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Mantle mixing and continental breakup magmatism

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Abstract

The frequent formation of large igneous provinces during the opening of the Atlantic Ocean is a surface manifestation of the thermal and chemical state of convecting mantle beneath the supercontinent Pangea. Recent geochemical and geophysical findings from the North Atlantic igneous province all point to the significant role of incomplete mantle mixing in igneous petrogenesis. On the basis of a whole-mantle convection model with chemical tracers, I demonstrate that sublithospheric convection driven by surface cooling can bring up dense fertile mantle without a thermal anomaly. When small-scale convection in the upper mantle breaks down into the lower mantle, strong counter upwelling takes place, entraining a large amount of dense crustal fragments accumulated at the base of the mantle transition zone. This multi-scale mantle mixing could potentially explain a variety of hotspot phenomenology as well as the formation of both volcanic and non-volcanic rifted margins, with a spatially and temporally varying distribution of fertile mantle.

Keywords: large igneous provinces; continental rifting; mantle convection; subducted crust

1. Introduction

The disintegration of the supercontinent Pangea was frequently marked by a burst of extensive and voluminous magmatism at rifting axes (Fig. 1) [1,2], including the US East Coast igneous province (the Central Atlantic; formed at ∼200 Ma), the Parana and Etendeka flood basalt provinces (the South Atlantic; ∼120 Ma), and the Greenland and British Tertiary igneous provinces (the North Atlantic; ∼56 Ma). The most popular hypothesis for such an unusual expression of terrestrial magmatism is the arrival of a mantle plume [3,4]. A starting mantle plume rising from a hot boundary layer, such as the core–mantle boundary, is characterized by a large mushroom-like head followed by a narrow conduit [5,6]. The impact of a plume head can nicely explain both the vast spatial extent (by the dimension of a plume head) and the large melt volume (by the high temperature of a plume material) of continental breakup magmatism. A succeeding plume tail can account for lingering smaller-scale igneous activities such as the Iceland hotspot in the North Atlantic. In essence, the plume impact hypothesis is a theory that attributes anomalous mantle melting to thermal anomalies in convecting mantle. For the melting of such hotter-than-normal mantle, we can make various geophysical and geochemical predictions, which can be compared with observations to validate the hypothesis.

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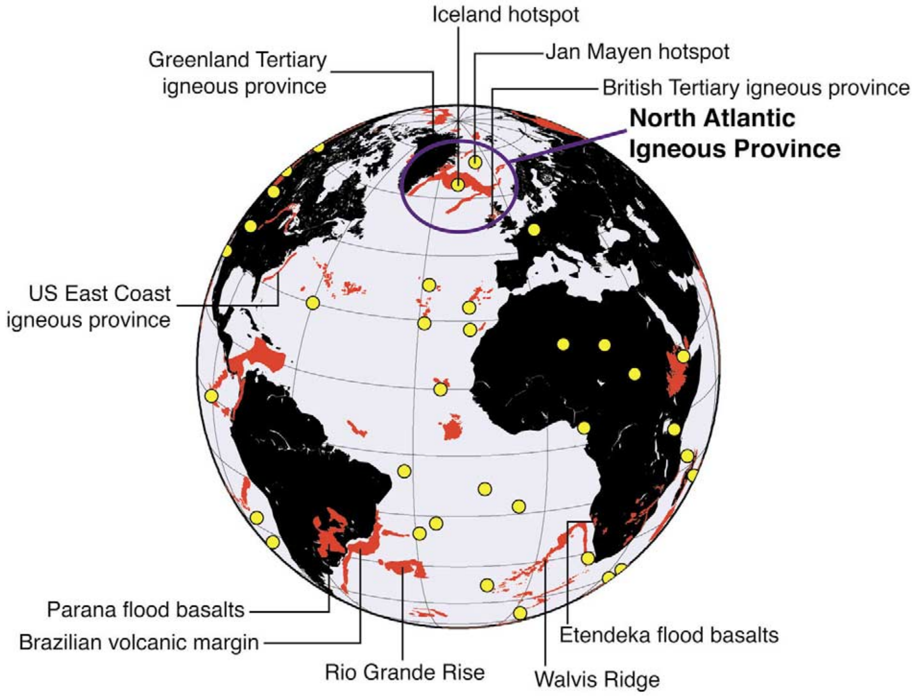

Fig. 1. Most Atlantic large igneous provinces are related to the dispersal of the supercontinent Pangea initiated ~200 million years ago. Red areas denote the distribution of large igneous provinces that have been recognized so far, and yellow circles denote currently active hotspots (after [2,55]).

