

Continental emergence and growth on a cooling earth

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Abstract

Isostasy considerations are connected to a 1-D model of mantle differentiation due to pressure release partial melting to obtain a model for the evolution of the relative sea level with respect to the continent during the earth secular cooling. In this context, a new mechanism is derived for the selective exhumation of exposed ancient cratons. The model results in a quantitative scenario for sea-level fall due to the changing thicknesses of the oceanic basaltic crust and its harzburgite residual layer as a function of falling mantle temperature. It is also shown that the buoyancy of the harzburgite root of a stabilized continental craton has an important effect on sea-level and on the isostatic readjustment and exhumation of exposed continental surface during the earth's secular cooling.

The model does not depend on the usual assumption of constant continental freeboard and crustal thickness and its application is not restricted to the post-Archaean. It predicts large-scale continental emergence near the end of the Archaean and the early Proterozoic. This provides an explanation for reported late Archaean emergence and the subsequent formation of late Archaean cratonic platforms and early Proterozoic sedimentary basins.

For a period of secular cooling of 3.8 Ga, corresponding to the length of the geological record, the model predicts a fall of the ocean floor of some 4 km or more. For a constant ocean depth, this implies a sea-level fall of the same magnitude. A formula is derived that allows for an increasing ocean depth due to either the changing ratio of continental with respect to oceanic area, or to a possible increase of the oceanic volume during the geological history. Increasing ocean depth results in a later emergence of submarine ancient geological formations compared to the case when ocean depth is constant. Selective exhumation is studied for the case of constant ocean depth. It is shown that for this case, early exposed continental crust can be exhumed to a lower crustal depth, which explains the relative vertical displacement of low-grade- with respect to high-grade terrain. Increasing ocean depth is not expected to result in diminished exhumation. © 2000 Elsevier Science B.V. All rights reserved.

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1. Introduction

Isotopic geochemistry of the present-day sedimentary record reveals that most of the continental upper crust had been fractionated from the mantle at the start of the Proterozoic (2.5 Ga). Estimates range from 60 to 80% or more (Moorbath, 1975;

Taylor and McLennon, 1985; Patchett and Arndt, 1986; Rudnick, 1995). Since then, most of the upper continental crust has been recycled by erosion and deposition. Continental growth was low during the early Archaean, but it experienced a strong growth pulse during the later Archaean, which generated most, probably some 5% or more, of the upper continental crust (Windley, 1977; Taylor and McLennon, 1985; Condie, 1989; Rudnick, 1995). At 2.5 Ga ago, probably most of

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the cratonic continental crust and its deep mantle root had been fractionated from the mantle and had been stabilized. The formation of late Archaean cratonic sedimentary platforms and vast early Proterozoic sedimentary basins requires part of the continental crust to have emerged above sea level during the middle to late Archaean to be eroded and exhumed subsequently (Windley, 1977).

Continental freeboard has been considered to be a crucial parameter in the discussion of continental growth. Phanerozoic changes in continental freeboard have been the subject of discussion in terms of marine transgressions and regressions since the 1930's (Kuenen, 1939; Umbgrove, 1939). The causes of 'the pulse of the earth' of eustatic changes were ascribed to unknown workings of the interior of the earth. Only after the advent of the hypothesis of plate tectonics could Phanerozoic changes of continental freeboard be related to plate tectonic cyclic activity (Hays and Pitman, 1973; Hallam, 1977; Vail et al., 1978). Oscillating sea-level changes with an amplitude of about 200 m and a wavelength of about 200 Ma are recorded to be superposed on a small secular fall during the Phanerozoic. As no direct observation of seafloor older than 200 Ma is available, the Precambrian record of eustatic sea-level changes connected to plate tectonic activity is unknown. Therefore, only secular changes of continental growth and freeboard have been the subject of investigation (Wise, 1973; Schubert and Reymer, 1985; Galer, 1991; Galer and Mezger, 1998) based on model calculations constrained by assumptions about model parameters. The volume of the ocean, continental freeboard, and continental thickness have been constrained to be constant. Also, the condition of Airy isostasy for continental and oceanic crusts has been assumed in these studies.

Wise (1973) considers a constant volume post-Archaean continental crust that is maintained by stationary recycling of continental material through the mantle. Airy isostatic equilibrium is required to hold for a constant thickness continental and oceanic crusts.

Schubert and Reymer (1985), in order to accommodate post-Archaean additions from the mantle to the continental crust to compensate for losses by erosion of the continental surface require

secular deepening of the oceanfloor in order to meet the constraints of constant freeboard and crustal thicknesses of continental and oceanic crusts. They ascribe the secular foundering of the ocean bottom to the decline of radioactive heat production in the earth's interior. However, they do not present a rationale connecting deepening of the ocean basin to the decline of heat sources. Moreover, neither do they clarify why the continental surface does not founder as a consequence of the same decline of heat production. Galer (1991) and Galer and Mezger (1998) try to overcome these difficulties by proposing a decreasing thickness of the ocean crust due to the secular cooling of the earth. Following McKenzie (1984), they assume that, like at present, new oceanic crust is generated at oceanic spreading centres by isentropic decompression melting. Moreover, he takes Airy isostatic equilibrium to be governed by the density contrast between crustal rocks and fertile mantle peridotite (fertile mantle peridotite means upper mantle rock from which no basalt has been segregated). However, it has been demonstrated by Vlaar and Van den Berg (1991) that on a cooling earth, the low-density residual harzburgite layer also has to be taken into account when comparing isostatic conditions at different mantle temperatures. Moreover, they showed that the segregated basaltic layer has a strongly reduced thickness compared to that which McKenzie (1984) deduced. This is due to the circumstance that decompression melting only proceeds to the base of the formed crust.

