Mantle overturn and thermochemical evolution of a non-plate tectonic mantle

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A R T I C L E I N F O

Article history:
Received 17 November 2020
Received in revised form 3 May 2021
Accepted 4 June 2021
Available online 1 July 2021
Editor: R. Bendick

Keywords:
Venus
g eo dynamic modelling
 mantle overturn
 mantle evolution
 early Earth
 planetary evolution and compositional stratification

A B S T R A C T

We use coupled thermomechanical and thermodynamic modelling to investigate the evolution of a heating mantle below a stagnant lithosphere that leads to overturn events as proposed for Venus. Eclogitized crust entrapped into the upper mantle accumulates in a gravitational trap at the lower-upper-mantle boundary, where basaltic compositions display an intermediate density between upper and lower mantle. While the convecting upper mantle remains at a stable mantle potential temperature, the lower mantle progressively heats. Over 900 Ma, the basalt-rich layer develops into a thermal boundary layer with an offset in potential temperatures as high as 300 °C. Occasional plumes breaching the layer from below cause local increased magmatism at the surface but do not induce complete mantle overturn. Once the basalt-rich layer, however, reaches a critical thickness, so that the crust in its lowermost part experiences a transition into assemblages denser than lower mantle a runaway process is initiated. The entire crustal-rich layer is dragged into the lower mantle, switching the convective mode to whole mantle convection. Fertile, superheated lower mantle replaces the depleted cold upper mantle and magmatic production increases by orders of magnitude for ~50 Ma. During this period, crustal production-rates are so high that not only a new surface in generated but also the delamination rate of lower crust into the upper mantle is increased dramatically with the hot low-viscosity crust now forming much larger delaminating bodies. About 50 Ma after the onset of the overturn, the magmatic production subsides and crust starts to accumulate at the lower-upper mantle boundary again. The overturn event creates a mantle that is chemically heterogeneous and intensely foliated at the model-resolution scale with intermingled crust, depleted and undepleted mantle. The phase-change induced collapse of the basalt-layer and subsequent global overturn happen at mantle temperatures considerably lower than predicted by previous models for Venus.

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1. Introduction

Despite great similarities Earth has plate tectonics and Venus does not. Earth presently releases heat at around 40 TW or ≈88 mW/m² (Davies and Davies, 2010; Jaupart et al., 2007), a rate considerably larger than heat production from radiogenic decay (≈23 TW, e.g. Jaupart et al., 2007). The net cooling of Earth is due to the creation and destruction of tectonic plates, i.e. the subduction of cold slabs into the mantle, and associated high surface heat flow at mid-oceanic ridges where partial melting of the asthenosphere produces new oceanic lithosphere that subsequently cools conductively. Without plate tectonics, i.e. with an insulating stagnant lid at the surface, bulk heat flow on a terrestrial planet is significantly lower. Total surface heat flow for Venus is poorly constrained, but estimates suggest it is around half of its radiogenic heat production (Nimmo and McKenzie, 1998), which implies net heating of the planet’s interior. It is widely accepted that this heat is eventually released in a catastrophic event of mantle up- and downwelling called mantle overturn or sometimes lid overturn (e.g. Armann and Tackley, 2012; Davies, 2008; Ogawa, 2000; Papuc and Davies, 2012). Such an event would result in massive global magmatism shaping a new surface. Inference from the impact crater density on the Venusian surface suggests a uniform age of about 300-600 Ma (e.g. Strom et al., 1994) implying the most recent resurfacing event occurred at that time. Some researchers have proposed that resurfacing is continuous on Venus today, suggesting magmatism is more or less spatially uniform under a stagnant lid (e.g. O’Rourke et al., 2014). However, we follow former idea that resurfacing is catastrophic.

Venus may also provide a useful analogy for Earth during the early Archaean. Without stable subduction-driven plate tectonics,
Earth’s mantle might have likewise heated beyond the range of mantle potential temperatures (MPTs) that the more recent petrological record suggests (Ganne and Feng, 2017). This scenario is sometimes referred to as the Archaean thermal catastrophe (Korenaga, 2006).

Here, we apply thermomechanical modelling to better understand the controversial mechanism of mantle overturn. On Venus, it is presently unclear if the entire mantle below the lid experiences progressive heating or only a deeper portion below yet another thermo-mechanical boundary layer (TMBL). One possibility is that the entire mantle heats at a similar rate. Geochemical data obtained by the Venera probes in the 1980’s, however, suggest that the major element composition of the crust is similar to mid-oceanic ridge basalt (MORB) on Earth and that the source is a garnet peridotite with a MPT of about 1330°C, similar to that of Earth’s mantle today (McKenzie et al., 1992). Another possibility is a TMBL at the core-mantle boundary (CMB) similar to what commonly assumed for Earth to explain large igneous provinces (Hofmann and White, 1982). This would imply that active volcanism on Venus releases most of the produced radiogenic heat. We follow a third option (Ogawa, 2003) that a TMBL develops along the lower-upper mantle boundary (LUMB) which is defined as the breakdown of ringwoodite into bridgmanite and periclase in typical ultramafic mantle rocks. This phase transition increases the mantle density by about 10% and has a negative Caleytron slope for typical MPTs (see Fig. 1). This suppresses convective flow across the boundary and has previously been considered to cause layer-lidded convection on its own in a one-component primitive mantle (Richter and McKenzie, 1981). However, the Caleytron slope required to maintain layer-lidded convection for typical mantle-Rayleigh numbers is higher than recent experimental and theoretical evidence suggests for the LUMB (e.g. Christensen and Yuen, 1985; Hirose, 2002; Cottaar and Deuss, 2016).