A recent discovery of major element heterogeneity in the source mantle for the North Atlantic igneous province (Fig. 1), however, sharply conflicts with the significance of anomalously hot mantle [7]. The source mantle for this province appears to be significantly more fertile (i.e., more prone to melt) than normal pyrolitic mantle [8], probably resulting from the entrainment of recycled oceanic crust. Isotope geochemistry suggests that the original formation age of the recycled crust must be relatively young, in the range of 0.5–0.7 Ga [7,9,10]. This estimated age is consistent with the closure of the Iapetus (proto-Atlantic) Ocean. More surprising is the volume fraction of the recycled crust, which is estimated to be as much as 20–30% of the source mantle [7]. Because the mixed-in crustal component should almost completely melt if brought up to the surface (by definition), in addition to the decompressional melting of the surrounding mantle matrix, the upwelling of this fertile mantle can result in a very high degree of melting even if the mantle is not notably hotter than normal. Indeed, the temperature of estimated primary mantle melts for this igneous province is almost indistinguishable from that of normal mid-ocean ridge basalts [7]. Furthermore, recent mantle tomography suggests that a low-velocity anomaly beneath the Iceland hotspot may be totally confined to the upper mantle: both S-wave [11] and P-wave [12] global tomography models clearly show that a strong low-velocity anomaly beneath Iceland ends sharply at the 660-km discontinuity (see also [13] for a critical review of other tomographic studies such as [14]).

2. Dynamical dilemma of fertile mantle

It is difficult to reconcile these geochemical and geophysical findings with the conventional notion of a thermal plume rising from the core–mantle boundary. The short-term recycling of subducted oceanic crust prefers upper-mantle dynamics to a deep-seated plume. Segregation of oceanic crust from subducting slab at the 660-km discontinuity may provide a starting point for such recycling [15] (Fig. 2). The crustal component of a subducting slab has higher viscosity than the lithospheric component as well as ambient mantle because of its higher garnet content, and it also becomes strongly buoyant below the 660-km discontinuity. Crustal segregation has been suggested to be physically plausible if these viscosity and density contrasts are taken into account [16,17]. One may claim that Pangea experienced a significant northward motion since the Caledonian orogeny and that this motion destroyed a direct link between subducted crust and the suture. It should be noted, however, that it is impossible to resolve the *relative* motion of a supercontinent with respect to the underlying mantle from paleomagnetic data because of the possibility of true polar wander. The geology of Pangea's leading edge suggests that true polar wander has most likely occurred during the Permo-Triassic [18], in which case the entire mantle rotated as a whole and no significant shear is expected between Pangea and sublithospheric mantle. Thus, the northward migration of the supercontinent does not pose any impediment to short-term recycling.

A crucial issue is what happens next. Because of the basalt–eclogite phase transformation, subducted crust is denser than normal pyrolitic mantle above the 660-km discontinuity [15]. It appears that segregated crust would continually lie at the discontinuity, which represents the level of neutral buoyancy, unless it is entrained by upwelling plumes from the core–mantle boundary [19]. Such a plume entrainment scenario is, however, inconsistent with mantle tomography, because the seismically imaged 'plume' (i.e., low-velocity

