Small-scale convection in the Earth's mantle

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Abstract

Small-scale convection (SSC) in the Earth's mantle contributes to intraplate deformation, heat flow and volcanism. In this review, I give an overview over the causes and effects of SSC. SSC is a boundary layer instability that is driven by a density inversion, and mostly restricted to low-viscosity layers such as the asthenosphere. The density inversion that supports SSC can be related to thermal and/or chemical stratification. SSC is thought to occur beneath mature oceanic basins to restrict their subsidence and stabilize geothermal heat flux. The onset of SSC is preferentially triggered near lateral heterogeneity such as fracture zones or other steps in lithospheric thickness. SSC may also occur beneath continents, and seismic evidence for related perturbations has indeed been found. Both in continental and oceanic environments, SSC can cause dynamic topography, intraplate deformation, as well as melting of mantle rocks. Mantle melting can boost SSC through a positive-feedback mechanism, and most importantly, feeds intraplate volcanism. While plate tectonics and related natural hazards are mostly caused by large-scale whole-mantle circulation, intraplate geologic activity may be sustained by SSC in the upper mantle.

Keywords

Small-scale convection, edge-driven convection, convection, mantle, buoyant decompression melting, melting, volcanism, heat flow, subsidence, seafloor flattening, uplift, plume-lithosphere interaction

1. Introduction

The transport of primordial and radiogenic heat from the deep Earth's interior to the surface is mostly accomplished by mantle convection. Large-scale mantle flow drives the motion of tectonic plates and is thus responsible for first-order geological activity on Earth − such as mountain building, continental breakup, mid-ocean ridge and subduction-related volcanism – as well as the related natural hazards. Smaller scales of mantle flow, mostly occurring in the low-viscosity upper mantle, are superimposed on these large scales of whole-mantle circulation, and sustain deformation within tectonic plates. Such deformation can be e.g. expressed as dynamic uplift and extension (*Göğüş and Pysklywec*, 2008b), or subsidence and sedimentation (*Petersen et al.*, 2010). Small-scale convection (SSC) upwellings may further feed mantle melting and intraplate volcanism, thus e.g. supporting seamount and ocean-island formation (*Ballmer et al.*, 2007; *Bonatti and Harrison*, 1976; *Bonatti et al.*, 1977; *King*, 2007). Moreover, SSC can drive the motion of microplates within mobile belts, such as in the Mediterranean (*Faccenna et al.*, 2013), and thus control first-order geologic activity along microplate boundaries. SSC might be even an important mechanism for the generation of plate boundaries, e.g. by initiating subduction zones (*Solomatov*, 2004). Although several articles and book chapters have been published about SSC in the mantle, and some review chapters have summarized some of the related issues (e.g., *Ballmer et al.*, 2015c; *Parmentier*, 2007), no comprehensive review has yet been dedicated to the topic. In this module, I will discuss the causes for, and consequences of, SSC in the Earth's mantle.

2. Physical Background

Convection across a fluid layer is an efficient mode of heat transport with heat being transported along with the flow of anomalously warm or cold matter. Generally, convection is driven by unstable density stratification. Density inversion e.g. occurs as a consequnces of heating a fluid from below and/or cooling from above, such that low-density warm fluid is overlain by high-density cold fluid.

On all scales, convection competes with other modes of heat transport. For the Earth's interior, the most relevant competing mode is conduction of heat. Thermal convection is a more efficient mode than conduction if the fluid layer's Rayleigh number $Ra = \alpha\rho_0\Delta T g d^3/\kappa\eta$ exceeds a critical value (i.e., on the order of ~1000) (*Bénard*, 1901; *Turcotte and Schubert*, 1982). In this case, even infinitesimally small thermal perturbations grow exponentially to sustain the formation of convection cells. As thermal expansivity α, gravity g, reference density ρ_0, and conductivity κ are well-constrained and/or near-constant across the mantle, the temperature jump across the fluid layer ΔT, the layer thickness d and the viscosity of the fluid η control the Rayleigh number with $Ra \sim \Delta T d^3/\eta$.