We explore the hypothesis that basaltic crust entrenched in the mantle convection system is accumulated at the LUMB and thus enforces its character as a TMBL (e.g. Petersen et al., 2018), a mechanism called “basalt barrier” by Ogawa (2003). Basalt has two relevant properties: Under water-free conditions, it transforms into garnet-rich assemblages already at crustal conditions and is altogether denser than peridotite for almost the entire mantle range. For basaltic compositions, the LUMB-defining transition into perovskite-rich assemblages occur at higher pressures (i.e.≈28 GPa) as compared to peridotite (≈24 GPa; see Fig. 1). In basalt, this transition is defined by the breakdown of garnet, not ringwoodite as basaltic compositions are present as garnetite in the mantle transition zone (MTZ). This pressure offset defines a depth interval in the uppermost lower mantle where crust is less dense than ambient peridotite and gravitationally stable (Fig. 1).

Based on thermomechanical modelling, Ogawa (2000) showed that internal heating in the mantle of Venus would eventually lead to major bursts of magmatism and vigorous mantle whole-mantle convection causing significant compositional stratification, with the lower mantle becoming highly enriched in basaltic components. Ogawa (2003), Davies (2008) and Papuc and Davies (2012) all proposed accumulation of crust at the LUMB and an associated formation of a barrier that suppresses convective flow of material across it. This is the layer that would break down during mantle overturn once overheated lower mantle overrides this mechanical barrier. Several questions, however, remains unclear: How efficiently can a basalt barrier be formed for realistic MTZ-conditions, i.e. freely floating delaminated pieces of crust in ambient mantle? How effectively is the resulting barrier suppressing whole-mantle convection, to what extremes can the TMBL develop and which processes control breakdown of the TMBL to create a global mantle overturn? Is mantle overturn characterized by a dramatic short-lived event or a longer period of whole mantle convection? What are the consequences of mantle overturn for long-term stratification of the mantle? In this study, we use coupled thermodynamic- and thermomechanical modelling to address these questions.

2. Numerical model

We employ a 2D thermomechanical model allowing for elastic, viscous and plastic deformation. The governing equations are solved numerically using conservative finite differences on staggered grids combined with a non-diffusive particle-in-cell technique for advecting properties along the velocity field, a method described in Gerya and Yuen (2003) and in more detail in Gerya (2019). The velocity field is obtained by solving the Stokes equation under the assumption of continuity, which in 2D yields the following set of equations:

\[
\frac{\partial \sigma_{xx}}{\partial x} + \frac{\partial \sigma_{xz}}{\partial z} - \frac{\partial P}{\partial x} = 0,
\]
where \( \sigma_{xx}, \sigma_{xz}, \sigma_{zx} \) and \( \sigma_{zz} \) are the deviatoric stress tensor components, \( P \) is pressure, \( \rho \) is the density and \( g_z \) is the vertical gravitational acceleration. The equations are solved using the iterative multigrid method (Wesseling, 1992), allowing for a high spatial and temporal resolution. The Stokes equations are solved using free-slip boundary conditions. To account for the visco-elastic-plastic rheology, the coupling between stress and strain rate is as follows:

\[
\dot{e}_{ij} = \dot{e}_{ij}^{(\text{diffusion})} + \dot{e}_{ij}^{(\text{dislocation})} + \dot{e}_{ij}^{(\text{elastic})} + \dot{e}_{ij}^{(\text{plastic})}
\]

\[
= \frac{1}{2\eta_{ij}^{(\text{diffusion})}} \sigma_{ij} + \frac{1}{2\eta_{ij}^{(\text{dislocation})}} \sigma_{ij} + \frac{1}{2\mu} \frac{D\sigma_{ij}}{Dt} + \frac{\chi}{2\sigma_{ij}}
\]

where \( \dot{e}_{ij} \) is the total strain rate tensor, \( \sigma_{ij} \) is the deviatoric stress tensor, \( \eta \) is viscosity, \( \sigma_0 \) is the second invariant stress, \( \mu \) is the elastic shear modulus and \( \chi \) is a plastic multiplier that is nonzero only if \( \sigma_0 = \sigma_{\text{yield}} \). With the pressure-dependent Mohr-Coulomb failure limit, \( \sigma_{\text{yield}} = C + P \sin \phi \) where \( C \) is cohesion and \( \phi \) is the friction angle.

