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On the formation of continental silicic melts in thermochemical mantle convection models: implications for early Earth

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Abstract

Important constituents of Archean cratons, formed in the early and hot history of the Earth, are Tonalite–Trondhjemite–Granodiorite (TTG) plutons and greenstone belts. The formation of these granite–greenstone terrains is often ascribed to plate-tectonic processes. Buoyancy considerations, however, do not allow plate tectonics to take place in a significantly hotter Earth. We therefore propose an alternative mechanism for the coeval and proximate production of TTG plutons and greenstone-like crustal successions. That is, when a locally anomalously thick basaltic crust has been produced by continued addition of extrusive or intrusive basalts due to partial melting of the underlying convecting mantle, the transition of a sufficient amount of basalt in the lower crust to eclogite may trigger a resurfacing event, in which a complete crustal section of over 1000 km long sinks into the mantle in less than 2 million years. Pressure release partial melting in the complementary upwelling mantle produces large volumes of basaltic material replacing the original crust. Partial melting at the base of this newly produced crust may generate felsic melts which are added as intrusives and/or extrusives to the generally mafic crustal succession, adding to what resembles a greenstone belt. Partial melting of metabasalt in the sinking crustal section produces a significant volume of TTG melt which is added to the crust directly above the location of 'subduction', presumably in the form of a pluton. This scenario is self-consistently produced by numerical thermochemical mantle convection models, presented in this paper, including partial melting of mantle peridotite and crustal (meta)basalt. The metamorphic p, T conditions under which partial melting of metabasalt takes place in this scenario are consistent with geochemical trace element data for TTGs, which indicate melting under amphibolite rather than eclogite facies. Other geodynamical settings which we have also investigated, including partial melting in small scale delaminations of the lower crust, at the base of a anomalously thick crust and due to the influx of a lower mantle diapir fail to reproduce this behavior unequivocally and mostly show melting of metabasalt in the eclogite stability field instead.

Keywords: Mantle convection; Numerical modelling; Archean; TTG production; Granite–greenstone terrains

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1. Introduction

The finding of detrital zircons up to 4.4 Gy old, less than 200 million years after the accretion of the Earth, provides evidence that continental crust was already produced at this time (Peck et al., 2001). Growth curves for the continental crust, derived from isotopic analyses of rocks from Archean to present age, also indicate that the formation of continental crust may have started before 4.0 Ga (e.g., Taylor and McLennan, 1985; McCulloch and Bennett, 1994). In the recent Earth, most continental growth is accommodated by the accretion of volcanic arcs, which are formed by partial melting of the mantle wedge above subducting slabs, and intraplate volcanism, associated with hotspots. The material thus produced is on average of basaltic composition, but the continental crust has an andesitic bulk composition (Rudnick, 1995). In Archean cratons, a large fraction of the material consists of Tonalite–Trondhjemite–Granodiorite (TTG) suites, either metamorphosed to amphibolite facies gneiss–migmatites, or as plutons (Goodwin, 1991). Melting experiments indicate that these TTGs may have been formed by partial melting of hydrous metabasalts (eclogites or amphibolites) at pressures between 8 and 32 kbar (optimally 22) (Rapp et al., 1991). The presence of some water is required to lower the melting temperature to plausible values. The required conditions for this process may be found in subduction zone settings, but not exclusively (Rapp et al., 1991). Recent results from Foley et al. (2002, 2003) have shown, on the basis of Nb/Ta and Zr/Sm ratios in TTGs, that for the formation of these rocks, melting must take place in the amphibolite rather than the eclogite stability field.

Their experiments show that komatiitic material, which is more mafic than the present basaltic oceanic crust, and which may be expected to be more abundant for higher mantle temperatures (Nisbet, 1982), would transform into pyroxenite instead of eclogite at high pressure, which they infer to produce basaltic melts upon partial melting, and not TTG-like melts.

Recently, however, Rapp et al. (2003) have questioned the conclusion of Foley et al. (2002) that TTGs with the correct trace element characteristics are not formed by partial melting of eclogites on the basis of their own experimental results. Rapp et al. (2003) question the validity of the partition coefficient values used by Foley et al. (2002) and state that eclogite partial melting better reproduces TTG major element compositions than melting in the amphibolite melting. However, generally the results of melting experiments aimed at reproducing TTG's show discrepancies in Al_2O_3, TiO_2, and Mg# (Rudnick, 1995). Presently, our knowledge on this subject has not advanced far enough to disqualify either hypothesis (Matthias Barth, personal communication). In this work, we will consider the implications of the hypothesis of Foley et al. (2002, 2003).

