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Melting-induced crustal production helps plate tectonics on Earth-like planets

Diogo L. Lourenço *, Antoine Rozel, Paul J. Tackley

Institute of Geophysics, Department of Earth Sciences, ETH Zurich, Sonneggstrasse 5, CH-8092 Zurich, Switzerland

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ABSTRACT

Within our Solar System, Earth is the only planet to be in a mobile-lid regime. It is generally accepted that the other terrestrial planets are currently in a stagnant-lid regime, with the possible exception of Venus that may be in an episodic-lid regime. In this study, we use numerical simulations to address the question of whether melting-induced crustal production changes the critical yield stress needed to obtain mobile-lid behaviour (plate tectonics). Our results show that melting-induced crustal production strongly influences plate tectonics on Earth-like planets by strongly enhancing the mobility of the lid, replacing a stagnant lid with an episodic lid, or greatly extending the time in which a smoothly evolving mobile lid is present in a planet. Finally, we show that our results are consistent with analytically predicted critical yield stress obtained with boundary layer theory, whether melting-induced crustal production is considered or not.

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1. Introduction

Although the effects of melting and crustal production for planetary evolution and dynamics are acknowledged as being very important (Stevenson, 1990; Xie and Tackley, 2004; Davies, 2007; Ogawa and Yanagisawa, 2011; Nakagawa and Tackley, 2012), their effects on the mobility of the lithosphere are poorly understood. The different convection regimes and the general effects of melting are reviewed next in this section. In this article, for the purpose of a better readability, we will use the acronym "MCP" to represent "melting-induced crustal production".

1.1. Lid regime

In the Solar System, Earth is the only planet to be in a mobile-lid regime (Tackley, 2000; Stein et al., 2004), whilst it is generally accepted that all the other terrestrial planets are currently in a stagnant-lid regime (Solomatov, 1995), showing little or no surface motion. A transitional regime between these two, showing episodic overturns of an unstable stagnant lid, has been reported and might apply to Venus (Moresi and Solomatov, 1998; Rozel, 2012; Armann and Tackley, 2012).

It has been shown that a convection regime similar to plate tectonics can be modelled using strongly temperature-dependent viscosity and plastic yielding (Fowler, 1993; Moresi and Solomatov, 1998; Tackley, 2000; Stein et al., 2004). In these models, a maximal stress is imposed in the lithosphere. If stresses exceed this value, named the yield stress, the viscosity is decreased to bring the stresses back to the critical value. This is sufficient to break the lithosphere into "plates". However, the exact value of the stresses reached in the lithosphere strongly depends on the rheology (Fowler, 1985; Solomatov, 1995; Reese et al., 1998; Solomatov, 2004), and also on the surface boundary condition used in numerical simulations (Crameri et al., 2012). The critical yield stress necessary to obtain mobile-lid (plate-like) behaviour in numerical simulations is much lower than what is expected from laboratory rock deformation experiments (Kohlstedt et al., 1995). An explanation for this may be related to the presence of water in nature (Regenauer-Lieb et al., 2001; Dymkova and Gerya, 2013).

Numerical models typically focus on purely thermal convection, whereas compositional variations in the lithosphere can alter the stress state, simply by thickening the lithosphere or by providing lateral density anomalies that in turn produce additional stresses, greatly influencing the likelihood of plate tectonics. For example, Rolf and Tackley (2011) showed that the addition of a continent can reduce the critical yield stress for mobile-lid behaviour by a factor of around two, while Armann and Tackley (2012) found that bursts of crustal production caused by partial melting may trigger lithospheric overturn events, suggesting that melting may also play an important role in facilitating plate tectonics. Complicating matters is the finding that the final state of the system (stag-

* Corresponding author.
E-mail address: diogo.lourenco@erdw.ethz.ch (D.L. Lourenço).

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nant or mobile lid) can depend on initial conditions (Tackley, 2000; Weller and Lenardic, 2012; Lenardic and Crowley, 2012).

1.2. Melting and crustal production

Melting plays a major role in the evolution of planets' interiors, both during their formation and during their subsequent long-term evolution over billions of years (Stevenson, 1990; Nakagawa and Tackley, 2012). On present-day Earth, melting mainly occurs in the shallow mantle below the tectonic plates (McKenzie and Bickle, 1988), and may also occur in deeper regions of the upper mantle and above the core–mantle boundary (CMB) (Williams and Garnero, 1996). The surface expression of melting is volcanism, which on present-day Earth occurs mostly as the formation of oceanic crust at mid-ocean spreading centres (Schubert et al., 2001).

Partial melting causes differentiation when crust is produced at mid-ocean ridges because the major element composition of the melt is different from that of the source rock, thus leaving a depleted residue. Trace elements are generally incompatible and enter the melt phase (Hofmann, 1997). Some of them contain heat-producing isotopes, which contribute to around 50% of the present-day heat loss from the interior (Davies, 2007). Also, partial melting modulates heat loss from the interior by transporting heat from the interior to the surface, where it erupts, solidifies and cools. The advected heat cools the planet, but at the same time, melting absorbs latent heat locally, which lessens the maximum mantle temperature.

