental area. Testing three different thermal evolution models, we show that constancy of the continental freeboard (±200 m) can be achieved if the upper mantle potential temperatures has never been 110–210 °C hotter than present. This temperature range is relatively large because freeboard models greatly depend upon thermal evolution models. For constant hypsometry, the area of emerged continental crust would not exceed 12% of the Earth's area by 2.5 Ga. Assuming less efficient mountain building processes in the Archaean further reduces this area to 2–3% of the Earth's area. Given the possibility that 80% of the present-day continental area was already produced by 2.5 Ga, this implies that the continents were mostly flooded throughout the Archaean. Their progressive emergence left in the sedimentological record the footprints of the coupling of a new geochemical reservoir (the felsic continental crust) with previously coupled reservoirs (the mantle and the hydrosphere). Continental emergence resulted in major changes in the Earth's environment, notably by contributing to the rise of atmospheric oxygen.

The models presented in this paper have some limitations. Firstly, there is a poor control of ocean volume over time. For instance, a secular change in the depth of mid-ocean ridges could modify the ocean volume by changing the efficiency of hydration of the oceanic crust (Kasting and Holm, 1992). Secondly, we assume a constant thickness of continental crust over time. The performed Monte Carlo analysis shows that the thickness of the continental crust is the key assumption, and that the continents could have been entirely emerged by 2.5 Ga if the continental crust was 10 km thicker than present. Although, this value is twice that suggested by previous studies Galer and Mezger (1998). Finally, we propose a model assuming isostatic equilibrium, but convective motions in the mantle induce dynamic topography (Husson and Conrad, 2006) and thus small-amplitude changes in sea level over short time scales.

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