Non-hotspot volcano chains originating from small-scale sublithospheric convection

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[1] Some oceanic volcano chains violate the predictions of the hotspot hypothesis for geographic age progressions. One mechanism invoked to explain these observations is small-scale sublithospheric convection (SSC). In this study, we explore this concept in thermo-chemical, 3D-numerical models. Melting due to SSC is shown to emerge in elongated features (∼750 km) parallel to plate motion and not just at a fixed spot; therefore volcanism occurs in chains but not with hotspot-like linear age progressions. The seafloor age at which volcanism first occurs is sensitive to mantle temperature, as higher temperatures increase the onset age of SSC because of the stabilizing influence of thicker residue from previous mid-ocean ridge melting. Mantle viscosity controls the rate of melt production with decreasing viscosities leading to more vigorous convection and volcanism. Calculations predict many of the key observations of the Pukapuka ridges, and the volcano groups associated with the Line, Cook-Austral, and Marshall Islands. **Citation:** Ballmer, M. D., J. van Hunen, G. Ito, P. J. Tackley, and T. A. Bianco (2007), Non-hotspot volcano chains originating from small-scale sublithospheric convection, *Geophys. Res. Lett.*, *34*, L23310, doi:10.1029/2007GL031636.

1. Introduction

[2] Although most of the intraplate volcanism in ocean basins is expressed in linear chains, not all of them are accountable to a fixed hotspot [*Morgan*, 1971]. The most fundamental prediction of the hotspot hypothesis is a linear age progression of the volcanic edifices along the chain; however, some ridges – such as the Marshall, Line, and Cook-Austral Islands – display highly irregular age-distance relationships [*Davis et al.*, 2002; *Koppers et al.*, 2003; *Bonneville et al.*, 2006] and therefore require another mechanism.

[3] The small-sized Pukapuka and neighboring ridges form another well-studied example. They are aligned by plate motion with ages unexplained by the hotspot hypothesis [*Sandwell et al.*, 1995], and are accompanied by topography and gravity lineations with a wavelength of ∼200 km [*Haxby and Weissel*, 1986].

[4] Several models have been put forward to account for the gravity lineations and the associated volcanism. Firstly, cracks in the lithosphere due to diffuse extension of the Pacific Plate [*Sandwell et al.*, 1995] or due to thermal cracking [*Gans et al.*, 2003] might sample pre-existing melt. Secondly, return flow of anomalously hot, wet and maybe partially molten peridotite from the Tuamotu and Marquesas hotspots towards the East Pacific Rise might undergo fingering and channelling [*Weeraratne et al.*, 2007]. Thirdly, small-scale sublithospheric convection (SSC) might dynamically produce melting and the associated gravity lineations [*Buck and Parmentier*, 1986; *Haxby and Weissel*, 1986; *Marquart*, 2001].

[5] The lithospheric cracking hypothesis presumes a partially molten reservoir in the asthenosphere. This was proposed to account for its rheological properties [*Anderson and Sammis*, 1970], something that has been recently challenged [*Karato and Jung*, 1998; *Faul and Jackson*, 2005]. Moreover, this model does not predict negative density anomalies in the asthenosphere as observed from gravimetry and local seismic tomography [*Harmon et al.*, 2006, 2007; *Weeraratne et al.*, 2007]. Even though the surface morphology of the chains might be influenced by tensional cracks [*Lynch*, 1999], it is unlikely that they control magma generation itself. Instead, channelized return flow and SSC may both account for the volcanism and associated density anomalies altogether [*Marquart*, 2001; *Harmon et al.*, 2006]. This paper explores the latter mechanism.

[6] In the Earth's uppermost mantle SSC is likely to develop due to instabilities of the thickened thermal boundary layer beneath mature oceanic lithosphere (Figure 1). It is characterized by convective rolls aligned by plate motion [*Richter and Parsons*, 1975]. Lower mantle viscosities or lateral density heterogeneity (thermal or compositional variations) cause SCC to begin at younger plate ages [*Huang et al.*, 2003; *Dumoulin et al.*, 2005] with the possibility of partial melting in the upwelling limbs of SSC [*Haxby and Weissel*, 1986].

