

# Plume —lid interactions during the Archean and implications for the generation of early continental terranes

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## Abstract

Many Archean terranes are interpreted to have a tectonic and metamorphic evolution that indicates an intra-crustal reorganization driven by lithospheric-scale gravitational instabilities. These processes are associated with the production of a significant amount of felsic and mafic crust, and are widely regarded to be a consequence of plume-lithosphere interactions. The juvenile Archean felsic crust is made predominantly of rocks of the tonalite-trondhjemite-granodiorite (TTG) suite, which are the result of partial melting of hydrous metabasalts. The geodynamic processes that have assisted the production of juvenile felsic crust, are still not well understood. Moreover, it is still unclear how their evolution can affect the petrogenesis of TTGs. Here, we perform 2D and 3D numerical simulations coupled with the state-of-the-art of petrological thermodynamical modelling to study the tectonic evolution of a primitive Archean oceanic plateau in order to understand how the pressure and temperature conditions of both mafic and felsic melts evolve with time. Aiming at doing that, we study the effects of a tailless plume interacting with the lithosphere with different upper mantle potential temperatures ( $T_P$ ). In our numerical simulations, the continuous emplacement of new mafic dry intrusions and the extraction of the felsic melts,

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generate an unstable lower crust which drips soon after the plume arrival. The gravitational instabilities promotes mantle cooling, and mixing between plume, crustal and asthenospheric material. Moreover, they limit the crustal thickness of the oceanic plateau promoting the generation of felsic melts at low to medium pressures. The subsequent tectonic evolution depends on the asthenosphere  $T_P$ . If the  $T_P$  is high enough ( $\geq 1500$  °C) the entire oceanic crust is recycled within 2 Myrs. By contrast at low  $T_P$ , the thin oceanic plateau slowly propagates generating plate-boundary like features.

*Keywords:*

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## 1. Introduction

Only less of 10% of the total volume of continental crust is of Eo to Meso Archean age (4.0 –3.0 Ga) (?). The preserved Archean continental crust is located at the top of old, thick, strong and buoyant cratonic lithosphere, forming  
5 different terranes. In many cases the recorded geological history of Archean terranes preserves features seen in Phanerozoic terranes —e.g. horizontal displacements consistent with opening of basins (West Pilbara) Van Kranendonk et al. (2007) or orogenetic processes (Itsaq) (Nutman et al., 2013). However, many of the Archean terranes testify that high grade metamorphism, production of  
10 felsic crust and plume-related magmatism happened simultaneously (Van Kranendonk et al., 2015; Chardon et al., 1996; Choukroune et al., 1995) with a vertical reorganization of the crust driven by gravitational instabilities.

Archean continental crust is a mixture between felsic and mafic crustal components (Kamber, 2015; Condie, 1981, 1993). The felsic crust is mainly composed of TTGs (Tonalite-Trondhjemite-Granodiorites) with lesser potassic granitoids (Moyen & Stevens, 2006; Moyen, 2011), while mafic crust is composed of  
15 several types of rocks spanning from ultramafic to iron tholeiites (Anhaeusser, 2014). Juvenile Archean TTGs may be produced by a plethora of processes spanning from fractional crystallization (Smithies et al., 2019; Arndt, 2013) to  
20 partial melting of mafic protolith (Johnson et al., 2017; White et al., 2017; Palin

et al., 2016b,a; Moyen & Martin, 2012; Moyen, 2011). Nevertheless, it is widely accepted that most of the juvenile Archean TTGs are produced by partial melting of hydrated metabasalts (Palin et al., 2016a; Johnson et al., 2017; ?), which, on the ground of geochemical affinity, must have been enriched Tholeiites (EAT) (Palin et al., 2016a; ?; Condie, 1981). TTGs’ trace and major element compositions suggest that the partial melting must have occurred at high pressure and temperature in the presence of garnet, rutile, amphibole  $\pm$  plagioclase (Palin et al., 2016a; Moyen & Martin, 2012). These metamorphic reactions produce a mafic residuum that is denser than the underlying mantle (Palin et al., 2016a) and thus more prone to be dripped off or delaminated. The site of production of felsic crust may have been a shallow subduction slab or the bottom of a thickened oceanic plateau (Palin et al., 2016a; ?; Moyen & Martin, 2012).

Many Archean terranes (e.g. East Pilbara) show a close relation between felsic melts production and vertical reorganization of the crust (partial convective overturn, P.C.O.) (Van Kranendonk et al., 2004; Chardon et al., 1998; Collins et al., 1998; Bouhallier et al., 1995). These processes result in the classical structure of greenstone belts: the dome-and-keel geometries, in which felsic granitoids are rising to shallow crustal depths while mafic crust sinks. These structures have been widely recognized as atypical for any subduction setting and are more consistent with plume-lithosphere interactions (Van Kranendonk, 2010; Van Kranendonk et al., 2004; Choukroune et al., 1995). The gravitational instabilities allow for an active exchange between shallow-surface environment and the mantle reservoir (Lee et al., 2006; Bédard, 2006; Zegers & van Keken, 2001). All these processes may feedback and influence the mafic crust petrogenesis and ultimately the juvenile TTGs production. These processes are driven mainly by magmatic processes, and ultimately depend on the compositional and thermal state of the mantle and its convective style.

The mantle potential temperature ( $T_P$ ), the radiogenic heat production and the convection style all exert a strong control on the plume and lithospheric dynamics (Jellinek et al., 2002; Schubert et al., 2001; Solomatov, 1995; Davaille & Jaupart, 1993). The absence of strong viscosity contrast within the convecting

mantle and the high production of radiogenic heat may suppress plume activity, reduce their strength, or change their dynamics making them tailless (thermals following Jellinek et al. (2002)’s nomenclature). It is still unclear if the mantle  
55 was fully convecting or not (??) and what range of values of  $T_P$  are representative for the whole Archean Eon, with estimates spanning from 1400 to 1550 °C (Aulbach & Arndt, 2019; Ganne & Feng, 2017; Kamber, 2015; Herzberg et al., 2010). If  $T_P$  ranges are in the lower end of these estimates ( $< 1500^\circ\text{C}$ ), then the inferred high  $T_P$  Herzberg et al. (2010) is only an artifact of overturn  
60 events rather than representing the effective  $T_P$  of the whole convecting mantle and the preserved Archean cratons formed at the top of overturn upwelling zones (Bédard, 2018). This imply that the early stage of the planet can show a punctuated history, in which the major craton forming events are associated with periodic and discrete mantle convection reorganization Bédard (2018). As  
65 a consequence plume dynamics and crustal forming processes may have been different during the stagnant lid or overturn stage of the planet.

Venus has been considered a suitable analogue for early Earth processes (Van Kranendonk et al., 2015). The surface of Venus is punctuated by annular structures and shows evidence of discrete horizontal displacement without  
70 plate boundaries (?Harris & Bédard, 2014). The coronae structure have been interpreted as plume-related geodynamic processes (Harris & Bédard, 2014; ?). However, Venus surface is inaccessible and there is still no clear evidence of felsic crust. Moreover, it is in a stagnant-lid convection mode (Solomatov, 1995) and may not feature plume dynamics like present day Earth, additionally the hot  
75 upwelling may be tailless and/or coming from shallow levels of the convecting mantle.

Geodynamic and petrological modelling has been progressively integrated the field observations with quantitative/qualitative data. The role of gravitational instabilities on the structural and compositional evolution of the early  
80 Earth continental crust has been constrained (Piccolo et al., 2019; ?; Fischer & Gerya, 2016; Sizova et al., 2015; ?). However, despite the advancements, there are still unexplored scenarios that deserve attention. For instance, many numeri-

cal experiments base the compositional evolution on simplified parametrizations (e.g. (Fischer & Gerya, 2016; Jain et al., 2019; Sizova et al., 2015, 2014)) and  
85 employed only 2D numerical experiments (e.g. (Piccolo et al., 2019; Sizova et al., 2015)). These numerical experiments provide powerful insights, however the results may be biased by the 2D approach or by the resolution at which the crust is resolved, which is particularly important for global scale numerical experiments (e.g. Jain et al. (2019); Rozel et al. (2017)). Many numerical experiments focus  
90 on high  $T_P$  scenarios ( $> 1500^\circ C$ ) (Sizova et al., 2015; Fischer & Gerya, 2016), that are only one of the several possible scenarios. The existing 3D numerical examples, instead, focused on the structural evolution, and do not explore the effect of felsic crust extraction (Fischer & Gerya, 2016; Gerya et al., 2015). In order to have a comprehensive view of the evolution of early Earth it is necessary  
95 to explore, in a self-consistent way, the compositional and structural evolution of the Archean terranes combining petrological forward modelling with geodynamic 3D numerical simulations.

Here, we present a step-forward towards integrating petrological phase diagram in a 3D numerical framework. We study the generation and short-term  
100 evolution of Archean terranes and the effects of the upper mantle  $T_P$  on the evolution of the primitive crust and the condition at which felsic melts are generated, taking a tailless plume as a reference scenarios. Our results show that as a function of the upper mantle  $T_P$  plume-lithosphere interaction yields different compositional and structural evolution of the Archean terranes and that many  
105 observed processes do not require an active plate-tectonics.

## 2. Methods

### 2.1. Numerical Modeling

The geodynamic simulations were performed using *LaMEM* a marker in cell, finite difference 3D petro-thermo-mechanical code (Reuber et al., 2018; Kaus  
110 et al., 2016), which solves the fundamental continuum mechanics conservation equations for mass, momentum and energy.

The mass conservation equation is solved assuming that the materials are incompressible:

$$\frac{\partial v_i}{\partial x_i} + S = 0 \quad (1)$$

115  $v_i$  is the velocity vector component along the  $x_i$  direction (i.e  $x, y, z$ ). In order to incorporate the compaction and expansion effects due to the melt extraction, we introduced a source/sink ( $\mathbf{S}$ , see Tab. S1 and below for further explanations) term in the non-linear residual calculations.  $\mathbf{S}$  has an indirect effect on the momentum equation, because it affects the pressure and velocities. Therefore, 120 the momentum equation retains its canonical form:

$$\frac{\partial \tau_{ij}}{\partial x_j} - \frac{\partial P}{\partial x_i} + \rho g_i = 0, \quad (2)$$

where  $\tau_{ij}$  is the deviatoric stress tensor,  $P$  is the pressure,  $\rho$  is the density and  $g_i$  is the gravity vector component along the  $i$  direction.