Hence, the oceanic crustal thickness used by Galer (1991) and Galer and Mezger (1998) is about a factor of two in excess of a more realistic scenario involving decompression melting at an oceanic spreading ridge (Vlaar and Van den Berg, 1991). As the two effects (neglecting the buoyancy effect of the low density harzburgite layer and taking the segregated basaltic layer to be largely overestimated, both by about a factor of two for the Archaean) oppose each other, by fortuity, the resulting sea-level fall does not deviate strongly from that derived in the present paper.

However, by strongly overestimating the thickness of the oceanic basaltic layer and assuming a constancy of the thickness and freeboard of continental crust from the middle Archaean onward,

Galer's conclusion (Galer, 1991; Galer and Mezger, 1998) that the mantle could not have been more than 150 K hotter than at present, cannot be validated as Airy isostasy of an oceanic crust, which is taken to be overestimated by about a factor of two, is required to be in isostatic equilibrium with respect to a constant-thickness continental crust.

In the present study, continental crustal thickness and freeboard are not constrained a priori. Instead, from alternative isostatic considerations, it is shown that the early Archaean proto-continental crust was probably mainly submerged and that the emergence of the continents above sea level from the middle Archaean onward was highly selective and caused by a falling mantle temperature. Late Precambrian emergence of larger parts of the Proterozoic continent cannot be excluded a priori (Hargraves, 1976).

Apart from sporadic findings of Proterozoic ophiolites, no clearly recognizable remnants of Precambrian oceanic crust have been observed. No direct evidence for modern style plate tectonic activity is available for the latter period, nor is the timing known of the advent of plate tectonics in the earth's history. Vlaar (1985, 1986a,b) presented theoretical arguments that the conditions under which plate tectonics operates at present are violated at higher mantle temperatures. The strongest argument put forward concerns the strongly increasing gravitational stability of the oceanic lithospheric layering with increasing mantle temperature. The conditions under which subduction of Precambrian and particularly Archaean oceanic lithosphere could have been initiated could not be easily reached when mantle temperatures were higher than at present.

In the present paper, a basic assumption regarding the ancient oceanic lithosphere is that it is formed by differentiation and segregation of mantle diapirs rising to the base of an existing oceanic crust like at present at oceanic spreading centres. However, in this study, spreading is not required. The basic mechanism is the same as that for the creation of a new lithosphere at a modern spreading ridge, that is, isentropic decompression melting to the pressure of the base of an existing crust. A scenario for recycling this lithosphere at higher mantle temperatures is described by Vlaar

et al. (1993). As no Precambrian oceanic crust has survived to the present, it must be assumed that this crust and its harzburgite residue have been destroyed and reabsorbed and remixed in the mantle at the same rate as they were formed. In a cooling earth, this results in evolving decreasing thicknesses of the oceanic crust and its residual harzburgite layer, and — as will be shown in the following — changing isostatic conditions and fall of sea level during the earth's history.

The presence of high-melting-temperature komatiites in Archaean greenstone belts suggests that the earth was some 300 K hotter at the beginning of the Archaean (3.8 Ga) than at present. Some authors suggest (e.g. Abbott et al., 1994) that the temperature indicated by komatiites is not representative for the temperature of the upper mantle from which oceanic crust is fractionated. They assume that komatiites found in greenstone belts represent the higher temperature of mantle hot spots and that this temperature is not representative for the middle Archaean average upper mantle. However, it still remains to be seen whether greenstone belts represent oceanic crust. Here, they are viewed to be the ancient precursors of modern intracontinental basins (Hunter, 1974; McKenzie et al., 1980). From petrological arguments, Abbott et al. (1994) conclude that the mean mantle potential temperature, T_p , has decreased from about 1600°C at the middle Archaean to 1380°C at present. This appears not to be much at odds with the values for T_p used in this paper: $T_p = 1338^\circ\text{C}$ at present and 1667°C at 3.8 Ga. Rapid cooling of the earth during the (early) Archaean (Vlaar et al., 1993) brings these values within the range considered by Abbott et al. (1994).

A separate argument is given by the mere existence of the deep and strongly depleted cratonic lithospheric root as such. This suggests high upper mantle temperatures when the crust was segregated from the mantle during the early geological evolution.

