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Numerical models of the magmatic processes induced by slab breakoff

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ABSTRACT

After the onset of continental collision, magmatism often persists for tens of millions of years, albeit with a different composition, in reduced volumes, and with a more episodic nature and more widespread spatial distribution, compared to normal arc magmatism. Kinematic modelling studies have suggested that slab breakoff can account for this post-collisional magmatism through the formation of a slab window and subsequent heating of the overriding plate and decompression melting of upwelling asthenosphere, particularly if breakoff occurs at depths shallower than the overriding plate.

To constrain the nature of any melting and the geodynamic conditions required, we numerically model the collision of two continental plates following a period of oceanic subduction. A thermodynamic database is used to determine the (de)hydration reactions and occurrence of melt throughout this process. We investigate melting conditions within a parameter space designed to generate a wide range of breakoff depths, timings and collisional styles.

Under most circumstances, slab breakoff occurs deeper than the depth extent of the overriding plate; too deep to generate any decompressional melting of dry upwelling asthenosphere or thermal perturbation within the overriding plate. Even if slab breakoff is very shallow, the hot mantle inflow into the slab window is not sustained long enough to sufficiently heat the hydrated overriding plate to cause significant magmatism. Instead, for relatively fast, shallow breakoff we observe melting of asthenosphere above the detached slab through the release of water from the tip of the heating detached slab. Melting of the subducted continental crust during necking and breakoff is a more common feature and may be a more reliable indicator of the occurrence of breakoff. We suggest that magmatism from slab breakoff alone is unable to explain several of the characteristics of post-collisional magmatism, and that additional geodynamical processes need to be considered when interpreting magmatic observations.

1. Introduction

The nature and distribution of post-collisional magmas differ markedly from those corresponding to the pre-collisional history of convergence zones. Oceanic subduction creates linear belts of arcs dominated by calc-alkaline magma series, whereas post-collisional magmatism is more diverse chemically, and more widespread spatially. It can extend 100s of kilometres from the suture zone and can be sustained for millions of years after initial collision of the continental plates, although short-lived pulses of more intense melting and longer quiescent periods have been observed, as well as systematic variations in space and time (e.g. Chung et al., 2005; Pearce et al., 1990). Magmatism across these areas appears to originate from a variety of different sources, displaying signatures characteristic of crustal, subcontinental lithospheric and asthenospheric material (e.g. Lee et al., 2012).

Many different mechanisms have been proposed to explain magmatism in continental collisional areas. These include processes involving solely the continental crust: shear heating along crustal faults (Harrison et al., 1998); melting of thickened crust (England and Thompson, 1986); and the exhumation of crust back to the surface (Harris and Massey, 1994). Additionally, processes may involve an influx of heat or melt from the underlying mantle: large scale lithospheric delamination (Bird, 1978); slab breakoff (Davies and von Blanckenburg, 1995); edge-driven convection (Kaislaniemi and van Hunen, 2014); and small-scale

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convection at the base of the lithosphere (Kaislaniemi et al., 2014). Slab breakoff after collision is often invoked to account for the changes in the nature and composition of magmas in collisional areas (e.g. Coulon et al., 2002; Mahéo et al., 2002; von Blanckenburg and Davies, 1995). Upon continental collision, the subducted slab may weaken due to thermal heating and/or increased extensional stresses between the more buoyant continental and denser oceanic lithospheres, resulting in slab breakoff and the formation of a slab window, and possibly leading to magmatism (van Hunen and Allen, 2011).

The dynamics of slab breakoff has been investigated extensively, and it has been shown that the strength of the subducting oceanic and continental lithospheres, in part influenced by the oceanic slab age, convergence velocity, continental crustal and lithospheric thicknesses, and the mechanism of detachment, all have a control on the depth of breakoff (Andrews and Billen, 2009; Duretz et al., 2011; Gerya et al., 2004). Models have shown a wide range in this depth, from 40 to over 500 km (Baumann et al., 2010; Duretz et al., 2011), and have shown the potential importance of a 3D geometry which allows for lateral progression of the breakoff process through slab tearing (van Hunen and Allen, 2011; Magni et al., 2014b; Menant et al., 2016; Capitanio and Replumaz, 2013).

Slab breakoff may be fairly well constrained dynamically, but its consequences in terms of magmatism remain unclear. Previous studies have suggested that hot asthenosphere, flowing up through the slab window, may have the potential to melt through decompression and/or heat both the subducting crustal material and the overriding continental lithosphere (Davies and von Blanckenburg, 1995; van de Zedde and Wortel, 2001). Melting of the overriding lithosphere is viable where breakoff occurs at depths shallower than the base of this, but, asthenospheric melting may be less likely, as it requires very shallow breakoff at depths less than 50 km, unless the mantle is volatile rich.