Viscosity is calculated using experimentally-derived parameters with an Arrhenius-type flow law:

\[
\eta = \frac{1}{2 A n^{n-1}} \exp \left( \frac{E_a + PV_a}{RT} \right)
\]

where \( A \) is a pre-exponential factor, \( n \) is a stress exponent, \( E_a \) is activation energy, \( V_a \) is activation volume, \( R \) is the ideal gas constant and \( T \) is the absolute temperature. Values for all used rheological parameters are found in Table S3.

The heat equation is stated from a Lagrangian frame of reference, here in 2D:

\[
\rho c_p \frac{DT}{Dt} = -\frac{\partial q_s}{\partial x} - \frac{\partial q_d}{\partial z} + H
\]

where \( c_p \) is heat capacity, \( q_s \) is conductive heat flux and \( H \) is the sum of all source terms: radiogenic, shear, adiabatic and latent heat. Due to the length of our simulation, radiogenic heat is reduced significantly over time as the concentration of heat-producing elements decay (see Supplements S1). Shear heating is calculated from obtained stresses and strain rates:

\[
H_s = \sigma_{ij} \dot{e}_{ij}
\]

Adiabatic and latent heat are function of entropy and are accounted for in a thermodynamically consistent manner, using repeated interpolation from the entropy look-up table. The numerical procedure is described in detail in Petersen et al. (2018). Density and entropy for the involved lithologies in the model are obtained by interpolating from high resolution pre-calculated lookup tables generated in Thériak-Domino (de Capitani and Petrukakis, 2010) using the database of Holland et al. (2013). Compositions used for generating look-up tables for peridotite and basaltic crust are found in Supplements S4. The numerical implementation for partial melting of the mantle and depletion effects on density is described in Supplements S2. The model also accounts for the compaction that occurs in the melting zone when a parcel of mantle undergoes partial melting and melt is extracted to the surface. This routine is described in Supplements S3.

3. Model setup

3.1. Model geometry and boundary conditions

The model domain is 6000 \( \times \) 3000 km rectangular box, spanning the entire mantle depth range along with 120 km sticky-air layer. The resolution is 1025 \( \times \) 385 with variable grid spacing, achieving a cell resolution of about 6 km in the upper mantle in a 1:1 aspect ratio. In the lower mantle the aspect ratio is 1.55:1. Particles are distributed evenly across the model domain ensuring each cell has at least 25 particles.

The initial crust has a thickness of 30 km and a linear geotherm through the lithosphere from 500°C at the surface to 1100°C at Moho. The entire asthenospheric mantle initially follows an isentropic adiabat calculated from the entropy table with a potential temperature of 1400°C. Initially, the model has a primitive mantle composition. We assume that the Venus’ mantle has a concentration of heat-producing elements (HPes) identical to Earth’s mantle (Supplements S1). The gravitational acceleration is 8.87 m/s², which affects pressure gradients and shifts phase transitions in the mantle to about 10% greater depth.

All boundaries have free-slip boundary conditions and the rock-air interface is kept approximately stress-free by using the “sticky-air” approach for the upper 120 km of the model domain (Cramer et al., 2012). The atmosphere in this upper layer has a visco-plastic rheology with a viscosity of \( 10^{21} \) and a finite plastic strength of 0.1 MPa. The initial condition for the model is shown in Fig. 2.

3.2. Rheology

A composite flow law is used for the crust to account for changes in rheology associated with the significant phase transformations occurring during burial and delamination. At pressures lower than 1.5 GPa the flow law for the crust is the one for plagioclase (An75) of Ranalli (1995). At higher pressures, garnet is stable and becomes increasingly abundant throughout the upper mantle to about 27 GPa, where it transitions into perovskite. In the garnet-stable pressure range, a flow law derived for garnet at high pressure is used (Mei et al., 2010; Xu et al., 2013). In the perovskite stability field at pressures higher than 27 GPa, the model rheology for crust is identical to that of the ambient mantle (Čížková et al., 2012). For peridotite in the upper mantle, a combined dry flow law of diffusion creep and dislocation creep is utilized (Karato and Wu, 1993). The diffusion creep assumes a grain size of 0.925 mm and a shear modulus of 67 GPa, resulting in a median viscosity in the weakest part of the asthenosphere (at about 3 GPa) of 3 \( \times 10^{19} \) Pa-s, increasing to \( 1 \times 10^{21} \) Pa-s in the lowermost MTZ, as seen in Fig. 2. This viscosity is calculated for an early stage of the model and is a result of combined diffusion- and dislocation creep. The initial contrast in viscosity across the LUMB is more than an order of magnitude, but this contrast decreases over time as the lower mantle heats. All rheological parameters are listed in Supplements S4.