Eclogitic xenoliths from cratonic regions often show a depleted nature, indicating that melt has been extracted, and it has been suggested that these are remnants of granitoid formation, complementary to these crustal rocks (e.g., Rollinson, 1997; Barth et al., 2001). Furthermore, oxygen isotope studies on coesite inclusions in diamonds from eclogite xenoliths from the Guyana Shield suggest interaction of the original rock with sea water similar to modern day hydrothermal alteration of oceanic crust at midocean ridges (Schulze et al., 2003). However, theoretical considerations and numerical models have shown that maintaining plate tectonics in a hotter Earth is quite difficult (Vlaar and Van den Berg, 1991; Van Hunen, 2001; Van Thienen et al., 2004b). Adakites, formed by partial melting of a young, hot slab in a subduction zone setting, are sometimes considered as a modern day analogue of TTGs (Martin, 1999), and interpreted as supporting early plate tectonics. Smithies (2000), however, showed that adakites and TTG's differ in Mg number and SiO_2 contents, indicating that adakitic melts underwent interaction with mantle peridotite whereas TTG melts (especially pre-3.0 Ga) did not. A more suitable setting for producing TTGs than a subduction zone is therefore melting at the base of a thickened crust (Smithies, 2000). The composition of Phanerozoic Na-rich granitoids formed in this way (e.g., Atherton and Petford, 1993; Johnson et al., 1997) more closely matches the range of composition of Archean TTGs (Smithies, 2000). Alternatively, the intrusion and crystallization of mantle melts in the lower crust may provide a heat source for partial melting of the lower crust (Huppert and Sparks, 1988; Petford and Gallagher, 2001). Other mechanisms proposed in the literature use similar settings. Campbell and Hill (1988) suggested that the heat of a hot asthenospheric upwelling interacting with an existing crust may generate felsic melts, thus producing continental material. Theoretical considerations suggest that both the maximum size and the excess temperature of mantle plumes become smaller in a hotter mantle (McKenzie and Bickle, 1988; Nisbet et al., 1993). However, a mantle plume originating from a separately convecting lower mantle entering the upper mantle may have a much greater size and higher excess temperature (Van Thienen et al., 2004a), possibly supplying sufficient heat for partial melting of the lower crust. Zegers and Van Keken (2001) proposed that delamination of an eclogitic lower crust and associated extension of the overlying middle and upper crust may allow warmer mantle material to rise to shallower levels and cause crustal melting to produce continental material. They used a numerical model to demonstrate the feasibility of the mechanism and compared results to geological evidence from the Pilbara craton.

In this work, we use numerical thermochemical mantle convection models to investigate different geodynamical settings in which partial melting of (meta)basalt may take place. This is an extension of earlier work (Van Thienen et al., 2004a), in which the dynamics of the self-consistent growth and recycling of thick Archean oceanic crust was investigated. Four different scenarios in which partial melting of (meta)basalt may take place are investigated in detail: (1) small-scale delamination of the eclogitic lower crust, (2) large-scale resurfacing by rapid episodic 'subduction' (though not in a plate tectonic sense), (3) heating of the lower part of a thickening crust by the convecting mantle, and (4) heating of the crust from below by a mantle plume. We compare the melting conditions in our models to those inferred for natural TTGs (Winther, 1996), and apply the geochemical constraints which were found for the metamorphic stability field where TTG generation takes place (Foley et al., 2002, 2003). We also investigate the production rates of continental material which is formed by means of this mechanism using numerical models. Finally, we compare the rock associations produced in our models and their geometry to the granite–greenstone associations which are common in Archean terrains and come up with a new model for the production of these associations which does not require the operation of plate tectonics.

2. Model description

The numerical model used is nearly identical to that of Van Thienen et al. (2004a). Therefore, the reader is referred to Van Thienen et al. (2004a) for a complete model description. A short overview of the model is presented below. A single feature has been added to the model: partial melting of basaltic material. This will be described in Section 2.2.

2.1. Convection model

The thermochemical convection model used in this study is based on conservation of energy, mass and momentum under the extended Boussinesq approximation (Steinbach et al., 1989; De Smet et al., 1998) for an infinite Prandtl number medium. Internal heating due to decay of radioactive elements is included and is a function of both local trace element concentration (trace elements are fractionated during partial melting) and time (half-life is 2.5 Gy). Partial melting is included in the model, controlled by the parameterization of the melting phase diagram under consideration (De Smet et al., 1998).

Two deformation mechanisms, which are diffusion creep and nonlinear dislocation creep, account for the creep flow deformation of material. The parameters of the rheology model are composition dependent, and the effect of dehydration during partial melting (Hirth and Kohlstedt, 1996) is included as well. Brittle failure is approximated by a stress limiter mechanism.

All motion is driven by density perturbations caused by (a) temperature variations, (b) compositional variations, and (c) phase variations.

Compositional variations are a result of mantle differentiation by partial melting, see below. Phase transitions around 400- and 670-km depth are included in the model, with composition dependent parameters (see Van Thienen et al., 2004a). Additionally, basalt is transformed into heavier eclogite upon reaching depths in excess of 30 km in our model (for a discussion on the choice of parameter values, see Van Thienen et al., 2004a). The kinetics of this transition are approximated assuming a constant relaxation time for the transition of 1.25 My.

2.2. Partial melting

Partial melting of fertile mantle peridotite may produce a spectrum of melt compositions, ranging from tholeiitic to komatiitic. In general, higher melting pressures and/or greater degrees of melting meaning higher temperatures produce melts which are more MgO-rich, so more to the komatiitic side of the spectrum (Philpotts, 1990). Partial melting of basaltic material may also produce a range of melt compositions, depending on the conditions of melting. This melting behavior is simplified in our models, where the composition field defines three different basic types of materials: mantle peridotite, basalt and felsic material. These terms are used here to describe larger groups of rocks than their specific geological definitions. *Basaltic* materials are understood here to include all melt products of mantle peridotite, regardless of melting conditions, and the term *felsic* materials is used to indicate all melt products of basaltic material. Mantle peridotite and basalt have an extra component to their composition, which is the *degree of depletion*. Partial melting is modeled as an (irreversible) increase in this degree of depletion F, which is defined here as the mass fraction of melt produced from an initially unmelted material control volume of mantle peridotite or basalt. Since no material is actually removed in our model, the volumes of depleted residual material are somewhat overestimated (De Smet, 1999). We use third order polynomial parameterizations of the solidus and the liquidus of mantle peridotite, based on Herzberg and Zhang (1996), down to a depth of 400 km (see Fig. 1), assuming that melt produced below this depth is not segregated. Our isobaric melting curve, which is based on data presented by Jaques and Green (1980), is linear (see De Smet et al., 1998). A similar approach is used for partial melting of basaltic material. Since in our models all melts produced are assumed to migrate to the surface (see Section 2.3), we can also assume that the crustal