2. Decompression melting

Decompression melting of undepleted mantle peridotite is the most important crust forming

process in the earth. It is held to be the cause of the present-day generation of basaltic oceanic crust at mid-ocean ridges. It is generally agreed upon that in a hotter mantle, the melting of a rising diapir starts at a deeper level and that a larger volume of basaltic magma is formed. The thickness of the formed basaltic crust is a function of the potential temperature, T_p , of the mantle. T_p is defined as the temperature obtained when a material volume in the mantle is decompressed under isentropic and metastable conditions to the pressure at the earth's surface. When, under supersolidus conditions, partial melting takes place, the temperature of the diapir decreases as a consequence of latent heat consumption. McKenzie (1984), Vlaar (1986b), and McKenzie and Bickle (1988) derive the thickness of the basaltic crust as a function of mantle potential temperature. An important effect that has not been implemented by these authors is that the amount of melting is strongly influenced by the pressure exerted by the crust already formed. This leads to a considerably reduced thickness of the segregated crust (Vlaar and Van den Berg, 1991).

In the following, it is assumed that throughout the geological evolution, oceanic crust is generated by 1-D isentropic decompression melting of a fertile mantle diapirs that rise to the base of the crust. At present, this process is associated with the spreading of lithospheric plates at oceanic ridges. The formalism developed by McKenzie (1984) is used but is corrected for the crustal pressure effect (Vlaar and Van den Berg, 1991). Calculations are based on the experimentally determined solidus and liquidus of fertile garnet peridotite by Takahashi (1980), and partial melting behaviour by Jaques and Green (1980).

Fig. 1 shows the thicknesses as a function of mantle potential temperature of the segregated basaltic crustal layer for two cases. In the lower curve, the pressure effect of the crust already formed has been accounted for (Vlaar and Van den Berg, 1991). The upper curve represents the case when all expelled melt or formed crust is removed and the harzburgitic residue is allowed to rise to the surface (McKenzie, 1984). The latter condition is not met in practice as the segregated oceanic crust is taken to be subaqueous and in

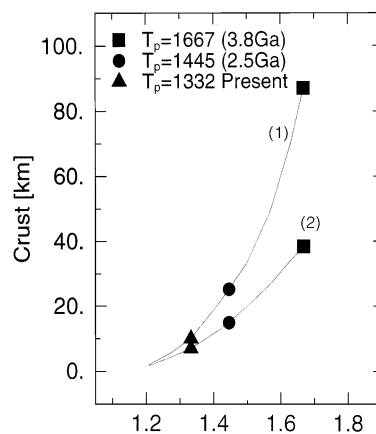


Fig. 1. Thickness of basaltic melt layer for two cases: (1) isentropic decompression melting proceeds to surface where melt is removed instantaneously until the top of the depleted layer reaches the surface; (2) melting proceeds only to the base of the crust. The Mantle potential temperature, T_p , is anchored at the early Archaean (3.8 Ga) (melting temperature of komatiite), an assumed temperature at the early Proterozoic (2.5 Ga) based on a high rate of cooling during the Archaean, and slow thereafter until the present.

general cannot be eroded before it is recycled in the mantle again. In the following, this condition will be simulated by replacing the basaltic crust by a lower density continental crust that emerges above sea level and is subject to erosion.

3. Cooling of the earth, falling sea level, and continental growth

At a higher mantle potential temperature, T_p , partial melting at the solidus starts at a deeper level. The thicknesses of the segregated and residual layers increase strongly with increasing T_p . In the 1-D model used in this paper, the degree of depletion of the residual harzburgite layer increases continuously from zero at the solidus to a maximum value at the base of the crust. This causes the composition of the harzburgitic root to decrease continuously relative to the underlying undepleted peridotite mantle and therefore compositional density too. Following Jordan (1997), his normalized density as a function of depletion is used in the present study. The densities of the basalt and the harzburgite are smaller than that of unde-

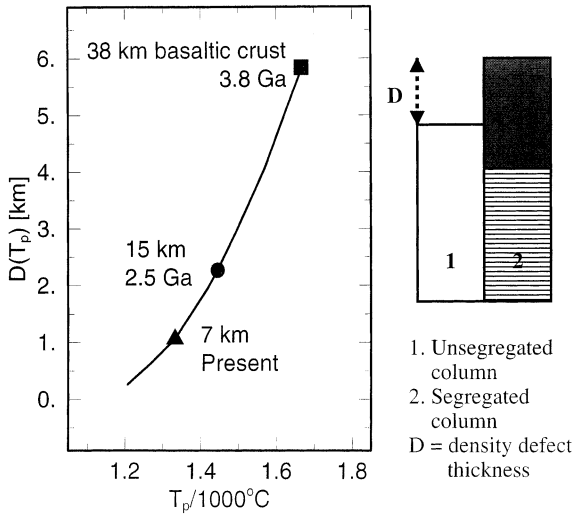


Fig. 2. The density of a segregated column is lower than that of a column of undepleted peridotite of equal mass and cross-section from which it has been derived. The lengthening of the column, the density defect thickness $D(T_p)$, is a strongly increasing function of mantle potential temperature. Segregation does not change the pressure at the base of the column.

pleted peridotite. Upon segregating an undepleted column, its length becomes larger (Fig. 2). This lengthening is termed density defect thickness (Oxburgh and Parmentier, 1977) and is obtained by integration of the density defect (Vlaar and Van den Berg, 1991). The density defect thickness, D , also increases strongly as a function of T_p (Fig. 2). In Fig. 3, the states of segregation at different mantle potential temperatures are compared ($T_{p1} > T_p$). As reference depths, the depth to the solidus at T_{p1} , $Z_{sol}(T_{p1})$, and the constant depth to the top of the non-segregated column, Z_{undepl} , are taken. The length of the undepleted column [$Z_{sol}(T_{p1}) - Z_{undepl}$] is constant for all T_p . When comparing at the temperature, T_p , the states of segregation at T_{p1} and T_p ($T_p < T_{p1}$), the effect of the fall in T_p on the shrinking of the earth has to be considered. As at $T_p = T_{p2}$, this effect is the same for all columns with $T_p > T_{p2}$, it can be neglected.