In a fluid heated from below and cooled from above, convection is usually driven by thermal boundary layer instability. Thin thermal boundary layers (TBL) with steep thermal gradients are sustained, because conduction is more efficient than convection for small d. Convective instability usually rises out of these TBL. For viscous instability developing from a cold TBL, over which the viscosity varies by several orders of magnitude, the concept of a local Ra has been shown to be useful (e.g., Parson and McKenzie, 1978), because only the negative buoyancy of the viscously deformable base of the TBL is available to drive convection. This situation needs to be accounted for to in determining the relevant parameters that control the *Ra*. For convection driven by unstable compositional stratification (Rayleigh-Taylor instability; see section 3.3), the relevant compositional $Ra_{comp} = \Delta\rho_{comp} g d^3/\kappa\eta$, where $\Delta\rho_{comp}$ is the compositional density contrast.

In a fluid that is rheologically layered (and compositionally heterogeneous), such as the Earth's mantle (e.g., *Mitrovica and Forte*, 2004; *Zindler and Hart*, 1986), convection can simultaneously occur on various scales (and be manifested as thermal plus compositional ─ or thermochemical ─ convection). Large-scale convective flow organizes across the whole mantle (i.e., on scales of several thousands of km), because *Ra* becomes super-critically large for large d (and large ΔT). Smaller scales of convection (i.e., of the order of hundreds of km) can organize within low-viscosity layers close to the global TBL at the top or bottom of the mantle, because *Ra* becomes sufficiently large for low η (and significant ΔT), even for rather small d (e.g., *Korenaga and Jordan*, 2003; *Solomatov and Moresi*, 2000). Thus, SSC may occur in the asthenosphere (*Richter and Parsons*, 1975), in low-viscosity regions near the core-mantle boundary (*Cizkova et al.*, 2010), or even across the whole upper mantle (*Korenaga and Jordan*, 2004). SSC may also develop near regional TBLs, such as the roofs of the large low shear-wave velocity provinces, where small-scale "plumelets" are thought to rise and feed hotspot volcanism (*Davaille*, 1999), or at the base of subducted slabs that stagnate in the mantle transition zone (*Motoki and Ballmer*, 2015).

The specific layer that undergoes convective instability must include a finite ΔT to maximize the relevant local *Ra*. So, in the case of top-down driven SSC from a cold TBL, the layer must include at least the relatively soft base of the TBL, and the relevant local *Ra* is limited by the viscosity of this base. In case of bottom-up driven SSC from a hot TBL, the local *Ra* is in turn restricted by the viscosity and thickness of the overlying layer.

3. Styles of SSC

Due to its applicability to surface geologic processes, the occurrence of SSC in the asthenosphere has been most closely studied. SSC in the asthenosphere is top-down driven by a density inversion with cool sublithospheric mantle overlying the warm asthenosphere, and facilitated by intrinsically low viscosities. The low viscosity of the asthenosphere is thought to be sustained by the abundance of small amounts of melt (*Anderson and Sammis*, 1970), a local dominance of dislocation creep (*Karato*, 1987), reduced mineral grain sizes (*Faul and Jackson*, 2005), relatively large temperatures ("plume-fed asthenosphere") (*Morgan et al.*, 1995), and/or relatively high water contents (*Karato and Jung*, 1998).

3.1 SSC beneath the oceanic lithosphere

A textbook case of SSC occurs beneath oceanic plates (*Richter*, 1973). Here, the sublithospheric TBL grows as the plate moves away from the mid-ocean ridge (MOR). As soon as the TBL exceeds a critical thickness, the local *Ra* becomes sufficiently large, and sublithospheric SSC initiates (*Fleitout and Yuen*, 1984; *Houseman and McKenzie*, 1982; *Parsons and McKenzie*, 1978; *Zaranek and Parmentier*, 2004). SSC is thought to initiate beneath oceanic lithosphere of age ~70 Ma, depending on regional conditions. Beneath significantly younger oceanic lithosphere, the TBL is thinner than the lithospheric harzburgite layer, the stiff depleted residue from MOR melting (*Ra* is small due to high η), and hence SSC should normally not occur. Beneath oceanic lithosphere of age ~70 Ma and older, the TBL instead extends through this stiff depleted residue and into the weak asthenosphere to drive SSC (*Afonso et al.*, 2008; *Ballmer et al.*, 2009; *Lee et al.*, 2005).