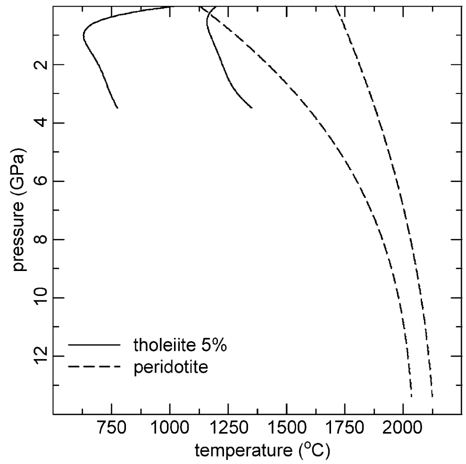


Fig. 1. Solidus and liquidus parameterizations used in the models for basalt/eclogite (solid curves) and peridotite (dashed curves). The basaltic curves are based on experimental results from a hydrated (5% water by weight) tholeiite (Green, 1982). The peridotite curves are based on Herzberg and Zhang (1996).

material produced by solidification at the surface becomes hydrated by contact with surface water. This is supported by isotope studies on coesite inclusions in diamonds in eclogite xenoliths from the Guyana Shield (Schulze et al., 2003), which indicate significant interaction with water. We therefore use a hydrous solidus and liquidus in our model experiments, based on data from Green (1982) (see Fig. 1). We use the same linear melting function for the partial melting of basalt as that used for peridotite.

2.3. Crustal growth

An important feature of the model is the self-consistent growth of crust. All melt produced in the model domain is assumed to migrate to the surface on time scales much shorter than the characteristic time scales of the model. We therefore let all melt which is produced flow into the model through the top boundary above the region where the melt was formed. When both basaltic and felsic crust is being produced, the inflowing material is a mixture of these two components, defined by different tracer composition values, in relative proportions corresponding to the respective amounts of melt being produced. The trace element concentrations of inflowing basalt and felsics are a function of the fractionation during partial melting of the respective source rocks.

2.4. Computational methods

The model domain of each experiment is square, with a size of 600 or 1200 km (see Table 1). This corresponds to either a quarter of the domain or the complete domain of the model which was used as the starting point of the present experiments (see Section 2.5), depending on the length scale of the process under consideration in the respective experiments. It is discretized using (a) finite elements with a boundary nodal point spacing of 6–12 km for the energy and momentum equations, and (b) 200 000 to 400 000 particle tracers, advected with the mantle flow, on which the development of depletion, trace element concentration, and metamorphic phase of basaltic material are evaluated. The vertical boundaries are periodic, and the bottom boundary has a free slip condition. On the top boundary, the vertical velocity is prescribed as a function of melt productivity, as will be described below, to allow the production of crust at the surface. The horizontal, tangential stress component is set to zero. The temperature on the top boundary is set to 0 °C. On the bottom boundary, a temperature consistent with the initial condition (see Section 2.5) is prescribed.

2.5. Initial conditions

The initial conditions for the present experiments are derived from model Ms in Van Thienen et al. (2004a). This original model was started with a cool (subsolidus) mantle with an extreme heating rate of 250×10^{-12} W kg^{-1}, in order to produce a hot and differentiated initial state for our experiments. This extreme internal heating value was reduced to a normal early Earth value of 15×10^{-12} W kg^{-1} after 30 km of basaltic crust had been produced, defined as t=0. From this situation, the model evolution was computed for several hundred million years. In model Ms of Van Thienen et al. (2004a), as well as in a range of similar models, each of the types of dynamics we investigate in this work is self-consistently produced during this latter several hundred million year long period. In the present experiments, we use snapshots of the entire domain (1200×1200 km) or part of the domain (600×600 km) at times when the process we would like to study is about to take place as starting conditions (see Table 1). Due to the prescribed periodic boundary condition on the vertical boundaries, the latter case results in inconsistencies, since the new boundaries were previously unconnected and therefore did not have the same temperature and velocities. However, these are small and do not significantly influence the solution. In contrast with the present experiments, the original model did not include partial melting of basalt. Therefore, the present experiments show an initial pulse of melting of basaltic material where the temperature is above the basalt solidus. In one experiment (Mctr, see Table 1), the generation of a lower mantle diapir, as it is about to take place in model Ms of Van Thienen et al. (2004a), is suppressed by increasing the density contrast over the olivine–perovskite phase transition. In every other respect, this model is the same as model Mdia.

Table 1
The range of models focussing on different processes and/or settings, with computational domain size and figure reference visualizing the process or setting

Model	Process/setting	Domain size (km)	Modification	Figure
Mdel	small-scale delamination	600×600	–	3a
Mres	resurfacing	600×600	–	3b and c
Mtcr	thickened crust	1200×1200	lower mantle diapirism suppressed	3d
Mdia	lower mantle diapirism	1200×1200	–	3e

Modifications of the model conditions relative to the original model Ms from Van Thienen et al. (2004a) (see text) are indicated as well.