Many previous slab breakoff studies used either fully dynamic models to study the complex mechanical process of breakoff and the associated mantle flow (e.g. Gerya et al., 2004), or used kinematic models of subduction to focus on the hydration state of the system and magmatic consequences (e.g. van de Zedde and Wortel, 2001). Recent numerical studies have started to use coupled petrological–thermomechanical numerical models to study the dynamics of collision and breakoff (Li et al., 2013; Menant et al., 2016); but a systematic investigation of the conditions required to generate breakoff-induced melting has not yet been conducted, which is undertaken in this study.

Continental collision involves a succession of complex, interacting dynamical and petrological processes, of which the contribution of slab breakoff to post-collisional magmatism is the least well understood and constrained. Therefore, we use coupled petrological–thermomechanical numerical models to examine conditions for post-collisional magmatism, focusing on the magmatic consequences of slab breakoff in particular, so that its effect can be better understood and dissociated from other processes. We will do so by varying a number of key parameters which are likely to affect the breakoff dynamics, and by tracking the occurrence of melting of a number of different materials over time. The dynamics in our models are in line with previous numerical studies and replicate many of the features that have been previously observed during collision and breakoff (Duretz and Gerya, 2013), and we thus use these models to explore the magmatic processes. We show that slab breakoff by itself is unlikely to explain all of the characteristics of post-collisional magmatism, and that additional processes may play an important role.

Table 1
Parameters, symbols, units and default values used within this study.

Symbol	Parameter and units	Default value
u	velocity [$\mathrm{m\,s^{-1}}$]	–
P	pressure [Pa]	–
η	viscosity [Pa s]	–
T	temperature [°C]	–
Q_0	radiogenic heat production [$\mu\mathrm{W\,m^{-3}}$]	see Table 3
Ra	thermal Rayleigh number [–]	2.8×10^6
Rb	compositional Rayleigh number [–]	–
C	compositional function [–]	–
e_z	vertical unit vector [–]	–
t	time [s]	–
α	thermal expansion coefficient [$\mathrm{K^{-1}}$]	3.5×10^{-5}
ρ_0	reference density [$\mathrm{kg\,m^{-3}}$]	3330
g	gravitational acceleration [$\mathrm{m\,s^{-2}}$]	9.8
h	model height [km]	660
κ	thermal diffusivity [$\mathrm{m^2\,s^{-1}}$]	8×10^{-7}
η_0	reference viscosity [Pa s]	2×10^{20}
$\delta\rho$	density difference	see Table 2
A	pre-exponential exponent [$\mathrm{Pa^{-n}\,s^{-1}}$]	see Table 2
x	distance [m]	–
ΔT	maximum temperature difference [°C]	1350
n	power-law exponent [–]	see Table 2
$\dot{\varepsilon}$	second invariant of strain rate [$\mathrm{s^{-1}}$]	–
E^*	activation energy [$\mathrm{J\,mol^{-1}}$]	see Table 2
V_{diff}	activation volume [$\mathrm{cm^3\,mol^{-1}}$]	0.7 (diff.), 8.0 (disl.)
R	gas constant [$\mathrm{J\,K^{-1}\,mol^{-1}}$]	8.3
$\eta_{\max}$	maximum viscosity [Pa s]	1×10^{24}
σ_y	yield stress	–
σ_0	surface yield stress [MPa]	40
$\sigma_{\max}$	maximum yield stress [MPa]	400
μ	friction coefficient	0.2
N	number of materials with differing densities	4

2. Method

2.1. Governing equations

We use the Cartesian version of the finite element code Citcom to model oceanic and continental subduction (Moresi and Gurnis, 1996; Zhong et al., 2000). This solves for the conservation of mass, momentum, energy and composition in a fluid, assuming incompressible flow and adopting the Boussinesq approximations. A standard non-dimensionalisation is applied to the governing equations (Table 1):

$$x = x'h \qquad t = t'h^2/\kappa \qquad \eta = \eta'\eta_0 \qquad T = T'\Delta T$$

Giving:

$$\nabla \cdot u = 0$$

$$-\nabla P + \nabla \cdot \left(\eta\left(\nabla u + \nabla u^T\right)\right) + \left(RaT - \sum_{i=1}^{N} Rb_i C_i\right) e_z = 0$$

$$\frac{\partial T}{\partial t} + u \cdot \nabla T = \nabla^2 T + Q_0$$

$$\frac{\partial C}{\partial t} + u \cdot \nabla C = 0$$

The thermal Rayleigh number, Ra, is defined as:

$$Ra = \frac{\alpha \rho_0 g \Delta T h^3}{\kappa \eta_0}$$

and similarly the compositional Rayleigh number of a given composition i, Rb_i:

$$Rb_i = \frac{\delta\rho_i g h^3}{\kappa \eta_0}$$

where $\delta\rho_i$ is the intrinsic density of composition i relative to mantle material.