*LaMEM* employs a viscous-elasto-plastic rheological constitutive equations 125 that connects the deviatoric stress tensor components ( $\tau_{ij}$ ) to the total strain rate tensor components ( $\dot{\varepsilon}_{ij}$ ) as follows:

$$\dot{\varepsilon}_{ij} = \dot{\varepsilon}_{ij}^{vis} + \dot{\varepsilon}_{ij}^{el} + \dot{\varepsilon}_{ij}^{pl} = \frac{\tau_{ij}}{2\eta_{eff}} + \frac{\overset{\diamond}{\tau}_{ij}}{2G} + \dot{\gamma} \frac{\partial Q}{\partial \tau_{ij}} \quad (3)$$

$$\overset{\diamond}{\tau}_{ij} = \frac{\partial \tau_{ij}}{\partial t} + \tau_{ik}\omega_{kj} - \omega_{ki}\tau_{kj} \quad (4)$$

$$\omega_{ij} = \frac{1}{2} \left( \frac{\partial v_i}{\partial x_j} - \frac{\partial v_j}{\partial x_i} \right) \quad (5)$$

130 Where  $\dot{\varepsilon}_{ij}$  is the total strain rate tensor, the superscript *el, vis, pl* indicate the elastic (*el*), viscous (*vis*) and plastic (*pl*) strain rate.  $\overset{\diamond}{\tau}$  is the Jaumann objective stress rate,  $G$  is the shear modulus.  $\omega$  is the spin tensor,  $\dot{\gamma}$  is the plastic multiplier,  $Q$  is the plastic flow potential, which is equal to the second invariant of the deviatoric stress tensor ( $\tau_{II}$ ), according to a dilatation-free non

135 associative Drucker-Prager flow rule.  $\eta_{eff}$  is the viscosity which is computed using the following equation:

$$\eta_{eff} = \frac{1}{2} B_n^{-\frac{1}{n}} \dot{\epsilon}_{II}^{\frac{1}{n}-1} \exp\left(\frac{E_a + PV_a}{nRT}\right) \quad (6)$$

$B_n$  is the pre-exponential factor,  $n$  is the stress exponent,  $\dot{\epsilon}_{II}$  is the second invariant of the strain rate tensor,  $R$  is the gas constant,  $E_a$  and  $V_a$  are  
140 respectively the activation energy and volume.

The plastic rheology is enforced by the application of the Drucker-Prager failure criterion (Drucker & Prager, 1952):

$$Y = \tau_{II} - \sin(\phi)P - C\cos(\phi) \leq 0 \quad (7)$$

$Y$  is the yield function,  $\phi$  is the friction angle and  $C$  is the rock's cohesion.  
145 During plastic deformation  $C$  and  $\phi$  can linearly decrease from the initial values ( $C_0, \phi_0$ ) to a final values ( $C_1, \phi_1$ ) simulating the progressive damage of a rock. The plastic weakening starts when materials accumulate at least 0.1 of total plastic strain and it stops when the material reaches a plastic strain of 1.0. The effective viscosity of the rocks is limited by an lower and upper cut-off ( $10^{18}$   
150 and  $10^{24} \text{ Pa} \cdot \text{s}$  respectively). Due to the high temperatures most of the area that are partially molten reaches the lower-cut off viscosity. As a consequence we simplify the creep viscosity by neglecting the melt-weakening factor (used for example by Piccolo et al. (2019)).

Conservation of energy is computed according to:

$$\rho C_p \frac{DT}{Dt} = \frac{\partial}{\partial x_i} \left( k \frac{\partial T}{\partial x_i} \right) + H_a + H_s + H_r \quad (8)$$

$$H_a = T\alpha \left( \frac{\partial P}{\partial x_i} v_i \right) \quad (9)$$

$$H_s = (\dot{\epsilon}_{ij} - \dot{\epsilon}_{ij}^{el}) \tau_{ij} \quad (10)$$

where  $C_p$  is the heat capacity,  $DT/dt$  is the objective temperature time derivative,  $k$  is the heat conductivity,  $H_r$  is the radiogenic heat productivity

160 ( $H_r = \rho A$ , where  $A$  is the amount of heat produced per unit of mass).  $H_a$  is the adiabatic heating and is computed according the eq. 9 (in which  $\alpha$  is the thermal expansivity) while  $H_s$  is the dissipative heating. We neglect latent heat of melting because its effect is negligible compared with shear and adiabatic heating.

## 165 2.2. Melt extraction and petrological modelling

### 2.2.1. Melt Extraction

In the simulations, melt extraction and emplacement are considered to occur instantaneously. Each timestep the effective melt fraction is computed ( $M^*$ ) by subtracting the total melt extracted from the melt fraction interpolated from the phase diagram (see Fig.1). Then we evaluate whether the volumetric melt  
170 fraction is sufficient to trigger melt extraction. If  $M^*$  is higher than a critical melt fraction quantity,  $M_{trs}$ , a fixed quantity of melt can escape from the source, and its volume ( $V_{ext}$ , extracted volume) is computed according to:

$$V_{ext} = \begin{cases} Ph_{Rat}dM dV, & \text{if } M^* - dM \geq M_{trs} \\ Ph_{Rat}(M^* - M_{trs})dV, & \text{if } M^* - dM < M_{trs} \end{cases} \quad (11)$$

175 Where  $dV$  is the volume of the current finite difference cell, and  $dM$  is the volumetric fraction of melt that can escape, and  $Ph_{Rat}$  is the volumetric fraction of the rock phase in the current cell. If the difference between  $M^*$  and  $M_{trs}$  is less than  $dM$ , the quantity of melt extracted is equal to it, and thus  $M_{trs}$  represents the minimum amount of melt that can be retained in a rock.

180  $dM$  represents the maximum volumetric fraction of melt that can be extracted from a source for a given timestep. This allow to explore different scenarios in which the extraction mechanism can be unable to compensate for the production of new melts within the source rocks, e.g. if  $dM$  is not able to compensate the production of the new melt, the average density of the partially  
185 molten area decreases, and thus its buoyancy increases. This numerical strategy allows us to reproduce the effect of sudden production of melt due to any major



hydrated breakdown reaction seen by the current phase diagram. Moreover using a fixed  $dM$  has two advantages: 1) it allows to not produce instantaneously high volume of crust ; 2) it prevents to induce strong volumetric deformation due to the source/sink term in the mass conservation equations.

The total amount of melt extracted from the current rock phase is computed by integrating the volume extracted from each cell along the  $z$  direction. Since the volume of melt is higher than the one of solid rocks, the total volume of extracted melt must be corrected by a factor that depends on the ration between the density of the melt and solid rock (  $\rho_{melt}/\rho_{solid}$ , as in Wallner & Schmeling (2016)):

$$V_{ext}^{Tot} = Vol_{cor} \sum_{k=0}^{n^{end}} V_{ext}(k) \quad (12)$$

$V_{ext}^{Tot}$  is the total amount of melt extracted from the current rock type along  $z$  direction at fixed  $x, y$  coordinates,  $Vol_{cor}$  is the volumetric correction (in average this correction is 0.9),  $k$ , is the  $k^{th}$  node along  $z$  direction,  $n^{end}$  is the last node of the numerical domain along  $z$  direction.

The total amount of melt extracted is converted into extrusions and intrusions:

$$\begin{cases} Vol_{eff} = (1 - I_R)V_{ext}^{Tot} \\ Vol_{int} = (I_R)V_{ext}^{Tot} \end{cases} \quad (13)$$

Where  $I_R$  is the relative proportion of intrusions,  $Vol_{eff}$  and  $Vol_{int}$  is the effusive and intruded volume respectively.  $I_R$  represents the volumetric amount of intrusion with respect to the total volume of melt extracted. The effusive volcanic rocks are treated as if they were sediments:  $Vol_{eff}$  is converted into an effective thickness dividing it for the current finite difference cell area, then is used to shift the (internal) free surface.

The intrusions are emplaced in a specific area of the crust (see Fig.1) (Wallner & Schmeling (2016)), the size and location of which depends on the current

crustal thickness:

$$\begin{cases} v_{int} = \left( \frac{V_{Int}}{D_{Min} - D_{Max}} \right) dz \\ D_{Min/Max} = Z_{Moho} + (D_{Int} \pm D_{Tol}) \delta_{crust} \end{cases} \quad (14)$$

215 Where  $D_{Min}$  and  $D_{Max}$  are the minimum and maximum depth of the intrusion interval,  $D_{Int}$  is the depth of intrusion and  $D_{Tol}$  is the relative dimension of half of the intrusion interval (which is an input parameter),  $\delta_{crust}$  and  $Z_{Moho}$  are the thickness of the crust and Moho depth at a given  $x$  and  $y$  coordinates. The volumetric source term used in the eq. 1 are computed by assembling a  
220 melt contribution to each rock type:

$$\mathbf{S} = \frac{1}{dt} \left( 1 - \frac{dV}{dV - V_{ext}} \right) + \frac{1}{dt} \left( 1 - \frac{dV}{dV + v_{int}} \right) \quad (15)$$

The source term is computed assuming that that extraction and injection locally shrink or expand the control volume during the timestep. It is an effective volumetric deformation (1/s).

### 225 2.2.2. Petrological Modeling

Each marker is associated with a rock-type which links it to calculated phase diagrams and the associated thermo-mechanical properties. The rock type of each marker may change as a function of the total melt extracted. Everytime that the phase diagrams is changed as a function of the total melt extracted, the  
230 amount of radiogenic heat is reduced by a factor of 2 (see Tab. S2). The phase diagrams store the melt quantity and the density of the solid fraction while the density of the melt is assumed constant (2700 and 2400  $\text{kgm}^{-3}$  for mantle and mafic phase diagrams respectively). Felsic crust formed as a consequence of crystallized TTGs magmas is modeled as a passive phase with a constant  
235 density (2700  $\text{kgm}^{-3}$ ), and it is not associated with any phase diagram.

The rock-types are divided into two sets: 1) Mantle ( $Ml$ ); 2) Mafic Crust ( $MC$ ) (as in (Piccolo et al., 2019)):

1. Mantle phase diagrams represent six different steps of mantle depletion ( $MD0\%$ ,  $MD10\%$ ,  $MD20\%$ ,  $MD30\%$ ,  $MD40\%$  and  $MD50\%$ ) and were

240 computed using *Perple\_X* (Connolly, 2009). We used the activity-composition  
model of Jennings & Holland (2015), and all calculations assume a buffered  
oxygen fugacity (all the compositions are listed in Tab.1, and all phase di-  
agrams are shown in Fig. S1). The first step of depletion (Mantle Step  
0%) is modeled using the pyrolite composition from McDonough & Sun  
245 (1995). The chemical evolution is assumed to be only a function of the  
amount of extracted melt. We compute the depleted phase diagrams us-  
ing the average composition of the residuum along the 10% melt isoline in  
every phase diagram (e.g. MD10% is computed using the residuum of the  
MD0%, MD20% is computed using the residuum composition of MD10%,  
etc.).

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2. Mafic phase diagram represents four depletion steps (*BS0%*, *BS15%*, *BS30%*  
and *BS45%*) (see Fig. S2). All phase diagrams were computed using  
THERMOCALC 3.45 (Powell & Holland, 1988), using the Holland &  
Powell (2011) dataset (ds62). The starting phase diagram has the com-  
position of a typical enriched Archean basalt (EAT). All solution model  
255 (*a-x* relations) used to produce these phase diagram are: epidote, olivine  
(Holland & Powell, 2011), silicate melt, augite, hornblende (Green et al.,  
2016), garnet, orthopyroxene, biotite, chlorite (White et al., 2014), mag-  
netite—spinel (White et al., 2002), ilmenite—hematite (White et al., 2000),  
260 Cbar-1 plagioclase, K -feldspar (Holland & Powell, 2003), and muscovite  
—paragonite (White et al., 2014)

Melt extracted from the mantle is converted into anhydrous intrusions and  
extruded hydrated EAT (*BS0%*). We assume that effusive rocks are always  
hydrated, consistent with the inferred Archean sub-marine environment (Kump  
265 & Barley, 2007).