3. Results

In this work, we investigate the partial melting of (meta)basalt in five different geodynamical settings, associated with different processes which may have been important in the production and recycling of oceanic crust in the early Earth, and which have been described in more detail by Van Thienen et al. (2004a). Before presenting the actual model results, we will first discuss the metamorphic phase diagrams which will aid the interpretation of the results.

3.1. Metamorphic phase stability fields

Trace element ratios (Nb/Ta and Zr/Sm) in TTGs indicate that these rocks have been formed by partial melting of a metabasalt in the amphibolite rather than the eclogite stability field (Foley et al., 2002, 2003). This is based on the modeling of trace element distribution, which indicates that amphibole must be present in the source rock and rutile, common in Archean eclogite xenoliths (Rudnick et al., 2000), must be absent. It is therefore important to monitor the pressure and temperature conditions under which partial melting takes place in our experiments. Fig. 2 shows the metamorphic phase diagrams for three different compositions. We have included three different compositions because of the uncertain nature and variability of the dominant rock type in early oceanic crust. Basaltic crust created at a midocean ridge will be thicker for higher temperatures (Vlaar and Van den Berg, 1991), and have a higher MgO content (Nisbet, 1982) than present MORB-like compositions. However, in an alternative geodynamic regime like that which is modeled in this work, production of crust does not take place due to partial melting of focused upwelling mantle in a ridge setting, but by stacking of extrusive units produced at many locations from partial melting of the convecting mantle. Due to the presence of preexisting crust and a depleted root, partial melting takes place at greater depths, which tends to increase the MgO content (Philpotts, 1990), and to lower degrees of melting, which counteracts this effect. Therefore, the dominant composition of crust produced by partial melting of mantle peridotite is expected to be somewhere in the wide range between a basaltic and a komatiitic composition, and we will include all three phase diagrams of Fig. 2 in the discussion of the results of the numerical models.

Fig. 2a shows the diagram for a MORB composition (simplified from Hacker et al., 2003). The results of Foley et al. (2002) show that partial melting of metabasalts in the (garnet) amphibolite stability field produces TTG melts. For a MORB composition, this is at depths less than 50 km at temperatures between approximately 450 and 800 °C.

For a gabbroic composition (Fig. 2b, from a diagram of Foley et al., 2003 based on data from Green and Ringwood, 1967), the amphibolite stability field, and therefore the TTG production field, is larger, extending up to temperatures of about 1000 °C and depths of about 75 km (note that the dashed lines are extrapolations of the phase boundaries indicated by solid lines).

No amphibolite is formed for picritic/komatiitic compositions (Fig. 2c, from Foley et al., 2003), and these rocks do not produce TTG-like melts under any conditions during partial melting, but basaltic liquids instead.

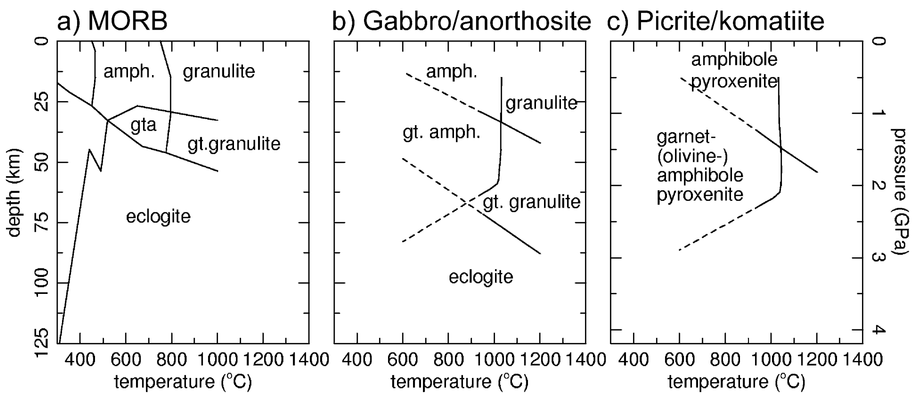


Fig. 2. Metamorphic phase diagrams for different magmatic rocks, which may be expected in Archean oceanic crust. (a) Mid-Ocean Ridge Basalt, from Hacker et al. (2003). amph. indicates amphibolite, gta is garnet–amphibolite, and gt. granulite is garnet granulite. (b) Gabbro/anorthosite (Green and Ringwood, 1967; Foley et al., 2003). (c) Picrite/Komatiite. Data from experiments on Gorgona komatiite by Foley et al. (2003).

3.2. Geodynamical settings

The left-hand side and middle frames of Fig. 3 illustrate all five geodynamic processes under consideration, showing the composition field. In the left-hand frames, the color scale from blue to orange indicates the degree of depletion of peridotite. Basaltic/eclogitic (black/red) and felsic (green) tracers are included. Depleted metabasalt has a lighter red color. Arrows indicate the instantaneous flow field. The middle frames show the contours of the crust in black, with regions of partial melting of peridotite in blue and partial melting of (meta)basalt in red and green.

The first process is a small-scale delamination of the lower crust (Fig. 3a).

