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PLUME –LID INTERACTIONS DURING THE ARCHEAN AND IMPLICATIONS FOR THE GENERATION OF EARLY CONTINENTAL CRUST

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ABSTRACT

Many Archean terranes are interpreted to have a tectonic and metamorphic evolution that indicates 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. 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 with particular regard on the condition of extraction of felsic melts. In our numerical simulations, the continuous emplacement of new, dry mafic intrusions and the extraction of the felsic melts, generate an unstable lower crust which drips into the mantle soon after the plume arrival. The subsequent tectonic evolution depends on the asthenosphere T_p . If the T_p is high enough (≥ 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.

1 Introduction

Archean continental crust preserved on Earth occurs above old, strong and buoyant cratonic lithosphere. In some cases, the geological features recorded in these Archean terranes resembles those commonly seen in Phanerozoic terranes, such as horizontal displacements consistent with opening of basins (e.g. West Pilbara) Van Kranendonk et al. (2007) or orogenic processes (e.g. Itsaq) (Nutman et al., 2013). In many others, Archean terranes testify that high-grade metamorphism, production of felsic crust and plume-related magmatism happened simultaneously (Van Kranendonk et al., 2015; Chardon et al., 1996; Choukroune et al., 1995).

8 Archean continental crust is a mixture of mafic and felsic components (Kamber, 2015; Condie, 1981, 1993). The
 9 felsic crust is composed of tonalite–trondhjemite–granodiorite (TTG) magmas with a minor contribution of potassic
 10 granitoids (Moyen, 2011; Moyen & Stevens, 2006), whereas mafic crust is composed of several types of rocks spanning
 11 from ultramafic to iron tholeiites (Anhaeusser, 2014). Juvenile Archean TTGs can be produced by many mechanisms,
 12 although partial melting of a hydrated metabasalt (Johnson et al., 2017; White et al., 2017; Palin et al., 2016b,a; Moyen
 13 & Martin, 2012; Moyen, 2011) is accepted as being most prevalent. The trace and major element profiles for TTG
 14 magmas suggest that the partial melting must have occurred at medium-high pressure and high temperature in the
 15 presence of garnet, rutile, amphibole \pm plagioclase (Palin et al., 2016a; Moyen & Martin, 2012). The extraction of
 16 felsic melts leaves behind a mafic residuum that is denser than the underlying mantle (Piccolo et al., 2019; Palin et al.,
 17 2016a) and more prone to be delaminated or dripped off.

18 Many Archean terranes (e.g. East Pilbara) show that a close relationship exists between felsic melt production and
 19 vertical reorganization of the crust (partial convective overturn) (Van Kranendonk et al., 2004; Chardon et al., 1998;
 20 Collins et al., 1998; Bouhallier et al., 1995). These processes result in the classical structure of greenstone belts:
 21 the dome-and-keel geometries, in which felsic granitoids rise to shallow crustal levels, while mafic crust sinks.
 22 These structures are atypical for convergent margin settings and they are more consistent with plume–lithosphere
 23 interactions (Van Kranendonk, 2010; Van Kranendonk et al., 2004; Choukroune et al., 1995). The gravitational
 24 instabilities allow for an active exchange between shallow-surface environment and the mantle reservoir (Lee et al.,
 25 2006; Bédard, 2006; Zegers & van Keken, 2001). All these processes may feed back and influence the mafic crust and
 26 the petrogenesis of juvenile TTGs. Their occurrence depends on the intensity of magmatic processes and on the
 27 thermal and mechanical state of the convecting mantle.

28 The mantle potential temperature (T_p), the degree of radiogenic heat production and the convection style exert a
 29 strong control on plume–lithosphere dynamics (Jellinek et al., 2002; Schubert et al., 2001; Solomatov, 1995; Davaille &
 30 Jaupart, 1993). The absence of strong viscosity contrast within the convecting mantle and the high production of
 31 radiogenic heat can reduce plume activity or change their form, making them tail-less (Jellinek et al., 2002). It is still
 32 unclear what range of values of T_p are representative for the Archean, with estimates spanning from 1400 to 1650 °C
 33 (Aulbach & Arndt, 2019; Ganne & Feng, 2017; Kamber, 2015; Herzberg et al., 2010). If T_p ranges are in the lower end
 34 of these estimates ($< 1500^\circ\text{C}$), then the inferred high T_p Herzberg et al. (2010) is only an artifact of overturn events
 35 and the preserved Archean cratons formed at the top of overturn upwelling zones (Bédard, 2018).

36 Geodynamic and petrological modelling has been progressively integrated with field observations with quantita-
 37 tive/qualitative data to provide new insight into the tectonics of Archean terranes. Many numerical experiments
 38 base the compositional evolution on simplified parametrizations (e.g. (Fischer & Gerya, 2016; Jain et al., 2019; Sizova
 39 et al., 2015, 2014)) and/or employed only 2D numerical experiments (e.g. (Piccolo et al., 2019; Sizova et al., 2015)).
 40 These numerical experiments provide powerful insights, however the results may be biased by the 2D approach or by
 41 the resolution at which the crust is resolved, which is particularly important for global scale numerical experiments
 42 (e.g. Jain et al. (2019); Rozel et al. (2017)). The existing 3D numerical examples, instead, focus on the structural
 43 evolution, and do not explore the effect of felsic crust extraction (Fischer & Gerya, 2016; Gerya et al., 2015). In
 44 order to have a comprehensive view of the evolution of early Earth, it is necessary to explore the compositional
 45 and structural evolution of the Archean terranes combining petrological forward modelling with geodynamic 3D
 46 numerical simulations.

47 Here, we present a new advance towards integrating petrological phase diagram in a 3D numerical framework. We
 48 study the generation and short-term evolution of Archean terranes and the effects of the upper mantle T_p on the
 49 evolution of the primitive crust. Our results show that as a function of the upper mantle T_p , plume–lithosphere
 50 interaction yields different compositional and structural evolution of the Archean terranes. Further, we show that
 51 vertical tectonic deformation is not exclusively found in plate-tectonic like regimes and we demonstrate several
 52 different mechanisms that may produce felsic crust in a stagnant lid geodynamic scenario.

53 2 Methods

54 2.1 Numerical Modeling

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

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

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

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 (S , see Tab. S1 and below for further explanations) term in the non-linear residual calculations. S has an indirect effect on the momentum equation, because it affects the pressure and velocities. Therefore, 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 τ_{ij} is the deviatoric stress tensor, P is the pressure, ρ is the density and g_i is the gravity vector component along the i direction.

LaMEM employs viscous-elasto-plastic rheological constitutive equations that connect the deviatoric stress tensor components (τ_{ij}) to the total strain rate tensor components ($\dot{\epsilon}_{ij}$) as follows:

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

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

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

Where $\dot{\epsilon}_{ij}$ is the total strain rate tensor, the superscript *el,vis,pl* indicate the elastic (*el*), viscous (*vis*) and plastic (*pl*) strain rate. $\overset{\circ}{\tau}$ is the Jaumann objective stress rate, G is the shear modulus. ω 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 (τ_{II}), according to a dilatation-free non associative Drucker-Prager flow rule. η_{eff} is the viscosity which is computed using the following equation:

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

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 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, ϕ is the friction angle and C is the rock's cohesion. During plastic deformation C and ϕ can linearly decrease from the initial values (C_0, ϕ_0) to a final values (C_1, ϕ_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 a lower and upper cut-off (10^{18} and 10^{24} $Pa \cdot s$ 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:

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

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

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

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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 ($H_r = \rho A$, 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 α 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.

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2.2 Melt extraction and petrological modelling

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2.2.1 Melt extraction

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In the simulations, melt extraction and emplacement are considered to occur instantaneously. In 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 fraction is sufficient to trigger melt extraction. If M^* is higher than a critical melt fraction threshold, M_{trs} , a fixed quantity of melt escapes from the source, and its volume (V_{ext} , extracted volume) is computed according to:

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

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Where dV is the volume of the current finite difference cell, and dM is the volumetric fraction of melt that escapes, 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.

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dM represents the maximum volumetric fraction of melt that can be extracted from a source for a given timestep. This allows exploration of the cases in which melt extraction cannot compensate for melt production, e.g. if dM exfiltration is less efficient than production of new melt, the average density of the partially molten area decreases, and thus its buoyancy increases. Moreover using a fixed dM has two advantages: 1) it prevents instantaneous production of large volumes of crust; and 2) it prevents inducing a strong volumetric deformation due to the source/sink term in the mass conservation equations.

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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. The total volume of extracted melt must be corrected by a factor that depends on the ratio between the density of the melt and solid rock (ρ_{melt}/ρ_{solid} , as in Wallner & Schmeling (2016)):

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

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V_{ext}^{Tot} is the total amount of melt extracted from a rock type along z direction at fixed x, y coordinates; Vol_{cor} is the volumetric correction, which on average 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.

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The total amount of melt extracted is converted into extrusive and intrusive igneous rocks:

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

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Where Vol_{eff} and Vol_{int} is the extruded (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 (see Fig. S1).

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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:

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

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

$$135 \quad \mathbf{S} = \frac{1}{dt} \left(1 - \frac{dV}{dV - V_{ext}} \right) + \frac{1}{dt} \left(1 - \frac{dV}{dV + v_{int}} \right) \quad (15)$$

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

138 2.2.2 Petrological Modeling

139 Each marker is associated with a rock-type that is linked to a calculated phase diagram showing the equilibrium
 140 mineral assemblage and melt proportion as a function of P and T conditions, and the rock's thermo-mechanical
 141 properties. The rock type of each marker can change as a function of the total melt extracted. Every time that the
 142 phase diagram is changed, the amount of radiogenic heat is reduced by a factor of 2 (see Tab. S2). The phase diagrams
 143 store the melt quantity and the density of the solid fraction while the density of the melt is assumed constant (2700 and
 144 2400 kgm^{-3} for mantle and mafic phase diagrams respectively). Felsic crust formed as a consequence of crystallized
 145 TTGs magmas is modeled as a passive phase with fixed density (2700 kgm^{-3}), and it is not associated with any phase
 146 diagram.

147 The rock-types are divided into two sets: 1) mantle (M); 2) mafic crust (MC) (as in (Piccolo et al., 2019)):

- 148 1. Mantle phase diagrams represent six different steps of mantle depletion ($MD0\%$, $MD10\%$, $MD20\%$, $MD30\%$,
 149 $MD40\%$ and $MD50\%$) and were computed using *Perple_X* (Connolly, 2009). We used the activity-composition
 150 model of Jennings & Holland (2015), and all calculations assume a buffered oxygen fugacity (all the composi-
 151 tions are listed in Tab.1, all phase diagrams are shown in Fig. S2). The first step of depletion (Mantle Step
 152 0%) is modeled using the pyrolite composition from McDonough & Sun (1995). The chemical evolution is
 153 assumed to be only a function of the amount of extracted melt. We compute the depleted phase diagrams
 154 using the average composition of the residuum along the 10% melt isoline in every phase diagram (e.g.
 155 $MD10\%$ is computed using the residuum of the $MD0\%$, $MD20\%$ is computed using the residuum composition
 156 of $MD10\%$, etc.).
- 157 2. Mafic phase diagrams represents four depletion steps ($BS0\%$, $BS15\%$, $BS30\%$ and $BS45\%$) (see Fig. S2). All
 158 phase diagrams were computed using THERMOCALC 3.45 (Powell & Holland, 1988), using the Holland &
 159 Powell (2011) dataset (ds62). The starting phase diagram has the composition of a typical enriched Archean
 160 basalt (EAT). All solution model ($a-x$ relations) used to produce these phase diagram are: epidote, olivine
 161 (Holland & Powell, 2011), silicate melt, augite, hornblende (Green et al., 2016), garnet, orthopyroxene, biotite,
 162 chlorite (White et al., 2014), magnetite-spinel (White et al., 2002), ilmenite-hematite (White et al., 2000),
 163 Cbar-1 plagioclase, K-feldspar (Holland & Powell, 2003), and muscovite-paragonite (White et al., 2014).

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

167 Mafic phases are represented by five rock-types: $BS0\%$, $BS15\%$, $BS30\%$, $BS45\%$ and anhydrous intrusive mafic crust,
 168 where $BS0\%$ is the protolithic metabasalt composition. $BS0\%$ is assumed to be fully hydrated, and that the molar
 169 volume of water is sufficient to minimally saturate the solidus at 1.0 GPa (Palin et al., 2016a). After 15% of melting
 170 and melt extraction, the rock type is changed to the next basaltic step (e.g. $BS15\%$). $BS45\%$ is most depleted lithotype
 171 and has the same composition of $BS30\%$ but is no longer able to produce melt. The mafic intrusions are assumed to
 172 be anhydrous, and unable to produce any melt, and are modeled as if they had the same composition of $BS45\%$ (as
 173 in (Piccolo et al., 2019)). They represent the composite underplated bodies expected to be found at the base of an
 174 evolving oceanic plateau (Palin & Dyck, 2018; Van Kranendonk, 2010; Cox, 1980). All the melt that is extracted from
 175 the mafic protolith is converted into felsic crust. We do not distinguish any subtype of felsic rocks, and we do not
 176 consider the felsic rocks to produce any subsequent melts.

177 3 Results

178 We focus on the effect of the melt extraction parameter of the mantle phases and the asthenospheric mantle potential
 179 temperature (T_P). We carried out both 2D and 3D numerical experiments. 3D numerical experiments have been

180 performed to explore the effect of T_P while 2D numerical experiments have been performed to test the effect of dM
 181 and M^{trs} , and the long-term evolution of the system.

182 We mainly focus on the P and/or T conditions at which both felsic and mafic melts are extracted because this
 183 information can be used as a compositional proxy. We classify the felsic melts extracted into two main categories:
 184 TTG-like melts (TTG-opt) and non-TTG-like melts. TTG-like melts are the melts extracted within a P - T range that
 185 produces a bulk composition that best fits the chemistry of the natural TTG (TTG-opt). We further split TTG-like
 186 melts into three categories following Moyen (2011) as a function of the pressure of extraction within the optimum field:
 187 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)
 188 and high-pressure TTGs (HP-TTGs, >1.8 GPa, 800–1000 °C). Non-TTG-like melts are all other felsic melts of a broadly
 189 granitic composition that did not form in the optimal P-T field identified above. Mafic melts are tracked using their
 190 extraction temperature, which can give insights on mantle source (plume or asthenosphere or contaminated mantle).