3.3. Temporal resolution

The model employs a dynamic time step allowing advection of a particle in a single time step of up to 10% of the minimum cell size, with a maximum allowed time step of 5000 years. The limited advection distance within a single time step means the model progresses highly non-linearly, with the bulk of the computational time spent on the overturn phase. The presented model is a single run with a total run time of 3 months on 2 Intel Xeon E5-2630v4 CPUs.
4. Model presentation and analysis

In the following paragraph, we first present the overall model evolution (Fig. 3 and 4) which clearly shows three phases (accumulation, overturn, and restoration). We analyse the first order mechanics and the thermal as well as the compositional evolution of the mantle including variations of magmatic production rates (Fig. 5a), the evolution of MPTs (Fig. 5b) and the average vertical distribution of crust in the mantle (Fig. 5c). After that, we focus on three particular processes using model snapshots (Fig. 5, 6): (1) The delamination and accumulation of crust, (2) the penetration of the crustal layer at the LUMB by small hot upwellings from the lower mantle, and (3) the gravitational collapse of the layer that leads to mantle overturn. The paragraph thereafter discusses caveats and finally the large-scale implications for planetary evolution and mantle heterogeneity also on Earth as well as some interesting details.

4.1. The accumulation phase (0-880 Ma)

Initially, the convecting sublithospheric mantle has an adiabatic temperature profile with a MPT of 1400 °C. It is thus above the solidus in the uppermost part (Fig. 3a). In our model, the predicted melt fraction is deposited as basaltic extrusions at the surface on top of the existing crust. When the crust reaches a critical thickness of about 40 km, the base of the crust undergoes a series of phase changes and ultimately becomes denser than the underlying mantle. The negative buoyancy of eclogite below the brittle-ductile transition forms a Rayleigh-Taylor instability (RTI) that drags surrounding crust into delaminating drops along with some surrounding depleted mantle. This process continuously thins the crust back to subcritical thickness. The resulting convection in the mantle causes upwelling of fresh, fertile mantle, which further promotes decompression melting and crust production (e.g. Bédard, 2006; Johnson et al., 2014; Piccolo et al., 2019; Sizova et al., 2015). In the depth interval of 740-840 kilometres below the LUMB, basaltic crust is buoyant relative to the ambient peridotite in the uppermost part of the lower mantle. The accumulating layer together with the negative slope of the phase change suppresses convection across the LUMB. After about 250 Ma, an initial phase of moderately intense magmatism gradually ceases as the upper mantle is depleted of incompatible elements (Fig. 5a). The increasing levels of depletion in the upper mantle support separation of convection across the LUMB as they are associated with a mantle density contrast once the crust is sequestered. During the accumulation phase, the upper mantle remains at an equilibrium potential temperature of around 1400 °C (Fig. 5b), as radiogenic heating is balanced by conduction through the lithosphere and latent heat released through partial melting, a process referred to as magmatic heat piping (e.g. Turcotte, 1989). The lower mantle, however, undergoes progressive heating from radioactive decay and a large temperature offset develops across the LUMB during this phase (Fig. 3b-d, 5b). Two distinct events of moderately higher magmatic production due to upwelling of hot lower mantle occur during this phase at 600 and 800 Ma (Fig. 5c) and we discuss these events in detail in Section 4.4.

4.2. The overturn phase (880-930 Ma)

During the accumulation phase the upper mantle remains at a remarkably constant potential temperature while the lower mantle heats. Temperature-pressure conditions in its uppermost part of the lower mantle raise above the slope break of the LUMB-defining phase transitions at ca. 1750-1800 °C (Fig. 1) where the Clapeyron slope of these reactions turns positive. This means that convection across the LUMB is no longer suppressed by phase changes in the mantle but instead promoted. It has been proposed that hot plumes actually should pass the boundary easily (e.g. Hirose, 2002). However, at this model stage a separation of upper and lower mantle reservoirs is efficiently supported by the growing amount of crustal nodules mixed into the peridotite matrix at the LUMB (Fig. 3b-c; 6b3), i.e. by the basalt barrier (Ogawa, 2003). Note that the rw-to-pws+per transition in the matrix peridotite is located inside this mixed layer (e.g. Fig. 6b2). The increasing thermal gradient across the crust-mantle layer (Fig. 3a-d; Fig. 6b) implies that the lower-mantle portion of the layer gradually heats up. In the uppermost lower mantle, the density decrease with temperature is more pronounced for peridotite compared to crust due
continuous garnet-producing reactions in peridotite. This causes the lighter garnet-rich crust suspended in peridotite to gradually lose buoyancy and sink slightly into the uppermost lower mantle as the TMBL heats up. At 858 Ma, the crust at the bottom of the TMBL gets below its garnet-perovskite transition at one site. The crust then becomes about 100 kg/m$^3$ denser than the ambient lower mantle (Fig. 3d, Fig. 6c). This initiates an avalanche of sinking crust, as the pull from the post-garnet crust drags additional crustal through the phase transition (Fig. 6c). As the crust-rich layer and the following cold depleted upper mantle sink downwards, return flow (upwelling) from lower mantle breaches the layer some 2000 kilometres away from the site of collapse (Fig. 3e). The upwelling lower mantle has a MPT more than 200$^\circ$C hotter than the ambient upper mantle. It rises to the top and experiences high degrees of partial melting. This causes the local crust to thicken and delaminate into the mantle at a much higher rate than during the accumulation phase. Also, delaminated pieces of crust during this phase are much larger (Fig. 3e, 4a), because the crust produced in this regime undergoes a different thermal evolution: The geothermal gradient in the rapidly produced crust is much higher and, hence, the crust overall much weaker. This enables easy horizontal flow and more crustal material is pulled into a single Rayleigh-Taylor instability before delaminating. The large bodies rapidly fall through the upper mantle and perturb the existing compositional stratification in the MTZ. This facilitates the formation of additional upwelling sites. After less than 5 million years, the TMBL at the LUMB is destroyed and hot lower mantle rises into the upper mantle throughout the model (Fig. 4a-c) while the downward collapse of crustal-rich material into the lower mantle is propagating laterally. The magmatic production rate during this event is orders of magnitude higher than during the accu-
mulation phase (Fig. 5a). With lateral spreading of collapse, the