SSC acts to remove the base of the TBL and replace it by warm mantle from below, thereby transporting heat to the base of the lithosphere and balancing the thickness of the plate. Thus, the occurrence of SSC can account for the observed flattening of seafloor topography and of heat flow on oceanic plates older than ~70 Ma (*Cazenave et al.*, 1988; *Crosby et al.*, 2006; *Doin and Fleitout*, 1996; *Hasterok and Chapman*, 2011; *Parsons and Sclater*, 1977; *Stein and Stein*, 1994a; b), as well as seismic estimates for the thickness of the oceanic lithosphere (*Priestley and McKenzie*, 2006; *Ritzwoller et al.*, 2004). These are the main observations that support the SSC model. An alternative mechanism to account for these observations involves the collective effect of the mantle plumes (*Crough*, 1975; *Hayes*, 1988; *Morgan et al.*, 1995).

Direct observations of SSC beneath oceanic plates instead remain controversial. SSC is predicted to organize as convection rolls that are mostly confined to the low-viscosity asthenosphere (e.g., *Hall and Parmentier*, 2003; *van Hunen et al.*, 2005) (Fig. 1). To minimize the interaction with asthenospheric shearing and large-scale flow, SSC rolls more-or-less strictly align with the direction of the overriding plate (*Richter and Parsons*, 1975), depending on plate velocity and the time since the last plate reorganization (*Marquart*, 2001; *van Hunen and Zhong*, 2006). Accordingly, SSC are predicted to be associated with lineations in heat flow, gravity, seismic anomalies and seafloor topography (*Buck and Parmentier*, 1986) with wavelengths similar to the vertical extent of the asthenosphere (but cf. *Lev and Hager*, 2008). However, heat flow measurements and seismic tomography remain challenging in oceanic basins. Also, dynamic topography associated with SSC on thick and old lithosphere is predicted to be too small to be resolved (*Sleep*, 2011). On the much younger oceanic lithosphere close to the East Pacific Rise, lineations in geophysical observables of wavelengths ~100 km have been detected, and in many aspects are consistent with the effects SSC (*Harmon et al.*, 2011; *Haxby and Weissel*, 1986). Because of their geographic patterns that extend to very close of the East Pacific Rise, however, these lineations are difficult to be reconciled at least with the textbook case of SSC, unless the asthenospheric viscosity is very low (~10^{18} Pa·s or smaller), and (interaction with) alternative mechanisms such as viscous fingering (*Ballmer et al.*, 2013; *Weeraratne et al.*, 2007), or off-axis melting instabilities (*Barnouin-Jha et al.*, 1997), need to be considered. The much more extensive lineations of wavelengths 1500~2000 that are evident in full-waveform S-wave tomography (*French et al.*, 2013) as well as the gravity field (*Hayn et al.*, 2012) are also not fully consistent with sublithospheric SSC (i.e., mostly confined to the asthenosphere), both because of their large wavelengths as well as their manifestation beneath young oceanic plates, even crossing MORs (also see discussion). Future studies are indeed required to understand the interaction of SSC with other geodynamic mechanisms (e.g., viscous fingering), as well as the much larger scales of whole-mantle circulation.

3.2 SSC related to mantle plume activity

Small-scale convection can also occur in regionally restricted settings, such as in mantle plumes that pond beneath the lithosphere as a “pancake” of hot material. In this specific case, the conditions mentioned above in terms of the age of the lithosphere required for SSC are relaxed, because the viscosity of pancake is sufficiently low (*Agrusta et al.*, 2013; *Moore et al.*, 1998; *Moore et al.*, 1999; *Thoraval et al.*, 2006). The manifestation of SSC in the Hawaiian plume pancake can explain the occurrence of rejuvenated-stage volcanism and off-axis volcanism, the decrease geoid-to-topography ratio of the swell to the WNW, as well as geochemical asymmetry of shield-building volcanism (*Ballmer et al.*, 2011; *Cadio et al.*, 2012; *Garcia et al.*, 2010). In turn, SSC far away from mantle plumes (section 3.1) can sustain “hot-line” volcanic chains that ─ in contrast to plume-fed hotspot volcanism ─ display coeval activity over >1000 km (Bonatti and Harrison, 1976; Bonatti et al., 1977; Ballmer et al., 2007; 2009) (Fig. 1).