Mafic phases are represented by five rock-types: *BS0%*, *BS15%*, *BS30%*,  
*BS45%* and anhydrous intrusive mafic crust, where *BS0%* is the protolithic  
metabasalt composition. *BS0%* is assumed to be fully hydrated, and that the  
molar volume of water is sufficient to minimally saturate the solidus at 1.0 GPa

(Palin et al., 2016a), taken to be the base of the crust. After 15% of melting and melt extraction, the rock type is changed to the next basaltic step (e.g. *BS15%*). *BS45%* is most depleted lithotype and has the same composition of *BS30%* but is no longer able to produce melt. The mafic intrusions are assumed to be anhydrous, and unable to produce any melt, and are modeled as if they had the same composition of *BS45%* (as in (Piccolo et al., 2019)). They represent the composite underplated bodies expected to be found at the base of an evolving oceanic plateau (Palin & Dyck, 2018; Van Kranendonk, 2010; Cox, 1980). All the melts that is extracted from the mafic protolith is converted into *felsic crust*. We do not distinguish any subtype of felsic rocks, and we do not consider the felsic to be produce any melts.

### 3. Results

We mainly focus on the effect of the melt extraction parameter of the mantle phases and the asthenospheric mantle potential temperature ( $T_P$ ). We carried out both 2D and 3D numerical experiments. 3D numerical experiments have been performed to explore the effect of  $T_P$  while 2D numerical experiments have been performed to test the effect of  $dM$  and  $M^{trs}$ , and the long-term evolution of the system.

We mainly focus on the P and/or T conditions at which both felsic and mafic melts are extracted because this information can be used as a compositional proxy. We classify the felsic melts extracted into two main categories TTG-like melts (TTG-opt) and non-TTG-like melts. TTG-like melts are the melts extracted within a T and P range that produce a composition that best fits the chemistry of the natural TTG (TTG-opt). We further split TTG-like melts into three categories following Moyen (2011) as a function of the pressure of extraction within the optimum field: low pressure TTGs (LP-TTG, 0.6-1.0 GPa, 800-1000 °C), medium pressure TTGs (MP-TTG, 1.0-1.8 GPa, 800-1000 °C) and high pressure TTGs (HP TTGs, >1.8 GPa, 800-1000 °C). Non-TTG-like melts are all other felsic melts of a broadly granitic composition that did not

form in the optimal P-T field identified above. Mafic melts are tracked using  
 300 their extraction temperature, which can give insights on mantle source (plume  
 or asthenosphere or contaminated mantle).

In order to describe the macroscopic structural features, we introduce the  
 following terminology: *external zone* represents the lithosphere that has been  
 initially prescribed (sometimes referred as old-lithosphere); *internal zone* is the  
 305 newly produced composite crust produced during the plume-lithosphere inter-  
 action and mainly affected by RTIs; *boundary zone* is the limit between internal  
 and external zone, mainly dominated by the delamination of the lower crust of  
 the external zone.

We differentiate delamination and RTIs. Although they are driven by the  
 310 same force (buoyancy force), they have different surface manifestation and defor-  
 mation pattern. Delamination processes are associated with peeling off of the  
 dense lower crust –they produce asymmetrical topography and stress pattern  
 (compressive in the direction of propagation and extensional in the opposite  
 direction) Göğüş & Ueda (2018); Beall et al. (2017); Le Pourhiet et al. (2006);  
 315 Morency & Doin (2004). RTIs are symmetric, and characterized by compres-  
 sive and subsequent extensional tectonics after the removal of the dense unstable  
 portion of the crust/lithosphere Beall et al. (2017); Molnar et al. (1998). We  
 further differentiate the thinning processes into two categories: gravitational  
 thinning is related with RTIs instabilities, and imply that the removal of mass  
 320 from the crust through RTIs is not compensated by the production of new crust;  
 syn-magmatic thinning is associated with the asthenosphere drag forces.

### 3.1. Initial Setup

We performed systematic experiments in 2D and 3D. The 2D numerical  
 setups are boxes whose  $y$  axis is 2 km (3 grid nodes) compared with 100's of km  
 325 for the  $x$  and  $z$  directions. All numerical experiments shares the same boundary  
 conditions being:

- Mechanical: free slip on all boundaries except for the top boundary, which  
 features an internal free surface with a stress free open top boundary;

- Thermal: the top and bottom faces are isothermal (Top=20 °C, while  
the bottom boundary temperatures are set to be 1700 °C in most of  
the experiments to increase the convective vigor, especially at low  $T_P$   
conditions), while the lateral boundaries have no flux boundary conditions;

The 2D numerical domain size is  $1000 \times 2 \times 520$  km with  $257 \times 3 \times 257$  nodes (along  $x$ ,  $y$  and  $z$  directions respectively), while the 3D numerical setup is  $800 \times 800 \times 520$  km in size with  $257 \times 257 \times 257$  nodes. The air layer is 20 km thick in all cases (see Fig.S3).

The initial compositional field features an initial 100 km thick lithosphere, divided into 30 km of mafic crust and 70 km of depleted harzburgite ( $MD30\%$ ). The ratio between crust and lithospheric mantle is constant in all the numerical experiments. The upper crust is composed of hydrous basalts ( $BS0\%$ ), whose thickness is 15 km while the lower crust is made up of dry and infertile mafic intrusions. The asthenosphere is composed of pyrolytic and fertile mantle after McDonough & Sun (1995) ( $MD0\%$ ). The plume head is purely thermal and placed at the centre of the model and has a radius that is varied from 150 to 200 km. Mantle lithologies are modelled with a dry olivine rheology flow law (Hirth & Kohlstedt, 2004). The hydrous mafic crust and the felsic crust are modelled with wet quartzite flow law, while the residual mafic composition ( $BS45\%$ ) and dry intrusions have a mafic granulite rheology (Ranalli, 1995). All geological units have the same initial friction angle ( $30^\circ$ ) and cohesions (10 MPa). The cohesion and friction angle varies linearly as function of the accumulated plastic deformation until they reach 30% of their original values.

All references scenarios assume that mantle intrusive/effusive ratio ( $I_R$ ) is equal to 0.6 (consistent with plume geodynamic estimates (Crisp, 1984; White et al., 2006)). The interval of intrusions is between the Moho depth and 0.6 of the total thickness of the crust, while mafic crust  $I_R$  is 1, with an interval of intrusion spanning from 0.5 to 0.9 of the total crustal thickness.

The  $T_P$  temperature of the experiments range from 1350°C to 1550 °C , while the potential temperature of the plume is kept constant at 1600 °C ( $T^{Pl}$ ).

For the lower  $T_P$  experiments ( $<1500^\circ\text{C}$ ), we assumed that the temperature  
of the lower boundary is higher than the value expected for the given  $T_P$  to  
increases the convective vigors. The initial geothermal gradient of the litho-  
sphere is a linear double stage geotherm, featuring a  $T_{Moho}$  of  $800^\circ\text{C}$  and a  
temperature at the base of the lithosphere that depends on the astenospheric  
 $T_P$ .

### 3.2. Effect of $dM$ and $M^{trs}$

It is possible to subdivide the experiments into two regimes: low melt ex-  
traction efficiency ( $dM < 0.01$ ) and high melt extraction efficiency ( $dM \geq 0.01$ ).  
We briefly show two representative experiments that describe the geodynamic  
evolution of the two regimes.

*Low melt extraction efficiency ( $dM = 0.001$ ,  $M^{trs} = 0.001$ ,  $T^{Pl} = 1600^\circ\text{C}$ ,  $R = 200$   
 $\text{km}$ ,  $T_P = 1600^\circ\text{C}$ )*

1st Stage: Plume arrival stage (see Fig. 2) During the initial stage, the plume  
starts ascending, causing convection of the upper mantle, with an upwelling  
localized along the mantle plume axis direction (see Fig. 2, 0.881 Myrs). The  
mantle upwelling erodes the lithosphere and induces the partial melting of the  
asthenosphere underneath it. Since the amount of melt extracted each timestep  
is low, the positive buoyancy of the partially molten asthenosphere increases  
(see Fig. 2, 0.881 Myrs, column b), allowing it to penetrate the lithospheric  
mantle.

After the plume arrives at the base of the lithosphere, it spreads along the x  
direction (see Fig. 2, 2.169 Myrs). During the spreading, mantle plume diapirs  
penetrate the lithospheric mantle. These diapirs have a mushroom shape, and  
although most of the melt is produced within their center, the melt extraction  
occurs mainly in their peripheral portion. This melt extraction pattern, that  
can be observed even in the main mantle plume, results in a generation of small  
magmatically thickened areas (50-100 km scale), that imposes a wavelength on  
the crustal thickness. The flux of mafic melts is insufficient to heat up the

hydrated crust, so no felsic crust is produced at this stage of the simulation. As a consequence of localized crustal thickening, the lower crust starts to be eclogitized, producing a negative and unstable volume of mafic crust (see Fig. 2 2.169 Myrs, column b).

2nd Stage: Syn-magmatic extension and lithosphere removal (see Fig 2, 2.765-2.813 Myrs). The lower crust to the right of the plume axis drips off, removing the lithospheric mantle and causing the return flow of the asthenosphere. The asthenospheric return flow induces the production of high volume of mafic melt. The positive buoyancy of the partially molten mantle increases, producing an active upwelling (see Fig 2, 2.765 Myrs, column b). Since the lithospheric mantle is removed during the initial drip, the asthenosphere comes in contact with the hydrated crust, causing the production of significant volumes of felsic melt. The partially molten mantle spreads beneath the crust and its drag force initiates syn-magmatic extension. Since the amount of melt extracted in the central area is low, the newly produced crust in its the central area is thin and tends to thicken laterally outward. Because of the dripping of the remaining anomalous thickened portion of the crust, the upwelling is localized at the centre of the numerical domain (see Fig 2, 2.765 Myrs, column b). Three distinct types of structural homogeneous area are generated: the *internal zone*, that represents the area initially affected by the plume that expands laterally and is characterized by a fast syn-magmatic extension; two *boundary zones*, which are mechanically thickened areas and characterized by the constant peeling off of the lower crust (see Fig 2, 2.765 Myrs, column b); and two *external zones*. At the end of the stage, the gravitational instabilities are still active and are mainly RTIs. This final process allows further mantle melting which produces new hydrated basalts that bury the felsic crust.

3rd stage: Quiescent stage (see Fig 2, 14.180 Myrs). After the removal of the old lithosphere, the simulation enters in the quiescent stage. Since the mantle critically cools, and a new compositional lithospheric mantle stabilizes, the production of new crust is limited and associated with the small scale drips occurring at the base of the crust. Felsic crust is located at Moho boundary



because of its burial and as a consequence of the removal of the lower residual  
 420 and intrusive mafic crust (see Fig 2, 14.180 Myrs, d).

We performed a small numerical experiment in 3D, imposing a lower plume  
 radius ( $R=150$  km) and using low  $dM$  value (0.001). There are no significant  
 first order differences w.r.t the 2D scenario (see Fig. S4). However, 3D ef-  
 fects allow the development of a small and transient subduction-like zone. This  
 425 subduction zone is short lived and evolved from the delamination of the lower  
 crust. Furthermore it shows few similarities with long-lived Phanerozoic sub-  
 duction zones. The subduction-like process is caused indirectly by the plume,  
 and it evolves from the delamination of the lower crust of the external zone.  
 This is an important difference to the plume induced subduction described in  
 430 Baes et al. (2016); Gerya et al. (2015); Ueda et al. (2008), which is mainly caused  
 by the penetration of the mantle plume in the lithospheric mantle.