[7] Partial melting changes the compositional buoyancy due to melt retention and additional depletion of the residue [*Oxburgh and Parmentier*, 1977; *Schutt and Lesher*, 2006]; therefore, it promotes upwelling and allows for further melting. This self-energizing mechanism is able to sustain melt production for a couple of million years once melting initiates [*Tackley and Stevenson*, 1993; *Raddick et al.*, 2002; *Hernlund et al.*, 2007].

[8] In this study, we take the step towards fully thermo-chemical 3D-numerical models of SSC in order to test quantitatively the SSC-hypothesis for intraplate volcanism. Therein, we explore the 3D-patterns of melting associated with SSC, the age of seafloor over which it occurs, and the rates of melt generation by varying the key parameters mantle viscosity and temperature. Both are only weakly constrained or may strongly vary through the mantle.

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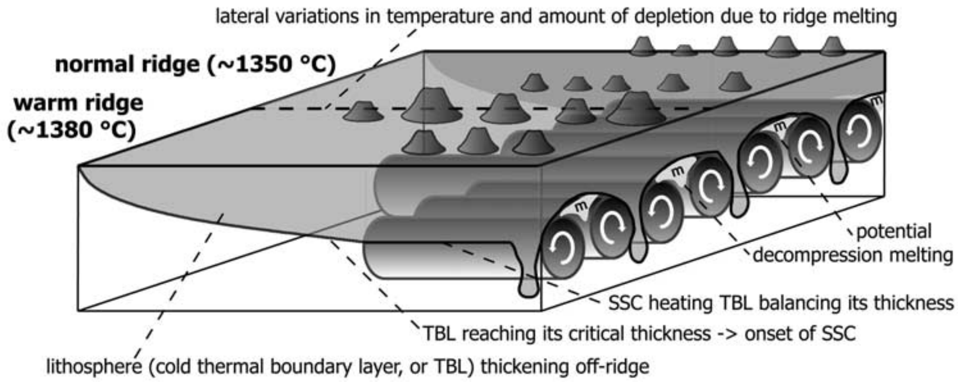

Figure 1. SSC-rolls develop after a critical thickness of the TBL is exceeded, and earlier if next to lateral density heterogeneity.

Finally, we discuss applications of the SSC-hypothesis on specific intraplate volcanic chains.

2. Method

[9] To explore 3D-numerical models we use a version of the finite element code CITCOM [*Moresi and Gurnis*, 1996; *Zhong et al.*, 2000; *van Hunen et al.*, 2005]. With this code, we solve the equations for conservation of mass, momentum and energy for an incompressible, infinite Prandtl number fluid with extended Boussinesq approximations.

[10] Density is both a function of temperature T and composition as modified by processes of melting and refreezing:

$$\rho = \rho_m(1 - \alpha(T - T_m)) + F\Delta\rho_{depl} + \varphi\Delta\rho_{melt}$$

where T_m, ρ_m, α, F, φ, $\Delta\rho_{depl}$, and $\Delta\rho_{melt}$ are reference temperature and density (3300 kg/m^3), thermal expansivity ($3{\cdot}10^{-5}$ K^{-1}), mass fraction of melt depletion, volume fraction of melt in the mantle, density change due to depletion (-72.6 kg/m^3 [*Schutt and Lesher*, 2006]) and melt (-500 kg/m^3), respectively.

[11] The melting model is from *Katz et al.* [2003] and is valid for hydrous peridotite melting in the shallow upper mantle. For this initial study, we assume a bulk water content for the starting source mantle of 0.0125 wt.-% yielding a $\sim$50°C solidus decrease compared to the anhydrous case. The equilibrium water content of the melt-solid mixture influences melting temperatures at each point in our melting model [*Katz et al.*, 2003], with the greatest reduction of the solidus occurring at the largest water content, where depletion F, is smallest [*Hirth and Kohlstedt*, 2003].