191 In order to describe the macroscopic structural features, we introduce the following terminology: *external zone*
 192 represents the initial lithosphere (sometimes referred as old-lithosphere); *internal zone* is the newly produced
 193 composite crust produced during the plume-lithosphere interaction and affected by RTIs; *boundary zone* is the limit
 194 between internal and external zone, dominated by the delamination of the lower crust of the external zone.

195 Although they are both driven by buoyancy, we differentiate delamination and RTIs, as both have different surface
 196 manifestations and deformation patterns. Delamination processes are associated with peeling off of the dense lower
 197 crust –they produce asymmetrical topography and stress pattern (compressive in the direction of propagation and
 198 extensional in the opposite direction) Göğüş & Ueda (2018); Beall et al. (2017); Le Pourhiet et al. (2006); Morency &
 199 Doin (2004). By contrast, RTIs are symmetric and characterized by compressive and subsequent extensional tectonics
 200 after the removal of the dense unstable portion of the crust/lithosphere Beall et al. (2017); Molnar et al. (1998). We
 201 further differentiate the thinning processes into two categories: gravitational thinning is related with RTIs instabilities,
 202 and imply that the removal of mass from the crust is not compensated by the production of new crust; syn-magmatic
 203 thinning is associated with the asthenosphere drag forces.

204 3.1 Initial Setup

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

- 208 • Mechanical: free slip on all boundaries except for the top boundary, which features an internal free surface
 209 with a stress free open top boundary;
- 210 • Thermal: the top and bottom faces are isothermal, while the lateral boundaries have no flux boundary
 211 conditions. The top boundary temperature is 20 °C, while the bottom boundary temperatures are set to be
 212 1700 °C in most of the experiments exploring low T_P scenarios, while the high T_P scenarios has a lower
 213 bottom temperature in agreement with the adiabatic gradient;

214 The 2D numerical domain size is 1000×2×520 km with 257×3×257 nodes (along x , y and z directions respectively),
 215 while the 3D numerical setup is 800×800×520 km in size with 257×257×257 nodes. The air layer is 20 km thick in all
 216 cases (see Fig.S4).

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

227 All references scenarios assume that mantle intrusive/effusive ratio (I_R) is 0.6, consistent with plume geodynamic
 228 estimates (Crisp, 1984; White et al., 2006). The interval of intrusions of the melt produced by mantle lithologies is
 229 between the Moho depth and 0.6 of the total thickness of the crust. The felsic melt produced by mafic crust is all
 230 converted into felsic intrusion ($I_R=1$), with an interval of intrusion spanning 0.5–0.9 of the total crustal thickness.

231 The T_P temperature of the experiments range from 1350°C to 1550 °C , while the potential temperature of the plume is
 232 kept constant at 1600 °C (T^{Pl}). In all the experiments the adiabatic gradient is 0.4 °C/km. For the lower T_P experiments
 233 (<1500°C), we assumed that the temperature of the lower boundary is higher than the value expected for the given T_P
 234 to increases the convective vigors. The initial geothermal gradient of the lithosphere is a linear double stage geotherm,
 235 featuring a T_{Moho} of 800 °C and a temperature at the base of the lithosphere that depends on the asthenospheric T_P .

236 3.2 Effect of dM and M^{trs}

237 It is possible to subdivide the experiments into two regimes: low melt extraction efficiency ($dM < 0.01$) and high melt
 238 extraction efficiency ($dM > 0.01$). We briefly show two representative experiments that describe the geodynamic
 239 evolution of the two regimes.

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

241 1st Stage: Plume arrival stage (see Fig. 2) During the initial stage, the plume starts ascending and forces upper mantle
 242 convection (see Fig. 2, 0.881 Myrs). The mantle upwelling erodes the lithosphere and induces the partial melting of
 243 the asthenosphere underneath it. The positive buoyancy of the partially molten asthenosphere increases due the
 244 low dM (see Fig. 2, 0.881 Myrs, column b), allowing the asthenosphere to penetrate the lowermost portion of the
 245 lithospheric mantle.

246 After the plume arrives at the base of the lithosphere, it spreads along the x direction (see Fig. 2, 2.169 Myrs). During
 247 the spreading, mantle plume diapirs penetrate the lithospheric mantle. These diapirs have a mushroom shape and their
 248 peripheral portions are composed of residual mantle. This results in the generation of small magmatically thickened
 249 areas (50-100 km scale), that imposes a wavelength on the crustal thickness. The flux of mafic melts is insufficient to
 250 heat up the hydrated crust, so no felsic crust is produced. As a consequence of localized crustal thickening, the lower
 251 crust starts to be eclogitized, producing a negative and unstable volume of mafic crust (see Fig. 2 2.169 Myrs, column
 252 b).

253 2nd Stage: Syn-magmatic extension and lithosphere removal (see Fig 2, 2.765–2.813 Myrs). The eclogitized lower
 254 crust drips off, removing the lithospheric mantle and causing the return flow of the astenosphere. The asthenospheric
 255 return flow induces the production of high volume of mantle melt. The positive buoyancy of the partially molten
 256 mantle increases, producing an active upwelling (see Fig 2, 2.765 Myrs, column b). The asthenosphere comes in
 257 contact with the hydrated crust, causing the production of significant volumes of felsic melt. The partially molten
 258 mantle spreads beneath the crust and its drag force initiates syn-magmatic extension. The newly produced crust is
 259 thin in the central area of the upwelling and tends to thicken laterally. Then after the removal of the all unstable
 260 portion of the lower crust, the upwelling is localized at the centre of the numerical domain (see Fig 2, 2.765 Myrs,
 261 column b). Three distinct structural homogeneous area are generated: the *internal zone* represents the area initially
 262 affected by the plume that expands laterally and is characterized by a fast syn-magmatic extension; two *boundary*
 263 *zones*, which are mechanically thickened areas and characterized by the constant peeling off of the lower crust (see
 264 Fig 2, 2.765 Myrs, column b); and two *external zones* that are forced to under-thrust the new composite crust. At the
 265 end of the stage, the gravitational instabilities are still active and are mainly RTIs. This final process allows further
 266 mantle melting which produces new hydrated basalts that bury the felsic crust.

267 3rd stage: Quiescent stage (see Fig 2, 14.180 Myrs). After the removal of the old lithosphere, the simulation enters in
 268 the quiescent stage. A new compositional lithospheric mantle stabilizes and the production of new crust is limited.
 269 Felsic crust is located at Moho boundary because of its burial and as a consequence of the removal of the lower
 270 residual and intrusive mafic crust (see Fig 2, 14.180 Myrs, d).

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

272 1st stage: Plume arrival (see Fig. 3, 1.962 Myrs): During plume ascent, the lithospheric mantle is stable. During
 273 the spreading of the mantle plume beneath the lithosphere, most of the melt is extracted in the plume axis area
 274 generating a oceanic plateau that extends from -200 to 200 km (see Figure 3, 1.962 Myrs column a). The lower crust of
 275 the thickened plateau becomes unstable (see Figure 3, 1.962 Myrs column b), and the hydrated crust starts melting
 276 producing small volume of felsic melts. The production of felsic melts and their extraction produces a considerable
 277 volume of residual mafic crust, that enhances the buoyancy contrast with the underlying and mantle.

278 2nd stage: Dripping, delamination and lithospheric mantle removal (see Figure 3, 2.062–3.207 Myrs). Due to the
 279 continuous magmatic activity the lower crust of the oceanic plateau drips off. The dripped lower crust locally removes
 280 the lithospheric mantle, causing an asthenosphere return flow. The ascending astenosphere partially melts and
 281 enhances the production of new mafic intrusions, that heat up the hydrated crust allowing the production of further

282 felsic melts (see Figure 3, 1.962 Myrs column a and b). The dripped off material interrupts the plume buoyancy flux and
 283 the normal asthenosphere replaces the removed mantle lithosphere. The oceanic plateau undergoes syn-magmatic
 284 extensions and gravitational thinning (see Figure 3, 2.186 Myrs). After the complete removal of the lower crust of the
 285 oceanic plateau, it is possible to identify internal, external and boundary zones. The internal zone is dominated by
 286 RTIs that thins the crust. The boundary zones are characterized by the discontinuous delamination of the lower crust.
 287 Delamination and RTIs assist the expansion of the internal zone until the external zone is completely removed.

288 3rd stage: Quiescent (see Figure 3, 14.180 Myrs). After the removal of the old lithosphere the simulation enters its
 289 quiescent stage. The amount of new felsic crust produced is small compared to the previous stages. Most of the felsic
 290 crust is buried to middle and lower crustal depths due to the addition of new hydrated basalt during the removal
 291 stage (see Figure 3, 14.180 Myrs column d). Due to the RTIs, the amount of felsic crust is not evenly distributed in the
 292 section, and the local felsic crust anomalies generate small domes similar to the dome-and-keel geometry.

293 We performed nine 2D numerical experiments in order to constrain the effects of dM and M^{trs} of the mantle phases.
 294 Although M_{trs} controls the melt extraction and the volumetric melt fraction that is retained in the source, it has no
 295 effects on the geodynamic evolution nor on the absolute volume of the felsic and mafic crust. dM exerts a strong
 296 control on the volume of extracted melts and on the geodynamic evolution of the system. To perform 3D numerical
 297 experiments, we choose to use the parameters of the high dM numerical experiments.

298 3.3 Effect of the potential temperature of the asthenosphere

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

302 3.3.1 High mantle potential regimes ($T_P \geq 1500^\circ C$)

303 **Reference experiment high potential temperature regime ($dM = 0.05$, M^{trs} , $T^{Pl} = 1600^\circ C$, $R = 200$ km, T_P
 304 $= 1500^\circ C$)**

305 1st stage: Plume Arrival (see Fig. 4, 1.308 Myrs): As soon as the mantle plume approaches sub-lithospheric depths
 306 (200–150 km), it starts to produce melts. The spreading velocity of the mantle plume is limited (~ 20 cm/yrs), forcing
 307 the melt extraction processes to occur in a limited circular area. The first drip occurs soon after the penetration
 308 of the plume into the bottom lithospheric mantle, while, ongoing mafic magma intrusion and extrusion results in
 309 magmatic thickening of the crust. Most of the TTG-like melts are produced at medium pressure condition (1.0–1.8
 310 GPa). After the occurrence of the first RTIs the extraction of MP-TTG like melts is focused on the boundary zone,
 311 while the internal zone is dominated by the extraction of low pressure felsic melt (see Fig. 4, 1.308 Myrs, column b).

312 2nd stage: Dripping, delamination and lithospheric removal (see Fig. 4, 2.108–2.593 Myrs): After the delamination
 313 of the first drip, the internal zone radially extends. In the internal zone, new felsic and mafic crust is continuously
 314 produced causing RTIs, while in the boundary zone the lower crust of the external zone is removed via delamination
 315 (see Fig. 4, 2.108 Myrs column a). The delamination of the lower crust has two effects: 1) it allows a syn-magmatic
 316 extension in the internal zone allowing its expansion; and, 2) it deforms the crust of the external zone. The delamination
 317 drives a burial of the hydrated mafic crust, resulting in the production of LP-MP-HP TTG like melts from the external
 318 to internal zones (see Fig. 4, 2.108 Myrs column b). The internal zone is characterized by the continuous extraction of
 319 LP-TTG melts, with some spots of MP-TTG melts associated with crustal drip. The internal zone is gravitationally
 320 thinned. The generation of residual mantle generates a stable density profile beneath the new composite crust. This
 321 regime is not stable, because as soon as the internal zone reaches a radius of 100–150 km, the delaminated crust in
 322 the boundary zones segments in a discrete *slab*, and during their foundering they become similar to RTIs (see Fig. 4,
 323 2.362 Myrs column a/c). The transition from slab-like to drip-like gravitational instabilities induces a significant
 324 removal of the lithospheric mantle resulting in a syn-magmatic thinning of the crust. The removal of the lithospheric
 325 mantle triggers mantle melting beneath the external zone which results in the production of high volumes of LP-TTG
 326 melts and in the weakening of the crust (see Fig. 4, 2.362 Myrs column a-b-c). The delamination/drip processes become
 327 extremely efficient and the boundary zone is destroyed. The outward propagation of the internal zone happens at
 328 high speed (~ 100 cm/yr) and the remnants of the boundary zone propagate radially and then they are removed by
 329 RTIs (see Fig. 4, 2.593 Myrs column a). The interference between different RTIs generates local felsic crust anomalies.

330 3rd stage: Quiescent (see Fig. 4, 8.301 Myrs column a): After the removal of the old lithosphere, the experiment
 331 enters in its final stage. The newly-generated crust is mostly composed of mafic components ($\sim 40\%F$) (see Fig. 5,
 332 d). However, the felsic intrusive bodies are not homogeneously distributed within the crust (see Fig 4 a-b-d). These
 333 felsic bodies are irregular and are surrounded by mafic crust. The felsic crust anomalies are correlated with positive

334 topographic ones, while the area composed of mafic lithologies are associated with negative topographic anomalies
 335 (see Fig 4 c). Since the numerical experiments runs only for 10 Myrs, we do not observe directly the partial convective
 336 overturn. However, the architecture of the new crust and the spatial distribution of the felsic components within its
 337 volume, predispose the occurrence of partial convective overturn.