*High melt extraction efficiency ( $dM = 0.05$ ,  $M^{trs}=0.001$ ,  $T^{Pl}=1600$  °C,  $R=200$   
 km,  $T_P=1600$  °C)*

1st stage: Plume arrival (see Fig. 3, 1.962 Myrs): During the plume ascent,  
 435 the lithospheric mantle is stable. During the spreading of the mantle plume  
 beneath the lithosphere, most of the melt is extracted in the plume axis area  
 generating a oceanic plateau that extends from -200 to 200 km, whose maximum  
 thickness is located at the axis of the mantle plume (see Figure 3, 1.962 Myrs  
 column a). The lower crust of the thickened plateau becomes unstable (see  
 440 Figure 3, 1.962 Myrs column b), and because of the newly emplaced intrusion,  
 the hydrated crust starts melting producing small volume of felsic melts. The  
 production of felsic melts and their extraction produces a considerable volume  
 of residual mafic crust, that enhances the buoyancy contrast with the underlying  
 and mantle.

445 2nd stage: Dripping, delamination and lithospheric mantle removal (see Fig-  
 ure 3, 2.062-3.207 Myrs). Due to the continuous magmatic activity the lower  
 crust of the oceanic plateau drips off. The dripped lower crust locally removes  
 the lithospheric mantle, causing an asthenosphere return flow. *The ascending*

*asthenosphere partially melts and enhances the production of new mafic intru-*  
*sions, that heat up the hydrated crust, allowing the production of further felsic*  
450 *melts* (see Figure 3, 1.962 Myrs column a and b). The dripped off material  
interrupts the plume buoyancy flux, and the normal asthenosphere replaces  
the removed mantle lithosphere. The oceanic plateau undergoes syn-magmatic  
extensions and gravitational thinning (see Figure 3, 2.186 Myrs). After the  
455 complete removal of the lower crust below the oceanic plateau, it is possible  
to identify internal, external and boundary zones. The internal zone is in stark  
contrast with the low dM experiment is mainly characterized by RTIs that thins  
the crust. The boundary zone, instead, is not thickened as in the previous ex-  
periment, and it is characterized by the discontinuous delamination of the lower  
460 crust. Delamination and RTIs assist the expansion of the internal zone until  
the external zone is completely removed.

3rd stage: Quiescent (see Figure 3, 14.180 Myrs). After the removal of the  
old lithosphere, and the generation of a new compositional mantle lithosphere  
and composite crust, the simulations enter its quiescent stage. The amount  
465 of new felsic crust produced is small compared to the previous stages, and is  
associated with small drips of residual and intrusive mafic material. As for the  
previous numerical experiments, most of the felsic crust is buried to middle  
and lower crustal level due to the addition of new hydrated basalt during the  
removal stage (see Figure 3, 14.180 Myrs column d). The absolute amount and  
470 volumetric fraction of felsic crust is higher than for the low dM experiments.  
Due to the RTIs, the amount of felsic crust is not evenly distributed in the  
section, and the local felsic crust anomalies generate small domes, reproducing  
the typical dome-and-keel geometry of Archean terranes. The average depletion  
of the asthenospheric mantle is higher than in the low dM experiment and the  
475 mantle is cooler and less fertile.

We performed nine 2D numerical experiments in order to constrain the ef-  
fects of dM and  $M^{trs}$  of the mantle phases. Although  $M_{trs}$  controls the melt  
extraction and the volumetric melt fraction that is retained in the source, it  
has no effects on the geodynamic evolution nor on the absolute volume of the

480 felsic and mafic crust.  $dM$  exerts a strong control on the volume of extracted melts and on the geodynamic evolution of the system. To perform 3D numerical experiments, we choose to use the parameters of the high  $dM$  numerical experiments.

### 3.3. Effect of the potential temperature of the asthenosphere

485 We explore the effect of the  $T_P$  of the asthenosphere using 3D (run for 10 Myrs) and 2D (run for 40 Myrs) numerical experiments. Our results can be divided into two main regimes: high mantle potential temperature (high  $T_P$ ) and low mantle potential temperature regimes (low  $T_P$ ).

#### 3.3.1. High mantle potential regimes ( $T_P \geq 1500^\circ C$ )

490 *Reference experiment high potential temperature regime ( $dM = 0.05$ ,  $M^{trs} = 0.001$ ,  $T^{Pl} = 1600^\circ C$ ,  $R = 200$  km,  $T_P = 1500^\circ C$ )*

1st stage: Plume Arrival (see Fig. 4, 1.308 Myrs): As soon as the mantle plume approaches sub-lithospheric depths (200-150 km), it starts to produce melts. The spreading velocity of the mantle plume is limited ( $\sim 20$  cm/yrs),  
 495 forcing the melt extraction processes to occurs in a limited circular area. The first drip occurs soon after the penetration of the plume into the lithospheric mantle, while, ongoing mafic magma intrusion and extrusion results in magmatic thickening of the crust. During this thickening most of the TTG-like melts are produced at medium pressure condition (1.0-1.8 GPa). After the occurrence of  
 500 the first RTIs the extraction of MP-TTG like melts is focused on the boundary zone, while the internal zone is dominated by the extraction of low pressure felsic melt (see Fig. 4, 1.308 Myrs, column b).

2nd stage: Dripping, delamination and lithospheric removal (see Fig. 4, 2.108-2.593 Myrs): After the delamination of the first drip, the internal zone  
 505 radially extends. In the internal zone, new felsic and mafic crust is continuously produced causing RTIs, while in the boundary zone the lower crust of the external zone is removed via delamination (see Fig. 4, 2.108 Myrs column a). The delamination processes, that is occurring in the boundary zone, has two effects:

1) it allows a syn magmatic extension in the internal zone allowing its expansion;  
 510 and, 2) it deforms the crust of the external zone facilitating its sub-accretion.  
 The delamination drives a burial of the hydrated mafic crust, resulting in the  
 production of LP-MP-HP TTG like melts from the external to internal zones (see  
 Fig. 4, 2.108 Myrs column b). The internal zone is characterized by the con-  
 tinuous extraction of LP TTG like melts, with some of MP TTG-like melts  
 515 extraction. The internal zone is gravitationally thinned, as the continuous RTIs  
 locally cool the mantle, and the generation of residual mantle generates a stable  
 density profile beneath the new composite crust, preventing efficient astheno-  
 spheric ascent. However, this initial regime is not stable, because as soon as  
 the internal zone reaches a radius of 100-150 km, the delaminated crust in the  
 520 boundary zones segments in a discrete *slab*, and during their foundering, they  
 become similar to a RTIs (see Fig. 4, 2.362 Myrs column a ). The transition  
 from slab-like to drip-like gravitational instabilities induces a significant volume  
 of lithospheric mantle to be removed from the external zone, producing a syn-  
 magmatic thinning of the crust in front of the boundary zone. The removal of  
 525 the lithospheric mantle triggers mantle melting beneath the external zone which  
 ultimately results in the production of high volumes of LP-TTG like melts and  
 in the weakening of the crust (see Fig. 4, 2.362 Myrs column a-b). The de-  
 lamination/drip processes become extremely efficient, and the boundary zone is  
 destroyed by gravitational processes. The outward propagation of the internal  
 530 zone happens at high speed ( $\sim 100\text{cm/yr}$ ) and the remnants of the boundary  
 zone propagate radially removed by RTIs (see Fig. 4, 2.593 Myrs column a).  
 The interference between different RTIs generates local felsic crust anomalies,  
 and the new composite crust thins due to gravitational processes. The result is  
 to promote the extraction of felsic melts at extremely low pressure compared to  
 535 the initial stage.

3rd stage: Quiescent (see Fig. 4, 8.301 Myrs column a): After the removal of  
 the old lithosphere, and after a transition period dominated by RTIs, the exper-  
 iment enters in its final stage. The newly-generated crust is mostly composed of  
 mafic components ( $\sim 40\%\bar{F}$ ) (see Fig. 5, d). However, the felsic intrusive bod-

ies are not homogeneously distributed within the crust (see Fig 4 a-b-d). These felsic bodies are irregular and are surrounded by predominantly mafic crust. The felsic crust anomalies are correlated with positive topographic anomalies, while the area mainly composed of mafic lithologies are associated with negative topographic anomalies (see Fig 4 c). Since the numerical experiments runs only for 10 Myrs, we do not observe directly the partial convective overturn. However, the architecture of the new crust and the spatial distribution of the felsic components within its volume, predispose the occurrence of partial convective overturn.

During the evolution of the experiments significant volumes of mafic melt are extracted from the asthenosphere and only a minor part from the plume(see Fig 5 column b). The temperature at which the mafic melts are extracted varies from a maximum imposed by the potential temperature of the plume to a temperature lower than the initial background asthenospheric  $T_P$  ( $\sim -100^\circ C$ ).  $T^{Gen}$  (temperature of generation) of the mafic melts varies with time, and decreases exponentially with time. The distribution of the temperature of extraction varies with time as well, with the initial stage of experiments featuring a wider distribution of  $T^{Gen}$  compared to the quiescent stage. The variability of the mafic  $T^{Gen}$  is related to the cooling of the mantle and suggests a strong interaction between cold crustal material and asthenosphere. The mafic growth curve (cumulative volume of mafic melts extracted vs time,  $P^{Mafic}$ ) show an initial slow production rate followed by an exponential production rate increase during the removal stage.

The felsic crust growth curve (cumulative volume of felsic melts extracted vs time,  $P^{Felsic}$ ) has the same shape of the mafic growth one (see Fig 6 column a) suggesting a strong relation between mafic and felsic melts production. The volumetric fraction of TTG-like melts (TTG-opt) varies with time. Nevertheless, most of the felsic melts produced are extracted outside the optimum field at lower pressure and temperature (see Tab. S3). The proportion of LP, MP and HP TTG like melts changes with time, with an initial stage featuring significant volumetric fraction of MP-TTG like melts, which decreases with time during

the dripping stage. The gravitational thinning associated with RTIs within the internal zone favour the production of LP-TTG (see Fig 4) and the destruction of the boundary zone prevents an efficient burial of the mafic hydrated basalt to great depth.

575 The experiment featuring higher mantle potential temperature ( $T_P = 1550^\circ C$ ) shows the same evolution (see Fig 6). The only difference that can be reported is related to the absolute quantity of new crust generated which is higher than the reference scenarios in the highest  $T_P$  tested. We additionally tested the effect of changing the interval and the relative depth at which the mafic intrusions are  
580 emplaced. The main geodynamic scenario is not changing but the pressure at which felsic melts are extracted is higher than the reference scenarios resulting in larger amounts of TTG-like and MP-TTG-like melts.  $D_{Tot}$  (the thickness of the intrusion interval normalized over the current crustal thickness) exerts a second order control on the condition at which TTG-like melts are produced (see  
585 Fig 6 column a), while the  $D_{int}$  has a primary role, promoting the extraction of MP and HP TTG-like melts.

The 2D experiment (see Fig 3) shows first order similarities with the 3D numerical experiment, but there are some important differences to highlight (see Fig 4 column a). The first main difference is related to the fact that  
590 the boundary zone in the 2D experiments is not destroyed during the removal stage. The second difference is related to the volumetric fraction of TTG-like melts produced and volumetric fraction of MP-HP TTG like melts, which are higher in 2D scenarios.

### 3.3.2. Low mantle potential temperature regime

595 *Reference experiment low potential temperature regime ( $dM = 0.05$ ,  $M^{trs} = 0.001$ ,  $T^{Pl} = 1600^\circ C$ ,  $R = 200$  km,  $T_P = 1400^\circ C$ )*

1st stage: Plume arrival (Fig 7, 0.560 Myrs ): The first stage of the experiments is similar to the high potential temperature reference experiments. The only difference is related to the efficiency of mantle plume spreading beneath  
600 the mantle lithosphere. It is possible to observe the same pattern of felsic melt

extraction, with a more prominent extraction of LP-TTG like melts within the internal zone and the extraction of higher pressure melts in the boundary zone.