[12] Melt is accumulated and passively advected with the viscous mantle flow, until a threshold porosity of 1% is reached, at which point basaltic melt in an interconnected network becomes mobile [*Faul*, 2001]. Then, any excess melt is instantaneously extracted to the surface to maintain φ = 1%, which is equivalent to assuming the timescale for melt extraction is much smaller than that for mantle flow.

[13] The rheology depends on temperature and depth, ignoring any compositional effects:

$$\eta = \eta_m \exp\left(\frac{E^* + \rho_m g z V^*}{RT} - \frac{E^*}{RT_m}\right)$$

where R, g, z, η_m, E^*, and V^* are the gas constant, gravitational acceleration, depth, reference mantle viscosity, activation energy (120 kJ/mol), and activation volume (5 cm^3/mol), respectively. The low value for E^* is applied to mimic the contribution of dislocation creep in the asthenosphere [*Christensen*, 1984]. Higher values would underestimate lithospheric erosion induced by SSC [*van Hunen et al.*, 2005] and overestimate flexural rigidity near seamounts [*Watts and Zhong*, 2000].

[14] Calculations are performed in a Cartesian box heated from below and cooled from above in 384 × 96 × 48 finite elements representing 3000 × 920 × 400 km. We apply free slip boundary conditions at the sides and no slip at the top and the bottom, where we impose +65 km/Myr and +10 km/Myr, respectively (i.e., plate motion relative to the lower mantle of 55 km/Myr). The inflow boundary represents a plane parallel to the ridge axis on 4 to 20 Myr-old seafloor. Inflow T- and F-profiles are self-consistently derived from 2D mid-ocean ridge models. Added to the inflow T-profile is a small thermal random noise (±2°C). Calculations are continued until a statistical steady-state is reached.

3. Results

[15] In our simulations, SSC develops spontaneously out of our initial thermal conditions. The flow self-organizes in rolls aligning with plate motion with a preferred wavelength of $\sim$200 km. SSC starts to erode and remove the harzburgite layer in downwelling sheets (Figure 2), which is replaced by hot and fertile peridotite from below. If the lithosphere is young (i.e., thin) enough for SSC to occur above the (hydrous) peridotite solidus, the upwelling limbs of SSC spawn decompression melting.

[16] Thermal buoyancy both triggers SSC and is the dominant factor that sustains it. The lateral temperature variations of 200–300°C at a typical SSC depth of melting (60–100 km) yield thermal buoyancy of about five times greater than compositional buoyancy. Once melting initiates, melt retention and depletion buoyancies further fuel the instabilities. Melt retention buoyancy is the more important factor; firstly, because $\Delta\rho_{melt} \approx 7{\cdot}\Delta\rho_{depl}$, and secondly, because melt buoyancy does not inhibit downwellings where melt refreezes, whereas depletion buoyancy does.

[17] The duration of melting is ultimately controlled by heat transfer due to SSC, since SSC continually feeds fertile