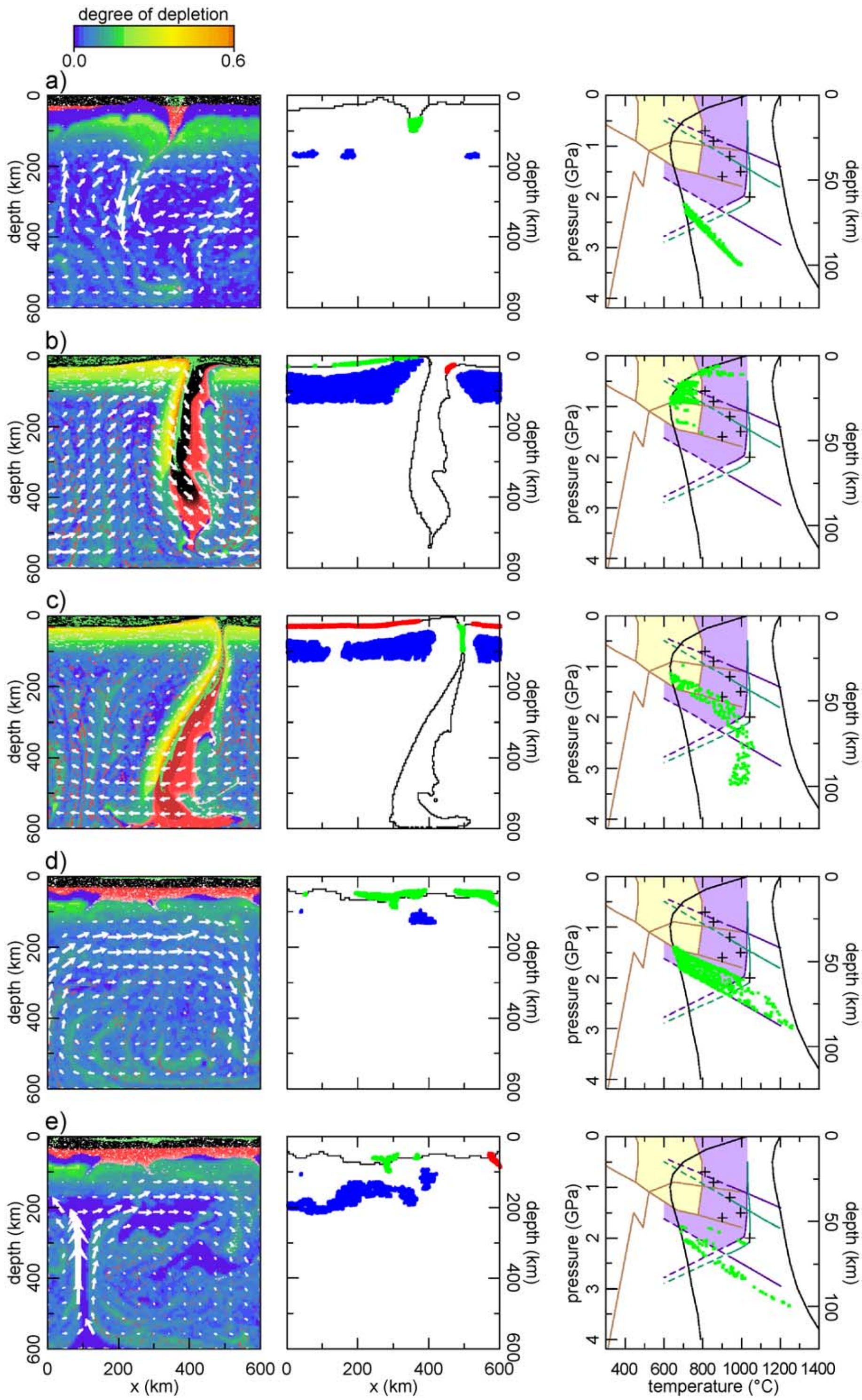
Partial melting in the upwelling limb of a small-scale mantle circulation may locally add material to the crust, which becomes thicker. The lower part of the crust may become involved in the circulation. At first, this is passive entrainment, but when sufficient basalt has turned into eclogite, the involvement becomes active. Partial melting of metabasalt takes place in the tip of the downwelling material, indicated by the green spot in the middle frame.

The second process is large-scale resurfacing, shown in Fig. 3b and c. In this process, a local anomalous thickening of the crust, for example caused by the addition of material by a mantle plume, may become gravitationally unstable when a large part of the (lower) crust has turned into eclogite. This may trigger the entire crust to sink into the mantle on a very short time scale (about 2 million years). A new crust is produced by partial melting of upwelling mantle material. Two settings for partial melting of (meta)basalt can be indicated here. Fig. 3b shows partial melting, indicated by the green color in the middle frame, at the base of the freshly produced crust which replaces the material which is sinking into the mantle. The required heat is supplied by upwelling mantle material associated with the resurfacing event. In Fig. 3c, partial melting takes place in the sinking crustal material itself, indicated by the green color in the middle frame.

Fig. 3d shows the third process, which is the partial melting of a locally thickened crust. Due to successive stacking of basaltic extrusives, the crust locally extends to about 80 km depth at temperatures above the hydrous basalt solidus. Penetration of a mantle diapir into the upper mantle as a heat source for melting at the base of the crust is shown in Fig. 3e.