2nd stage: Delamination and sub accretion (Fig. 7, 1.139-8.524 Myrs): After the occurrence of the first RTIs, it is possible to recognize the internal, boundary  
605 and external zones. The internal zone expands thanks to the delamination of the lower crust of the external zone, and it gravitationally thins due to the RTIs (Fig. 7, 1.139 Myrs column a). The felsic crust produced within the internal zone is drifted apart towards the boundary zone, because of drag forces exerted by the astenospheric return flow. The main effect is promoting the extraction  
610 of LP-TTG in the internal zone and crustal thickening in the boundary zone (and thus MP TTG like extraction ) (Fig. 7, a). The boundary zone matures as a function of the amount of the felsic crust that is accreted at the rim of the internal zone (see Fig. 7, 3.395 Myrs), which additionally facilitates the external zone sub-accretion. The result of the maturation of the boundary zone  
615 is to create an annular continental terrain and forebasin. The pressure of felsic melt extraction within the external zone increases towards the internal zone mimicking a subduction-like process (see Fig.7 b).

The experiment does not reach the quiescent stage (Fig 7, 8.524 Myrs). And the processes described above stabilize the boundary zone generating a  
620 plate-like boundary feature. Within the boundary zone, the same extraction pattern is observed (LP-MP-HP TTG like melts) and the width of the annular continental terrain increases. At the end of the time span explored, the crust can be subdivided into three areas: the internal zone whose crust is mostly mafic with felsic intrusive bodies that are emerging locally due to the RTIs (see  
625 Fig. 9, a, b d); an annular continental terrain at the rim of the internal zone over-thrusting the external zone crust; a annular foreland basin whose depth is  $-1.0$  km (Fig.9 c). The annular continental terrains have a topography that spans from  $2.0$   $-3.0$  km (Fig.9 c) and internal zone featuring a relative high topography ( $0.6$ – $1.5$  km). The topographic gradient imposed by the annular  
630 continental terrain is high and facilitates the over-thrusting of the external zone mafic crust due to the gradient of gravitational potential energy between the

annular continental terranes and fore-land basin (similar to the results of Rey et al. (2014)). In the internal zone the topographic heights are associated with the amount of felsic crust, and the pattern of small basins is like those that have  
635 been observed in the reference high  $T_P$  scenarios (Fig.9 c-d).

The amount of new mafic crust generated increases linearly with time (see Fig 10, column b, first row) and  $T^{Gen}$  of the mafic melts varies from the mantle potential temperature to a temperature lower than the initially prescribed asthenosphere  $T_P$ . The variability of the mafic melt extraction temperature  
640 decreases with time indicating a progressive cooling of the asthenosphere due to the continuous delamination/RTI processes.

The volumetric fraction of TTG-like melts is higher than high  $T_P$  experiments, and the proportion of LP-MP-HP TTG like melts is variable. The volumetric fraction of LP TTG-like melts decreases with time, while the MP  
645 and HP TTG like melts proportion increases (see Fig 10, column a, first row). The decrease of LP-TTG like melts is a consequence of the stabilization of the boundary zone.

A higher initial  $T_P$  (1450 °C) does not change the main geodynamic processes but allow a faster propagation of the internal zone. Therefore, the internal  
650 zone reaches the boundary of the numerical domain, generating similar processes to those in the high potential—temperature experiments. However, in this case, rather than being a physical transition, it is a numerical artifact that is a consequence of the limited size of the numerical domain. The removal of the old lithosphere can be observed in both mafic and felsic growth curves, because they  
655 shift from linear to super-exponential growth rate (see Fig 10 a,b, second row). Lowering the initial mantle potential temperature (1350 °C) results in a twofold decrease in the volume of both mafic and felsic crust respect with the low  $T_P$  reference scenario. The volumetric fraction of TTG-like melts is low, and the proportion of HP and MP TTG melts are low. Decreasing the initial thickness  
660 of the lithosphere does not affect the previous results but reduces the proportion of MP and HP TTG-like melts and is associated with a higher production of felsic and mafic melts.



We tested the same setup using a 2D numerical experiment. The main difference is related to the volumetric fraction of TTG-opt, which is higher. However, LP TTG like melts are proportionally the most common type of melt at low  $T_P$ , this result is in stark contrast with the high  $T_P$  one. Exploiting the fact that the low  $T_P$  2D and 3D numerical experiments are similar, and do not show the oddities observed in the high  $T_P$  ones, we performed additional low  $T_P$  experiments to observe the long term effect of the thermal boundary condition and radiogenic heating and initial lithospheric thickness (see Fig. S6). Lowering only the thickness and the temperature at the base of the numerical domain in 2D numerical experiments yields a two-stage experiment, in which the expansion of the internal zone slows down with time and shuts down for a 3-4 Myrs (see Fig ?? and S7). However, due to the high radiogenic heat production (which decreases as a function of the depletion step, see Methods and Tab S2), the mantle heats up with time, producing a new melting stage allowing the reinitiation of the expansion until all the old-lithosphere is removed. If all mantle rocks feature a low radiogenic heat productivity (similar to the Phanerozoic one,  $A = 0.33 \cdot 10^{-5} \text{ W/kg}$ ) the outward expansion of the internal zone simply stops, and the experiments remains in a quiescent stage during its whole duration.

### 3.3.3. *Difference between 2D and 3D numerical experiments*

2D and 3D numerical experiments have important differences, which are significant in high asthenosphere  $T_P$  scenarios. These differences are reflected into the production of felsic crust, which is mostly extracted at low pressure in 3D numerical experiments in contrast with higher pressures in the 2D experiments. These differences emerge because of the RTIs are more strongly developed in 3D cases than in 2D numerical experiments. The differences arise from the limited spread velocity of any upwelling in 3D cases (see Fig 11). The spreading velocity (the x and y velocity components of velocity) controls the distribution of the newly generated crust, which results in a more localized crustal thickness anomaly. The spread velocity beneath the lithosphere depends on the

temperature contrast between normal asthenosphere and plume, therefore, at high mantle potential temperature the spread is limited, generating thickened plateau, and promoting an unstable crust. In the 2D numerical experiments the lateral spreading velocity of the plume is higher than 3D cases, allowing the generation of a moderately thickened and wider oceanic plateau: this promote the melting of hydrated crust at moderate pressure, and generates a more stable configuration. 2D cases featuring a lower asthenosphere  $T_P$  yield a thinner oceanic plateau that promotes lower pressure TTG generation. In 3D cases, most of the mafic-melts are extracted along the plume axis direction, yielding an ultra-thickened and narrow plateau, which is more unstable, ultimately resulting in stronger RTIs and thus efficient gravitational thinning.

## 4. Discussion

### 4.1. Limitations

The resolution of the numerical experiments may affect the geodynamic evolution and the estimates of the felsic/mafic crust produced. We performed the 2D numerical experiment shown in Fig. 3, with different grid sizes (see Tab S3) to assess the impact of this effect. The described geodynamic processes do not show any significant first order differences as a function of the resolution. The only remarkable difference is related to the RTIs, as their size changes as function of the ability of the numerical code to resolve the crust. The numerical resolution seems to not affect the total amount of mafic crust volume and the final crustal thickness at the end of the experiments. The normalized total mafic productivities of all simulations with respect to the average ranges from -0.04 to 0.06, with no clear trend as a function of the resolution ( $\sim \pm 13 \cdot 10^3 km^3$ ). The final crust thickness shows similar variability, decrease with resolution ranging from -0.08 (high resolution) to 0.08 (low resolution) ( $\sim \pm 1.5 km$ ). The quantity that is most affected is the absolute volume of felsic crust produced, which increases with the resolution. But instead of being caused by the resolution dependency of our melt extraction algorithm, it is associated with the resolution

of the crust. The volume of felsic crust and volumetric fraction of TTG-like melts increase with the resolution. The relative proportion of each LP, MP and HP is also affected, with MP-TTG are becoming dominant at extremely high  
725 resolution (i.e. 513x513 nodes per direction), largely at the expense LP-TTG like melts and minorly at the expense of HP-TTG like melts. Nevertheless, the distribution of the P-T conditions are consistent within these numerical experiments (see Fig. S8). In a broad sense our results still hold at low and high resolution, although, it is necessary to consider the uncertainty related with the  
730 resolution of the experiments.

Most of the numerical simulations enters in a quiescent stage. This stage is influenced by the size of the numerical box and by the mantle melt productivity, so it can be interpreted as a saturation stage. Despite this limitation, which hinders an extrapolation of our the results to a wider numerical domain, they  
735 key physical processes are not affected.

Our initial scenario depicts a thermal (small) plume without tail. This plume is not fed by any buoyancy flux coming from the lower mantle. This is done to model a plume that was active during a stagnant lid convection modes, in which long lived large plumes are unlikely to remain stagnant for long periods  
740 of time. At the end of the numerical experiments the crustal thickness is thin ( $\sim 28km$ ) with respect to the inferred scenarios ( $35 - 50km$ ) (Van Kranendonk et al., 2015; Van Kranendonk, 2010; Smithies et al., 2009) and most likely a long-lasting mantle plume would yield a thicker crust. However, most of the thinning processes are associated with gravitational instabilities, that in case of  
745 a long-lasting plume are expected to be more effective.

We introduce a novel approach to deal with the compositional variation due to magmatic processes. However, it introduces several issues such as stepwise variations of properties. The stepwise variations of density are reduced by the interpolation scheme employed and their effect is minor compared to the actual  
750 phase reactions observed within the same diagram. Radiogenic heat production is handled discretely by decreasing its value as a function of the depletion step. This may affect the long-term results of our numerical simulation, which (to-

gether with the extremely high computational cost of each simulations) is one of the reasons that we opt for short-term experiments. The most suitable way to handle radiogenic heat production is to enforce a partition coefficient of radionuclides as a function of the melt produced. However, in order to do that, it is necessary to assume a partition coefficient and the partition of radionuclides on the fly computed as a function of mineralogical composition (e.g. Rummel et al. (2018)) further increases the computational cost of the simulation. Moreover, we assume that the average composition of the residuum along a melt isoline is representative of all the chemical complexity. Previous work (White et al., 2017; Palin et al., 2016a,b) state that within the compositional range that is explored here, the bulk composition has a minor effect on the topology of the phase diagram and on the composition of the relative melts, so densities and volumetric melt fractions are not affected. The last limitations is associated with the usage of the phase diagrams. We assume that the thermodynamic equilibrium is always reached and we neglect the effects of metastability and the kinetic of reactions that can control the timescales of the processes.

#### 4.2. Mafic crust production

The primary mafic melt composition depends on the thermal state of the mantle. Consequently, the basalts that feature a composition close to the primary melts have been widely employed to estimate Archean mantle  $T_P$  (Ganne & Feng, 2017; Herzberg et al., 2010). The debate concerning the upper mantle potential temperature is still open with estimates spanning from 1400 to 1600 °C (Aulbach & Arndt, 2019; Herzberg et al., 2010). Moreover, these estimates show considerable variations at any one point in time (Ganne & Feng, 2017; Herzberg et al., 2010) suggesting that this variability may be due to the interaction of a relatively cold asthenosphere and mantle plumes (Kamber, 2015; Arndt, 2013). Our results suggest that the mafic crust composition may be variable with high and low #Mg mafic rocks coexisting consistently with many Archean terranes such as Pilbara or Kapavaal (Van Kranendonk et al., 2015).