3.3 Compositional SSC

While the textbook case of SSC is exclusively driven by an inversion of thermal density, convective instability may be alternatively fueled by an inversion of compositional density (i.e., Rayleigh-Taylor instability (*Rayleigh*, 1913)) with high-density fluid underlain by low-density fluid. Such a situation can e.g. occur near subduction zones, where the slab carries a layer of low-density materials such as sediments or serpentinized basalt into the mantle. Dehydration of serpentinite leads to hydration and/or partial melting of overlying mantle rock. The resulting mélange of low-density rocks becomes convectively unstable as soon as it is juxtaposed to the high-density peridotitic mantle wedge above (*Gerya and Yuen*, 2003). In this case, a strongly unstable compositional density inversion can even overcome stable thermal layering (i.e., the hot mantle wedge overlying the cool slab). Similarly, compositional instability can rise out of the buoyant harzburgitic underbelly of a subducted slab that stagnates in the transition zone (*Motoki and Ballmer*, 2015) or near the core-mantle boundary (*Tackley*, 2011).

Another form of compositional SSC is buoyant decompression melting (BDM) instability. A layer of rock that is very close to or at its solidus may undergo localized melting due to small lateral thermal variations and/or passive upwelling. Any such localized melting induces focused upwelling, because melt (as well as the residue of melting) is less dense than rock. Since this upwelling in turn causes further decompression melting in a positive-feedback loop, short-lived small-scale convection cells emerge (*Tackley and Stevenson*, 1993). BDM usually ceases after one full overturn, because it runs out of fuel; the depleted residue of magmatism cannot continue melting without additional heat input (*Raddick et al.*, 2002) (Fig. 2). Nevertheless, the process of BDM can assist other mechanisms such as SSC in sustaining magmatism (*Ballmer et al.*, 2009). Episodes of BDM can also occur near a MOR, as initial melts are produced on-axis and subsequently undergo instability off-axis (*Barnouin-Jha et al.*, 1997; *Sparks et al.*, 1993). In this case, BDM is boosted by a cessation of extension and divergence as the partially molten material moves away from the ridge axis (*Hernlund et al.*, 2008a). A cessation of extension can also trigger BMO within the partially molten asthenosphere in intraplate settings (*Hernlund et al.*, 2008b).

3.4 Edge-driven convection

Convective instability is generally assisted by lateral heterogeneity. The presence of lateral heterogeneity, such as fracture zones or cratonic margins, strongly reduces the timescales for the onset of convective instability, and thereby acts to trigger SSC (*Dumoulin et al.*, 2005; *Huang et al.*, 2003). Thus, the most vigorous and stable downwellings of SSC occur near steps of lithospheric thickness, such as the edges of cratonic keels or orogenic roots (Fig. 3). Upwelling return flow is in turn focused at a distance of several hundreds of km away from the step to induce uplift and magmatism (*Kaislaniemi and van Hunen*, 2014; *King*, 2007; *Missenard and Cadoux*, 2012; *Shahnas and Pysklywec*, 2004; *Till et al.*, 2010; *van Wijk et al.*, 2010). This variant of SSC has been dubbed "edge-driven" convection (*King and Anderson*, 1998). The edge-driven convection model implies that downwellings consistently emerge along cratonic margins with uplift and volcanism along a belt parallel to the margin at a distance of a few 100s of km. It has been used to account for intraplate volcanism as well as seismic observations mainly on the African and South American plates (*King and Ritsema*, 2000; *King*, 2007).