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

347 The felsic crust growth curve (cumulative volume of felsic melts extracted vs time, P^{Felsic}) has the same shape of the
 348 mafic growth one (see Fig 6 column a) suggesting a strong relation between mafic and felsic melts production. The
 349 volumetric fraction of TTG-like melts (TTG-opt) varies with time. Nevertheless, most of the felsic melts produced are
 350 extracted outside the optimum field at lower pressure and temperature (see Tab. S3). The proportion of LP, MP and
 351 HP TTG like melts changes with time, with an initial stage featuring significant volumetric fraction of MP-TTG like
 352 melts, which decreases with time during the dripping stage. The gravitational thinning associated with RTIs within
 353 the internal zone favour the production of LP-TTG (see Fig 4). The destruction of the boundary zone prevents an
 354 efficient burial of the mafic hydrated basalt to great depth.

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

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

368 3.3.2 Low mantle potential temperature regime

369 **Reference experiment low potential temperature regime ($dM = 0.05$, M^{trs} , $T^{Pl} = 1600^\circ\text{C}$, $R = 200\text{ km}$, $T_P =$
 370 1400°C)**

371 1st stage: Plume arrival (Fig. Fig 7, 0.560 Myrs): The first stage of the experiment is similar to the high potential
 372 temperature reference experiment. The only difference is related to the efficiency of mantle plume spreading beneath
 373 the mantle lithosphere. It is possible to observe the same pattern of felsic melt extraction, with a more prominent
 374 extraction of LP-TTG like melts within the internal zone and the extraction of higher pressure melts in the boundary
 375 zone.

376 2nd stage: Delamination and sub accretion (Fig. 7, 1.139–8.524 Myrs): After the occurrence of the first RTIs, it is
 377 possible to recognize the internal, boundary and external zones. The internal zone expands thanks to the delamination
 378 of the lower crust of the external zone, and it gravitationally thins due to the RTIs (Fig. 7, 1.139 Myrs column a). The
 379 felsic crust produced within the internal zone is drifted apart towards the boundary zone, because of drag forces
 380 exerted by the asthenospheric return flow. The main effect is promoting the extraction of LP-TTG in the internal
 381 zone and crustal thickening in the boundary zone (and thus MP TTG like extraction) (Fig. 7, a-c). The boundary
 382 zone matures as a function of the amount of the felsic crust that is accreted at the rim of the internal zone (see
 383 Fig. 7, 3.395 Myrs), which additionally facilitates the external zone sub-accretion. The result of the maturation of the
 384 boundary zone is to create an annular felsic terrain and forebasin. The pressure of felsic melt extraction within the
 385 external zone increases towards the internal zone mimicking a subduction-like process (see Fig.7 b).

386 The experiment does not reach the quiescent stage (Fig 7, 8.524 Myrs). And the processes described above stabilize
 387 the boundary zone generating a plate-like boundary feature. At the end of the time span explored, the crust can
 388 be subdivided into three areas: the internal zone whose crust is mostly mafic with felsic intrusive bodies that are
 389 emerging locally due to the RTIs (see Fig. 9, a, b d); an annular continental terrain at the rim of the internal zone
 390 over-thrusting the external zone crust; a annular foreland basin whose depth is -1.0 km (Fig.9 c). The annular felsic
 391 terrains have a topography that spans from $2.0 - 3.0$ km (Fig.9 c) and internal zone featuring a relative high topography
 392 ($0.6 - 1.5$ km). The topographic gradient imposed by the annular felsic terrain is high and facilitates the over-thrusting
 393 of the external zone mafic crust due to the gradient of gravitational potential energy between the annular felsic
 394 terranes and fore-land basin (similar to the results of Rey et al. (2014)). In the internal zone the topographic heights
 395 are associated with the amount of felsic crust, and the pattern of small basins is like those that have been observed in
 396 the reference high T_P scenarios (Fig.9 c-d).

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

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

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

415 We tested the same setup using a 2D numerical experiment. The main difference is related to the volumetric fraction
 416 of TTG-opt, which is higher. LP TTG like melts are the most common type of melt at low T_P , this result is in stark
 417 contrast with the high T_P one. We performed additional low T_P experiments to observe the long term effect of the
 418 thermal boundary condition and radiogenic heating and initial lithospheric thickness (see Fig. S6). Lowering only
 419 the thickness and the temperature at the base of the numerical domain yields a two-stage experiment, in which the
 420 expansion of the internal zone slows down with time and shuts down for 3–4 Myrs (see Fig S6 and S7). However,
 421 due to the high radiogenic heat production, the mantle heats up with time producing a new magmatic pulse that
 422 results in the expansion of the internal zone. If all mantle rocks feature a low radiogenic heat productivity (similar to
 423 the Phanerozoic one, $A = 0.3310^{-5}$ W/kg) the outward expansion of the internal zone stops, and the experiments
 424 remains in a quiescent stage.

425 3.3.3 Difference between 2D and 3D numerical experiments

426 The 2D and 3D numerical experiments have important differences primarily related to the production of felsic
 427 crust, which is extracted at low pressure in 3D numerical experiments in contrast with higher pressures in the 2D
 428 experiments. These differences emerge because of the RTIs are more developed in 3D cases than in 2D numerical
 429 experiments. The differences arise from the limited spreading velocity of any 3D upwellings (see Fig 11). The
 430 spreading velocity (the x and y velocity components of velocity) controls the distribution of the new crust, which
 431 results in anomalous localizations of crustal thickness. The spreading velocity beneath the lithosphere depends on
 432 the temperature contrast between normal asthenosphere and the plume. In the 2D numerical experiments, the lateral
 433 spreading velocity of the plume is higher than in 3D, allowing generation of a thickened and wider oceanic plateau:
 434 this promotes melting of hydrated crust at moderate pressure, and generates a more stable configuration. In 2D cases
 435 featuring a lower asthenosphere T_P , a thinner oceanic plateau is formed that promotes LP TTG generation. In 3D
 436 cases, most of the mafic-melts are extracted along the plume axis direction, yielding an ultra-thickened and narrow
 437 plateau, which is more unstable with stronger RTIs resulting in a more efficient gravitational thinning.

438 4 Discussion

439 4.1 Limitations

440 The resolution of the numerical experiments can affect the geodynamic evolution and the estimates of the felsic/mafic
 441 crust produced. We performed the 2D numerical experiment shown in Fig. 3 with different grid sizes (see Tab S4)
 442 to assess the impact of variable resolution; however, the key geodynamic processes do not show any significant
 443 differences. The most obvious change is related to the size of RTIs, which depends on the ability of the code to
 444 resolve the crust. The numerical resolution seems to not affect the total volume of mafic crust and the final crustal
 445 thickness at the end of the experiments. The normalized total mafic productivities of all simulations with respect to
 446 the average spans from -0.04 to 0.06 ($\sim \pm 13 \cdot 10^3 km^3$). The final crust thickness shows similar variability, decreasing
 447 with resolution and ranging from -0.08 (high resolution) to 0.08 (low resolution) ($\sim \pm 1.5 km$). The quantity that is most
 448 affected is the absolute volume of felsic crust produced, which increases with the resolution. But instead of being
 449 caused by the resolution dependency of our melt extraction algorithm, it is still associated with the resolution of the
 450 crust. The volume of felsic crust and volumetric fraction of TTG-like melts increase with the resolution. The relative
 451 proportion of each LP, MP and HP is also affected, with MP-TTG becoming dominant at extremely high resolution
 452 (i.e. 513x513 nodes per direction), largely at the expense of LP-TTG like melts and minorly at the expense of HP-TTG
 453 like melts. Nevertheless, the distribution of the P - T conditions are consistent within these numerical experiments
 454 (see Fig. S8). Our results still hold at low and high resolution, although, it is necessary to consider the uncertainty
 455 introduced by the numerical resolution.

456 Most of the numerical simulations enter a quiescent stage. This stage is influenced by the size of the numerical box
 457 and by the mantle melt productivity, so it can be interpreted as a saturation stage. Despite this limitation, they key
 458 physical processes are not affected.

459 Our initial scenario depicts a thermal (small) tail-less plume. This plume is not fed by any buoyancy flux coming from
 460 the lower mantle. This is done to model a plume in stagnant lid convection mode, in which long lived plumes are
 461 unlikely to happen. At the end of the numerical experiments the crustal thickness is thin ($\sim 28 km$) with respect to
 462 the inferred scenarios ($35 - 50 km$) (Van Kranendonk et al., 2015; Van Kranendonk, 2010; Smithies et al., 2009) and
 463 most likely a long-lasting mantle plume would yield a thicker crust. However, most of the thinning processes are
 464 associated with gravitational instabilities, that in case of a long-lasting plume are expected to be more effective.

465 We introduce a novel approach to deal with the compositional variation due to magmatic processes; however, it
 466 introduces several issues, such as stepwise variations of properties. The stepwise variations of density are reduced
 467 by the interpolation scheme employed and their effect is minor compared to the actual phase reactions observed
 468 within the same diagram. Radiogenic heat production is handled discretely by decreasing its value as a function of the
 469 depletion step. This may affect the long-term results of our numerical simulation, which (together with the extremely
 470 high computational cost of each simulations) is one of the reasons that we opt for short-term experiments. The most
 471 suitable way to handle radiogenic heat production is to enforce a partition coefficient of radionuclides as a function
 472 of the melt produced. In order to do that, it is necessary to assume a partition coefficient on the fly as a function of
 473 mineralogical composition (e.g. Rummel et al. (2018)) which is computationally unfeasible. Moreover, we assume that
 474 the average composition of the residuum along a melt isopleth is representative of all the chemical complexity. The
 475 last limitations is associated with the usage of the phase diagrams. We assume that the thermodynamic equilibrium is
 476 always reached and we neglect the effects of metastability and the kinetic of reactions that can control the timescales
 477 of the processes.

478 4.2 Mafic crust production

479 The primary mafic melt composition depends on the thermal state of the mantle. Consequently, the basalts that
 480 feature a composition close to the primary melts have been widely employed to estimate Archean mantle T_p (Ganne
 481 & Feng, 2017; Herzberg et al., 2010). The debate concerning the Archean upper mantle potential temperature is still
 482 open with estimates spanning from 1400 to 1650 °C (Aulbach & Arndt, 2019; Herzberg et al., 2010). Moreover, these
 483 estimates show considerable variations at any one point in time (Ganne & Feng, 2017; Herzberg et al., 2010) suggesting
 484 that this variability may be due to the interaction of a relatively cold asthenosphere and mantle plumes (Kamber,
 485 2015; Arndt, 2013). Our results suggest that the mafic crust composition can be variable with high and low #Mg mafic
 486 rocks coexisting consistently with many Archean terranes such as Pilbara or Kapavaal (Van Kranendonk et al., 2015).

487 The temperature of extraction of mafic melt is variable at any timestep. The colder mafic melts indicate a strong
 488 interaction with the delaminated crust. Although we do not consider chemical exchange, degassing and magmatic
 489 hybridization in our simulations, this thermal exchange must result in at least one of these processes (Bédard, 2006).
 490 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, hybridization between mantle and crustal related melts has been invoked to produce the mafic source of East Pilbara TTGs (Smithies et al., 2009). 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 for TTG petrogenesis with significant differences between 2D and 3D experiments: 2D experiments generally show a higher proportion of TTG-like melts (40-50%) with respect to 3D (20-30%).

The felsic melts are mostly produced at relatively low-pressure and low-temperature conditions ($T^{Gen} < 700^\circ \text{C}$) ensuring 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 to ~ 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 slow 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 ($T < 700 - 800^\circ \text{C}$ and $P < 1.2 \text{ GPa}$), outside the TTG optimum $P-T$ field. The production of generally LP felsic melts has been identified even in global scale 2D numerical experiments (Jain et al., 2019). To 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 $BS_0\%$, which is computed using an EAT composition, and assuming that is hydrated enough to be minimally saturated at 1.0 GPa (Palin et al., 2016a). This composition represents a likely "typical" source for the generation of felsic crust, as the degree of hydration maximizes the felsic melt generation. This saturated composition has a low wet 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 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.

Most of the TTG-like melts are LP-TTG type, 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 dominated by LP-TTGs. However, 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), which is the best preserved classic dome and basin terrain developed through primarily vertical tectonic processes. The dominance of LP-TTGs in our simulations may also be a compositional artifact, as the bulk composition of the mafic protolith has not directly been addressed to produce the LP, MP and HP TTGs classification (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 an iron-rich composition that can 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, 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 classification 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 modelling without accounting for the variability of mafic source protolith will produce misleading results. This implies that any classification is arbitrary and a more deep connection between petrological modelling and geodynamic numerical simulation is required (cf. Palin et al. (2020)).

4.4 Structural evolution and implication for subduction initiation

We identify two main regimes: low and high T_p . In both scenarios, with a surface dominated by mafic lithologies, which is consistent with the inferred Archean's continental crust surface (Kamber, 2010; Condie, 1993).

High T_p regimes ($\geq 1500^\circ \text{C}$) We observed structures that resemble dome-and-keel geometry which has been recognized as the effect of partial convective overturn (Van Kranendonk, 2010; Van Kranendonk et al., 2004). However

544 partial convective overturn occurs after a long thermal maturation of the gneiss dome due to the high radiogenic
 545 heat production (François et al., 2014; Bodorkos et al., 2006), or because of the emplacement of hot intrusions
 546 (Van Kranendonk et al., 2015). In our numerical experiments, the apparent rising of felsic domes is a consequence of
 547 RTIs and delamination induced deformation. These anomalies predispose the felsic crust to rise in the subsequent
 548 mantle-related magmatic events. The East Pilbara terrane has a long-lasting structural evolution that started during
 549 the Eo/Paleo-Archean and ended during the Meso-Archean (3.65–3.22 Ga) (Van Kranendonk et al., 2015; Hickman
 550 & Van Kranendonk, 2012). The evolution of this Archean terranes has been divided into three main stages. The
 551 first related to the generation of a primitive and mostly undifferentiated mafic crust (our initial conditions), the
 552 second related to plume-related magmatism and generation of the first TTGs and the last one with the stabilization
 553 of the cratonic lithosphere (Van Kranendonk et al., 2015). Our results represent the second stage, in which the hot
 554 mantle upwelling starts interacting with the proto lithosphere. During this stage, a primitive continental crust is
 555 slowly generated, with a composite mafic volcanic sequence composed of komatiites, high Mg basalts, and tholeiite
 556 (Van Kranendonk et al., 2015). The felsic crust generation is associated with delamination and dripping events, that
 557 catalyse the production of an ultra-depleted mantle and felsic crust production. In this stage, the mantle is potentially
 558 hydrated and contaminated by the delaminated crust Bédard (2006).