Fig. 3. Time evolution of the model. The first column shows lithology with green shades representing peridotite, purple is crust and orange are regions of active partial melting. The second row shows temperature, and the third row shows degree of depletion for peridotite. Crust is not remelted in the model and shown as black. See main text for description of model evolution. High-resolution versions of lithology- and depletion figures are found in Supplementary Figs. SF1a-j. Viscosity maps for corresponding stages are shown in Supplementary Figs. SF2a-e. Continued in Fig. 4.
simultaneously formed upper-mantle melange of rising primitive mantle and downwelling depleted mantle plus crust is dragged back into the lower mantle. This results in a heavily sheared mixture of variously depleted mantle and crust (Fig. 4c-e; Fig. 7). After about 30 Ma, a large portion of the hot lower mantle has risen into the upper mantle and parts of it have returned as this complex plum-pudding melange. During overturn, the timespan from lower mantle material reaching the melting zone and returning back to the lower mantle is in the order of a few to tens of millions of years. It takes another 20 Ma for magmatic production to decrease back to lower than pre-overturn levels (Fig. 5a).

4.3. The restoration phase (930-1300 Ma)

At 930 Ma, intense magmatism associated with mantle overturn has tapered off, and the entire mantle has much more homogeneous MPTs than the highly stratified mantle before the overturn. Most of the crust produced throughout the model run has sunk into the lower part of the lower mantle. The overall mantle temperatures around the LUMB are below the slope break (Fig. 1; Fig. 5b). Some accumulation of remaining crust in the depth-inversion window below the LUMB (Fig. 5c), suggest the reestablishment of the TMBL. The second generation TMBL, however, is not nearly as well-defined as the initial layer, in part due to numerical diffusion of crustal droplets, but primarily because we do not consider refertilization of depleted mantle or melting of crust. Hence, depletion of original mantle is irreversible in our model and there is no real cyclic behaviour supported. In fact, when temperatures in the uppermost lower mantle cross the slope break of the LUMB again, the limited amount of crust at the LUMB is not sufficient to protect the TMBL.

During this phase, a portion of hot primitive mantle gets trapped in the mid-MTZ (Fig. 4d, 5b), where the temperature-dependent transition from wadsleyite to periclase plus garnet causes a density increase with increasing temperature at high temperatures as this continuous reaction has a highly negative Clapey-
ron slope. At regular MPTs, mantle rocks in the mid-MTZ experience the expected thermal expansion, hence, there is a density minimum just before the onset of the above-mentioned continuous reaction at about 1800 °C. Subsequent cooling of the hot MTZ and associated warming of the colder mantle rocks above decreases the density of both and the reservoir is finally emptied in two minor plume events also expressed in small peaks in magmatic productivity (Fig. 4e, 5a).

4.4. Three focus processes

In this paragraph, we document and discuss three critical processes in the model evolution more closely, namely (1) delamination of crust and convection in the upper mantle during the accumulation phase, (2) breaching of the LUMB by small plumes during the accumulation phase, and (3) the onset of the turnover phase. Close-ups from three respective stages presented in Fig. 3a,c,d are shown in Fig. 6 which features density instead of temperature.

Fig. 6a illustrates the upper mantle during the early accumulation phase after 81 Ma (Fig. 3a). The crust at this stage is about 40 kilometres thick. At downwelling sites, the cooled lower crust consists of garnet-rich mafic granulites with a density similar to the upper mantle. The real eclogite transition (plagioclase-out) leading to considerably higher densities happens actually just below the Moho in the dropping string of crust (Fig. 6a2). This is possible, because the lower crust is soft and flows horizontally to downwelling sites, which are several hundred kilometres apart from each other. A rather consistent melting depth of ca. 100 kilometres reflects almost uniform MPTs in the convecting system. Adjacent to downwelling sites, return-flow of mantle creates an increased production of crust. There is some variation in density contrast due to phase changes, but crust remains denser all the way to the bottom of the MTZ. Immediately below the Moho, the uppermost mantle is formed by a more or less stable layer of depleted mantle, which is less dense than ambient mantle and does not show partial melting (Fig. 6a3). This depleted mantle is only mixed into the upper mantle as it is dragged down together with the delaminating crust. Hence, it is the delamination of crust that breaks the compositional stratification of the upper mantle. The viscosity structure controlling model behaviour in the uppermost mantle is actually controversial and we address this issue in the discussion below and further in the supplemental material.