3.5 SSC beneath continents

Beneath the continents, SSC is most likely caused by a combination of mechanisms. As beneath the oceans, TBL instability remains an important driving mechanism for continental SSC (see section 3.1) (*Houseman et al.*, 1981), but variations in composition (section 3.3) (*Houseman and Molnar*, 1997; *Neil and Houseman*, 1999) as well as steps in lithospheric thickness (section 3.4) (*King and Anderson*, 1998) are common ingredients beneath continents that can trigger SSC. Sublithospheric topography is common beneath continents, and likely controls the geometry of SSC beneath the slow-moving continental plates (*Fourel et al.*, 2013; *Milelli et al.*, 2012). For example, a combination of horizontal flow, edge-driven convection and viscosity heterogeneity in the asthenosphere can give rise to vertical flow and magmatism (*Ballmer et al.*, 2015a; *Conrad et al.*, 2010; *Kaislaniemi and van Hunen*, 2014; *Till et al.*, 2010) (Fig. 3). Entrainment of warm plume material or enriched slab-derived material into such upwellings may provide the conditions for mantle melting (e.g., *Duggen et al.*, 2009). Intraplate extension acts to advance small-scale convective instability due to the induction of mantle upwelling as well as the creation of sublithospheric topography (e.g., along rifts) (*Boutilier and Keen*, 1999; *Buck*, 1986; *van Wijk et al.*, 2008; *van Wijk et al.*, 2010). Alternatively, underplating of dense plutonic rocks may drive top-down instability (*Zhai et al.*, 2007). Any related SSC has been suggested to be sufficient to even destroy stable cratonic roots (e.g., *Gao et al.*, 2009). In turn, compositional variations may also stabilize cratonic roots, e.g. by muting SSC as a consequence of decreased lithospheric viscosity and/or density. As the viscosity structure in continental plates is usually layered, deformation during convective instability may be focused along weak horizons and therefore lead to the "delamination" of elongated chunks of the lower crust (*Göğüş and Pysklywec*, 2008b; *Kay and Kay*, 1993). Accordingly, delamination of the lower crust may be regarded as a variant of SSC in the presence of complex rheology (*Burov and Molnar*, 2008), in which the local *Ra* is controlled by the viscosity of the weak zone, and strongly anisotropic viscosity controls the geometry of downwellings. Delamination and dripping are two geometrical end-members of SSC beneath continents.

Subcontinental SSC has various geological and geophysical implications. Intraplate uplift and subsidence is commonly related to SSC up- and downwellings. Hence, SSC controls erosional patterns as well as sustains the formation of sedimentary basins (*Petersen et al.*, 2010). For example, delamination may induce rapid uplift and erosion just after detachment of the lower crust, followed by persistent subsidence due to focusing of downwelling flow (*Göğüş and Pysklywec*, 2008a). Beneath continents with dense seismic networks, it is possible to detect isotropic and anisotropic seismic velocity anomalies that can be directly related to SSC (*Alsina and Snieder*, 1995; *Makeyeva et al.*, 1992; *Schmandt and Humphreys*, 2010; *West et al.*, 2009; *Yang and Forsyth*, 2006). Intraplate seismicity, deformation, mountain building and volcanism have also been related to SSC. While large-scale convection drives the motion of tectonic plates and major geologic activity along plate boundaries, SSC can account for a wide range of intraplate geological processes.

4. Discussion

A key issue regarding the "theory" of SSC in the mantle is the observability of the process. From a fluid dynamics point of view, SSC is well-established and almost an inevitable product of convection in a high-*Ra* fluid with significant rheological layering, as is the Earth's mantle. Even though any methods for determining the radial viscosity structure of the Earth suffer fundamental trade-offs, at least a relatively weak asthenosphere and viscosity jump somewhere near 660~1,000 km depth are robust features of glacial-rebound and geoid inversions, respectively (e.g., *Rudolph et al.*, 2015). Also, the predictions of the theory of SSC on a broader scale (e.g., seafloor flattening) are well consistent with observations, and a wide range of circumstantial evidence supports the theory (see above). However, from a phenomenon point of view, there are few direct observations of SSC in the mantle, and the patterns of convection remain poorly constrained.

In recent years, at least "mid-scale" convective patterns become well in reach of proper characterization by geophysical observations. Here, I consider mid-scale convection as SSC with wavelengths on the order of ~1,200 to 2,000 km. For simple rheology, such mid-scale wavelengths are expected for convection across the entire upper mantle, or even down to ~1,000 km depth. Such mid-scale convection is expected to be sustained by progressive cooling of the asthenosphere by sublithospheric SSC, which accordingly tends to break down into larger convection cells down to the base of the transition zone (*Korenaga and Jordan*, 2004). Mid-scale convection may further rise from a second-order TBL at the base of the transition zone (*Motoki and Ballmer*, 2015), for example due to limited material exchange between the upper and lower mantles (*Ballmer et al.*, 2015b; *Ballmer et al.*, 2017; *Tackley et al.*, 1993), which is also evident by slab stagnation at various depths in the mid mantle (*Fukao and Obayashi*, 2013; *Goes et al.*, 2017). Alternatively, SSC on wavelength of 1,000-2,000 may be mostly confined to the asthenosphere if the rheology, for example due to lattice-preferred orientation beneath the moving plates, is significantly anisotropic (*Lev and Hager*, 2008).