For $T_p < T_{p1}$, only part of the undepleted column [$Z_{sol}(T_{p1}) - Z_{undepl}$] above the solidus $Z_{sol}(T_p)$ is subject to segregation. For all $T_p < T_{p1}$, this results in a lengthening of the total undepleted column by the amount of lengthening of only the

segregated part. This is equal to $D(T_p)$, the density defect thickness at T_p . The ocean bottom for $T_p < T_{p1}$ is falling, therefore, as $D(T_p)$. Taking also oceanic water depth to be a function of T_p , $d(T_p)$, the falling ocean bottom results in a fall of sea level relative to a continent stabilized at T_{p1} (or to the earth's centre) by:

$$D(T_{p1} - D(T_p) + d(T_{p1}) - d(T_p)) = D_1 - D + d_1 - d \quad (1)$$

where $D_1 = D(T_{p1})$, $D = D(T_p)$, $d_1 = d(T_{p1})$ and $d = d(T_p)$.

The secular change in depth of the ocean depends on the change in the ratio of continental versus oceanic area, thus on a lateral continental growth, or on the change of ocean volume. As only a secular change of sea level is considered here, the effect on sea level due to short-term geodynamic processes like present-day plate tectonics is neglected. In the following calculations, ocean depth is taken either to be constant or to increase by 1 km during the earth's evolution. The latter case results in later emergence of a once submerged crust and is compatible with lateral growth of the continent. Still, later emergence results if an additional increase in ocean volume had taken place. For the sake of simplicity, it is assumed here that the earth's volatile content had been outgassed prior to 3.8 Ga and that therefore the ocean water volume can be taken to be constant in time. The depth of the ocean then is mainly determined by the area of the ocean.

Lateral growth of the ancient continent and consequent decrease of the ocean area therefore results in deepening of the ocean with time. As the evolution of continental growth and the shrinking of the ocean area is not known in detail, only schematic values for ocean depth are used.

Taking $d_1 = d$, in Eq. (1), hence constant ocean depth over time for $T_p < T_{p1}$, Eq. (1) reduces to $(D_1 - D)$. The fall in sea level is then equal to the fall in density defect thickness $D(T_p)$.

For $T_{p1} = 1667^\circ\text{C}$ at 3.8 Ga, and $T_p = 1332^\circ\text{C}$ at present, Fig. 2 gives a sea-level fall of $(5.8 - 1.1)$ km = 4.7 km over this period. Throughout the Archaean, when T_p is assumed to drop from 1667 to 1445°C , the sea-level fall is 3.5 km. For the

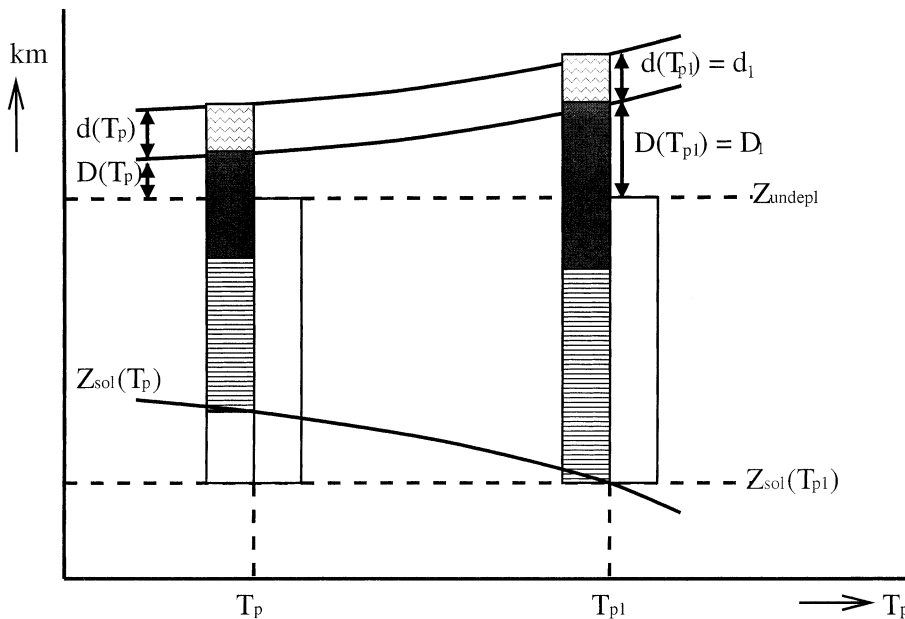


Fig. 3. (Not to scale) Depth to the solidus $Z_{sol}(T_p)$, density defect thickness $D(T_p)$, and ocean depth $d(T_p)$. The state at T_{p1} is taken as the state of reference for $T_p < T_{p1}$. $[Z_{undepl} - Z_{sol}(T_{p1})]$ is the length of the column before segregation, subject to a decreasing degree of depletion for $T_p < T_{p1}$. After segregation, the pressure at depth $Z_{sol}(T_{p1})$ for $T_p < T_{p1}$ is constant and can be taken as the depth of isostatic compensation. The length of the segregated column above the constant reference depth, Z_{undepl} , including the ocean depth is $D(T_p) + d(T_p)$. The small isostatic effect of variable ocean depth is not taken into account. The situation can be viewed as one in which columns of variable length with their base on the solidus and their top at $D(T_p)$ at variable T_p are floating in a sea of undepleted mantle.