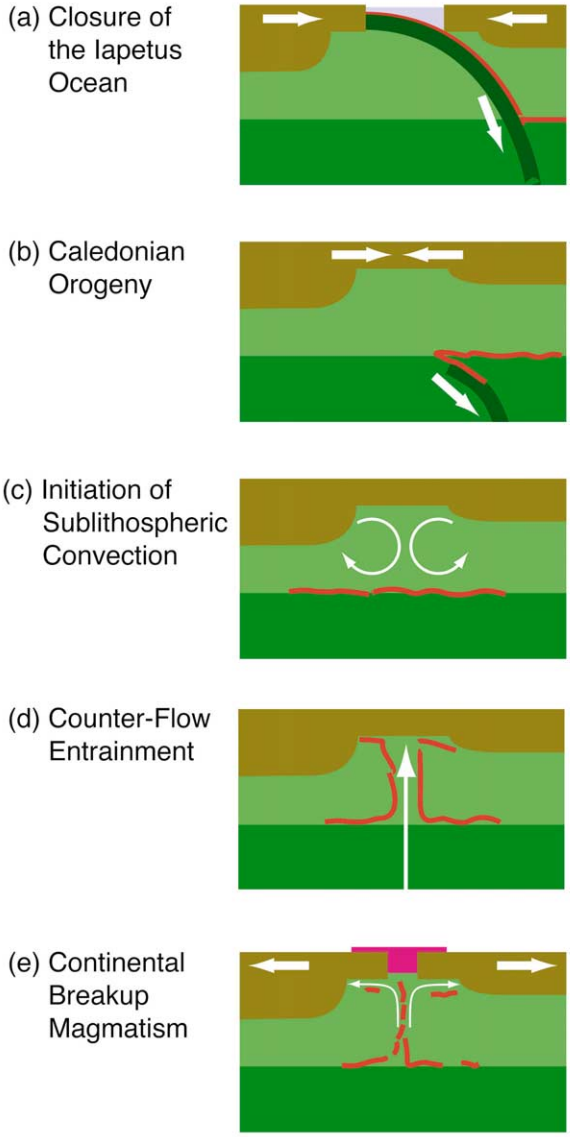

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Fig. 2. Schematic illustration of a possible scenario for the formation of a large igneous province during the breakup of Pangea. (a) The closure of the Iapetus Ocean took place prior to ~500 Ma, and during this period, a layer of oceanic crust was segregated from subducting slab at the base of the mantle transition zone. (b) The Iapetus Ocean diminished, and the Caledonian Orogeny took place during ~500–400 Ma. (c) The supercontinent Pangea was assembled during ~320–170 Ma, and sublithospheric convection was initiated beneath suture zones. Though convective motion is depicted here two-dimensionally, it is most likely three-dimensional, with strong convection confined in the out-of-plane direction [30,31]. (d) Sublithospheric convection was confined in the upper mantle for a while, but eventually this layering broke down, and mass-conserving counter upwelling entrained crustal components to the base of lithosphere. (e) Plate-driven flow during continental rifting (e.g., ~56 Ma for the North Atlantic) further assisted this entrainment process, and the circulation of fertile mantle to shallow depths resulted in a high degree of melting, which is observed as flood basalt volcanism. Mantle dynamics corresponding to stages c and d is considered in detail in this paper.

anomalies beneath Iceland) does not extend into the lower mantle [11,12]. It may be possible to tweak the entrainment scenario to be consistent with seismological observations, by assuming that the Iceland plume is actually dying out and that its tail end happens to be right at the base of the transition zone, though it would be quite a coincidence. The melting of hot plume mantle is also irreconcilable with surface lava geochemistry [7] and the crustal structure of the Greenland Tertiary igneous province [20–22]. How can we, then, bring up the dense crust without a thermal anomaly? This positive density anomaly associated with fertile mantle is a long-standing problem in plume dynamics [23] though it has usually been neglected in geodynamic modeling [24,25]. Here I propose that sublithospheric convection driven by surface cooling can naturally resolve this dynamical dilemma.

The dynamics of sublithospheric convection has been studied by a number of geophysicists in the last three decades or so [26–29], but its importance in transient geological phenomena has been realized only recently [30,31]. For an initial-value problem like continental breakup magmatism, in which surface divergence evolves from null to some finite value, it is essential to understand the most likely initial state. Though isothermal and static asthenosphere has long been implicitly assumed as the standard initial condition [1,3,4], such static mantle is physically unrealistic. The natural state of sublithospheric mantle beneath continental lithosphere is small-scale convection driven by cooling from above. The three-dimensional pattern of such sublithospheric convection is probably modulated by variable lithosphere thickness [30,31], being most intense beneath pre-rifting axes where lithosphere is expected to be thinner than old cratonic lithosphere. Another important point that has also recently been raised is that the spatial scale of sublithospheric convection is likely to evolve from the upper-mantle scale to the whole-mantle scale about 100 million years after the onset of convection [32,33]. Though an endothermic phase transition with a Clapeyron slope of -2 to -3 MPa/K [34] and $\sim$30-fold viscosity increase [35–37] are expected at the base of the mantle transition zone, both of them could impede material flux into the lower mantle *only temporarily*, even for small-scale temperature anomalies associated with sublithospheric convection. The breakdown of temporally layered, small-scale convection is characterized by large-scale downwelling, like a well-known 'mantle avalanche' [38–40], and, as required by the conservation of mass, upwelling from the lower mantle to the upper mantle. I suggest that this counter upwelling can be strong enough to bring up dense eclogite crust from the 660-km discontinuity. The liquidus of the crustal component is lower than the solidus of the surrounding mantle matrix [41]; adiabatically upwelling mantle with a potential temperature of 1623 K, for example, intersects the solidus and liquidus of anhydrous mid-ocean ridge basalt at $\sim$150 km and $\sim$70 km, respectively. Once brought up to shallow depths, therefore, the crustal component starts to melt, and resulting melt retention buoyancy can further enhance upwelling.