Fig. 6b zooms in on the lower MTZ and uppermost lower mantle during a local upwelling event after 600 Ma model time at ca. 4000 kilometres model location. This event is reflected in a period of higher magmatic production (Fig. 3c1; Fig. 5a). A layer with up to 50% of crust has formed all along the LUMB creating an intermediate-density layer that acts as a TMBL (Fig. 3c1; Fig. 5c). A pronounced thermal discontinuity across the LUMB has developed (Fig. 3c2) with the lower mantle having temperatures above the slope break of the LUMB phase transitions. A local upwelling has
breached the crust-rich layer and brought hot lower mantle into the transition zone and further up (Fig. 3c2). The local removal of the crust-rich layer has led to a rather stable upward stream since the LUMB is now depressed by a few kilometres at the site (Fig. 6b2) thus promoting upward flow. Such an event might in some circumstances trigger global mantle overturn, particularly if
there is a higher temperature offset across the TMBL, but in this case, the increased local production rate of crust and the associated increased accumulation (rain) of crust is sufficient to plaster over the hole in the TMBL. Note that the local increase of crustal production rate, i.e. the rapid local thickening of crust leads to a somewhat increased size of a couple of delaminating droplets, an effect that is later observed massively at a global scale during the overturn event. We cannot exclude that with different model parameters such an event would lead to run-away and global overturn. But in contrast to the runaway event described below this process causes a counteracting effect.

Fig. 6c shows the same domain 250 Ma model time later when global mantle overturn initiates at the same site. The crust-mantle mixture is somewhat thicker that model average because of the increased crustal rain during the earlier upwelling period. Some crust at the bottom of the layer has crossed the garnet-out transition and has become considerably denser than ambient mantle (Fig. 6c2). This material now starts to collapse into the lower mantle. This drags more crust through the phase transition and thus starts a runaway effect that leads to overturn. The initial instability (the first crust crossing the garnet-out phase transition) grows quite slowly for almost 20 Ma but then rapidly takes off initiating overturn over only another 30 Ma. The two previous episodes of lower-mantle upwelling (Fig. 5a) penetrated the TMBL but these bottom-up instabilities faded away as the crust-mantle mixture closed the breach. This instability keeps growing as there is no retarding effect.

5. Mantle overturn, caveats and implications for Earth

In the following, we first cover some numerical caveats, potential large uncertainties in boundary conditions and other simplifications. We then discuss implications for the evolution of Venus and Earth.

5.1. Uncertainties and limitations

Several parameters of the presented model are subject to discussion and/or large uncertainties. Properly addressing the thermal and material budget would require a model geometry and thermal boundary conditions suited for that purpose. Previous studies have addressed these issues on Venus using more comprehensive approaches e.g. spherical 3D or circular 2D (e.g. Armann and Tackley, 2012 and references therein). Thermal budgeting is not a primary target in this study and we instead employ a rectangular 2D model that for example overestimates the fraction of the lower mantle. The behaviour of our model after the first complete mantle overturn is severely affected by the lack of crustal melting or chemical homogenisation in our model, which develops into a rather stable compositional stratification. Including such mechanisms would be necessary for modelling multiple cycles of mantle overturns. Using our 2D rectangular model setup, on the other hand, permits tracking a high spatial and temporal resolution that properly resolves phase transitions and does not smear out short-wavelength compositional and density contrasts. We will address the importance of this when discussing the results.

If Venus has a mantle similar to Earth's in terms of composition in both major elements and heat-producing elements (HPEs), the mantle should have similar rheological properties, warm up at a similar rate and be subject to the same phase transitions. The concentration of heat producing elements used in this study is extrapolated back in time from Earth's mantle today. The silicate Earth may have experienced events that significantly altered the distribution of HPEs, namely the production of continental crust during the Archaean, where about or more than 40% of Earth's HPEs may be concentrated today (Korenaga, 2008). The strong differentiation of HPEs through formation of continental crust on Earth is hypothesized to require water (Moyen and Martin, 2012) and the presence of water on Venus in the past is speculative. We do not employ multi-stage melt evolution similar to Piccolo et al. (2019), but in doing so, their models produce significant volumes of mafic resites that delaminate in a behaviour similar to what is observed in our model and would likewise accumulate at the LUMB, potentially forming a TMBL.

The initial conditions for the mantle, and especially the lower mantle, are also subject to large uncertainties. In terms of composition we assume a primitive mantle composition throughout the mantle range (Supplements S4). The initial thermal conditions are adiabatic, which is appropriate for a convecting upper mantle. This is an idealized model fundamentally targeting the event of mantle overturn, and any variation in temperature across the LUMB will naturally evolve from the initial condition as radiogenic heat and convective cooling of the upper mantle compete.