3.3. Conditions of partial melting

The p, T conditions under which partial melting of metabasalt takes place in the different geodynamical settings with respect to the metamorphic phase boundaries (see Fig. 2) are indicated in the right-hand side frames of Fig. 3. The p, T conditions of melting basaltic tracers are indicated in green, and the corresponding locations are indicated in green in the middle frames (melting tracers not indicated in the p, T diagrams are shown in red). These results show that if we assume a MORB-like composition (brownish phase boundaries in Fig. 3) for the basaltic crust which is partially melting in our experiments, TTG-like melts formed in the amphibole stability field consistent with observations of Nb/Ta and Zr/Sm ratios in natural TTGs can be produced in two of the geodynamical settings, both associated with large-scale resurfacing [see Fig. 3b and c, where the yellow region indicates the (garnet) amphibolite stability fields]. The melting conditions at the base of the newly produced crust correspond to the shallow conditions inferred for natural TTGs by Winther (1996), indicated by black crosses in the right hand frames. In the sinking crust (Fig. 3c), conditions correspond to the deepest conditions for natural TTGs inferred by Winther (lowest two crosses in the right hand frame). In the other settings, melting takes place at levels which are too deep for amphibolite stability, well into the eclogite field. For a material with gabbroic/anorthositic composition (purple phase lines), the stability field of (garnet) amphibolite (purple shaded area, partially overlapping with the yellow MORB region) is extended relative to MORB. The two settings associated with the resurfacing event again produce TTG melts. Partial melting of the lower thickened crust in either the presence (Fig. 3d) or absence (Fig. 3e) of a mantle diapir may also produce TTG melts, but the uncertain location of the extrapolated


degree of depletion
0.0
0.6
a)
b)
c)
d)
e)
depth (km)
x (km)
pressure (GPa)
temperature (°C)

boundary of the amphibole stability field makes this questionable. The small-scale delamination setting does not produce TTG melts in the amphibole stability field consistent with natural TTGs. More mafic material, represented by the greenish phase boundaries in Fig. 3 for picrite/komatiite, show most melting in the (garnet) pyroxenite fields, producing basaltic melts.

3.4. Production rates of continental material

Fig. 4 shows the 2-D volume of felsic material produced in the different geodynamical settings listed in Table 1 and shown in Fig. 3. The values for 1200×1200 km models (Mtcr and Mdia) have been normalized to correspond to the 600×600 km domain of the other models. It is clear from this figure that the crustal production rates range over orders magnitude for the different processes. A single small-scale delamination (see Fig. 3a) produces about 30 km^2 of crust in 10 million years, whereas partial melting of metabasalt associated with a resurfacing event (see Fig. 3b and c) produces nearly 2000 km^2 in less than 2 million years. This corresponds to equivalent felsic layers spread out over the entire domain width of 50 m and 3.3 km, respectively. Another important point of Fig. 4 is the near coincidence of the curves of models Mtcr (thickened crust) and Mdia (lower mantle diapirism). This indicates that although much hot lower mantle material is introduced into the upper mantle in model Mdia and not in model Mtcr, it does not pass this heat on directly to the lower crust to allow it to melt. The diapir head spreads out in the upper mantle and does not reach depths less than 100 km (consistent with the maximum depth of 90 km attained by a numerical model diapir in similar thermal and identical rheological conditions in Van Thienen et al., 2003), and does not reach shallower levels where crustal material is present.

3.5. Distribution of felsic material in the crust

In order to link the results of our numerical experiments to field observations on Archean cratons, we examine the distribution of felsic material in the crust and its association with basaltic material. Fig. 5 illustrates the distribution of basaltic and felsic material in the crust of model Mtcr (see Fig. 3c) by showing all individual particle tracer compositions (basaltic: black, felsic: green, peridotitic: blue) in different regions of the domain. Zoom-ins of two settings in which TTG-melts are produced by partial melting of (meta)basalt, indicated by red squares in the top panel, are presented in the lower frames. The left panel shows a crustal section associated with the partial melting of young lower crust in a resurfacing setting. Although basaltic material dominates the crust, a significant amount of felsic material is interspersed throughout the crust. In our model, this is produced by the successive stacking of basaltic and felsic extrusives, but in a similar setting in the Earth, intrusives are also possible. The lower right panel of Fig. 5 shows a zoom-in of a crustal section associated with 'subducting' basaltic crust. This image is quite different from the previous, since in the downwelling part only partial melting of metabasalt is taking place, and not of peridotite. This results in the exclusive production of felsic crust without any associated basaltic crust. In our model, a felsic body of about 20-km width and 10-km depth is produced.

Fig. 3. Partial melting of metabasalt in five different geodynamical settings. The left-hand side frames show the composition field, with the colour scale from blue to orange indicating the degree of depletion of peridotite. Black indicates basalt, red is eclogite, and the colour range in between indicates material in the transition stage. Eclogite becomes lighter red as it is more depleted. Greenish crustal material is the product of (meta)basalt melting. The middle frames show basaltic particle tracers which undergo partial melting in red and green, and melting peridotite in blue. The green tracers are also shown in the right hand side frames, which show the location of partial melting of metabasalt in p, T space. The phase diagrams represented in Fig. 2 are included, and the regions which produce Nb/Ta and Zr/Sm ratios consistent with natural TTGs are indicated in yellow (MORB) and purple (gabbro/anorthosite). The inferred melting conditions of some natural TTGs (Winther, 1996) have been included as black crosses, and solidus and liquidus used are indicated by black curves. Note that the frames in the right column have a different depth scale than the other frames. (a) Small-scale delamination of the lower crust (Mdel). (b) Partial melting of a newly produced lower crust associated with a resurfacing event (Mres). (c) Partial melting in a 'subducting' crust during a resurfacing event (Mres). (d) Partial melting of the base of a thickened crust (Mtcr). (e) Lower mantle diapirism (Mdia). Frames d and e only show 1/4 of the complete computational domain of 1200×1200 km.