Another effect of mantle melting is that it dehydrates and stiffens the shallow part of the mantle (Hirth and Kohlstedt, 1996). It introduces viscosity and compositional stratifications in the shallow mantle, because viscosity increases with the loss of hydrogen upon melting (Faul and Jackson, 2007; Korenaga and Karato, 2008). Also, the presence of melt along grain edges influences the deformation rate of grains, and therefore the viscosity (Zimmerman and Kohlstedt, 2004; Scott and Kohlstedt, 2006). While these effects may be significant, we do not consider them in this work.

Therefore, melting has a major role in the long-term evolution of rocky planets, enhancing heat loss, causing chemical differentiation of the interior and influencing geochemical signatures. Although it has been taken into account in some thermal evolution studies, its effect on plate tectonics has never been systematically studied.

1.3. This study

In this article, we present a set of 2D spherical annulus simulations of mantle convection (Hernlund and Tackley, 2008) considering MCP. We focus on the question of whether MCP changes the critical yield stress required to obtain mobile-lid behaviour as a function of governing parameters, particularly the reference viscosity. We first describe our model in section 2. In section 3 we present our results, which are discussed in section 4. Finally, in section 5 we present the conclusions of this study.

2. Numerical model and physical model

The numerical model used here is based on the one described by Armann and Tackley (2012) and Tackley et al. (2013), although with parameters adjusted to the case of the Earth. The model incorporates realistic parameter values and physics descriptive of planet Earth, and thus includes compressibility, phase transitions, pressure–temperature-dependence of viscosity, time-dependent internal and basal heating and plasticity. Diffusion creep, with the assumption of homogeneous grain size, is the assumed deformation mechanism. The values used for the standard physical parameters are given in Table 1. Throughout our study we varied the reference viscosity and the yield stress in the model, as well as whether or not MCP is included. Within this framework, the effect of MCP was systematically tested.

Table 1
Physical properties. (UM = Upper Mantle (dry olivine); PV = Perovskite; PPV = Post-Perovskite; Act. en. stands for Activation energy, Act. vol. for Activation volume.)

Property	Symbol	Value	Units
Surface temperature	T_{surf}	300	K
Init. potential temp.	T_{P0}	1600	K
Specific heat capacity	C_p	1200	J/kg/K
Gas constant	R	8.3145	J/K/mol
Latent heat of melting	L	600	kJ/kg
Internal heating rate	H	$18.77 \cdot 10^{-12}$	W/kg
Half-life	t_{half}	2.43	Ga
Act. en. - UM	E_{ol}	300	kJ/mol
Act. vol. - UM	V_{ol}	5.00	cm^3/mol
p_{decay} - UM	p_{decay_ol}	∞	GPa
Act. en. - PV	E_{pv}	370	kJ/mol
Act. vol. - PV	V_{pv}	3.65	cm^3/mol
p_{decay} - PV	p_{decay_pv}	200	GPa
Act. en. - PPV	E_{ppv}	162	kJ/mol
Act. vol. - PPV	V_{ppv}	1.40	cm^{3}mol
p_{decay} - PPV	p_{decay_ppv}	1610	GPa
Thermal expansivity	α	$5 \cdot 10^{-5}$	K^{-1}
Density	ρ	3300	kg/m^3
Gravity	g	9.81	m/s^2
Mantle thickness	h	2890	km
Thermal diffusivity	κ	$7.6 \cdot 10^{-7}$	m^2/s
mid-UM pressure	P_{UM}	30	GPa
Density difference	$\Delta\rho$	200	kg/m^3
Phase trans. eclogite	D_{ecl}	60	km

2.1. Rheology

In our models, the viscous deformation mechanism is diffusion creep, which is assumed to follow a temperature- and pressure-dependent Arrhenius law:

$$\eta_{diff}(T, p) = \eta_0 \exp\left(\frac{E + pV}{RT} - \frac{E}{RT_0}\right), \tag{1}$$

where η_0 is the reference viscosity at zero pressure and reference temperature T_0 (= 1600 K), E is the activation energy, p is the pressure, V is the activation volume, T is the absolute temperature and R is the gas constant. Different values for E and V are used for the upper and lower mantle (Karato and Wu, 1993; Yamazaki and Karato, 2001) and can be seen in Table 1. The activation volume decreases with increasing pressure according to the formula:

$$V(p) = V_0 \exp\left(-\frac{p}{p_{decay}}\right). \tag{2}$$

p_{decay} is given in Table 1. A viscosity jump of 10 is imposed at the transition between the upper and lower mantle (Čížková et al., 2012 and references therein). A second viscosity jump of 10^{-3} (compared with the above material) is imposed at the transition to post-perovskite at lowermost mantle depths, as suggested by mineral physics experiments and theoretical calculations (Ammann et al., 2010; Hunt et al., 2009). The reference viscosity η_0 is varied in our simulations, ranging from $5 \cdot 10^{19}$ Pa·s to 10^{21} Pa·s.

In order to obtain plate-like behaviour, plastic yielding is employed (Moresi and Solomatov, 1998; Tackley, 2000). It is assumed that the material deforms plastically after reaching a yield stress:

$$\tau_y = \tau_{duct} + \tau'_{duct}\, p, \tag{3}$$

where τ_{duct} is the ductile yield stress and τ'_{duct} is the vertical gradient of the ductile yield stress. In practice, this last parameter prevents yielding in the deep mantle. The parameter with the greatest influence in the previous equation is τ_{duct}, and is therefore

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