post-Archaean until the present, the sea-level fall is 1.2 km. When the schematic ocean depth is increased by 1 km at the Archaean–Proterozoic boundary, the sea-level fall for the post-Archaean is decreased to 0.2 km over this period. When considering matters related to continental freeboard in connection with continental growth, a post-Archaean (gradual) increase in sea level due to shrinking of the oceanic area, even on a cooling earth, should be taken into account.

In the present paper, it is assumed that oceanic lithosphere is continuously recycled through the upper mantle. Cratons and their mantle roots, which have been stabilized early in the continental evolution, have not been destroyed and recycled. Their stabilization can be shown to be due to thermal blanketing by the low-density continental crust laying on top of the mantle root (de Smet et al., 1999). As the oceanic lithosphere exhibits a strongly increased gravitational stability with increasing T_p (Vlaar, 1986b), its stability can

become permanent when it is covered by a thick layer of sediments causing thermal blanketing. I suggest that a substantial contribution to continental growth of volume and areal extent is due to stabilization of oceanic crust, which has been subject to thermal blanketing by sediments in epicontinental basins. In this case, fractionation of a new continental crust from the mantle is effected by incorporating oceanic crust into the continent.

A sedimentary basin formed in this manner is subaqueous at the time of its formation and stabilization. A sea-level fall may cause the newly formed addition to the continent to emerge above sea level at some time later during the geological history.

As an example, a sedimentary basin formed at 2.5 Ga is considered. If its surface was at a depth of h km below sea surface, its emergence above sea level takes place [Eq. (1)] when $D_1 - D = h$, if the ocean depth, d , does not change in time, or $D_1 - D = h + 1$ when the ocean depth increases by 1 km. Taking $T_{p1} = 1445^\circ\text{C}$, $D_1 = 2.3$ km at 2.5 Ga,

and $h=2$ km, this geological structure would emerge at $D=1.9$ or 0.9 km, depending on the constant or increasing ocean depth. This means that emergence took place at some point during the first half of the Proterozoic or that the basin would still be submerged at present. In the present model, a late Precambrian emergence of late Archaean and early or later Proterozoic platforms cannot therefore be excluded (Hargraves, 1976) unless T_{p1} at 2.5 Ga has been strongly overestimated. In this case, it must be assumed that since 2.5 Ga, the mantle has cooled considerably less than is indicated in the present model. It has been shown by Vlaar (1986b) that even a slight increase in mantle potential temperature results in a strongly increased stability of the oceanic lithospheric layering. An 50 K increase gives an increase in the age of transition from stable to unstable oceanic lithosphere from the present 30 Ma to about double this value. Therefore, the possibility of creating stable epicontinental basins, even at moderately elevated mantle temperatures compared to the present, is plausible. Even at present, this may be the case in large deltaic sediment fans.

4. Selective exhumation of the continental crust

The existence of the deep depleted harzburgite mantle root of the Archaean craton must be the result of extensive depletion and the segregation of a thick (basaltic) crust. It, appears, therefore to be a paradox that the cratonic crust is continental. Consequently, it must be inferred that the present cratonic crust is allochthonous with respect to the mantle root. This felsic continental crust therefore must have replaced an earlier oceanic mafic crust, and its function has been to stabilize the cratonic structure in its history by thermal blanketing and its low density (de Smet et al., 1999). Lateral translation of floating proto-continental fragments must have been facilitated by the low strength of the lower part of a thick basaltic crust (Hoffman and Ranalli, 1988).

Early Archaean growth and addition of continental rocks to the crust have probably occurred at a slow rate (Taylor and McLennan, 1985;

Rudnick, 1995). Small proto-continental crustal fragments must, therefore, have been floating in a thick basaltic layer. Isostatic conditions were determined by the small density contrast between felsic and mafic rocks. Vlaar (1986a) has shown that under these conditions, protracted loading of the continental surface with felsic rocks or their detritus results in protracted burial of continental crustal material. The small density difference and resulting isostatic conditions favour replacing the basaltic layer by felsic rocks under subaqueous conditions. The early continental proto-crust therefore must have remained subaqueous until it emerged above sea level due to crustal growth and to the sea-level fall described in this paper.

Sediments in the oldest terrains, however, are evidence that small tonalite islands must have been subaerial to provide sediments for high- and low-grade terrains.

The Archaean is characterized by the existence of high-grade gneissic terrains consisting of rocks of the T.T.G. suite, low-grade granite-greenstone belts, and late Archaean cratonic platforms.