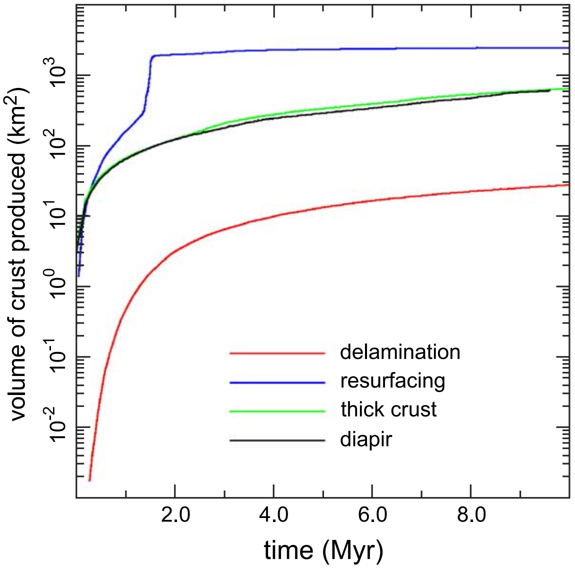


Fig. 4. 2-D volumes of felsic material produced by partial melting in the different models. Volumes are normalized to a 600-km domain for larger models.

4. Discussion

The results of our modeling experiments as shown in Fig. 3 indicate that the production of TTG lavas in the amphibole stability field in a hot Archean mantle is more likely to take place in settings associated with a resurfacing event than in association with a mantle plume or small scale delamination of the lower crust. However, due to the simplifications made in our numerical model, one possibly important agent of heat transfer is not included in our experiments. The emplacement of a basaltic liquid (both the high temperature and the latent heat effect of solidification) may cause significant amounts of partial melting in the lower amphibolitic crust (Huppert and Sparks, 1988; Petford and Gallagher, 2001). Therefore, specifically mantle diapirism may still be important because it may generate large amounts of basaltic melt (see the blue tracers in the middle frame of Fig. 3e, indicating peridotite melting), which may in this way cause partial melting in the lower crust. It may also play an

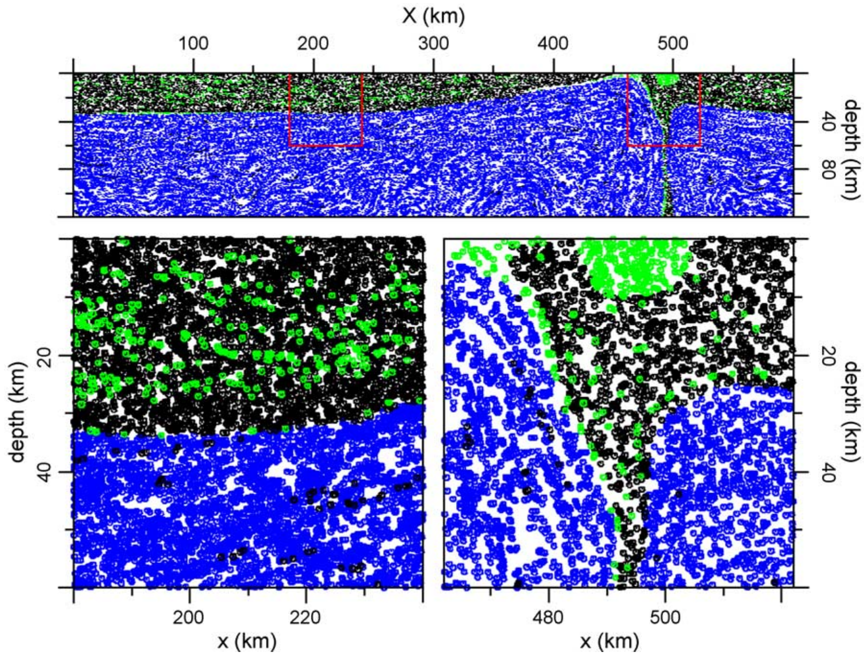


Fig. 5. Zoom-in on the crust of model Mtcr as shown in Fig. 3c. (Meta)basaltic tracers are indicated in black, felsic tracers in green, and (both depleted and fertile) peridotitic tracers in blue. Two red squares in the top panel indicate the locations of the zoom-ins of the lower frames.

additional role in the partial melting of the newly produced crust associated with a resurfacing event in Fig. 3b. However, because of the smaller length scales of porous two-phase flow compared to flow of mantle material (Scott and Stevenson, 1984, find magma soliton wavelengths of a few kilometers for plausible estimates of parameter values, which require a spatial resolution of an order of magnitude smaller to be accurately resolved numerically), inclusion of realistic melt segregation through multiphase flow in upper mantle scale numerical models is a major computational challenge at the moment.

Because the resurfacing events as shown in Fig. 3b and c are episodic, the production of felsic material in the associated settings is episodic as well. This is consistent with geological evidence, which indicates that the addition of felsic material to cratons was an episodic process, e.g., in the Pilbara craton (Smith, 2003) and the Kaapvaal craton (Anhaeusser and Walraven, 1999).