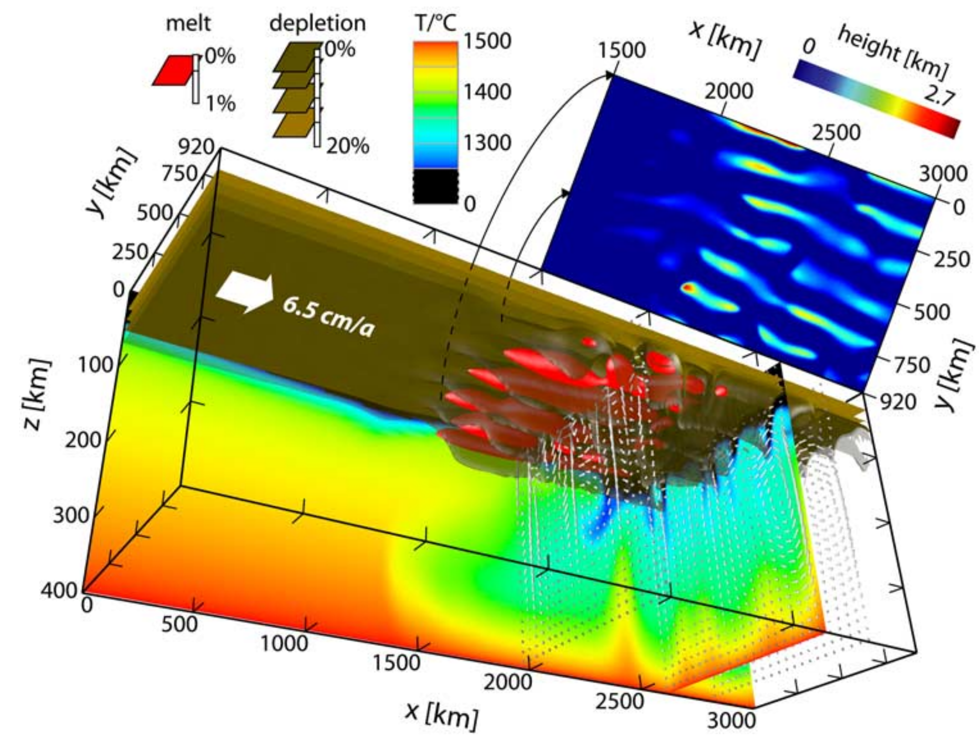

Figure 2. Isosurfaces of melt fraction and depletion, and cross-sections of the temperature and velocity field, for an example calculation (T_m = 1380°C and $\eta_{eff} = 1.6 \cdot 10^{19}$ Pa·s). Lifted from the top is a plan view of the thickness of melt accumulated onto a plate moving over the box.

mantle to the melting zone (unlike that given by *Raddick et al.* [2002]). The underlying asthenosphere experiences a net heat loss because of latent heat carried by melting, and more importantly due to heat advected by SSC to the lithosphere. Since the asthenosphere progressively cools, melting sustained by intrinsic density variations begins to fade such that it rapidly ceases.

[18] For each calculation, Figure 3 shows the volumetric flux of extracted melt per kilometer of plate in the direction of plate motion versus the age of the seafloor; for T_m = 1350°C (1410°C) eruption rate is predicted to peak at a seafloor age of ∼30 Myr (∼50 Myr), and to span a broad range of 8 to 9 Myr. This duration changes little over the range of parameters examined, and corresponds to a length scale of about 750 km for Pacific plate-motion.

[19] We find that the onset time and amount of volcanism (extracted melt) are most sensitive to T_m and η_{eff}. The onset of melting, which succeeds the onset of SSC, is later for both increasing effective mantle viscosity η_{eff} [*Huang et al.*, 2003] and increasing reference temperature T_m, since a higher T_m provides a thicker harzburgite layer and thus a more stably stratified system [*Zaranek and Parmentier*, 2004]. Despite melting occurring beneath a thicker lithosphere, the amount of melting still correlates positively with T_m, and therefore, a higher T_m enables more volcanism on older seafloor. The amount of volcanism (represented by the area beneath the curves in Figure 3) correlates positively with T_m and particularly strongly with η_{eff}. Generally, significant melting due to SCC requires η_{eff} to be no larger than $\sim 2.5 \times 10^{19}$ Pa·s ($\sim 1.5 \times 10^{19}$ Pa·s) for T_m = 1410°C (T_m = 1350°C). If it is only slightly lower than that, large volumes of volcanism are to be expected, because of an earlier onset of SSC beneath a thinner lithosphere and a more vigorous mantle flow. For the cases presented here, predicted crustal thicknesses of seamounts range from 1.7 to 5.9 km.