The specific patterns of mid-scale convection (whether confined to the asthenosphere or throughout the upper mantle) may be related to geophysical observations. Viable candidates are low-viscosity fingers as retrieved from seismic tomography (*French et al.*, 2013; *Katzman et al.*, 1998), as well as undulations of the residual (dynamic) topography (*Hoggard et al.*, 2016) and gravity field (Hayn et al., 2012) on the relevant scales. It may further be mirrored by the patterns of slab sinking (and related return flow (*Faccenna et al.*, 2013)) near subduction zones, and by that of microplates in mobile belts such as the Mediterranean (*Faccenna et al.*, 2010). Along these lines, we may already be able to map the patterns of mid-scale convection over vast regions of our planet, although continental regions without sufficient seismic instrumentation remain problematic (due to lithological heterogeneity and tectonic complexity). First attempts to map vertical upper-mantle flow in geophysically well-characterized continental regions such as the western US are indeed promising (*Afonso et al.*, 2016; *Schmandt et al.*, 2014).

5. Conclusion and Outlook

Convection is an efficient mechanism for the transport of heat across the mantle. It is generally driven by a density inversion due to the combined effects of temperature and composition. In addition to large-scale convection on the order of thousands of km, small-scale convection (SSC) is thought to occur near thermal and compositional boundary layers, particularly across regions of reduced mantle viscosity, such as the asthenosphere. SSC should be advanced by the presence of lateral heterogeneity, for example by steps of lithospheric thickness along cratonic margins or fracture zones. Beneath continents with their complex geologic structures, SSC is influenced by various factors, including sublithospheric topography, horizontal flow, compositional heterogeneity as well as the related effects on density and viscosity structure. While only indirect evidence for the occurrence of SSC beneath mature oceanic basins is provided by the flattening of seafloor topography (and heat flow), direct evidence for the occurrence of SSC beneath continents comes from seismic observations.

Future work is required to better understand the interplay between large-scale and small-scale flow in the global context. Large-scale whole-mantle convection should influence the patterns of SSC by driving shear flow in the asthenosphere, and by sustaining large-scale thermal pertubations that will delay or advance SSC through their effects on viscosity. Also, large-scale flow may set up boundary layers in the mid-mantle or transition zone, from which SSC instabilitty may develop, for example near stagnant slabs (*Fukao and Obayashi*, 2013) or ponding/deflected plumes (*French and Romanowicz*, 2015; *Kumagai et al.*, 2008). Furthermore, the effects of mineral grain-size and fabric on mantle rheology and SSC, and vice-versa, require detailed further study. Perhaps most importantly, future efforts should be focused on detecting the signals and pattens of SSC. For example, direct seismic evidence for the occurrence of sublithospheric SSC beneath the oceans on the expected wavelength of several 100s of km remains elusive. Systematic comparison of model predictions and observations will lead to great insight into the underlying mantle dynamics and dominant rheological mechanisms.