Some thermal plume models advocate the arrival of a starting plume head beneath continental lithosphere as the driving force for the rifting of a supercontinent [3,4], thereby providing a single 'plume origin' explanation for both continental breakup and breakup magmatism. Available geological records suggest, however, that the dispersal of Pangea was mostly controlled by changes in plate boundary forces with the minor role of mantle plumes (if any) [42]. In the working hypothesis depicted in Fig. 2, therefore, counter upwelling and continental breakup do not have to be synchronous. Counter upwelling could continue for more than tens of millions of years (in a manner similar to so-called plume incubation models). Once rifting is initiated by (mostly) far-field stress, plate-driven flow would further assist the entrainment by upwelling, and the circulation of fertile mantle to shallow depths results in extensive melting and thus the formation of a flood basalt province.

3. Numerical models

To test the feasibility of this counter-flow entrainment model, I investigate the potential evolution of mantle beneath a pre-rifting axis of a supercontinent, with finite element convection modeling [43]. My whole-mantle model is 5800 km wide and 2900 km deep, which is discretized by 256×128 2-D variable-size quadrilateral elements. Note that this model is set up to study mantle dynamics *along* a pre-rifting axis, whereas cartoons in Fig. 2 depict mantle evolution *across* the axis. All boundaries are free-slip. The (potential) temperatures of the top and bottom boundaries are fixed to 273 K and 1623 K, respectively. The vertical boundaries are adiabatic. No internal heating is considered because I am interested in transient dynamics driven by surface cooling. Internal mantle temperature is initially set to 1623 K plus 0.1% random perturbations. Random perturbations are necessary to trigger the convective instability of the top thermal boundary layer, and after the onset of convection, the nearly isothermal initial condition is quickly destroyed by small-scale convection. Thus, modeling results