559 **Low T_P regimes ($\leq 1450^\circ\text{C}$)** In these experiments, there are plate-boundaries-like features that are associated
 560 with the production of middle- to high-pressure felsic melts. Within the same settings coexist different structural
 561 patterns and modes of felsic melts production. The lithospheric mantle is rigid and less prone to be captured by the
 562 delaminating crust preventing the asthenosphere from partially melting outside the internal zone. The propagation of
 563 the internal zone generates annular predominantly felsic terranes that assist the sub-accretion of the initial lithosphere.
 564 Itsaq gneiss complex (West Greenland) is an old Archean terranes mainly composed of felsic gneiss that records a
 565 potential long phase of horizontal tectonics (Nutman et al., 2013, 2009). The horizontal tectonic phases have been
 566 interpreted as a consequence of subduction-related processes (Nutman et al., 2013, 2009). Sizova et al. (2015) using 2D
 567 numerical experiments shows that subduction-like processes arise as a consequence of the chemical evolution of the
 568 crust, and associated the generation of predominantly felsic crust terranes with transient subduction events. Our
 569 results show that vertical tectonic processes generate significant horizontal displacement, generating stacking of
 570 felsic crust slivers due to the co-operation of RTIs, delamination and mantle drag. Although, we do not explicitly
 571 simulate the felsic crust reworking, these thickened continental terranes would eventually undergo partial melting
 572 generating further evolved continental crust. It is widely believed that continental crust forming processes evolve
 573 with the secular cooling of the planet (Moyen & Martin, 2012; Ganne & Feng, 2017; Palin et al., 2020). Several lines of
 574 research suggest that the felsic crust produced become progressively potassic and that the pressure of generation of
 575 the felsic melts increased (Laurent et al., 2014). Our low T_P regimes show behaviour that is consistent with these
 576 observations and importantly show that these contrasting P – T conditions may coexist together.

577 In our numerical experiments, the interaction between the transient plume and lithosphere generates a semi-circular
 578 structure that radially expands with velocities that depends on the background T_P . In both explored regimes the
 579 propagation of the internal zone is accompanied by disintegration of the newly-produced crust, and its lateral
 580 displacement over several 100s km. These structural feature resembles the coronae structure on the surface of Venus
 581 (Harris & Bédard, 2014; Stofan et al., 1992). Most of the deformation is governed by three main processes: the drag
 582 force exerted by the asthenosphere that rises as a consequence of the RTIs and delamination processes; pulls forces
 583 exerted by the foundering drips and delamination of the lower crust at the boundary zones.

584 Our results can be compared with numerical experiments exploring plume-induced subduction (Ueda et al., 2008; Baes
 585 et al., 2016; Gerya et al., 2015). The main target parameters of these previous studies has been the plastic and viscous
 586 strength of the lithosphere and the buoyancy of the plume. These works discovered four regimes through which
 587 a simulation evolves: 1) underplating plume; 2) crustal fragmentation; 3) transient subduction; 4) self-consistent
 588 subduction. Our results characterize the first regime, and soon evolve into crustal fragmentation Ueda et al. (2008) due
 589 to crustal gravitational instabilities. In the 3D numerical experiments, a strong crustal fragmentation occurs at high
 590 T_P , that prevents the formation of any localized deformation zone. By contrast in the low T_P cases, the combination
 591 of gravitational instabilities and compositional evolution of the crust allows the emergence of plate-like boundaries
 592 features.

593 We observe subduction-like processes that are related to inefficient melt extraction and which emerge due to
 594 asthenospheric return flow caused by gravitational instabilities. This inefficient melt extraction does not reduce the
 595 melt-induced compositional buoyancy, causing active upwellings that pushes the newly generated crust towards the
 596 old-lithosphere forcing under-thrusting (see Fig. 2). These behaviours have also been observed in 3D models (see Fig.
 597 S5), but are only a consequence of inefficient melt extraction. The mantle plume in these experiments triggers the
 598 gravitational instabilities, but does not cause subduction to initiate.

599 Our results show that several plate-like processes occur in a geodynamic scenario dominated by vertical displacements.
 600 As noted by Sizova et al. (2015), the emergence of plate-like processes is most likely due to the compositional evolution
 601 of the oceanic crust, which also has direct implications for the rock types diagnostic of subduction – such as blueschists
 602 – that may form and be preserved in geological terranes through time (Palin & White, 2016). The generation of
 603 plate-like boundaries in low T_P experiments provides clear evidence that the compositional evolution of the crust and
 604 the plume-induced dynamics generates lateral and vertical heterogeneities, creating weak zones that may ultimately
 605 become plate boundaries. It is arguable that the presence of these distinctive compositional and structural domains
 606 can evolve into modern-style subduction zones. If these zones connect and become a global network, the transition
 607 from a stagnant lid regime to a mobile lid (plate tectonic) geodynamic regime may proceed.

608 5 Conclusion

609 Our 2D and 3D numerical experiments predict several fundamental differences in the styles of plume–lid interaction
 610 as a function of initial lithosphere thickness and initial T_P . At high T_P most of the deformation and production of
 611 new mafic crust is mediated by asthenospheric mantle flow resulting from gravitational instabilities ($T_P \geq 1500^\circ\text{C}$).
 612 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
 613 temperature at which basalt is generated supports the idea that upper mantle T_P may have been lower, and indicate
 614 that mafic crust is composed of different mafic lithologies. As a consequence, it is necessary to address the effect of
 615 this variability to constraint the condition at which natural Archean TTGs are generated. Our numerical experiments
 616 predicts that significant amount of new felsic crust are produced, even though, the P – T conditions do not match the
 617 most previously suggested ranges that are thought necessary to produce high amounts of TTGs-like melts, indicating
 618 that further investigation is required.

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 625 significantly improved thanks to comments of J. Bedard, J. Van Hunen and an two anonymous reviewers.

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

Rock Type	SiO ₂	Al ₂ O ₃	Cr ₂ O ₃	FeO	MgO	CaO	Na ₂ O
MD0%	48.809	8.195	1.066	10.438	27.501	3.590	0.401
MD10%	49.235	7.047	1.126	10.241	29.232	2.843	0.276
MD20%	49.782	5.463	1.156	10.161	31.385	1.924	0.128
MD30%	50.328	3.815	1.163	10.023	33.930	0.723	0.018
MD40%	50.202	3.327	1.070	9.517	35.782	0.102	0.000
MD50%	49.309	3.608	0.994	8.812	37.276	0.001	0.000

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.

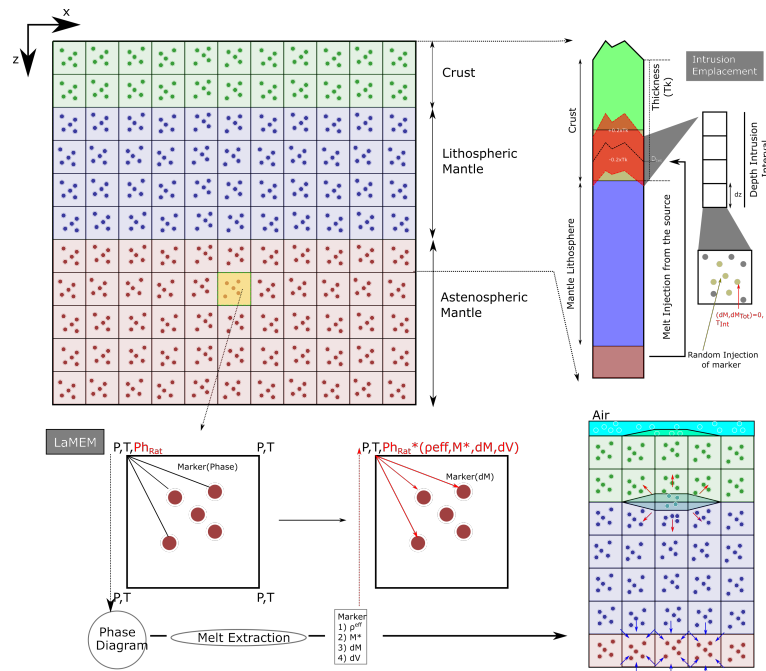


Figure 1: Schematic representation of melt-extraction algorithm. For each timestep all the rock-type properties are interpolated into the main grid. Then from pressure and temperature, the density and melt quantity are interpolated from the phase diagrams and scaled accordingly to the Ph_{Rat} . After 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

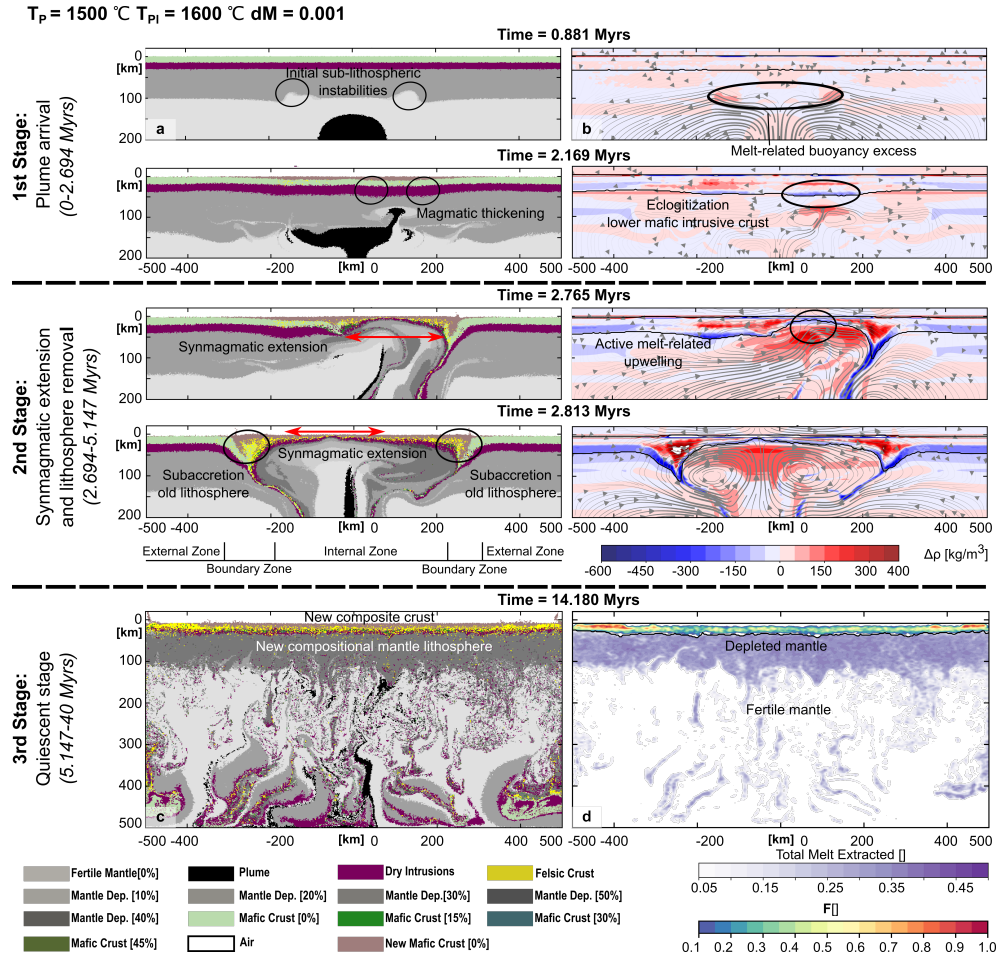


Figure 2: Main evolutionary stages of the low dM numerical experiment. The sub-plots of the 1st and 2nd stage zoom in on the lithospheric scale processes (-20 -200 km). Column a) features sub-plots represents the compositional field, while column b) shows several snapshots of the buoyancy contrast that controls the evolution of the numerical experiment, the black thick lines represent the internal free surface and the Moho topography, while the grey arrows represent the streamline of the velocity field. $\Delta\rho$ is computed by subtracting the density field along x direction for a given depth from the average density of the considered depth ($\Delta\rho(x, z) = \rho(x, z) - \rho_{mean}(z)$, 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.