6. References

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7. Figures

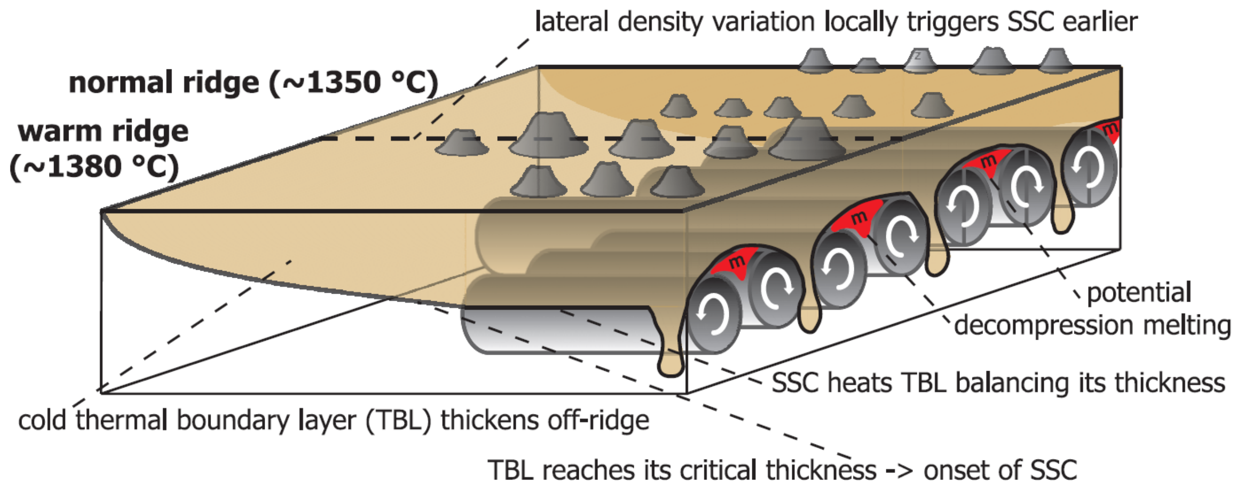

Fig. 1: Cartoon of sublithospheric SSC beneath the oceanic lithosphere that grows as a function of seafloor age. At a critical seafloor age, SSC rolls develop and may potentially sustain decompression melting and coeval volcanism over large along-plate distances. Figure is reproduced from *Ballmer et al.* (2009) reprinted with the permission of the *American Geophysical Union*.

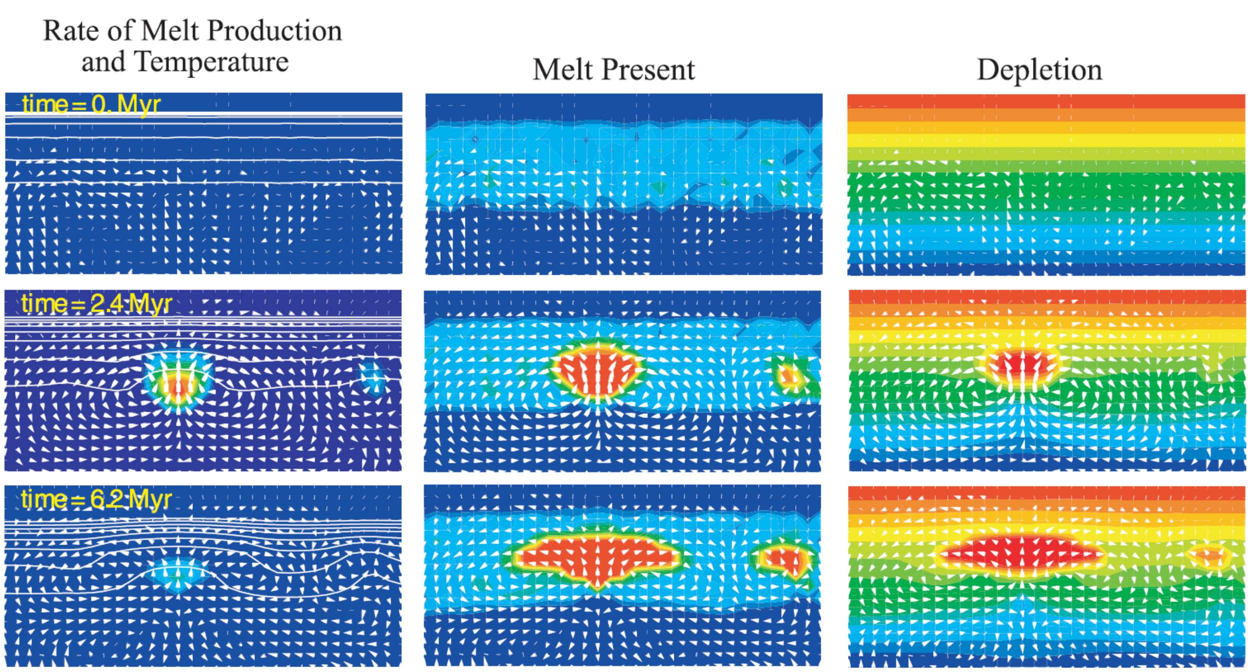

Fig. 2: Numerical model predictions of buoyant decompression melting (BDM). Colors show melt production rate (left column), melt retention (center) and depletion (right); arrows mark direction and speed of mantle flow; lines show isotherms (left only). From top to bottom, panels