4. Discussion and Conclusion

[20] We have shown from 3-D numerical simulations of SSC that melting is expected in the upwelling limbs of SSC-rolls for η_{eff} of $1.3–2.3 \cdot 10^{19}$ Pa·s. These viscosities are within the range estimated for the oceanic asthenosphere [*Cadek and Fleitout*, 2003; *Hirth and Kohlstedt*, 2003]. Volcanism is predicted to occur along lineaments aligned with plate motion and to span seafloor ages of 25–50 Myr, positively correlating with mantle temperature T_m. Melt volume strongly depends on η_{eff} and can be sufficient to generate kilometer-high seamounts. What is key is that because the melting zone is elongated (∼750 km) and not a well-defined spot, the absolute locations of volcanic events are predicted to scatter over a broad range of distances and times (8 to 9 Myr). Simple geographic age progressions are therefore not predicted.

[21] Such a behavior could explain key observations of some, previously enigmatic, volcanic chains in the Pacific. The Pukapuka ridge, for instance, consists of seamounts rising ∼2.5 km over the surrounding seafloor. This height and the estimated seafloor ages during volcanism (20–30 Myr) at least in the western portion of the Pukapuka ridges are consistent with our model predictions for T_m = 1350°C and $\eta_{eff} \approx 1 \cdot 10^{19}$ Pa·s. Also, the possible age progression that has been suggested along Pukapuka [*Sandwell et al.*, 1995] is much too fast to be explained by a fixed Pacific hotspot.

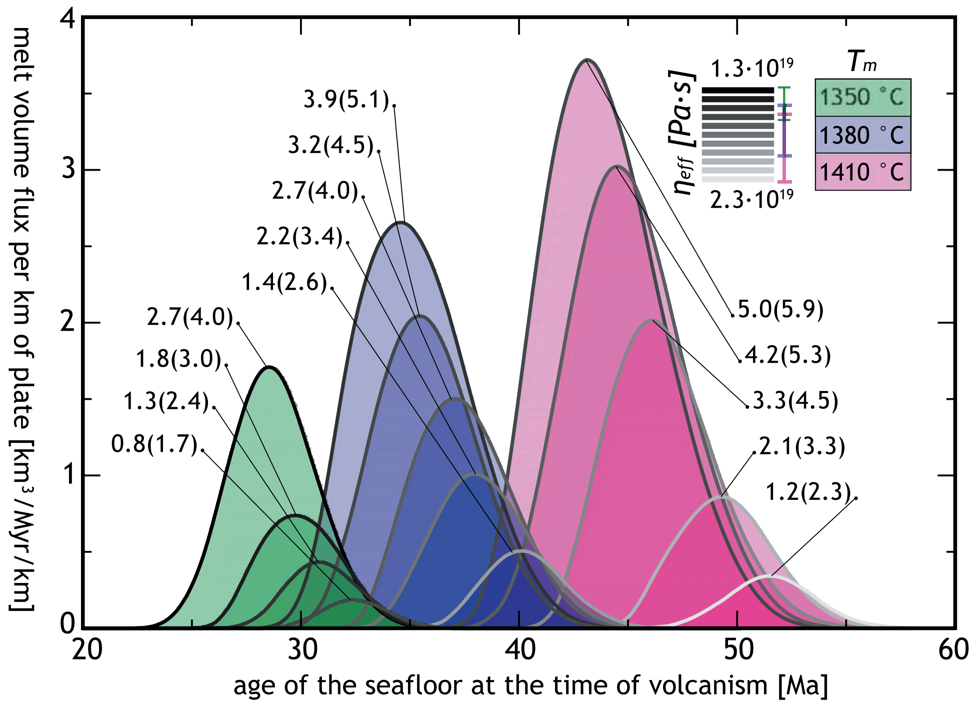

Figure 3. Amount of volcanism vs. age of the underlying seafloor for different model runs (varying the parameters T_m and η_{eff}). Vertical axis is the total volumetric rate of melt production in vertical (y–z) cross-section of our model divided by the number of SSC cells. It represents the average volume of melt produced by a single SSC cell per unit time and per kilometer in the direction of plate motion. Curves are shaded according to η_{eff} as indicated in the scale above. Colors indicate different reference temperatures. Numbers indicate heights (in km) of volcanic edifices with a slope of 10° that could be created if all of the melt extracted above a SSC upwelling accretes as continuous volcanic ridges or, in parentheses, as chains of circular volcanoes spaced 100 km apart. Viscosity predominantly controls volcano height, whereas T_m controls both the age of seafloor during magmatism and volcano height.