Whether melts end up as extrusives or intrusives will have an impact on the thermal evolution of the crust. The simplification that all magma ends up as extrusives is an endmember, but likely a simplification that suppresses delamination rather than promoting it. Extrusive crust will cool more efficiently due to exposure to the atmosphere. During periods of low to moderate melt production, the crust will equilibrate and reach thermal steady-state, i.e. an approximately constant thermal gradient through the lithosphere. During the overturn when melt production is high, new crust is formed so rapidly that crust is buried before cooling, insulating it from the cooling effect of the atmosphere, enabling it to retain some of its heat. The net result is analogous to intrusive melt emplacement, where the intrusive crust will retain much of its initial heat, as it is thermally insulated from formation. Our model develops higher geothermal gradients in the crust during periods of high magmatic activity compared to lower magmatic activity. It is still colder, though, compared to if intrusive melt emplacement was included, as extrusives will experience some degree of cooling from the atmosphere before burial by younger crust.

The viscosity model employed in this study (see above) is very similar to ones that have been used in previous studies (e.g. Armann and Tackley, 2012) with the addition of dislocation creep that might become relevant at high stresses. As pointed out in Armann and Tackley (2012), there is a pronounced dependency of diffusion creep viscosities on grain size in the pre-exponential term. Grain size in the flow law is a more or less arbitrary parameter, as the range of petrologically reasonable grain sizes is larger than the range yielding reasonable viscosities. Additionally, most models including ours do not utilize strain weakening by grain-size reduction, something that would be expected in nature. The model presented here assumes a grain size of 0.925 mm in the upper mantle. For depths below 300 kilometres, predicted average viscosities in the upper mantle are slightly lower than but generally in line with ones inferred geophysically by gravity-topography relations (Rolf et al., 2018; Supplementary Figs. SF3; SF4). Supplementary Fig. SF4 presents a model using a pre-exponential term corresponding to average viscosities slightly above the ones proposed in Rolf et al. (2018) in the mentioned depth range. This higher viscosity model shows an evolution very similar to the one presented here (Supplements S5 and Supplementary Figs. SF4-SF5). Experimentally inferred flow laws, however, predict very thin lithosphere and average viscosities down to $3 \times 10^{19}$ Pa s below the lithosphere for relatively hot adiabats - a scenario not favoured by gravity-topography based estimations (Rolf et al., 2018; Supplementary Fig. SF3). Certainly, the relatively weak asthenosphere promotes delamination of crust in our model. This process might be more complex, but in a world where the crust thickens progressively by magmatic activity and turns into eclogite at its base,
delamination is bound to happen. Some recent models associate delamination of eclogitic crust with thermal erosion of the lithosphere by upwelling mantle and further weakening of the mantle by partial melting, on early Earth (Johnson et al., 2014) and on Venus (Gülich et al., 2020).

5.2. Implications for Venus and Earth

The presented model results suggest for a Venus-like scenario that (1) delaminated pieces of crust can compact into a basalt barrier with >50% crust in the lower MTZ, (2) this TMBL can support temperature differences of more than 200 °C and pressure-temperature conditions may beyond the slope-break of the phase change in the uppermost lower mantle, (3) the collapse of the basalt barrier and global mantle overturn are phase-change controlled and initiated by crust crossing the garnet-out transition, (4) global mantle overturn happens in about 50 millions of years as dramatically increased crustal production rates associated with upwelling of superheated lower mantle accelerate the collapse of the stratification at the LUMB. Previous studies have addressed global mantle evolution on Venus using thermomechanical modelling (e.g. Papuc and Davies, 2012; Armann and Tackley, 2012). These models consider the entrenchment of crust into the mantle and also its enrichment in the density inversion channel below the MTZ and are partly more comprehensive in terms of model space and model time. Our model, however, shows very high resolution in space and time. It tracks, for example, coherent pieces of delaminated crust through the entire model time and can thus handle sinking and compaction in the MTZ more realistically. The preservation of sharp material boundaries together with small grid elements allows also for more efficient tracking of stress concentrations and the associated activation of dislocation creep in local shear zones. The behaviour of large-scale tectonics is likely controlled by exploitation of weaknesses and strain localization. Inclusion of strain-induced grain-size reduction would further promote their behaviour. Compared to previous studies on the basalt barrier (e.g. Ogawa, 2003; Davies, 2008; Papuc and Davies, 2012; Armann and Tackley, 2012), our model predicts much more dramatic behaviour: The high resolution allows for closer packing of delaminated crustal nodules at the LUMB. Overturn happens at much shorter time scale, affects the entire model and is associated with magma-production rates some orders of magnitude higher than the background rate. For most of the model time, heat flow at the surface is much lower than radiogenic heat production (the model heats) and that accumulated heat is lost during overturn. Overturn is initiated by runaway growth of a small, but pronounced weakness, the densification of basalt when crossing the garnet out reaction. The capacity of tracking such weaknesses causes overturn in the model here to happen at considerably lower average temperatures than other models. We predict Venus’ interior to be on average much colder than other models do, i.e. not much hotter than Earth after overturn.