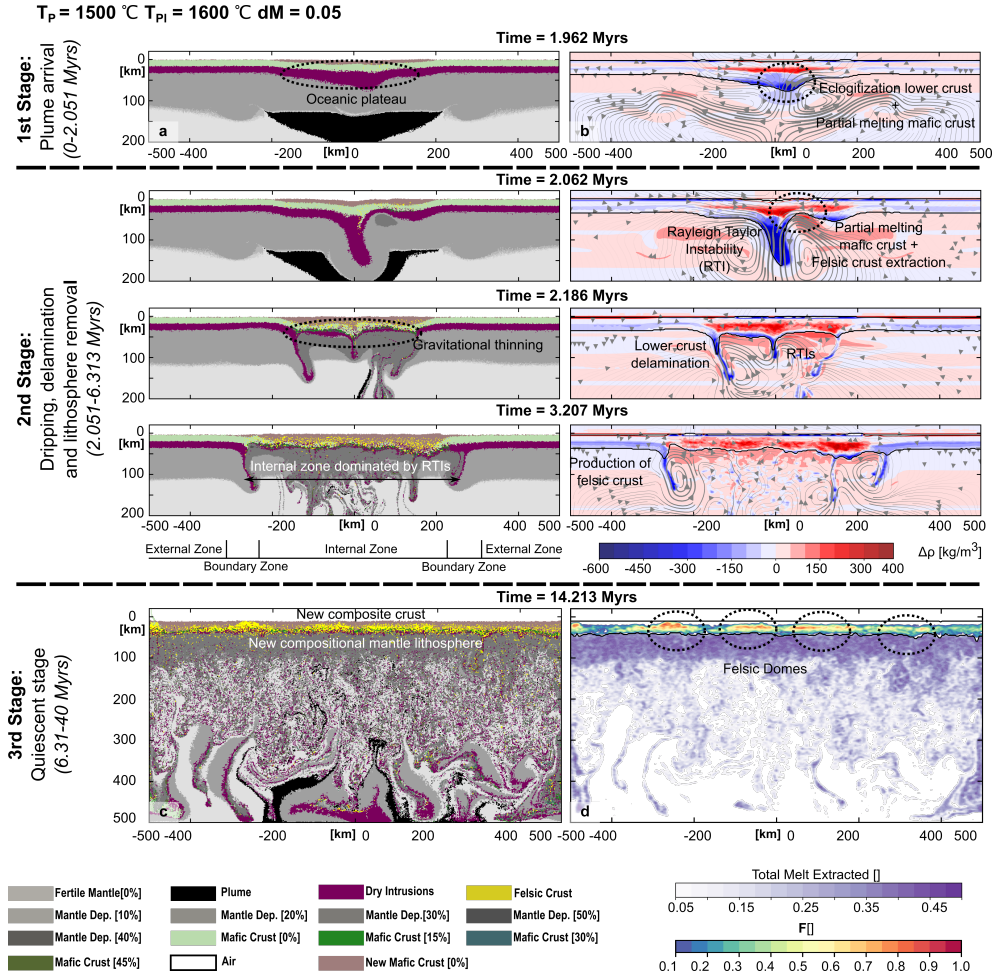


Figure 3: Main evolutionary stages of the high dM numerical experiment. The sub-plots of the 1st and 2nd stage zoom in on the lithospheric scale processes (-20 -200 km). Column a) features sub-plots represents the compositional field, while the column b) shows several snapshots of the buoyancy contrast (see the caption of Fig.2 for further explanations), with the streamline of the velocity field, 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.

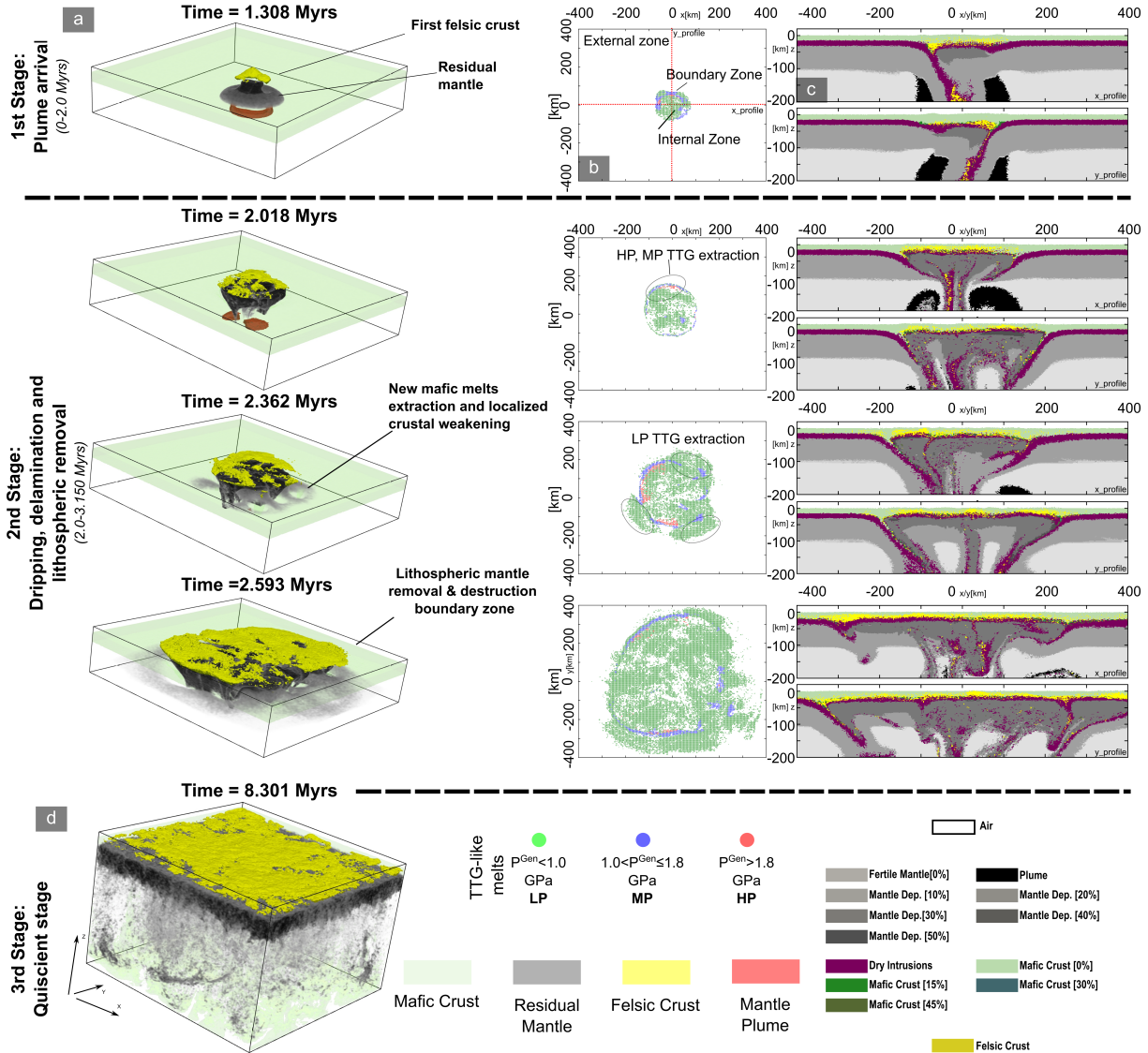


Figure 4: Temporal evolution of the high potential temperature reference scenario ($T_p = 1500$ °C, $R = 200$ km, $T^{Pl} = 1600$ °C, $dM = 0.05$, $M^{TTS} = 0.001$, $I_R = 0.6$). The 1st and 2nd stage zoom in on the first 200 km of the numerical domain. The column a) represents a simplified compositional 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) This represent slices of the compositional field taken along x and y direction; d) Represents simplified compositional field of the whole numerical domain.

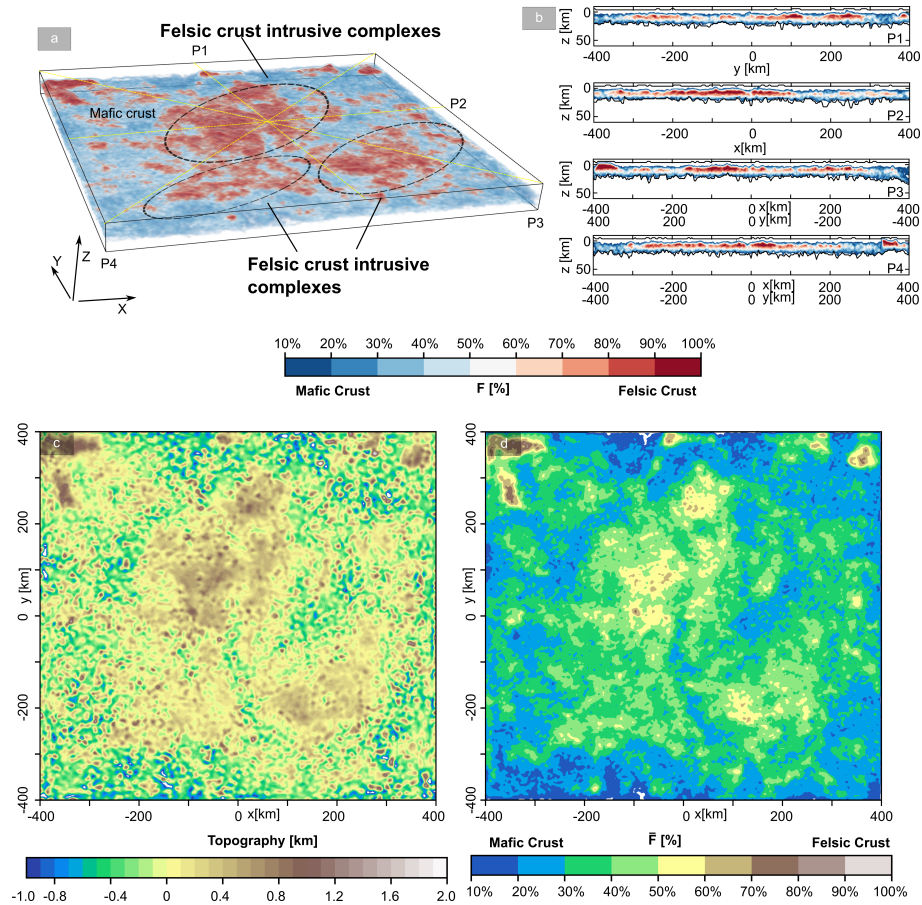


Figure 5: Final composition of the crust, and the effect of crustal material on the topography (the topography is computed using as a reference the average depth of the internal free surface). a) The plot represents the fractional volume of felsic components (F) within the crust; the yellow lines represent 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 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.

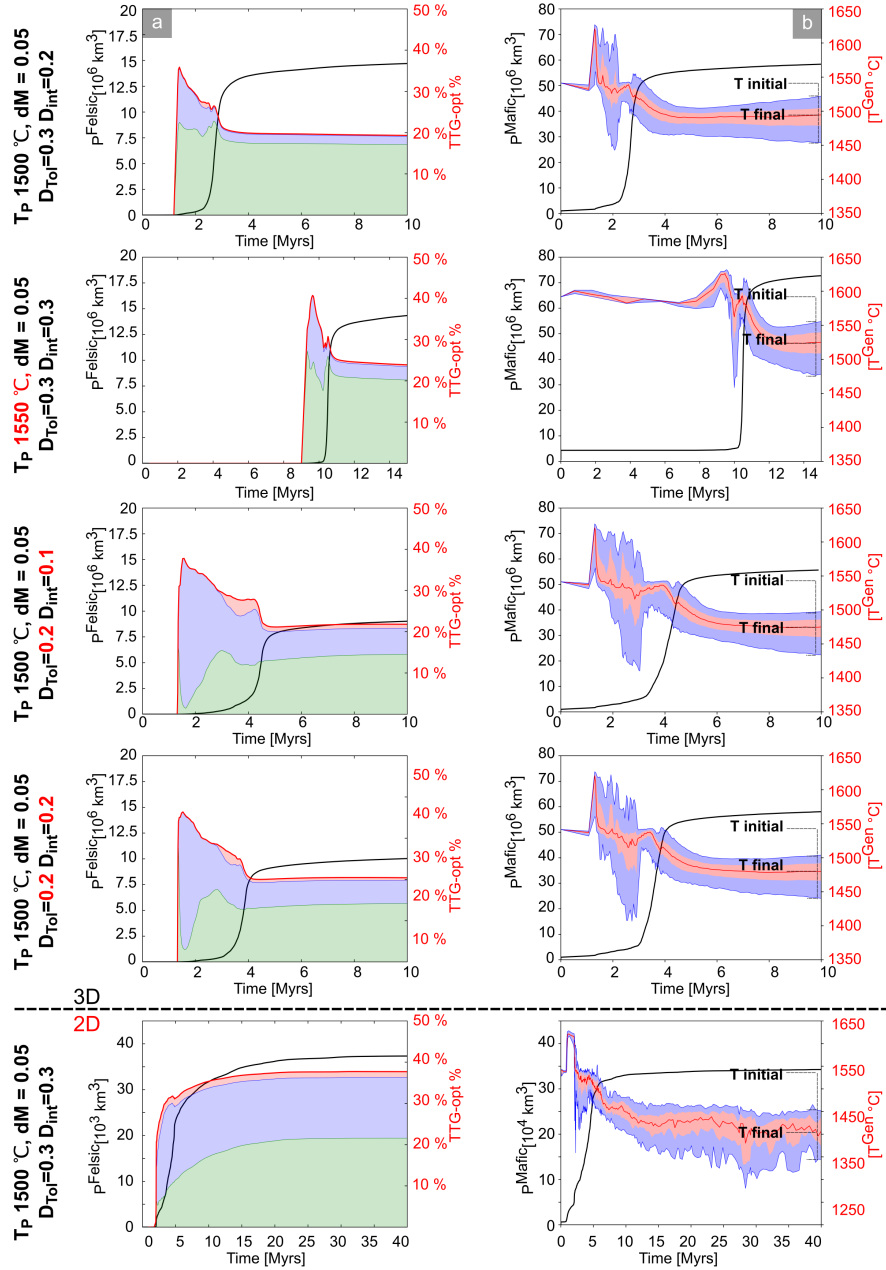


Figure 6: Summary of the felsic and mafic melt extraction P-T conditions of the high T_p regimes experiments. Each row represents different experiments. The parameters that have been changed respect the high T_p reference scenarios shown in Fig. 4 are highlighted in red. The last row is the 2D numerical 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.