At present, the high-grade terrains are outcropping at the surface, and exhibit deep crustal metamorphism. They appear to have been buried to lower crustal depths (20–40 km) in the earlier part of the Archaean and subsequently eroded, resulting in crustal exhumation subsequently, and appear to have stabilized in the late Archaean. This process can be understood in terms of a protracted increase in emerging continental mass above sea level and its denudation to deep crustal depths. In the present paper, this is attributed to a rapid fall in sea level due to strong cooling of the earth during the Archaean.

Granite-greenstone belts are assumed here to be basin-like structures with alternating layering of mainly basaltic magmatic outflows and volcanoclastic sediments. The sediments display a deep-to shallow-water deposition. This layering is intruded by later granitic intrusive plutons. The greenstone belts are emplaced on the continental basement and can probably be seen as ancient analogues of modern graben-like basins (Hamilton, 1998). In the present paper, they are supposed to be generated by mantle plumes impinging from below on an already existing sub-

aqueous continental (proto)crust. This gives rise to magmatism leading to volcanic outflows and lower crustal anatexis in order to generate granitic plutons. The low-pressure metamorphic grade of granite greenstone belts ranges from greenschist to amphibolite facies. This implies that they have been buried to pressures of up to 2–3 kb or depths of up to 8–12 km after their formation under shallow- to deep-water conditions (Condie, 1989; Galer and Mezger, 1998). As, in the present paper, they are assumed to be caused by mantle thermal events, thermal relaxation after their formation is assumed to have caused subsidence to deeper water levels and coverage by sedimentary layers that have now disappeared.

Adjacent high- and low-grade terrains have undergone considerably different exhumation histories indicating large relative vertical displacements prior to their cratonic stabilization in the later Archaean. In the following, it will be shown

that selective exhumation can be explained by protracted subaerial erosion and exhumation of the high-grade terrains, and a subaqueous position of the granite greenstone belts until the late Archaean.

In order to simulate erosion and exhumation of the high-grade terrains, the basaltic oceanic crust is replaced by a continental crust of lower density in order to create a positive continental freeboard. For the sake of simplicity, the ocean depth is kept constant in time at 3 km.

The increase in ocean depth over time is not expected to result in largely different conclusions as exhumation is mainly governed by the buoyant harzburgite residual layer.

At T_{p1} , the oceanic basaltic crust and water layer are replaced by a lower density continental crust, resulting in a continental freeboard, h_1 , without changing the depth of isostatic compensation at $Z_{sol}(T_{p1})$ (Fig. 4).

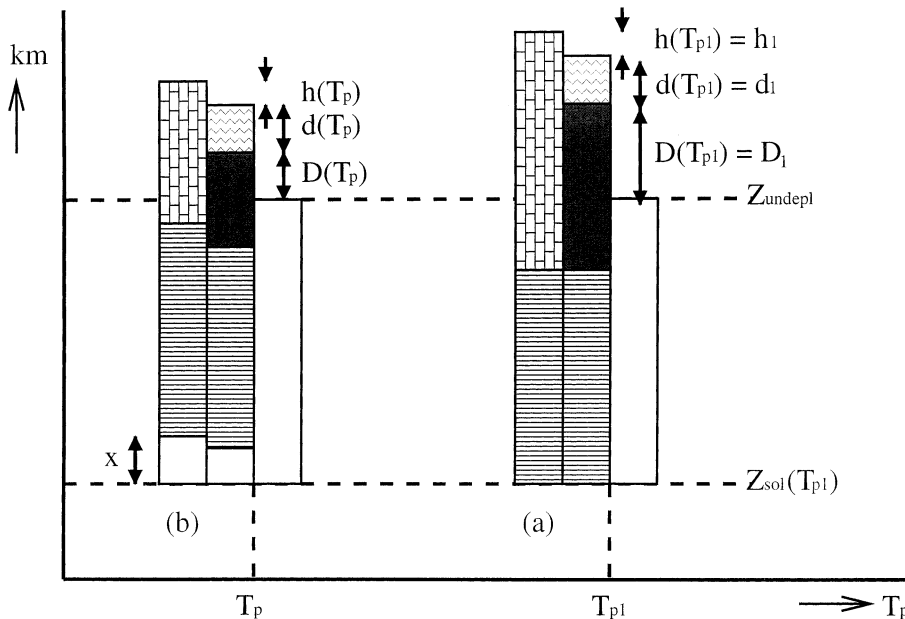


Fig. 4. (Not to scale) At T_{p1} , the oceanic basaltic crust and water layer are replaced by a lower density continental crust, resulting in a continental freeboard, h_1 , without changing the depth of isostatic compensation: $d_1\rho_w + d_{b1}\rho_b = (D_1 + d_1 + h_1)\rho_c$. d_1 and d_{b1} are the layer thicknesses of the ocean and oceanic crust at T_{p1} , and ρ_w , ρ_b , and ρ_c are the densities of water, basalt, and continental rocks, respectively. At $T_p < T_{p1}$, the sea level has fallen by $(D_1 + d_1 - D - d)$. If no erosion takes place, the continental surface would stand at $(D_1 + d_1 + h_1 - D - d)$ above sea level at T_p . If erosion results in a continental freeboard, h , at T_p , isostatic readjustment displaces the continental column upward over the distance, x . The depth of crustal exhumation then is $(D_1 + d_1 + h_1 - D - d - h + x)$ and is compensated at the base by replacing depleted by undepleted peridotite. Isostatic balance is governed by $x = (D_1 - D + d_1 - d + h_1 - h)\rho_c / (\rho_p - \rho_c)$. ρ_p is the density of undepleted peridotite.