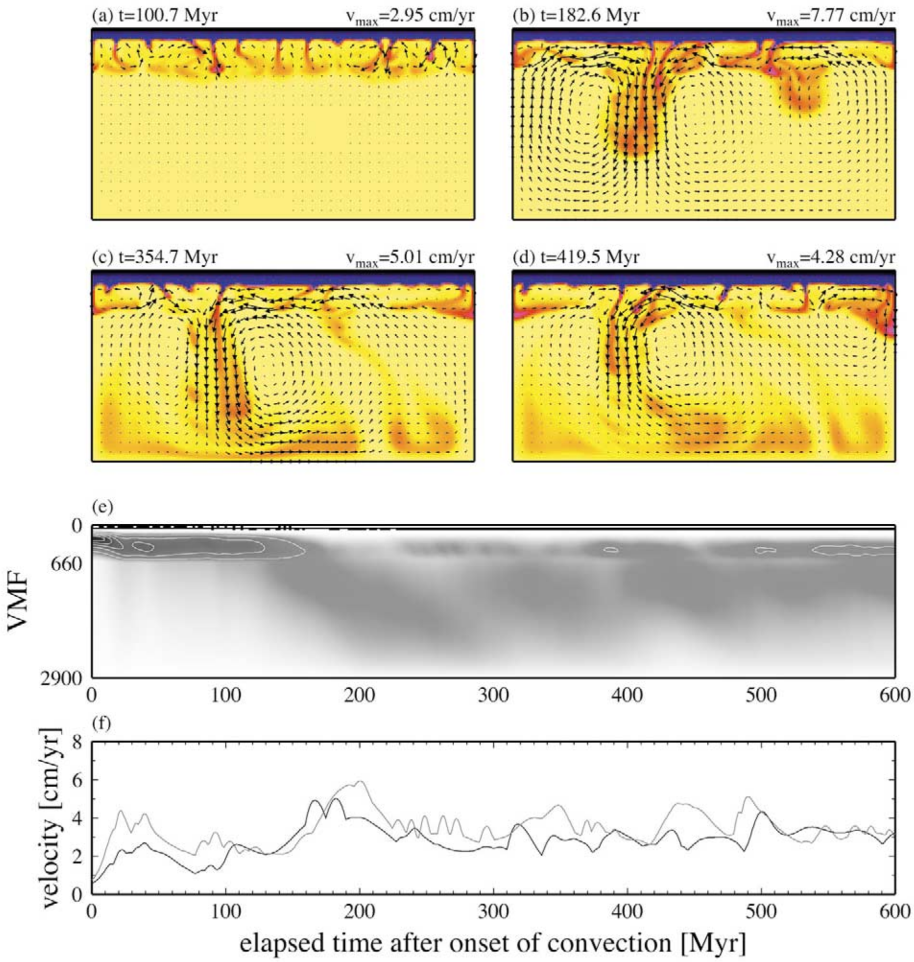

Fig. 3. Numerical solutions for isochemical whole-mantle transient cooling. Snapshots of temperature and velocity fields are shown at (a) t = 100.7 Myr, (b) t = 182.6 Myr, (c) t = 354.7 Myr, and (d) t = 419.5 Myr. Velocity arrows are normalized by maximum velocity, which is denoted at every snapshot. Elapsed time shown here is measured from the onset of convection, after which lithosphere is thicker than ∼100 km. (e) Vertical mass flux diagnostic with contour interval of 2.0. It can be seen that most of the vertical mass flux is confined above 660 km for the first 150 Myr. (f) Maximum upwelling (solid) and downwelling (dotted) velocities observed in the model domain.

do not strongly depend on this particular initial condition. The state of sublithospheric mantle beneath a supercontinent could have been more dynamic, including hot plumes rising from the core–mantle boundary. The role of such deep-mantle plumes is, however, quite uncertain as explained earlier, and at least for the North Atlantic igneous province, such a complication in the initial condition is not demanded by available geophysical and geochemical observations.

I employ the following form of temperature- and depth-dependent viscosity:

$$\mu(T,z) = \mu_0(z)\exp\left(\frac{E}{RT} - \frac{E}{RT_0}\right) \quad (1)$$

where E is activation energy, R is the universal gas constant, and T_0 is the reference temperature of 1623 K. The reference viscosity profile, $\mu_0(z)$, is set at 10^{20} Pa s for 0–660 km depth and 3×10^{21} Pa s for 660–2900 km depth. With a reference density of 3300 kg/m^3, a thermal expansion coefficient of 3×10^{-5}/K, gravitational acceleration of 9.8 m/s, and thermal diffusivity of 10^{-6} m^2/s, the 'bottom' Rayleigh number, which is defined with the viscosity of the lower mantle, is approximately 10^7. A relatively low activation energy of 120 kJ/mol is used based on [32,44]. The combination of higher activation energy with lower reference viscosity, however, can produce similar kinematics (i.e., the same quasi-steady-state thickness of lithosphere) with faster convection velocity and thus a shorter time scale [32,33]. The endothermic phase transition at 660 km depth is assigned with a Clapeyron slope of −2 MPa/K and a density jump of 10%.

First, an isochemical case is shown in Fig. 3. To measure the time scale of layering breakdown, vertical mass flux [45] is monitored, which is defined as:

$$R(z) = [\int w^2 \mathrm{d}x]^{\frac{1}{2}} / \int [\int w^2 \mathrm{d}x]^{\frac{1}{2}} \mathrm{d}z \quad (2)$$

where w denotes the vertical component of velocity. After the onset of convection, sublithospheric convection is confined in the upper mantle for about 150 million years. The actual layering period beneath Pangea was probably shorter than this, because 'initial' sublithospheric mantle (i.e., when Pangea was assembled) was unlikely in a static and isothermal state as assumed in the model. During this period, cold thermal anomalies above 660 km depth gradually develop their horizontal scale, to the point where phase boundary buoyancy can no longer support them. Convection velocity dramatically increases upon the sudden breakdown of the initially layered state, from ∼3 cm/yr to ∼7 cm/yr (Fig. 3a,b). Strong upwelling of lower-mantle material across the 660-km depth is observed. Because the bulk of the lower mantle is sequestered from surface cooling during the layered stage, the advent of large-scale circulation results in replenishing the upper mantle with uncooled material.