The temperature of extraction of mafic melt is variable at any timestep.

The colder mafic melts indicate a strong interaction with the delaminated crust. Although we do not consider chemical exchange, degassing and magmatic hybridization in our simulations, this thermal exchange must result in at least one of these processes (Bédard, 2006). The mantle may be hydrated and contaminated by the dripped crustal material. These processes are feasible and could modify the composition of the mafic melt opening new perspectives for TTGs petrogenesis. For instance, the hybridization between the mantle and crustal related melts has been invoked to produce the mafic source of East Pilbara TTGs (Smithies et al., 2009; ?; Johnson et al., 2017). If the mantle is sufficiently hydrated, the mafic intrusions can carry enough water to yield TTG-like melts during fractional crystallization or subsequent partial melting events Smithies et al. (2019); Van Kranendonk et al. (2015); Macpherson et al. (2006). Despite the fact that our simplified approach cannot reproduce all these processes of contamination, our results indicate that they are possible and need to be explored in a consistent geodynamic and petrological way.

#### 4.3. Felsic crust production

Most of the felsic melts are extracted outside the optimum field with significant differences between 2D and 3D experiments: 2D experiments generally show a higher proportion of TTG-like melts (40-50%) respect 3D (20-30%).

The felsic melts are produced at low pressure and low-temperature conditions ( $T^{Gen} < 700^{\circ}C$ ) entailing that most of the complementary mafic residuum is produced at low depths see Tab. S3. The production of complementary mafic residuum facilitates the gravitational instabilities (Sizova et al., 2015; Piccolo et al., 2019), and in our experiments limits the maximum thickness of the crust ( $\sim 28$  km). The combination of the emplacement of mafic dry intrusions and low crustal thickness yields a thin and hot crust. In Fischer & Gerya (2016) the dripping stage of the simulation is associated with the thinning of the crust and they observed a quiescent stage characterized by magmatic thickening. The dripping stage is the most productive in terms of felsic crust production, it is associated with low pressure and low-temperature magmatism, outside the

TTG optimum field. The production of generally LP felsic melts has been identified even in global scale 2D numerical experiments (Jain et al., 2019). To  
815 promote high temperature/pressure partial melting, it is necessary to reconsider the productivity of the mafic protolith or to increase the viscosity of the crust. All the mafic crust phase diagrams stem from BS0%, which is computed using an EAT composition, and assuming that is hydrated enough to be minimally saturated at 1.2 GPa (Palin et al., 2016a). This composition represents the  
820 best scenarios for the generation of felsic crust, as it maximizes the felsic melt generation. This saturated composition has a low solidus temperature close to 650 °C, and it makes the mafic protolith susceptible to thermal perturbation introduced by mafic and felsic intrusive bodies. For instance, the volume of TTG-like melts increases if the depth of intrusions is close to the Moho depth  
825 and the depth interval of emplacement is narrow (see Tab. S3). Therefore, one potential solution to these issues is to consider the mafic protolith to be hydrated but undersaturated. Lower water content shifts the temperature of the solidus to higher temperatures (Palin et al., 2016a,b), promoting high-temperature melting (within the optimum field) and may inhibit crustal-scale RTIs.

830 Most of the TTG-like melts are produced at low-pressure conditions, with a variable amount of MP-TTG like crust as a function of the depth of intrusion and the initial mantle potential temperature. Our results are not consistent with the world-wide TTGs data set (Moyen, 2011), firstly because we do not produce enough TTG-like melts, secondly because our TTG-like melts are dom-  
835 inated by LP-TTGs. Despite that, our results are broadly consistent with the current proportion of LP and MP TTGs of the East Pilbara Paleo-Archean TTGs (Johnson et al., 2017). The dominance of LP-TTGs in our simulations may be an artifact, as the bulk composition of the mafic protolith has not directly been addressed to produce the LP, MP and HP TTGs classification  
840 (Moyen, 2011). For instance, it has been inferred that the mafic protolith of Paleo-Archean East Pilbara TTGs are the Coucal basalts (Johnson et al., 2017; Smithies et al., 2009). These basalts have a composition that can potentially bear the required mineralogical assemblage to generate MP-TTGs at a lower

pressure (0.7-0.8 GPa) than that proposed by Moyen (2011). In particular,  
845 all the three types of TTG melts were suggested by (Johnson et al., 2017) to  
form at a pressure lower than 1.4 GPa. The LP, MP and HP TTGs classifica-  
tion reflect the occurrence of the required mineralogical assemblages and does  
not represent a rigid set of P-T conditions. Thus, simulations that specifically  
aim to create the "right" proportion of LP, MP and HP TTGs in geodynamic  
850 modelling without accounting for the variability of mafic source protolith will  
produce misleading results.

#### 4.4. Structural evolution

We identify two main regimes: low and high  $T_P$ . In both the scenarios,  
the upper crust of this composite crust is dominated by mafic effusive rocks,  
855 consistently with the inferred Archean crustal composition and greenstone belts  
(Kamber, 2010; Condie, 1993).

*High  $T_P$  regimes ( $\geq 1500^\circ\text{C}$ ).* Most of the tectonic evolution is occurring due  
to the RTIs. We observed structures that resemble dome-and-keel geometry  
which has been widely recognized as the effect of partial convective overturn  
860 (Van Kranendonk, 2010; Van Kranendonk et al., 2004). However partial con-  
vective overturn occurs after a long thermal maturation of the gneiss dome due  
to the high radiogenic heat production (François et al., 2014; Bodorkos et al.,  
2006), or because of the emplacement of hot intrusions (Van Kranendonk et al.,  
2015). In our numerical experiments, the apparent rising of felsic domes is a  
865 consequence of RTIs and delamination induced deformation. Thus, rather than  
being the product of a partial convective overturn, these anomalies predispose  
the felsic crust to rise in the subsequent mantle-related magmatic events. East  
Pilbara has a long-lasting structural evolution that started during the Eo/Paleo-  
Archean and ended during the Meso-Archean (3.65-3.22 Ga) (Van Kranendonk  
870 et al., 2015; Hickman & Van Kranendonk, 2012). The evolution of this Archean  
terrane and the relative subcontinental mantle lithosphere (SCLM), has been  
divided into three main stages. The first related to the generation of a primi-  
tive and mostly undifferentiated mafic crust (our initial conditions), the second

related to plume-related magmatism, and generation of the first TTGs and the  
875 last one concerned with the stabilization of the cratonic lithosphere (Van Kra-  
nendonk et al., 2015). Our results, especially at high  $T_P$ , represents the second  
stage, in which the hot mantle upwelling starts interacting with the proto litho-  
sphere. During this stage, a primitive continental crust is slowly generated, with  
a composite mafic volcanic sequence composed of komatiites, high Mg basalts,  
880 and tholeiite (Van Kranendonk et al., 2015). During this stage, the felsic crust  
generation is associated with delamination and dripping events, that catalyze  
the production of an ultra-depleted mantle and felsic crust production. In this  
stage, the mantle is potentially hydrated and contaminated by the delaminated  
crust.

885 *Low  $T_P$  regimes ( $\leq 1450^\circ C$ ).* In these experiments, there are structural fea-  
tures that resemble plate boundaries that are associated with the production of  
middle- to high-pressure felsic melts. Within the same settings coexist differ-  
ent structural patterns and modes of felsic melts production. The lithospheric  
mantle is rigid and less prone to be captured by the delaminating crust pre-  
890 venting the asthenosphere from partially melting outside the internal zone. The  
propagation of the internal zone generates annular continental terranes that as-  
sist the sub-accretion of the old-prescribed lithosphere. Itsaq gneiss complex  
(West Greenland) is an old archean terranes mainly composed of felsic gneiss  
that records a potential long phase of horizontal tectonics (Nutman et al., 2013,  
895 2009). The horizontal tectonic phases have been interpreted as a consequence of  
subduction-related processes (Nutman et al., 2013, 2009). Sizova et al. (2015)  
using 2D numerical experiments shows that subduction processes can arise as  
a consequence of the differentiation processes, and associated the generation of  
predominantly felsic crust terranes with transient subduction events. Our re-  
900 sults, for the first time, show that vertical tectonic processes generate significant  
horizontal displacement, generating stacking of felsic crust slivers due to the co-  
operation of RTIs, delamination and mantle drag. Our results show that it is  
not necessary neither to have a long-lasting nor a transient subduction to gen-



erate these high grade felsic terranes. Although, we do not explicitly simulate  
905 the felsic crust reworking, these thickened continental terranes would eventually  
undergo to partial melting generating further evolved continental crust. It is  
widely believed that continental crust forming processes evolve with the secular  
cooling of the planet (?). Several lines of research suggest that the felsic crust  
produced become progressively potassic and that the pressure of generation of  
910 the felsic melts increased (?Laurent et al., 2014). Our low  $T_P$  regimes show  
behavior that is consistent with these observations and importantly show that  
these contrasting P-T conditions may coexists together.

In our numerical experiments, the interaction between the transient plume  
and lithosphere generates a semi-circular structure that radially expands with  
915 velocities that depends on the background  $T_P$ . In both explored regimes the  
propagation of the internal zone is accompanied by disintegration of the newly-  
produced crust, and its lateral displacement over several 100s km. These struc-  
tural feature resembles the coronae structure on the surface of Venus (Harris  
& Bédard, 2014; Stofan et al., 1992). Most of the deformation is governed by  
920 three main processes: the drag force exerted by the asthenosphere that rises as  
a consequence of the RTIs and delamination processes; pulls forces exerted by  
the foundering drips and delamination of the lower crust at the boundary zones.  
Our results support that a limited mobile lid is possible without requiring plate  
tectonics.

## 925 5. Conclusion

Our 2D-3D numerical experiments predict several fundamental differences  
of the plume-lid interaction as a function of the initial thickness and initial  $T_P$ .  
At high  $T_P$  most of the deformation and production of new mafic crust is medi-  
ated by the asthenospheric mantle flow resulting from gravitational instabilities  
930 ( $T_P \geq 1500^\circ\text{C}$ ). At low  $T_P$  ( $T_P \leq 1450^\circ\text{C}$ ) most of the newly generated crust is  
closely related to plume activity. This variability of temperature at which basalt  
is generated supports the idea that upper mantle  $T_P$  may have been lower, and

indicate that mafic crust is composed of different mafic lithologies. As a consequence, it is necessary to address the effect of this variability to constraint  
935 the condition at which natural Archean TTGs are generated. Our numerical experiments predicts that significant amount of new felsic crust are produced, even though, the P-T conditions do not match the most suitable ones to produce high amounts of TTGs-like melts, and further investigation are required.

## 6. Acknowledgments

940 DFG grant, SPP 1833 Building a Habitable Earth has supported this research. B.J.P.K. has been supported by MAGMA Consolidator Grant (ERC project 71143). The data are accessible upon request to the first author. We acknowledge Adina E. Püsk and Tobias S. Baumann for the useful suggestion that improved the visualization of our numerical experiments, Anton A. Popov  
945 for the useful suggestions and tip provided during the implementation of melt extraction algorithm, and N. Berlie, N. Riel and E. Moulas for the discussions and insights provided. The manuscript has been significantly improved thanks to comments of J. Bedard, J. Van Hunen and an anonymous reviewer.

Table 1: Composition of mantle phases diagram. All the compositions are listed as mole % oxide.