Isostasy pertains to the base of the oceanic and the continental crust and requires:

$$d_1\rho_w + d_{b1}\rho_b = (D_1 + d_1 + h_1)\rho_c \quad (2)$$

where d_1 and d_{b1} are the thicknesses of the oceanic layer and the oceanic crust at T_{p1} , and ρ_w , ρ_b , and ρ_c are the densities of water, basalt, and continental rocks, respectively.

Taking $\rho_w = 1 \text{ g cm}^{-3}$, $\rho_b = 3 \text{ g cm}^{-3}$, $d_1 = 3 \text{ km}$, and requiring the continental freeboard to be $h_1 = 0.75 \text{ km}$ at T_{p1} , Eq. (2) gives:

$$3(1 + d_{b1}) = (d_{b1} + 3.75)\rho_c \quad (3)$$

d_{b1} and ρ_c are therefore interdependent due to the assumption that continental and oceanic crusts have their base at the same depth.

For $T_{p1} = 1667^\circ\text{C}$, and $d_{b1} = 38 \text{ km}$ at 3 Ga, the density of continental rocks becomes $\rho_c = 2.8 \text{ g cm}^{-3}$. This value appears to be reasonable for a crust of intermediate composition.

For $T_{p1} = 1445^\circ\text{C}$, and $d_{b1} = 15 \text{ km}$ at 2.5 Ga, $\rho_c = 2.56 \text{ g cm}^{-3}$. Though marginal, this is still an acceptable value for a late Archaean, more felsic crust. Therefore, with acceptable values for continental density, a freeboard of 0.75 km can be obtained, departing from the specific type of isostasy used here. At a lower post-Archaean T_{p1} , however, a continental freeboard of 0.75 km or more can only be realized by assuming crustal thickening subject to Airy isostasy. Airy isostasy is then superposed on the isostatic state presented in this paper. In the following, only the effects of the latter, in particular for the Archaean exposed continental surface, are investigated.

If no erosion had taken place, the continental surface with freeboard h_1 at T_{p1} would stand at $(D_1 + d_1 + h_1 - D - d)$ above sea level at T_p . Taking the constant depth of the ocean with time, $T_{p1} = 1667^\circ\text{C}$, and $h_1 = 0.75 \text{ km}$, in agreement with $\rho_c = 2.8 \text{ g cm}^{-3}$, the early Archaean exposed continent would stand at $(D_1 + h_1 - D)$ above ambient sea level at T_p , 4.25/5.45 km above sea level at 2.5 Ga/present day.

As erosion did in fact take place, the continental surface at T_{p1} with freeboard, h_1 , is assumed to have been eroded to h above sea level at T_p . Erosion and exhumation proceed from T_{p1} to T_p under conditions of isostatic balance in a cooling and changing mantle. The base of the column

formed at T_{p1} thereby is shifted upward over the distance, x . Isostatic balance at T_p requires:

$$x = (D_1 - D + d_1 - d + h_1 - h)\rho_c / (\rho_p - \rho_c) \quad (4)$$

where ρ_p is the density of undepleted mantle rock.

Erosion then has resulted in a crustal exhumation of the amount:

$$x + (D_1 - D + d_1 - d + h_1 - h). \quad (5)$$

For the entire Archaean period (3.8–2.5 Ga), constant ocean depth, ($d_1 = d$), and erosion to sea level ($h = 0$), by substituting $D_1 - D = 3.5 \text{ km}$, $h_1 = 0.75 \text{ km}$, $\rho_c = 2.8 \text{ g cm}^{-3}$, and $\rho_p = 3.33 \text{ g cm}^{-3}$ in Eqs. (4) and (5), this results in $x = 22.88 \text{ km}$, and a crustal exhumation of 27.13 km. If erosion of the 3.8 Ga continental surface had proceeded to the present sea level, the crustal exhumation would be 34.24 km. These values are consistent with the observed erosion of the early Archaean high-grade terrains to a lower crustal depth.

In order to explain the large relative vertical displacement of high-grade terrains and low-grade granite-greenstone belts, in the present model, greenstone belts must have become aurally exposed only later in the geological history and should not have been subject to the large vertical displacements that high-grade terrains underwent. These conditions can only be met when, after their formation, greenstone belts that were formed under shallow to deep water conditions had foundered to a deeper, but low-grade, crustal depth below the sea surface. The greenschist to amphibolite metamorphic grade indicates shallow crustal burial. One way to achieve this is by thermal relaxation of the lithosphere after formation of the greenstone belts by a thermal anomaly underneath the crust.

It is now assumed that at 3.8 Ga, a greenstone belt existed side by side to a future high-grade terrain that had not yet been exhumed. The high-grade terrain had a positive freeboard, and the greenstone belt was buried at a low metamorphic grade in a crust whose top was 3 km below sea level.