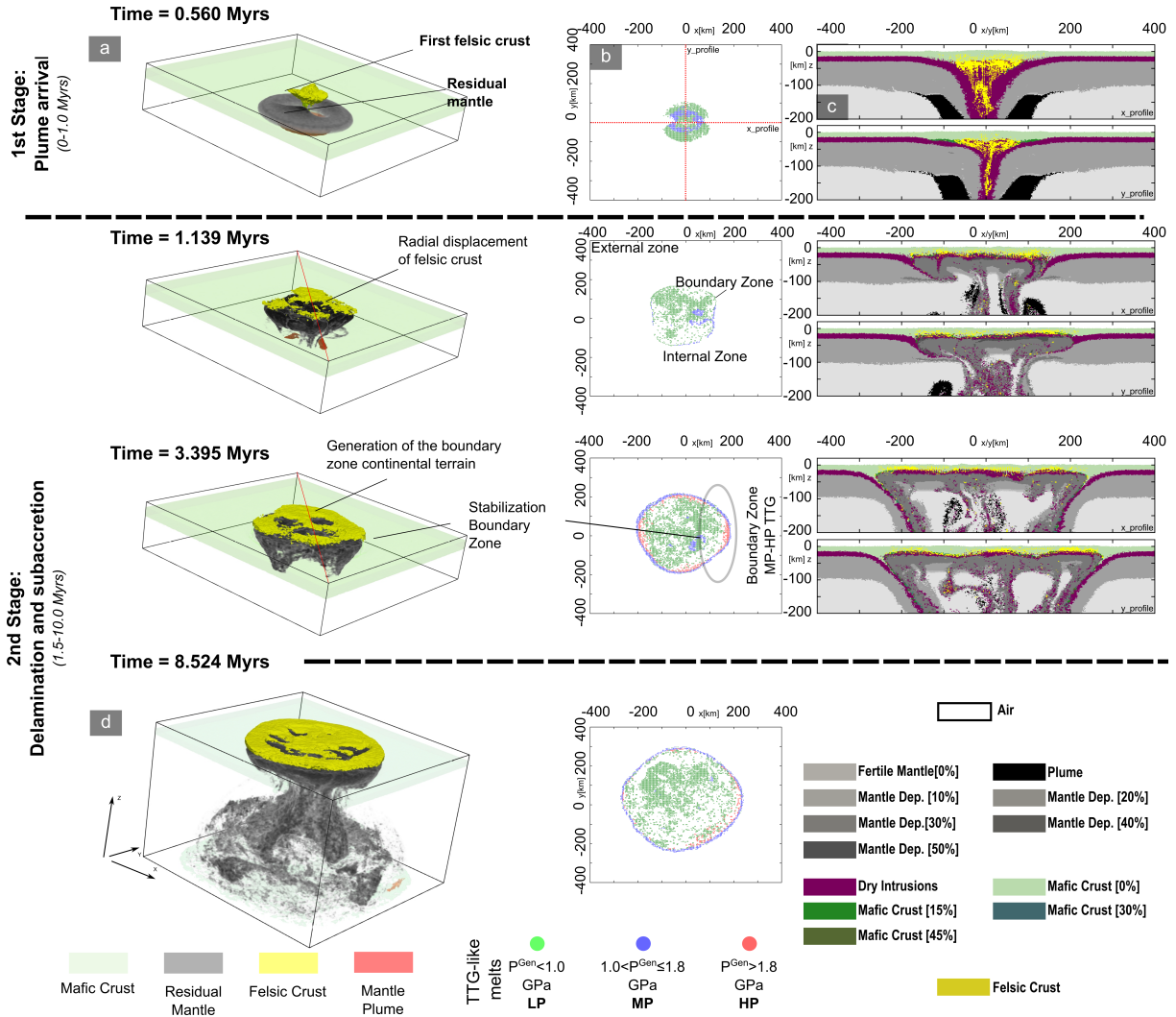


Figure 7: Temporal evolution of the low T_p reference scenario ($T_p = 1400$ °C, $R = 200$ km, $T^{Pl} = 1600$ °C, $dM = 0.05$, $M^{trs} = 0.001$, $I_R = 0.6$). The column a) represents a simplified compositional field (mafic crust collects all the mafic protolith phases, the residual mantle comprises 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) This represent slices of the compositional field taken along x and y direction; d) Represents simplified compositional field of the whole numerical domain.

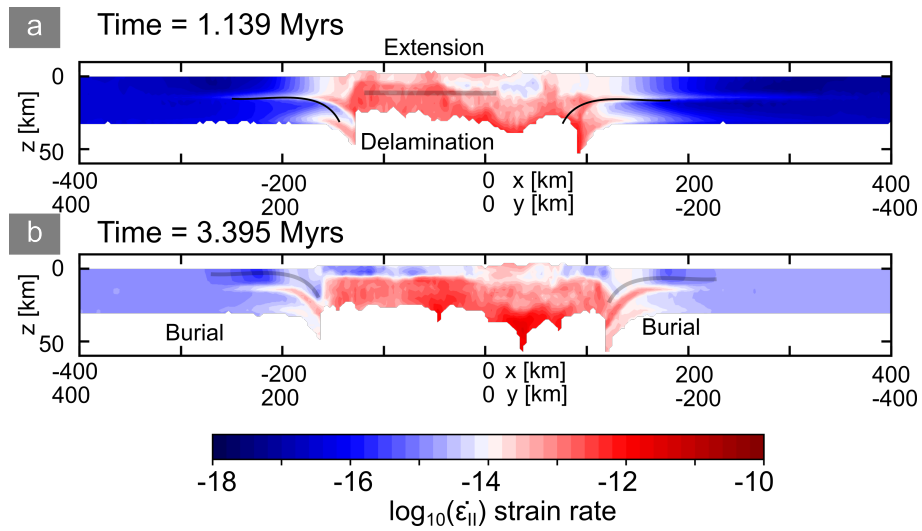


Figure 8: Two crustal scale slices of the second invariant of the strain rate. The profiles are taken along the two thick red lines shown in the compositional plot at 1.139 and 3.395 Myrs in Fig. 7. These profiles focus on the evolution of the boundary zone, that thickens with time. The thickening is 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.

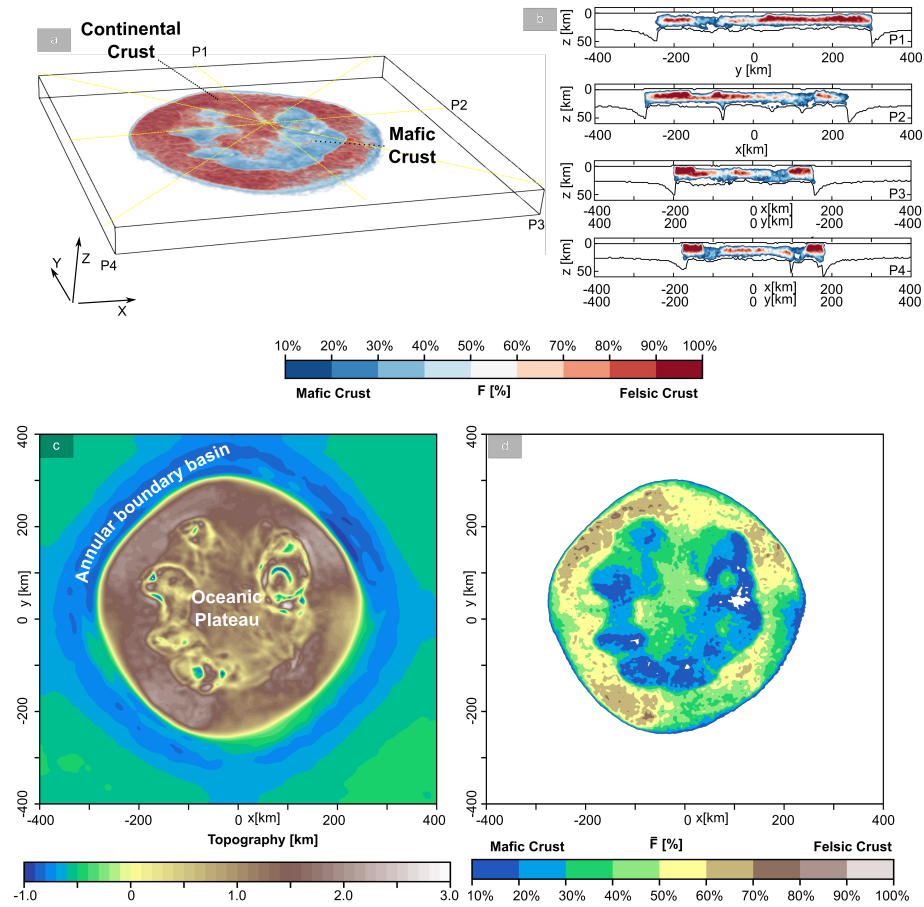


Figure 9: Final composition of the crust, and the effect of crustal material on the topography (the topography is computed using as a reference the average depth of the internal free surface) of the low potential temperature reference scenario. a) Represents the volume fraction of felsic crustal component within the crust. The thick yellow line represents the direction of the slices that are shown 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} .

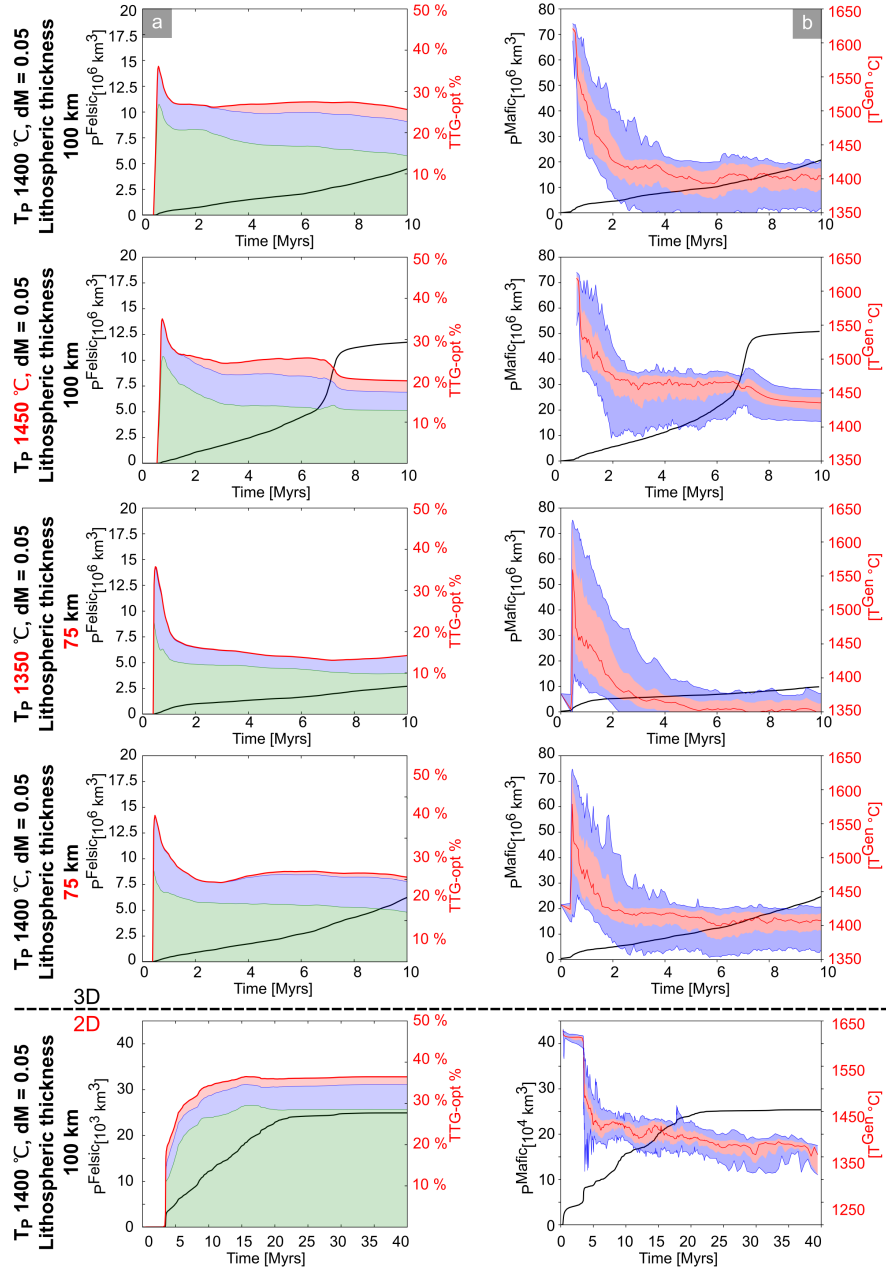


Figure 10: Summary of the felsic and mafic melt extraction P-T conditions of the low T_p regimes experiments. Each row represents different experiments. The parameters that have been changed respect the low T_p reference scenarios shown in Fig. 8 are highlighted in red. The last row is the 2D numerical 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.

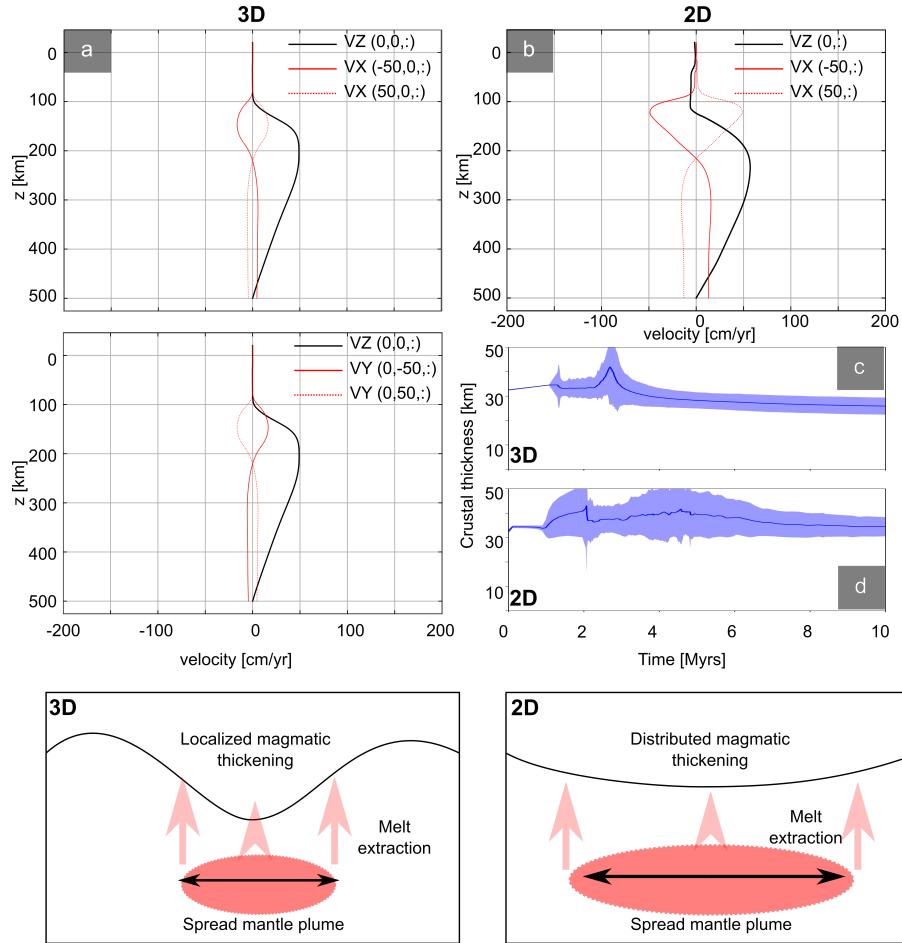


Figure 11: Velocities profiles of both 2D and 3D numerical test shortly before the removal stage, and the evolution of the crustal thickness. At the bottom of the figure, two simple sketches highlight the main difference of 2D and 3D numerical experiments. a) These two pictures represent the velocity 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 ± 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 ± 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.