Next, I elaborate my model by incorporating crustal fragments initially floating above the 660-km depth (Fig. 4). Following [46], tracers are advected with a fourth-order Runge–Kutta scheme. Position error for tracer advection along one circuit in a steady-state convection has been verified to be of order 10^{-5}. Tracer distribution is converted to a chemical buoyancy field using finite element shape functions [47]. The intrinsic density contrast of the crustal component with respect to the ambient mantle is specified as follows: 0% for 0–60 km depth, −15% to +3% for 60–200 km depth, +3% for 200–660 km depth, and −6% for 660–2900 km. For 0–60 km depth, melt retention buoyancy is not expected because the surrounding mantle matrix starts to form a porous melt network at ∼60 km [48,49] (model results are, however, insensitive to this specification because upwelling is limited by ∼100-km-thick lithosphere). For 60–200 km depth, I assume that melt produced within entrained crust is completely retained. As subducted crust is unlikely to be completely dry, the initial depth of melting is probably greater than 150 km. Here I choose 200 km as a conservative estimate. The temperature of the entrained component is not significantly lowered by the latent heat of fusion, because of heat exchange with the surrounding matrix [50], thus complete melting is expected below 70 km. Melt density is based on the calculation by [51]. Relative buoyancy for 200–660 km depth is from [15]. For rapid downwelling associated with the breakdown of layered convection,