On the average, Archean cratons consist of 60% granitoid gneiss, 30% massive granitoid plutons and 10% greenstone belts (Goodwin, 1991). These greenstone belts generally consist of successions of major (ultra)mafic and minor felsic supracrustal rocks, and have *preserved* thicknesses of 5–20 km (Condie, 1994). The massive granitoid plutons, on the other hand, are units of tens to over 100 km diameter which consist of granitoids, the early and mid-Archean cases typically belonging to the TTG suite and later granitoids being dominated by potassic granites (Goodwin, 1991). The greenstone association may correspond well to the type of crust produced by partial melting of the lower part of newly produced crust in association with a resurfacing event, as can be seen in the lower left panel of Fig. 5. The granitoid plutons correspond well to the large volume of felsics which are produced above a sinking crust in a resurfacing setting without any associated basalts, as shown in the lower right panel of Fig. 5. Zegers et al. (1999) did radiometric dating of felsic intrusives and supracrustal (greenstone belt) rocks from the eastern Pilbara (>3 Ga), and found two distinct periods of about 50–80 My in which both large batholiths and supracrustal rocks were produced. This shows that the coeval production of these two types of rock associations may be consistent with geological observations. Although greenstone belts appear to have been formed in a variety of settings, a number of them show oceanic assemblages (Kimura et al., 1993; Nijman et al., 1998; Kerrich et al., 1999). The setting in which the greenstone-like succession is produced, forming the new crust in a resurfacing event, can also be considered an oceanic environment. The short time scale on which the crust interpreted as greenstone belt in our model is produced contrasts with the fact that the age of individual units within a greenstone belt may show an age range of 300 My (Thurston, 1994). Therefore, it may be more realistic to interpret the crustal segment produced in the resurfacing event as a single unit of a greenstone belt, to which new material may be added later by similar or other processes.

On the basis of these results, we propose that resurfacing events may have played an important role in the early formation of Archean cratons. The mechanism coevally forms both greenstone-like successions and massive volumes of felsic melt, which may form large plutons. The TTGs produced have low Nb/Ta and high Zr/Sm ratios as a result of melting in the amphibolite stability field, consistent with analyses of Archean TTG rocks (Foley et al., 2002). Furthermore, this mechanism does not require the operation of plate tectonics, which is often invoked to explain Archean geology in general (e.g. De Wit, 1998) and greenstone belt formation specifically (see Table 5 of Polat et al., 1998), but against which strong physical arguments have been published for a hotter mantle (Vlaar, 1986; Vlaar and Van den Berg, 1991; Van Thienen et al., 2004b).

The prominent role in our model results of resurfacing dynamics as an agent in the generation of continental type crust raises the question of the implications of these findings for our sister planet Venus. This is because massive resurfacing events have been suggested to have occurred on Venus less than a billion years ago. Crater count studies have been interpreted as an indication for a global resurfacing of Venus around 500 million years before present (Schaber et al., 1992), and it has been suggested that resurfacing may have taken place episodically on Venus (Turcotte, 1995; Fowler and O'Brien, 1996), possibly by periodic plate tectonics (Solomatov and Moresi, 1996). This observation led us to speculate that the episodic resurfacing behavior, self-consistently produced in thermochemical mantle convection models applied to the early Earth, may have been active on

Venus as well (Van Thienen et al., 2004a). If resurfacing was indeed a common type of mantle dynamics on Venus, our model results would suggest that silicic melts and continental type crust might have formed on a significant scale on Venus. Felsic material in the Venusian crust has been inferred from morphology of dome-shaped volcanoes (McKenzie et al., 1992), and a local intermediate composition has been interpreted from rock analyses on Venus (see Kargel et al., 1993). Felsic material should be observable in near-infrared thermal radiation if it forms a significant part of large Venusian highland terrains (Hashimoto and Sugita, 2003). Because of such (possible) observations and inferences, the question whether the dynamics discussed in Van Thienen et al. (2004a), and studied for its ability to produce felsic material here, would work on Venus as well is an important one. However, silicic melts to build continental type crust form under modest temperature conditions when the peridotite solidus is lowered by a sufficiently high water content. Since Venus is generally considered to be a dry planet (Campbell and Taylor, 1983; Nimmo and McKenzie, 1998, although it may have been somewhat wetter in its earlier history, possessing a fraction of $x \times 10^{-3}$ of the terrestrial oceans), conditions have probably been unfavorable for the production of granitoids as it takes place in our models applied to Earth (Campbell and Taylor, 1983).

5. Conclusions

The model results show that partial melting of metabasalt in two distinct settings associated with rapid resurfacing, at the base of newly produced basaltic crust and in the necking part of a 'subducting' old crust, takes place in the amphibolite stability field when assuming a basaltic or gabbroic crustal composition. In the other settings which were investigated, which are small scale delamination of the lower crust, melting at the base of a anomalously thick crust and melting associated with a lower mantle diapir, partial melting of metabasalt generally takes place under uncertain metamorphic conditions or in the eclogite stability field. Since the former type of melting produces TTG melts with Nb/Ta and Zr/Sm ratios consistent with geochemical observations, and the latter does not, we suggest that partial melting of metabasalt associated with resurfacing events may have been an important agent in the production of continental material during the early history of the Earth. Furthermore, the configuration of the rock associations produced in our model resembles Archean granite–greenstone terrains, and we therefore propose that the resurfacing mechanism may have played an important role in the generation of (some of) these granite–greenstone associations found on Archean cratons.

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