Therefore, from the above, a scenario can be imagined whereby, at 2.5 Ga, the greenstone belt would just have emerged above sea level, and the high-grade terrain would have been exhumed to a

crustal depth of 27 km. Selective exhumation of high-grade terrain probably took place mostly during the early Archaean when the earth's cooling was most rapid. A secular sea-level fall, however, also continued thereafter, and protracted erosion of the stabilized craton proceeded. This gave rise to the formation of early Archaean cratonic platforms and early Proterozoic epicontinental sedimentary basins.

Exhumation causes thinning of the continental crust. A 3.8 Ga crust of 41.75 km is reduced to about 15 km at 2.5 Ga. Therefore, 26.75 km of crust have been eroded away.

The present Archaean crustal thickness is about 40 km, and this had been in place when the craton had stabilized in the late Archaean. In order to compensate the loss of the upper crust caused by its exhumation, new mantle-derived mafic material must have been added to the crust before its stabilization. Exhumation and isostatic readjustment as such cause decompression and renewed melting in the harzburgite mantle root. If this additional decompression follows shortly after isentropic segregation of the crust from the mantle, the mantle root can be regarded to generate melt under the same isentropic conditions. The generated volume of extra melt is an upper bound. If decompression takes place at a later time, the amount of melt is less because of conductive cooling of the shallower mantle layers. A rough estimate of the maximum amount of additional melt can be made from Fig. 1. The difference between the upper and lower curves is the additional melt generated when all pre-existing crust is removed. This volume of melt is roughly equal to the crustal volume. For example, at 3.8 Ga, some 25 km of newly added crust have to be accounted for. This is approximately half the difference between curves (a) and (b). This implies that renewed isentropic decompression of the mantle root can easily satisfy the required crustal growth. After conductive cooling of the mantle root, however, additional melt generation takes place at a deeper level in the harzburgite root. The amount of melt generated is then smaller. However, it has been demonstrated by de Smet et al. (1999) that convection in the hot Archaean upper mantle also leads to protracted melt generation. It could be

assumed, therefore, that sufficient additional melt can be generated to make up for the eroded crust.

5. Discussion and conclusions

Primary basaltic crust is assumed to have been generated by supersolidus isentropic decompression melting in a convective shallow upper mantle. When a low-density, more felsic, continental secondary crust replaces the basaltic crust, it advances stabilization of the underlying harzburgite root by thermal blanketing (de Smet et al., 1999). The oceanic basaltic crust and its harzburgite residue, not being shielded in this way, are assumed to be continuously generated and destroyed.

On a secularly cooling earth, the thickness of the oceanic compositional lithospheric layering decreases as the mantle temperature falls. This causes changing isostatic conditions and a considerable decrease in sea level. The Archaean craton, however, has escaped destruction, and has been stabilized during the later Archaean and early Proterozoic. The regularly falling sea level caused parts of the stabilized continent to emerge above water. For the Archaean craton, this took place mostly during the later or latest Archaean, whereas this may have occurred in early Proterozoic basins during the later Precambrian. Erosion of the exposed continental surface leads to its exhumation by isostatic readjustment. The particular circumstances that determine isostasy on a cooling earth — the decreasing thickness of the oceanic compositional layering, secular deepening of the ocean floor and a corresponding fall in sea level, and also the substantial buoyancy of the low-density harzburgite root — favour exhumation of the Archaean high-grade terrain to deep crustal levels.

However, the model presented in this paper requires also that granite-greenstone belts, which have experienced low-grade metamorphism only, should have been buried at a shallow depth in a crust that did not emerge above sea level until the late stages of high-grade terrain exhumation. The large relative vertical displacements between high- and low-grade terrains must have been settled before the final stabilization of the craton.

Buick et al. (1995) observed that greenstone belts can be buried to deeper crustal levels. They reported a 3.5 Ga old greenstone belt that had been formed on top of an existing older basement of greenstone belt. Their inference was that this structure emerged repeatedly around 3.5 Ga. The present model, however, requires that shortly after its emplacement, the structure must have foundered to larger crustal depths and remained subaqueous until their late Archaean emergence.

The model relates the geological history to the falling temperature of the mantle. As such, it can be considered to be robust under the suppositions stated. Mantle temperature is secularly falling. As the decrease in mantle temperature, however, is dependent on the geodynamic mode of cooling, the cooling history of the earth as a function of time is strongly dependent on the cooling mechanism. As the geodynamic processes by which the earth cools are not known for the entire Precambrian, a time record for secular sea-level fall and its consequences are not available. Cooling of the earth could have been episodic and concentrated in shorter periods between longer periods of stagnancy and stabilization. Therefore, the cooling of the earth as a function of time, and therefore its effect on sea level and selective exhumation, cannot be directly related to the smooth fall of T_p used in this investigation. In particular, the temperature at 2.5 Ga as used here may have been over- or underestimated. This can affect considerations concerning post-Archaean continental emergence and freeboard and growth, and should be taken into account in future research.

Short-term non-secular events comparable to the modern plate tectonic cycle, orogenic processes, magmatic additions to the crust by plume activity, and associated crustal thickening together with Airy isostatic readjustment and resulting emergence have not been taken into consideration. Only the dominant effects of the earth's secular cooling on crustal segregation, sea-level change and their consequences have been treated here.

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