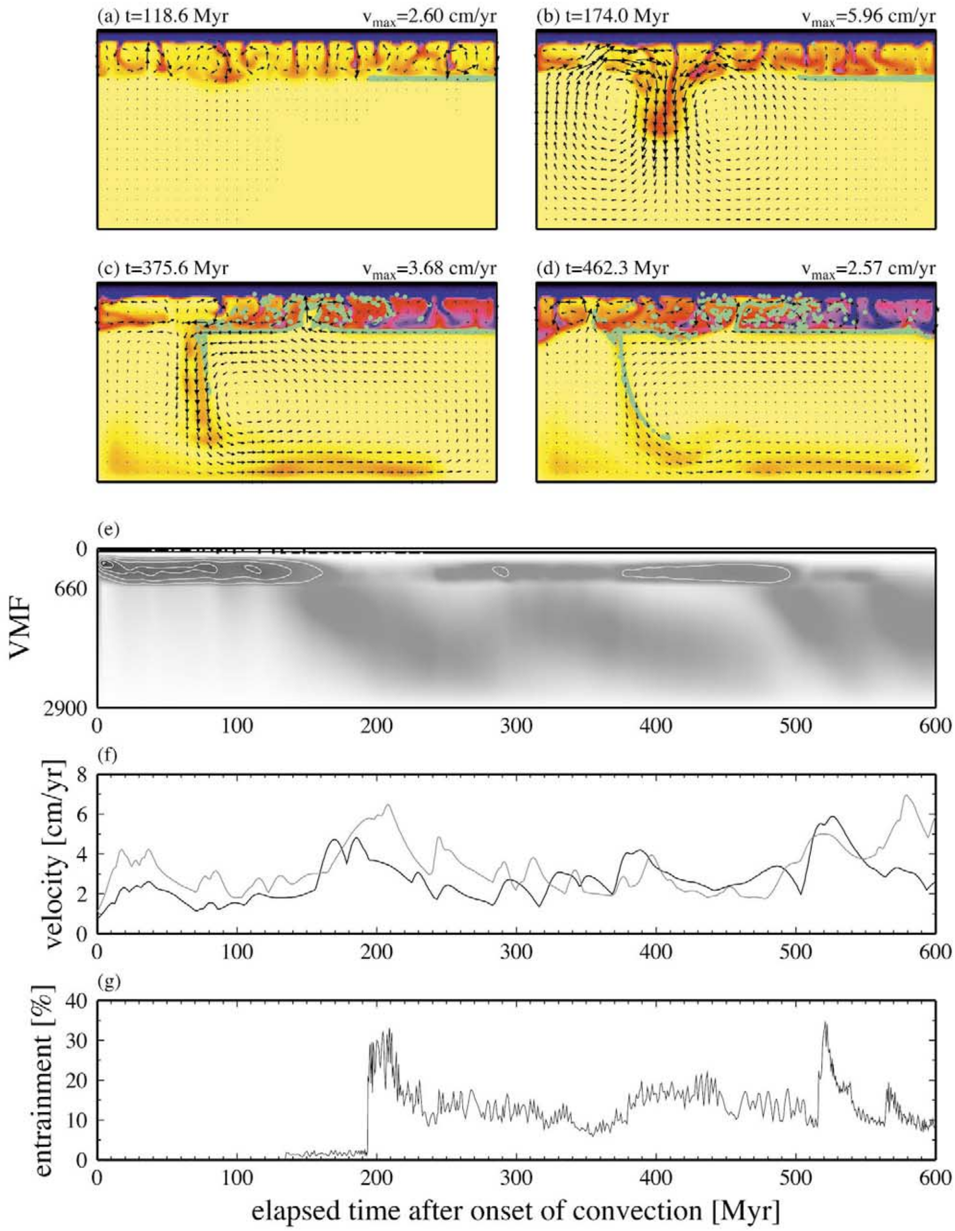

Fig. 4. Numerical solutions for whole-mantle cooling with crustal fragments at the 660-km discontinuity. Snapshots are shown at (a) $t=118.6$ Myr, (b) $t=174.0$ Myr, (c) $t=375.6$ Myr, and (d) $t=462.3$ Myr. Tracers for crustal fragments are denoted by green circles. (e) Vertical mass flux diagnostic. (f) Maximum upwelling (solid) and downwelling velocities (dotted). (g) Maximum fraction of crustal fragments entrained into the upper mantle shallower than 410 km.

transformation kinetics is likely to be important, so I use the results of [52] to specify crustal buoyancy in the lower mantle.

Since it would be unrealistic to expect perfect crustal segregation over the entire suture zone, chemical tracers for crustal material are initially distributed over only the right one-quarter of the model width and within a 30-km-thick layer.

About 11 000 chemical tracers are used, and elements in the initial layer have ∼70 tracers each. I would like to point out a semantic issue related to the evolution of this 'layer'. Accurately tracking the deformation of such a thin layer in a convection system is known to be numerically challenging, for which the chemical tracer method offers only a crude approximation. My intention is, however, not to model the evolution of a homogeneous layer. In reality, the crustal layer is expected to be composed of small fragments as a result of crustal segregation processes, and each of the chemical tracers can be thought to represent such a crustal fragment. In this case, the chemical tracer method is not an approximation; it rather fortuitously achieves nearly one-to-one correspondence between the model and the realistic complexity. A convergence test with the number of tracers varying from $\sim 10^3$ to $\sim 10^5$ suggests that modeling with 11 000 tracers should be able to capture the gross characteristics of counter-flow entrainment.

Because of its positive buoyancy below the 660-km depth, the presence of this crustal layer stabilizes the layered state (Fig. 4b). Large-scale downwelling, therefore, takes place first where the phase boundary is not covered by crustal fragments, and as a result, the region underlain by dense eclogite becomes the most likely locus to be exposed to counter upwelling from the lower mantle. The maximum entrainment of crustal component is observed to be ∼15–30% (Fig. 4g). After the layering breakdown at ∼150 Myr, partial layering gradually develops owing to spreading of crustal fragments along the 660-km discontinuity (Fig. 4c,d), which again breaks down at ∼500 Myr (Fig. 4e). This intermittent breakdown is similar in nature to what has been observed in previous whole-mantle convection models with an endothermic phase boundary [38,39,53]. For these recurrent breakdown events, crude correlation among vertical mass flux, maximum vertical velocity, and entrainment can be observed (Fig. 4e–g). I note that it is difficult to predict the timing of counter upwelling. The initial (dynamic) state of sublithospheric mantle corresponding to Fig. 2c is unknown, and I expect that it can be anywhere in the first 100 million years in my convection model. The onset of convection is known to have ∼10% standard deviation owing to the random nature of thermal noise [54]. The breakdown of layered convection is more random than the onset of convection (> 50% standard deviation) [33], and so is the mass-conserving counter flow. This randomness is further enhanced by a large degree of freedom in the initial distribution of crustal fragments. At this point, therefore, the best quantification would be that the counter-flow entrainment becomes plausible within a few hundred million years after the aggregation of a supercontinent. The convection model presented here is designed to study the pre-rifting stage. Once continental rifting is initiated, plate-driven flow can assist this entrainment process further, and excess magmatism will last until the crustal inventory is drained up.