show snapshots as model time (as annotated) increases. BDM develops from an instability of an initial melt layer (top center). It proceeds for several million years, and is shut off as depletion in the convection cell progressively increases (bottom right). Figure is reproduced from *Raddick et al.* (2002), reprinted with the permission of the *American Geophysical Union*.

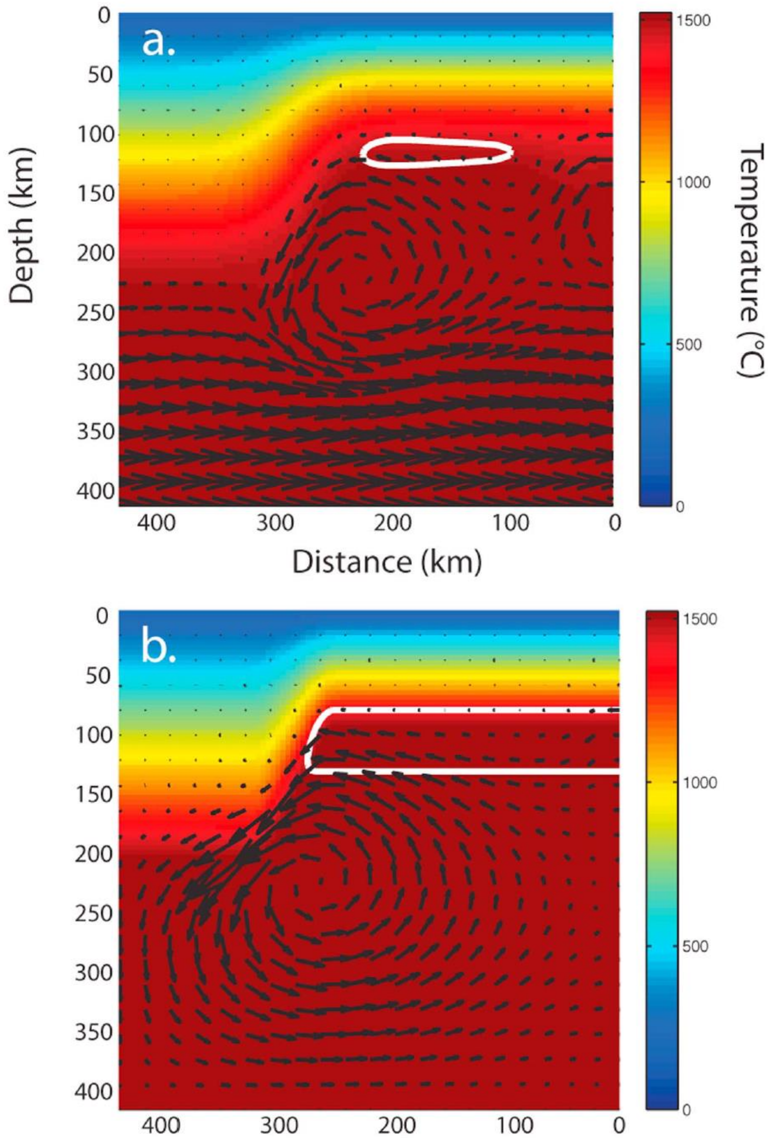

Fig. 3: Edge-driven convection (EDC) with (a) and without (b) asthenospheric shear-flow in the backgound mantle. Colors and arrows reflect mantle temperature and flow, respectively. The white contour (solidus) outlines the zone of potential mantle melting. EDC alone, or an interaction of EDC with “shear-driven upwelling” (*Bianco et al.*, 2011; *Conrad et al.*, 2010) due to the upward deflection of horizontal flow at a step of lithospheric thickness can sustain mantle melting. Figure is reproduced from *Till et al.* (2010), reprinted with the permission of the *American Geophysical Union*.