Table 2: Composition of basalts. All the compositions are listed as mole % oxide. (\*) Basalt Step 45% has the same phase diagram of Basalt Step 30 %, but, after the extraction, the predicted composition is the one listed. Basalt Step 45% is used to describe the density of the dry Intrusions.

Figure 1: Schematic representation of melt-extraction algorithm. For each timestep all the rock-type properties are interpolated into the main grid. Then from pressure and temperature, the density and melt quantity are interpolated from the phase diagrams and scaled accordingly to the Phase ratio ( $Ph_{Rat}$ ). After the computation of the total melt extracted, the negative and positive source term are applied to the grid. The intrusion emplacement depth interval is chosen according to the relative depth to the Moho, and the injected volume is evenly distributed along this depth interval

Figure 2: Main evolutionary stages of the low dM numerical experiment. The sub-plots of the 1st and 2nd stage zoom in on the lithospheric scale processes (-20 -200 km). Column a) features sub-plots represents the compositional field, while column b) shows several snapshots of the buoyancy contrast that controls the evolution of the numerical experiment, the black thick lines represent the internal free surface and the Moho topography.  $\Delta\rho$  is computed by subtracting the density field along x direction for a given depth from the average density of the considered depth ( $\Delta\rho(x, \mathbf{z}) = \rho(x, \mathbf{z}) - \rho_{mean}(\mathbf{z})$ , where x represents the x coordinate, and z represents the given depth). Therefore, this parameter represents the excess (negative values) and the deficiency (positive) of mass with respect to the background density. The snapshots of the 3rd stage represents the full numerical domain during the quiescent stage. c) represents the full compositional domain, while d) represents both the volumetric fraction of melt extracted from the mantle during the whole simulation, and the volumetric fraction of felsic crust components within the crust. The black thick lines represent the free surface and the Moho depth.

Figure 3: Main evolutionary stages of the high dM numerical experiment. The sub-plots of the 1st and 2nd stage zoom in on the lithospheric scale processes (-20 -200 km). Column a) features sub-plots represents the compositional field, while the column b) shows several snapshots of the buoyancy contrast (see the caption of Fig.2 for further explanations), the black thick lines represent the internal free surface and the Moho topography. The snapshots of the 3rd stage represents the full numerical domain during the quiescent stage. c) represents the full compositional domain, while d) is a plot that represents both the volumetric fraction of melt extracted from the mantle during the whole simulation, and the volumetric fraction of felsic crust (**F**) components within the crust. The black thick lines represent the internal free surface and the Moho topography.

Figure 4: Temporal evolution of the high potential temperature reference scenario ( $T_P = 1500$  °C,  $R=200$  km,  $T^{Pl} = 1600$  °C,  $dM=0.05$ ,  $M^{trs} = 0.001$ ,  $I_R=0.6$ ). The 1st and 2nd stage zoom in on the first 200 km of the numerical domain. The column a) represents a simplified compositional field (mafic crust collects all the mafic protolith phases, the residual mantle comprises all the mantle residual phases generated during the ongoing simulation, felsic crust represents the new granitoids emplaced, and red colour represent the initial plume). b) 2D maps that represent the distribution of LP (green dots), MP (blue dots) and HP (red dots) TTG like melts extracted; c) Represents simplified compositional field of the whole numerical domain.

Figure 5: Final composition of the crust, and the effect of crustal material on the topography (the topography is computed using as a reference the average depth of the internal free surface). a) The plot represents the fractional volume of felsic components ( $\mathbf{F}$ ) within the crust; the yellow lines represent the direction of the profiles that are shown in b). b) Slices of the figure a). The volumetric fraction of felsic components is not homogeneously distributed within the volume of the crust. The highest values of  $\mathbf{F}$  are associated with felsic domes, with irregular shapes. The black thick lines represent the free surface and the Moho depth. c) 2D contour of topography. The highest elevations are associated with the felsic domes seen in a), while the basins are associated with a relatively mafic component enriched crust; d) 2D map representing the crustal felsic crust composition ( $\bar{F}$ ).  $\bar{F}$  represents the volumetric fraction of felsic components within a column of material.

Figure 6: Summary of the felsic and mafic melt extraction P-T conditions of the high  $T_P$  regimes experiments. Each row represents different experiments. The parameters that have been changed respect the high  $T_P$  reference scenarios shown in Fig. 4 are highlighted in red. The last row is the 2D numerical experiments, which feature the same input parameter of the high  $T_P$  reference scenarios. The evolution of this experiment is shown in Fig. 3 .a) Felsic melt extraction conditions: left axis and black thick line represents the cumulative volume of felsic melts produced during the simulation (i.e. the felsic crustal growth curve); the right axis and the red thick line represents the cumulative volumetric fraction of TTG-like melts produced during the simulation. The shaded green, blue and red area represents LP-TTG, MP-TTG and HP-TTG like melts cumulative volumetric fraction respectively. b) Mafic melts extraction conditions: left axis and black thick line represents the cumulative volume of mafic melts extracted (i.e. the mafic crustal growth curve) during the simulation; right axis and red thick line represent the temperature of extraction of mafic melts ( $T^{Gen}$ ). The blue shaded area represents the temperature conditions in which 5-25% and 75-95% of melts are extracted. While the red shaded area represents the standard deviation of the  $T^{Gen}$  during the simulation evolution. The initial temperature and final temperature are highlighted in the right of each plot to highlight the effective cooling of the mantle during the whole simulation.

Figure 7: Temporal evolution of the low  $T_P$  reference scenario ( $T_P = 1400$  °C,  $R=200$  km,  $T^{Pl}= 1600$  °C,  $dM=0.05$ ,  $M^{trs} = 0.001$ ,  $I_R=0.6$ ). The column a) represents a simplified compositional field (mafic crust collects all the mafic protolith phases, the residual mantle comprises all the mantle residual phases generated during the ongoing simulation, felsic crust represents the new granitoids emplaced, and red colour represent the initial plume). All the figures represents the compositional field evolution in the first 200 km of the numerical domain, except for the figure in the bottom row that represents the whole numerical domain. The column b) are a set of 2D map that represent the distribution of LP (green dots), MP (blue dots) and HP (red dots) TTG like melts extracted; c) Represents simplified compositional field of the whole numerical domain.

Figure 8: Two crustal scale slices of the second invariant of the strain rate. The profiles are taken along the two thick red lines shown in the compositional plot at 1.139 and 3.395 Myrs in Fig. 7. These profiles focus on the evolution of the boundary zone, that thickens with time. The thickening is associated with the over thrusting of the annular continental terrain into the external zone mafic crust. The over thrusting processes are associated with a burial of the mafic crust and allows the production of MP and HP TTG like melts.

Figure 9: Final composition of the crust, and the effect of crustal material on the topography (the topography is computed using as a reference the average depth of the internal free surface) of the low potential temperature reference scenario. a) Represents the volume fraction of felsic crustal component within the crust. The thick yellow line represents the direction of the slices that are showed in b). b) Slice of volumetric fraction of felsic crust material. c) Topography of the experiments at the end of the explored timescale. The elevation distribution is correlated with the anomalous volume of felsic material. The strongest topographic gradient is observed in the boundary zone due to the over thrusting of the annular continental terranes into the external zone mafic crust. d) Felsic crustal composition map  $\bar{F}$ .

Figure 10: Summary of the felsic and mafic melt extraction P-T conditions of the low  $T_P$  regimes experiments. Each row represents different experiments. The parameters that have been changed respect the low  $T_P$  reference scenarios shown in Fig. 8 are highlighted in red. The last row is the 2D numerical experiments, which feature the same input parameter of the low  $T_P$  reference scenarios .a) Felsic melt extraction conditions: left axis and black thick line represents the cumulative volume of felsic melts produced during the simulation (i.e. the felsic crustal growth curve); the right axis and the red thick line represents the cumulative volumetric fraction of TTG-like melts produced during the simulation. The shaded green, blue and red area represents LP-TTG, MP-TTG and HP-TTG like melts cumulative volumetric fraction respectively (the proportion of each category at the end of the numerical experiments are listed in Tab. S3). b) Mafic melts extraction conditions: left axis and black thick line represents the cumulative volume of mafic melts extracted (i.e. the mafic crustal growth curve) during the simulation; right axis and red thick line represent the temperature of extraction of mafic melts ( $T^{Gen}$ ). The blue shaded area represents the temperature conditions in which 5-25% and 75-95% of melts are extracted. While the red shaded area represents the standard deviation of the  $T^{Gen}$  during the simulation evolution.

Figure 11: Velocities profiles of both 2D and 3D numerical test shortly before the removal stage, and the evolution of the crustal thickness. At the bottom of the figure, two simple sketches highlight the main difference of 2D and 3D numerical experiments. a) These two pictures represent the velocity field observed in the 3D numerical simulation shown in Fig.4. The black thick lines represent the z component of the velocity along the plume axis, while the red lines represent the x or y velocity component took at  $\pm 50$  km along x or y direction from the plume axis. b) This picture represents the velocity field observed in the 2D numerical experiments shown in Fig.3. The black line represents the z component of the velocity along the plume axis, while the red line the x components of the velocity field taken at  $\pm 50$  km along the x direction from the plume axis. c) Average crustal thickness evolution with time of the 3D numerical experiments; d) Average crustal thickness evolution of the 2D numerical experiment.

## 7. Supplementary Material

### 950 8. Post-processing and hardware information

LaMEM is a finite difference staggered-grid code. Melt extraction is performed in the central nodes, which are the one used to compute the continuity equation and where temperature and pressure are defined. The output grid used for visualization and data analysis is defined on the corner nodes of the staggered-grid, and all the variables are interpolated on the visualization grid. 955 This interpolation might affect melt extraction temperatures, as the interpolation slightly smoothen the temperature variation. Furthermore, most of the data are picked each time the output is created. This strategy was necessary to minimize the amount of memory needed to store the numerical experiments. 960 As a consequence, the melt extraction events that we used in our analysis are affected by both interpolation scheme and the number of output steps that are produced. The data concerning TTGs melting condition represents a small population of the total conditions of extractions. 2D experiments are less memory demanding than 3D, and we saved the output every 50 timesteps, while 3D experiments 965 outputs were saved every 200 timesteps. 2D numerical experiments required 10-20 thousands timesteps to be completed ( $\sim 40$  Myrs model time, 2-3 weeks per run), while 3D experiments required 30-40 thousands timesteps ( $\sim 10$  Myrs model time, 2-3 months per simulations on 512 cores).

#### 970 8.1. Moho identification and internal free surfaces

We divided the rock-types into three main categories: 1) Mantle ( $MI$ ); 2) Mafic crust ( $MC$ ); 3) Continental crust ( $CC$ ). At each timestep, the phase proportion of the three categories are computed by summing the phase ratio of mantle, mafic and continental rock-types respectively. To constraint the Moho 975 depth LaMEM sums the contribution of  $CC$  and  $MC$ , after which it loops from bottom to top saving the  $z$  coordinates of the nodes featuring  $CC + MC < 0.8$  (during this operation the air phases is considered part of the crust as well).