4. Discussion and conclusion

My convection model is relatively simple, i.e., a cooling whole-mantle system with segregated crust accumulated at the 660-km discontinuity. Yet, this incomplete state of mantle mixing can induce quite complex dynamics, with profound implications for the origin of anomalous magmatism. Voluminous breakup magmatism may simply reflect compositional heterogeneity in the upper mantle, not excess mantle temperature associated with a deep thermal boundary layer. It is less problematic, for example, to account for the proximity of the Iceland and the Jan Mayen hotspots, which are only ∼800 km apart [55], by upper-mantle heterogeneity than two closely located plumes rising from the depth of ∼3000 km. The mobility of the Iceland hotspot with respect to other Indo-Atlantic hotspots [56] is also no more surprising because counter-flow entrainment is not strongly anchored in the mantle. In addition, the amount of crustal fragments embedded in the mantle matrix is unlikely to be constant over time, considering the random nature of entrainment, so it is natural to expect time-varying melt productivity, which we may observe today as the V-shaped Reykjanes Ridge [57–59].

Though my model may be best characterized by the chemically heterogeneous upper mantle, it is not strictly an upper-mantle model. The lower mantle is also involved; counter upwelling is supplied from the lower mantle. This may explain why the Iceland hotspot has high ^{3}He/^{4}He ratios, which are usually considered to be a signature of relatively undegassed deep-mantle reservoir [60,61], whereas mantle tomography exhibits low-velocity anomalies only in the upper mantle. Sublithospheric mantle is likely to be in a fully dynamic state with various spatial scales, which is clearly seen even in the isochemical case (Fig. 3). Rifting above downwelling mantle may give rise to a non-volcanic rifted margin like the Newfoundland Basin and the Iberia margin.

Because entrained crustal fragments start to melt deeper than surrounding mantle, melt retention buoyancy becomes important. This contrasts with the melting of homogeneous mantle beneath normal mid-ocean ridges, for which retention buoyancy is probably small owing to the formation of a porous melt network [62,63]. Thus, a recent seismic study suggesting buoyancy-driven mantle upwelling beneath Iceland [64] is consistent with the melting of chemically heterogeneous mantle. Rapid lateral spreading in response to the basal topography of lithosphere is also possible for upwelling mantle with high melt retention buoyancy; such dynamics is not limited to plume–lithosphere interaction [65]. Eventually, the surrounding mantle matrix starts to melt at shallow depths ($< \sim 60$ km), and melt from the entrained crust will escape into the porous network developed in the matrix. Understanding the resultant chemical interaction between the melt and the matrix will be necessary to correctly predict the characteristics of the final melt product, especially in terms of the sodium content [66].

The thermal plume hypothesis implicitly assumes that the mantle is homogeneous in terms of major element composition and attributes excess magmatism to elevated mantle temperature only. This assumption is in stark contrast to well-accepted trace element and isotope heterogeneities in the mantle. One must realize that, even if we had a perfectly homogeneous mantle at some point, it is not obvious how to maintain such homogeneity because plate tectonics continuously introduces heterogeneity by chemical differentiation at mid-ocean ridges. It is still unresolved how quickly and thoroughly undepleted mantle is recovered from the mixture of oceanic crust and depleted lithosphere. Remixing could well be incomplete sometime somewhere in a complex convection system. Beyond the North Atlantic igneous province, however, the number of data available to distinguish between the thermal and chemical origins is surprisingly small. Though it would be extreme to suppose that all large igneous provinces are caused by chemical anomalies, it is equally unlikely that all of them have a purely thermal origin. An interdisciplinary approach focused on the product of mantle melting, with igneous geochemistry [7] and deep-crustal seismology [20,22], is a promising direction, being able to provide observational constraints on the thermal, chemical, and dynamic state of convecting mantle in space and time.

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