626 7 Supplementary Material

627 8 Post-processing and hardware information

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

640 8.1 Moho identification and internal free surfaces

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

646 The internal free surface is defined at the beginning of the simulation. The resolution of this surface is the horizontal
 647 resolution of the numerical experiments (e.g. 3.125 km in the simulation described in the current work). After the
 648 system of equations is solved, the velocity field is linearly interpolated from the main grid to the actual position of the
 649 internal free surface (see Fig.S1 1)). Then the free surface is advected in accordance of all the velocity component (in
 650 the simplified sketch of Fig.S1, we show only the advection as a function of the *z* component of the velocity). Then the
 651 new advected topography is interpolated back to the main horizontal grid. After the advection of the free surface, the
 652 particles above and below the free surface are eventually corrected, and their phase is changed to be air or rock phase
 653 as a function of their relative position to the new free surface position. After all these corrections, that are necessary
 654 to prevent that rocks phase are above the actual free surface, the volume of effusion is converted into an *effusive*
 655 *thickness*. The computation of the total volume of melt extracted is the cumulative summation of all the volumetric
 656 contribution along a column whose area is equal to the finite difference cell area (e.g. 9.765 m²). To compute the
 657 effective thickness of new crust added at the top of the free surface, this volume of extracted material is conceived to
 658 be equivalent to the volume of a rectangular prism whose base is represented by the finite difference cell area and its
 659 height is the unknown effective thickness of new crust. The effusive thickness (δ_{eff} [*m*]) is computed (see S13):

$$660 \quad \delta_{eff}(x, y) = \frac{Vol_{eff}}{dx \cdot dy} \quad (S1)$$

661 However, the effective thickness is defined in the cell center of the main numerical grid. Therefore this values must
 662 be linearly interpolated into the main grid. After the interpolation, the free surface is shifted accordingly to the local
 663 δ_{eff} and the area between the previously computed free surface and the new one is filled with effusive rocks (see S14).

664 The cumulative melt production is computed by summing the total melt extracted along *z* direction, at every timestep:

$$665 \quad V_{cum}^{Tot}(i, j, MI/MC, t) = \sum_{dts=1}^{ts} V_{ext}^{Tot}(i, j, MI/MC)(i) \quad (S2)$$

666 *i, j* are the grid indexes, V_{cum}^{tot} is the total volume cumulated during the whole simulation and computed for *MC* or *MI*,
 667 *ts* is the current time-step (*ts*), while *dts* is the time-step increment. Generation temperatures (T^{Gen}) are computed
 668 each time-step, averaging the temperature at which the mafic melts are produced along *z* direction and considering
 669 the volumetric contribution of each rock-type.

670 *F*, the relative amount of felsic crust, is computed by integrating all the *CC* phase ratio along *z* direction divided by
 671 the local thickness (*Th*) of the crust. The integration is within the local Moho depth and the topography (*Topo*):

$$\bar{F} = \frac{\sum^{k_{Topo(i,j)}} CC(k)\Delta z(k)}{\delta_{crust}(i,j)} \quad (S3)$$

672

673 8.2 Hardware information and solver option

674 The simulations were performed using the supercomputers offered by Johannes Gutenberg University, MOGON and
 675 MOGONII (hpc.uni-mainz.de). LaMEM is a PETSc based code (Balay et al., 2018), and the solving of the non linear
 676 system of equation were performed thanks to the PETSc infrastructure. 2D experiments were performed using 8
 677 cores and using the direct solver *mumps*, the internal linear solver is *fgmres* with a maximum of 30 linear iteration,
 678 and relative tolerance of 10^{-6} ; while the external non linear solver were allowed to iterate maximum 30 times, and
 679 the relative tolerance was set to be 10^{-6} . 3D numerical experiments were performed using 512 processors and using
 680 Eisenstadt-Walker methods, with a tolerance spanning from 10^{-1} to 10^{-4} ; the internal linear solver *fgmres* and the
 681 number of iteration allowed spans from 4 to 180.

682 9 Supplementary table and figures

683 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
S	s^{-1}	Negative and positive volumetric source terms
τ_{ij}, τ_{II}, P	Pa	Deviatoric stress component, 2nd invariant of the deviatoric stress tensor, Pressure
ρ	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
$\dot{\tau}$	$Pa s^{-1}$	Objective deviatoric stress component time derivative
ω_{ij}	s^{-1}	Spin tensor component
G	GPa	Shear modulus
η	$Pa s$	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
ϕ, ϕ_0, ϕ_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	μWkg^{-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
V_{eff}, V_{int}	m^3	Effusion and intrusion volume
V_{lcor}	n.d.	Volumetric correction
v_{int}	$m^3 m^{-1}$	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}, \bar{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 $C_1 = 3MPa$); 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 $1600 \text{ }^\circ\text{C}$. (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 (≤ 1.0 GPa), middle pressure condition ($1.0 \leq P \leq 1.8$ GPa) and high pressure (≥ 1.8 GPa). \overline{LT} , \overline{MT} and \overline{HT} represents the volumetric weighted proportion of felsic melts produced at low temperature condition ($\leq 800 \text{ }^\circ\text{C}$), middle temperature condition ($800-1000 \text{ }^\circ\text{C}$) and high temperature condition ($\geq 1000 \text{ }^\circ\text{C}$) 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 (δ_{crust}^{av}). 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)

Table S4: 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 (δ_{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)

684 9.2 Supplementary Figures

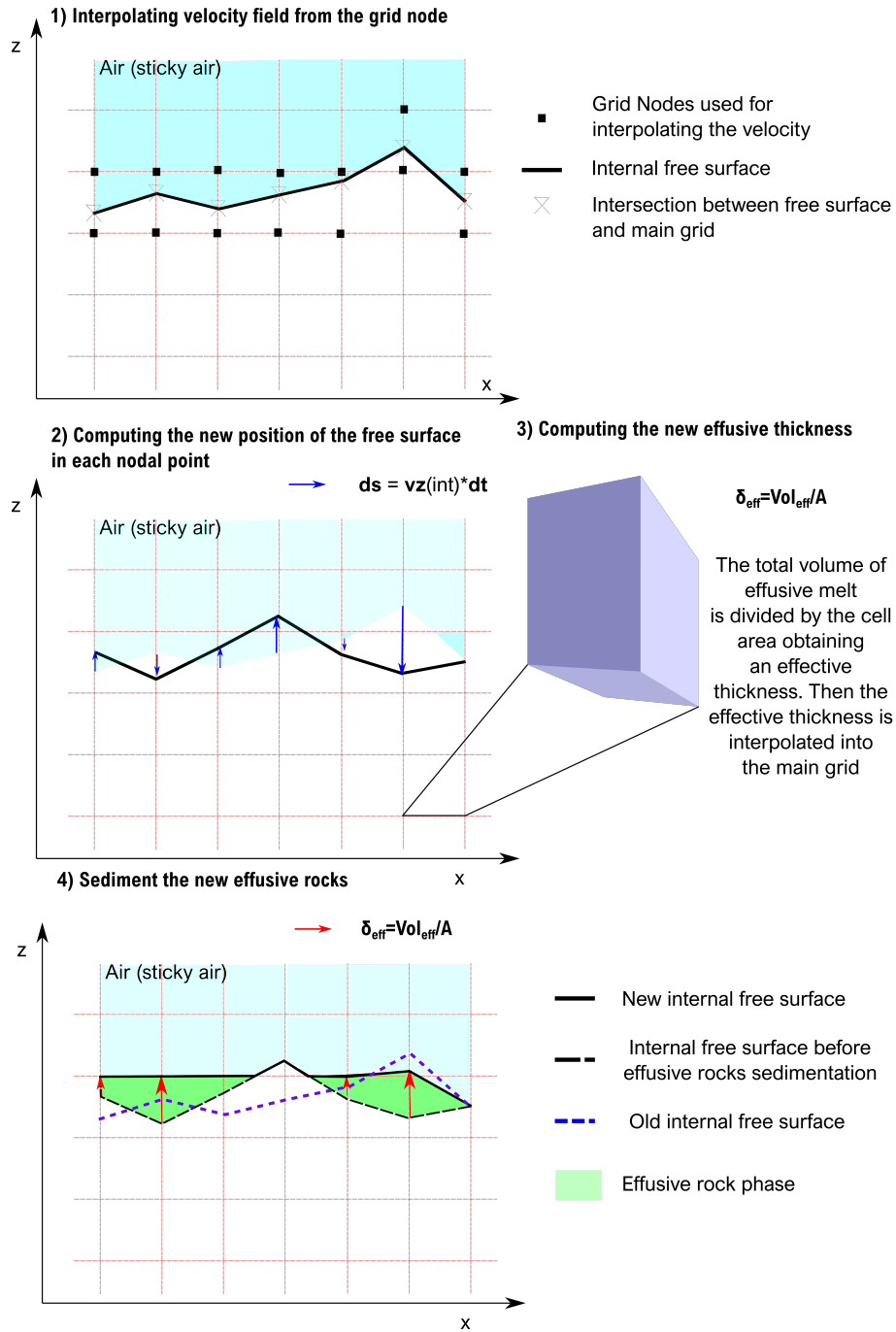


Figure S1: Simple representation of the free surface. 1) All the velocity's component are interpolated from the main grid node to the actual z coordinate of the free surface; 2) The z component is multiplied for the actual time-step (dt), then the free surface coordinate are shifted accordingly to this displacement; 3) The total volume of effusion is defined in the cell center. To retrieve the effective thickness Vol_{eff} is divided by the cell area; 4) The free surface is again shifted according to δ_{eff} and the difference between the previous surface and the new one is filled with the effusive phase (e.g. BS0%).

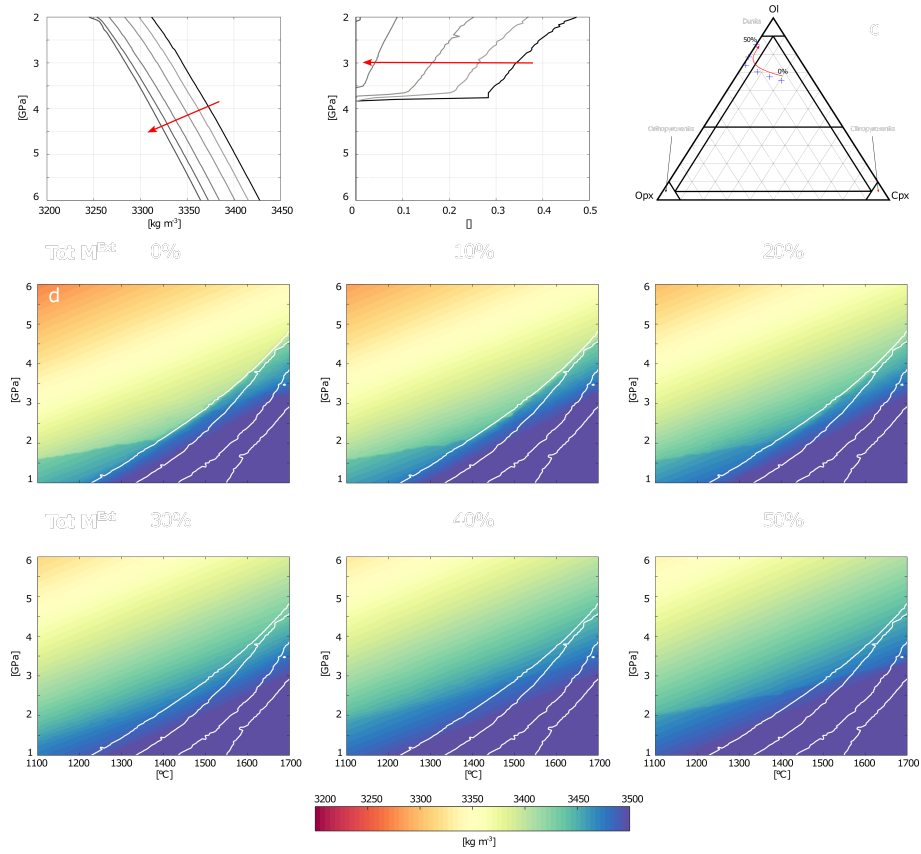


Figure S2: **a**): Density as a function of pressure at constant T_p (1200 °C), red arrow indicate increase depletion; **b**): Melt quantity as a function of pressure at constant T_p (1550 °C), red arrow indicates the increase depletion; **c**): Mode amount of clinopyroxene, orthopyroxene and olivine; **d**): Density of each depletion step.