At the beginning of each simulation the free surface is defined and shifted accordingly to the velocity field at the end of each timestep. The topography is  
980 corrected if its angle with respect to the horizontal plane is higher than  $30^\circ$ . The topography is measured using as reference the average depth of the free surface. The thickness is computed before solving the system of equations, and it is used to inject particles as a function of melt extraction parameters. The cumulative melt production is computed by summing the total melt ex-  
985 tracted along  $z$  direction, at every timestep:

$$V_{cum}^{Tot}(i, j, Ml/MC, t) = \sum_{dts=1}^{ts} V_{ext}^{Tot}(i, j, Ml/MC)(i) \quad (S1)$$

$i, j$  are the grid indexes,  $V_{cum}^{tot}$  is the total volume cumulated during the whole simulation and computed for  $MC$  or  $Ml$ ,  $ts$  is the current timestep ( $ts$ ), while  $dts$  is the timestep increment. Generation temperatures ( $T^{Gen}$ ) are computed  
990 each timestep, averaging the temperature at which the mafic melts are produced along  $z$  direction and considering the volumetric contribution of each rock-type.  $F$ , the relative amount of felsic crust, is computed by integrating all the  $CC$  phase ratio along  $z$  direction divided by the local thickness ( $Tk$ ) of the crust. The integration is within the local Moho depth and the topography ( $Topo$ ):

$$\bar{F} = \frac{\sum_{k=Topo(i,j)}^{k_{Moho(i,j)}} CC(k) \Delta z(k)}{\delta_{crust}(i, j)} \quad (S2)$$

## 8.2. Hardware information and solver option

The simulations were performed using the supercomputers offered by Johannes Gutenberg University, MOGON and MOGONII (hpc.uni-mainz.de). LaMEM is a PETSc based code (Balay et al., 2018), and the solving of the  
1000 non linear system of equation were performed thanks to the PETSc infrastructure. 2D experiments were performed using 8 cores and using the direct solver *mumps*, the internal linear solver is *fgmres* with a maximum of 30 linear iteration, and relative tolerance of  $10^{-6}$ ; while the external non linear solver were allowed to iterate maximum 30 times, and the relative tolerance was set to be

1005  $10^{-6}$ . 3D numerical experiments were performed using 512 processors and using  
Eisenstadt-Wanker methods, with a tolerance spanning from  $10^{-1}$  to  $10^{-4}$ ; the  
internal linear solver *fgmres* and the number of iteration allowed spans from 4  
to 180.

## 9. Supplementary table and figures

### 1010 9.1. Supplementary Tables

Table S1: List of symbols and relative unit of measure

Symbol	S.I. unit	Definition
$v_i$	$ms^{-1}$	velocity vector component
$\mathbf{S}$	$s^{-1}$	Negative and positive volumetric source terms
$\tau_{ij}, \tau_{II}, P$	$Pa$	Deviatoric stress component, 2nd invariant of the deviatoric stress tensor, Pressure
$\rho$	$kgm^{-3}$	Density
$g$	$ms^{-2}$	Gravity acceleration
$\dot{\epsilon}_{ij}, \dot{\epsilon}_{ij}^{el,pl,vis}, \dot{\epsilon}_{II}$	$s^{-1}$	Strain rate component, elastic, visco, plastic strain rate component, 2nd invariant strain rate tensor
$\dot{\sigma}$	$Pa s^{-1}$	Objective deviatoric stress component time derivative
$\omega_{ij}$	$s^{-1}$	Spin tensor component
$G$	$GPa$	Shear modulus
$\eta$	$Pa s$	Viscosity
$B$	$MPa^n s^{-1}$	Pre-exponential factor
$n$	n.d.	Stress exponent
$E_a$	$kJmol^{-1}$	Activation energy
$V_a$	$JMPa^{-1}mol^{-1}$	Activation volume
$\phi, \phi_0, \phi_1$	//	friction angle, initial friction angle, final friction angle
$C, C_0, C_1$	$MPa$	cohesion, initial cohesion, final cohesion
$T$	$K, ^\circ C$	Temperature
$k$	$WK^{-1}$	Heat conductivity
$A$	$\mu Wkg^{-1}$	Radiogenic heating per unit of mass
$H_a, H_s, H_r$	$Wm^{-3}$	Adiabatic, Shear and Radiogenic heating per unit of volume
$Ph_{Rat}$	n.d.	Phase proportion
$M_{ext}$	n.d.	Melt fraction extracted
$V_{ext}$	$m^3$	Extracted volume
$V_{ext}^{tot}$	$m^3$	Total volume extracted along $z$
$Vol_{Eff}, Vol_{Int}$	$m^3$	Effusion and intrusion volume
$Vol_{cor}$	n.d.	Volumetric correction
$v_{int}$	$m^3m^{-1}$	Volume of intrusion per unit of length
$D_{min}, D_{max}$	$m$	Extreme of depth interval of intrusion
$D, D_{int}$	n.d.	Relative distance from Moho, and relative half interval of intrusion
$\delta_{crust}, \delta_{Tkl}$	$km$	Crust and lithosphere thickness
$\bar{F}, \mathbf{F}$	n.d.	Average amount of felsic crust components along $z$ direction and proportion of felsic crust components at $x, y, z$ coordinates
$P_{Mafic/Felsic}$	$[km^3]$	Total amount of mafic or felsic crust produced.

Table S2: List of the petrophysical properties and phase diagrams. The symbol and relative unit of measures are listed Tab. S1. All rock types share the same: shear modulus, ( $G = 40GPa$ ); initial and final friction angle ( $\phi_0 = 30^\circ$  and  $\phi_1 = 9^\circ$ ); initial and final cohesion ( $C_0 = 10MPa$  and  $C_1=3 MPa$ ); heat capacity ( $C_p = 1200 J/K/Kg$ ); thermal conductivity ( $k = 3$ ). (a)Hirth & Kohlstedt (2004); (b) (Ranalli, 1995). Mantle phase diagrams are produced for the current work, while the mafic crust phase diagrams are taken from Piccolo et al. (2019).

Table S3: Test names and list of parameters.  $D$ ,  $D_{int}$ ,  $dM$  and  $M^{trs}$  are the melt extraction parameter of mantle phases. The mafic crust extraction parameter are equal for most of experiments( $dM = 0.01$ ,  $M^{trs} = 0.08$ ,  $D = 0.7$  and  $D_{int} = 0.2$ ). (1)  $A$  of all mantle phases is  $0.33 \times 10^{-5} \mu W/kg$  and the lower thermal boundary condition is  $1600^\circ C$ . (2) Mafic crust melt extraction parameters  $dM$  and  $M^{trs}$  are 0.08 and 0.15 respectively.  $\overline{TTG-opt}$  is the total proportion of TTG-like melts.  $\overline{LP-TTG}$ ,  $\overline{MP-TTG}$  and  $\overline{HP-TTG}$  are the normalized total volume of low, middle and high pressure TTG-like melts.  $\overline{LP}$ ,  $\overline{MP}$  and  $\overline{HP}$  are the volumetric weighted proportion of felsic melts produced within low pressure conditions ( $\leq 1.0$  GPa), middle pressure condition ( $1.0 \leq P \leq 1.8$  GPa) and high pressure ( $geq 1.8$  GPa).  $\overline{LT}$ ,  $\overline{MT}$  and  $\overline{HT}$  represents the volumetric weighted proportion of felsic melts produced at low temperature condition ( $\leq 800^\circ C$ ), middle temperature condition ( $800-1000^\circ C$ ) and high temperature condition ( $\geq 1000^\circ C$ ) respectively.  $P^{Felsic}$  and  $P^{Mafic}$  are the cumulative amount of felsic and mafic melts produced within the experiments. The resolution tests lists only the final proportion of the TTG, and the final cumulative volume of both mafic and felsic melts and the final average crustal thickness ( $\delta_{crust}^{fav}$ ). The input parameters are the same of the **Test3L**, except for the number of nodes employed, that results into a different  $dx$  and  $dz$  and finite difference cell area (A)

## 9.2. Supplementary Figures

Figure S1: **a)**: Density as a function of pressure at constant  $T_P$  (1200 °C), red arrow indicate increase depletion; **b)**: Melt quantity as a function of pressure at constant  $T_P$  (1550 °C), red arrow indicates the increase depletion; **c)**: Mode amount of clinopyroxene, orthopyroxene and olivine; **d)**: Density of each depletion step.

Figure S2: All these figure represents density and melt fraction as a function of P-T conditions for the crustal phase diagram. The first row represents the density of the solid fraction, while the second represents the volumetric fraction of melt.

Figure S3: Initial setup.

Figure S4: **Inefficient melt extraction, plume radius:150 km, dM=0.001** All the picture are ordered chronologically from top to bottom. **a)**: Compositional profile taken along a1-a2  $[400\ x, -400\ y]$   $[-400\ x, 400\ y]$ ; **b)**: The transparent plot represents the strain rate state of the crust, while the gray underlying volumes is the contour of the volume of depleted mantle (the total melt extracted is  $\geq 30\%$ ); **c)**: Planar view of the strain rate invariant.

Figure S5: Snapshots representing the effects of low melt extraction efficiency ( $dM=0.001$ , for further details, see Tab. S3, *Test1D<sup>r150</sup>*). The radius of the plume is reduced with respect to the 2D numerical cases to highlight the effect of this parameter on the overall 3D dynamics. The low melt extraction efficiency results in a fast and strong mantle plume. The mantle plume mechanically remove the old lithosphere generating a thin and fairly felsic crust. The extension induced by the partially molten mantle to the newly generated crust induce plate-like behavior at the rim of the plume structure. **a)**: Evolution of the numerical experiment from the initial stage to the subduction stage. The picture depict the lower crust (orange color with transparency), the plume (black color with transparency) and felsic crust (yellow). The bottom left picture represents the subduction slab (green); **b)**: Topography contour map taken at 7.758 Myrs. Topography is measured respect the average depth of the free surface.; **c)**: Normalized amount of felsic crust component along z direction taken at 7.758; **d)**: Profile along the direction drawn in c). The rock type coloring is the same of Fig. 2, except for the mafic dry intrusions which are in orange.

Figure S6: All experiments features the same  $dM$  (0.05) and  $M^{trs}$  (0.001). **a)**: Initial  $T_P = 1400$  °C, lithospheric thickness,  $\delta_{tkl} = 100$ km (*Test3L<sub>1400</sub>* in Tab. S3); **b)**: Same as *a* except for the lithospheric thickness,  $Tkl = 75$ km (*Test3L<sub>1400</sub>LT* in Tab. S3); **c)**: Same as *b*), but lower radiogenic heating and with a bottom lower thermal boundary equal to 1600°C (*Test3L<sub>1400</sub>LR-LT* in Tab. S3).

Figure S7: All the experiments are shown in Fig.S6, the list of input parameter are listed in Tab.S3. **a)**: Each sub-plot has two *y axes*. *Left y axis*:  $P^{Felsic}$  against time. The black tick line refers to the left axis and depict the evolution of the total volume of felsic melt extracted with time; *Right y axis*:  $TTG-opt$  against time.  $TTG-opt$  represents the amount of felsic melts extracted within the optimum field with time. The red thick line represents the proportion of TTG-like melts. The shaded colored area represents the relative proportion of the three different categories: *green*: LP-TTG like melts; *blue*: MP-TTG like melts; *red*: HP-TTG like melts.

Figure S8: This figure represents the condition at which the felsic melts have been extracted during the whole simulation. We subdivided the P-T space in small elements ( $dP=0.02$  GPa,  $dT=50$  °C), and we compute the logarithmic frequency of felsic melts extracted in each of these small elements. We performed this operation for the whole resolution experiments (see Fig. S8). These figures testify that in all the numerical experiments there are not dramatic differences in melt extraction pattern, what is changing effectively is the frequency of occurrence of the felsic melt extraction. As a function of the resolution the number of points that we collect increases (we highlight the number of points at the upper left corner of each sub-figure). The proportion of LP-MP-HP TTGs (listed in Tab. S4) are affected by the resolution of the crust.

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