Venus has been considered a case study on how Earth-like planets without subduction-driven plate tectonics may evolve over time, and may therefore also provide insight into the processes Earth’s mantle may have experienced during the Archaean (e.g. Hansen, 2018). Komatitites on Earth are generally viewed to be a product of partial melting of peridotite at MPFs in excess of 1600 °C for protracted periods (e.g. Nisbet and Walker, 1982) and can for that reason not be generated from the upper mantle during layered convection. Bédard (2018) suggests the conditions for producing komatitites would be present during protracted periods of lower mantle upwelling such as mantle overturns. The occurrence of komatitites on Earth is also restricted to the Archaean, suggesting that Earth has since then switched into a convective mode that limits the extend of large-scale mid-mantle thermal boundary layer, where these hot reservoirs can form. While we propose the global build-up of this layer through delamination is an episodic feature for Venus that may have happened several times throughout its existence, it has likely not happened on Earth since the onset of plate tectonics, as subduction will break a global basalt barrier, and induce whole mantle convection (Davies, 2008) or perhaps rather “leaky layered convection” (e.g. Tackley, 2008). The notion of leaky layered convection reconciles a world with slabs subducting into the lower mantle while simultaneously permitting the existence of crust-enriched patches at the LUMB, far away from subduction zones, acting as TMBLs. Typically, this is also where LIPs form. LIPs might be a result of lower mantle patches undergoing heating for protracted periods below TMBLs. Because primitive or enriched mantle has a higher concentration of heat producing elements, a domain of primitive mantle in the lower mantle will heat at a higher rate than its surroundings, and eventually thermal expansion will outpace the depletion effect on density for the surrounding mantle. These pockets of superheated mantle could persist until today. As they rise they pick up remnants of crust at the LUMB before ascending through the upper mantle and melt due to decompression (e.g. Petersen et al., 2018). This provides a setting for LIP formation in a scenario alternative to upwelling rooted at the core-mantle boundary.

The complex architecture of the mantle at the late stage of the model suggests that Earth’s mantle could have been quite heterogeneous already before the onset of plate tectonics. At any point through the model evolution, the upper mantle remains more depleted than the lower mantle, which is likely a feature that has survived until today. While Venus may have undergone several cycles of overturns and subsequent quiescence, modelling this would require a form of resetting; i.e. refertilizing the mantle and/or redistribution of heat producing elements, mechanisms that are poorly understood and are not incorporated into the presented model. For Earth, the implications of this model is that significant heterogeneity can be introduced to the mantle by just a single overturn. After the onset of plate tectonics, radiogenic heat is vented out continuously at mid-oceanic ridges. Simultaneously, a source of cooling is introduced for the lower mantle, thus preventing future overturns from occurring. For Venus, the concentration of heat producing elements will decay and the time span between overturns increase until they cease completely.

Finally, we point to an exciting consequence of mantle overturn destroying a basalt barrier: A planet shrinks during an overturn by a few kilometers in diameter. An overturn as the one in the presented model transports crust from the density inversion channel below the LUMB deeper into the lower mantle. Garnet in the crust transforms into perovskite and periclase at around 27 GPa, increasing the density of the crustal assemblage significantly. This crust is replaced by upwelling peridotite that is present in the lower-mantle assemblage and remains so below the LUMB. In other words, during the overturn, crust in the density inversion channel undergoes dramatic reaction-related compression, whereas the replacing mantle does not undergo a corresponding expansion. Hence, such an event decreases the volume of the planet. A rough estimate of the change in diameter can be calculated by assuming that ~50% of the thermomechanical boundary layer in the uppermost lower mantle consist of crust. The layer of crust-enriched mixture below the LUMB is estimated to range from ~740-780 km in Venus gravity (e.g. Fig. 6), corresponding to ~660-700 km in Earth gravity, and the density increases by around 10%, leading to a total change in diameter of about 3 km. If crust gets refertilized into the mantle and becomes a part of the convective system, the volume of Venus is likely to grow over hundreds of millions of years, only to collapse rapidly again during the relatively short time scale of an overturn/resurfacing event. A planet with periodic
overturns would experience repeated contraction and expansion ("breathing").

CRediT authorship contribution statement

A.L. Vesterholt: Conceptualization, Data curation, Formal analysis, Methodology, Visualization, Writing – original draft, Writing – review & editing. K.D. Petersen: Software, Writing – review & editing. T.J. Nagel: Funding acquisition, Resources, Supervision, Writing – review & editing.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Acknowledgement

The authors thank Paul Tackley and an anonymous reviewer for careful and comprehensive reviews.

Appendix A. Supplementary material

Supplementary material related to this article can be found online at https://doi.org/10.1016/j.epsl.2021.117047.

References