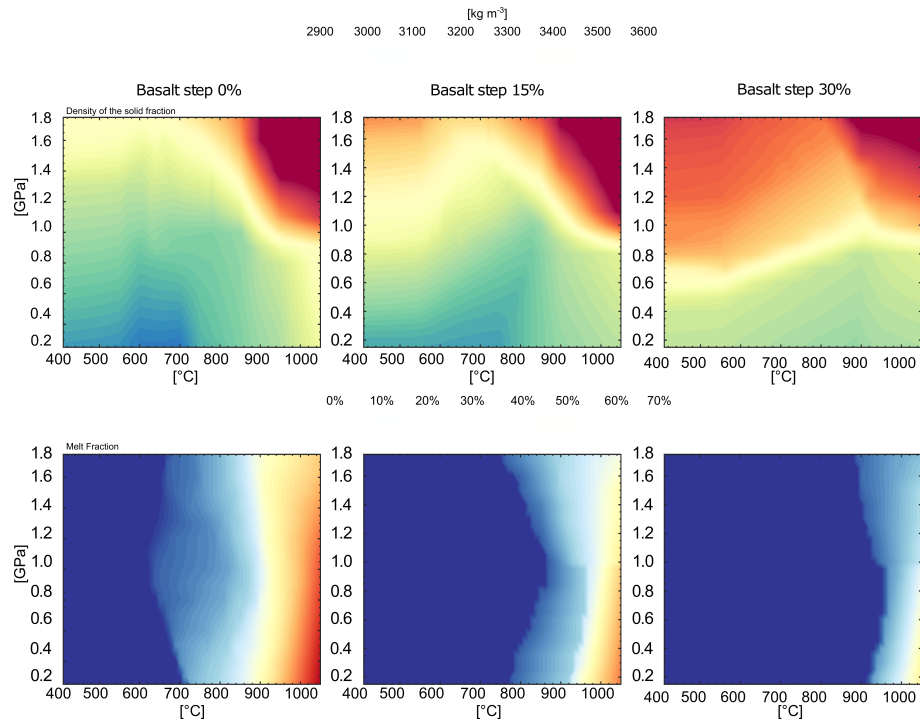


Figure S3: All these figure represents density and melt fraction as a function of $P-T$ conditions for the crustal phase diagram. The first row represents the density of the solid fraction, while the second represents the volumetric fraction of melt.

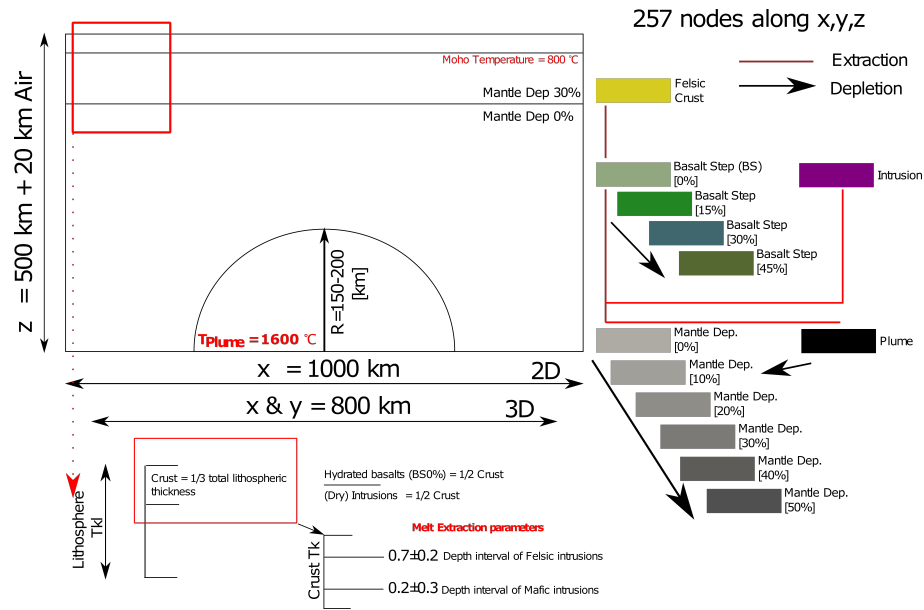


Figure S4: Initial setup.

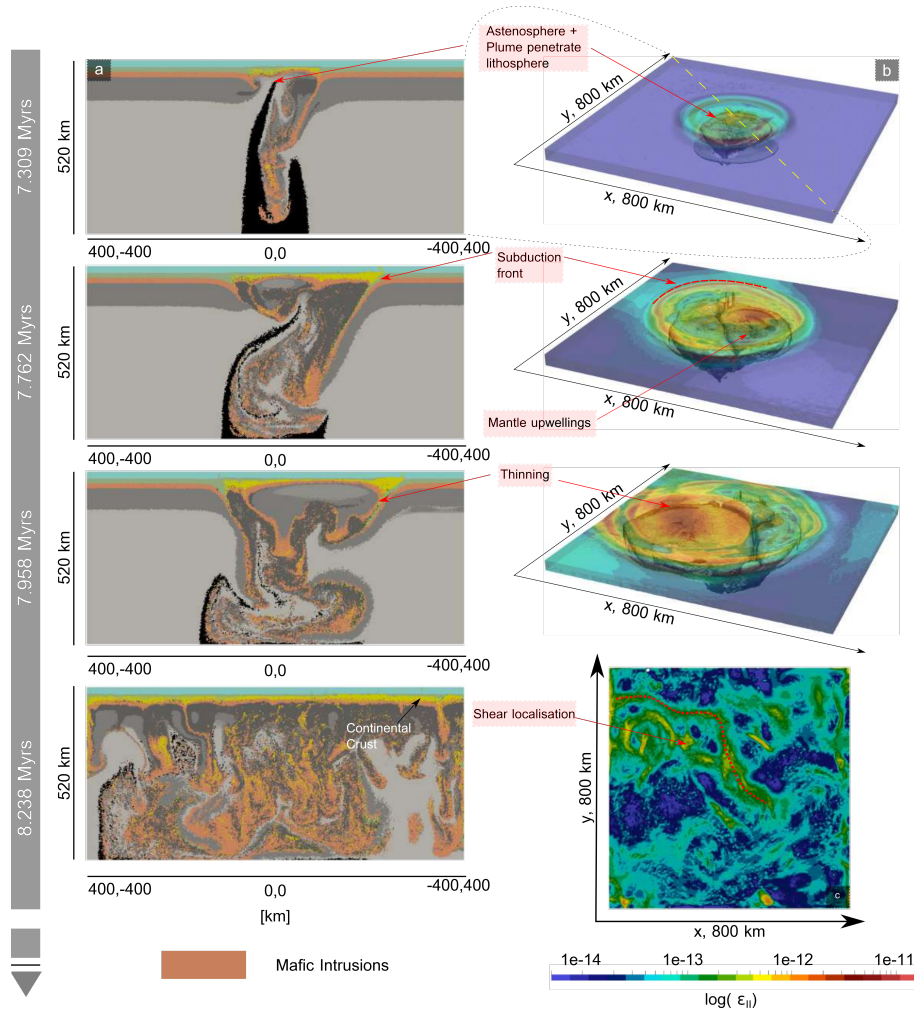


Figure S5: **Inefficient melt extraction, plume radius: 150 km, $dM = 0.001$** All the picture are ordered chronologically from top to bottom. **a)**: Compositional profile taken along a1-a2 $[400\ x, -400\ y] [-400\ x, 400\ y]$; **b)**: The transparent plot represents the strain rate state of the crust, 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.

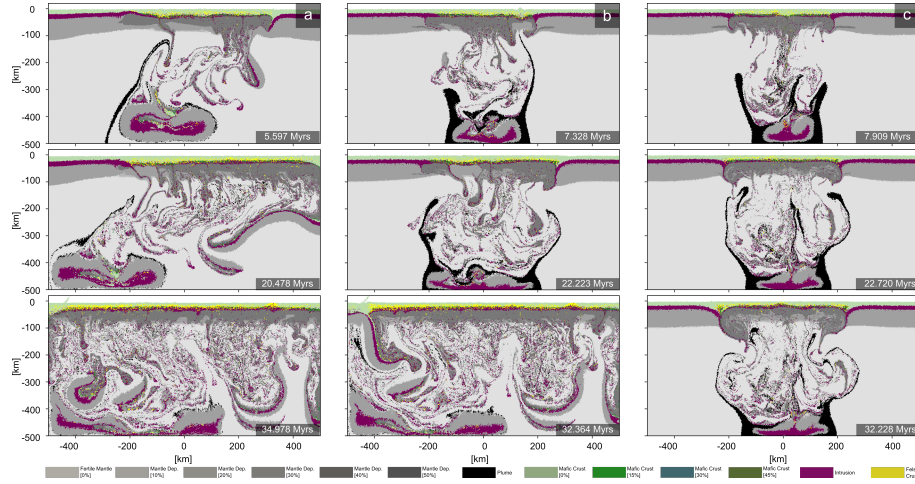


Figure S6: All experiments features the same dM (0.05) and M^{Trs} (0.001). **a)** Initial $T_p = 1400$ °C, lithospheric thickness, $\delta_{tkl} = 100$ km (*Test3L₁₄₀₀* in Tab. S3); **b)** Same as *a* except for the lithospheric thickness, $T_{kl} = 75$ km (*Test3L₁₄₀₀LT* in Tab. S3); **c)** Same as *b*, but lower radiogenic heating and with a bottom lower thermal boundary equal to 1600 °C (*Test3L₁₄₀₀LR-LT* in Tab. S3).

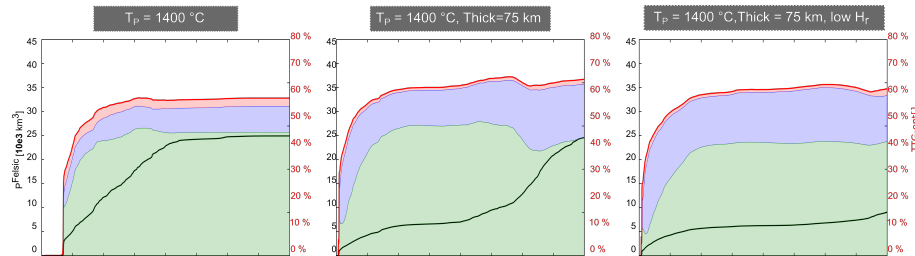


Figure S7: All the experiments are shown in Fig.S6, the list of input parameter are listed in Tab.S3. **a)** Each sub-plot has two *y axes*. *Left y axis*: P^{Felsic} against time. The black tick line refers to the left axis and depict the evolution of the total volume of felsic melt extracted with time; *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 TTT-like melts. The shaded coloured 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.

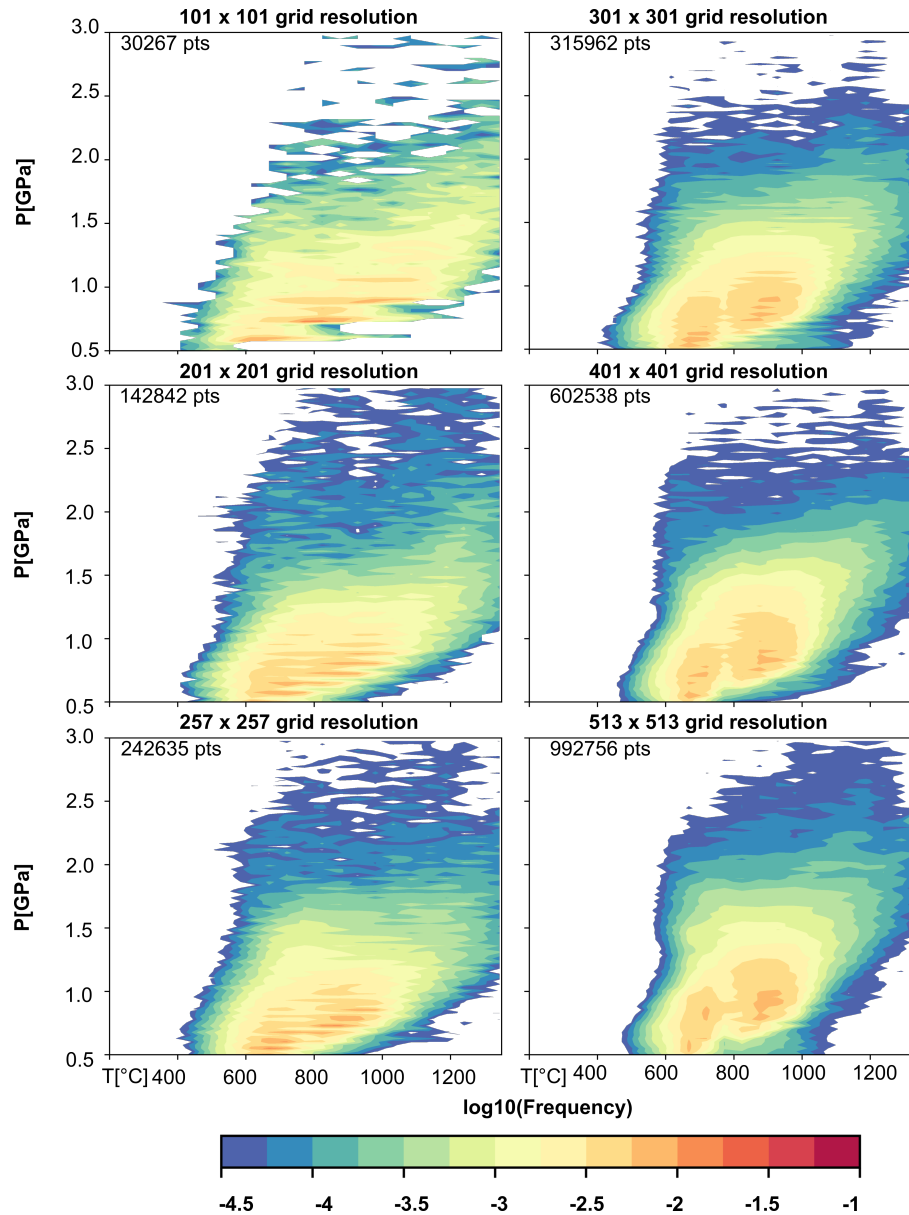


Figure S8: This figure represents the condition at which the felsic melts have been extracted during the whole simulation. We subdivided the P - T space in small elements ($dP = 0.02$ GPa $dT = 50$ °C), and we compute the logarithmic frequency of felsic melts extracted in each of these small elements. We performed this operation for the whole resolution experiments. These figures testify that in all the numerical experiments there are not dramatic differences in melt extraction patterns, 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 proportions of LP-MP-HP TTGs (listed in Tab. S4) are affected by the resolution of the crust.

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