Volcanism along the lineament could therefore rather have occurred over an elongate zone such as predicted by our models. Our current calculations, however, do not predict the formation of the eastern portion of the Pukapuka on seafloor younger than ~20 Myr.

[22] The Marshall Islands group consists of multiple short parallel volcano chains spaced ~200 km apart from one another with large individual guyots (once up to 4–5 km high). It erupted on seafloor ages of 50–90 Myr [*Koppers et al.*, 2003] within the Darwin Rise, which is interpreted as the Cretaceous' precursor of the South-Pacific Superswell [*McNutt*, 1998; *Smith et al.*, 1989]. The Cook-Australs are two parallel ridges (with an offset of ~300 km) of equally large volcanoes and formed on 40–100 Myr old lithosphere [*Bonneville et al.*, 2006, and references therein] over the Superswell. In both cases, the spacing between the chains and the lack of a simple age progression are consistent with a SSC-origin. Explaining the large seafloor ages would, however, require higher mantle temperatures than simulated in this study. Alternatively, thermal rejuvenation of the lithosphere and thinning of its thermal boundary layer may yield a late onset of convection and melting beneath old seafloor. Rejuvenation has been proposed for both the Darwin Rise [*Smith et al.*, 1989] and the Superswell [*McNutt*, 1998]. Correspondingly, η_{eff} must have been low (10^{19}–$2{\cdot}10^{19}$ Pa·s) to account for the large volume of volcanism.

[23] The Line Islands, at the western edge of the Darwin Rise, are reasonably well explained by SSC. These volcanoes display two events of quasi-synchronous volcanism that erupted laterally over ~2000 km and on seafloor ages of 30–55 Myr [*Davis et al.*, 2002]. This range of seafloor ages is well predicted by models with T_m of ~1380–1410°C. A relatively small horizontal thermal gradient in the mantle (possibly related to the Superswell) could account for the larger seafloor age span of the Line Islands compared to predictions of any single of our current calculations. The volcanoes reach a height of up to ~3.5 km above the surrounding seafloor, which, when considering the subsurface volcanic crust, represent melt thicknesses of the same order as predicted by models with low η_{eff}.

[24] Our simple, homogeneous models are able to reproduce the general characteristics observed at the above ridges. Particularly, they provide a framework accounting for volcanism with non-hotspot like age-progressions. But they have limitations in predicting durations of volcanism >10 Myr and ages of the underlying seafloor <25 Myr and >55 Myr, respectively.

[25] However, small off-axis thermal anomalies or greater melting depths of some lithologies [*Ito and Mahoney*, 2005] would likely enable volcanism also on seafloor of ages >55 Myr. Lateral density heterogeneity would locally reduce the onset age of SSC [*Huang et al.*, 2003; *Dumoulin et al.*, 2005] allowing significant melting in at least one upwelling limb (Figure 1), and potentially leading to volcanism on seafloor of ages <25 Myr (like at the eastern Pukapuka ridge). Since the distribution of such heterogeneities is

expected to be irregular, the onset age of convection would likely vary through time. This prediction would further complicate any possible age progression and could account for a larger apparent duration (like at the Line Islands).

[26] Volcanism due to SSC ultimately requires either η_{eff} on the lower bound or T_m on the upper bound of what is predicted for oceanic asthenosphere. Small η_{eff} are sufficient for volcanism on young seafloor, whereas slightly elevated T_m ($\sim$1400°C) are needed to produce large volumes of volcanism on middle-aged seafloor. However, significant volcanism due to a plume or lithospheric cracking invokes much greater thermal anomalies (>100°C). Along these lines, several other volcano chains might as well have their origin from SSC.

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