

The influence of plume head–lithosphere interaction on magmatism associated with the Yellowstone hotspot track

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ABSTRACT

Although commonly attributed to a mantle plume, time-transgressive magmatism of the Snake River Plain–Yellowstone (SRPY) province differs in important ways from that associated with typical oceanic hotspots. A fundamental question concerns the relative contributions of lithosphere vs. upwelling sub-lithospheric mantle to formation of SRPY basaltic magmas. Specifically, association of this province with initially thick and cold Archean lithosphere (Wyoming craton) poses a problem in that this lid will hinder and possibly prevent melting of rising plume material. Assuming an anhydrous peridotite mantle, melting can only occur if (1) the lid can be substantially thinned over geologically reasonable time and/or (2) the upwelling material is exceptionally warm, or (3) the lid was suitably thin to begin with. Petrologic modeling indicates that SRPY primitive basalts last segregated from mantle at conditions (<1500 °C and ~ 100 km depth; Leeman, W.P., Schutt, D.L., Hughes, S.S., 2009. Thermal structure beneath the Snake River Plain: implications for the Yellowstone hotspot. *J. Volcanol. Geotherm. Res.*) only slightly warmer than MORB-source mantle and significantly cooler than sources of many oceanic hotspot magmas.

In this study, geodynamic models were developed to evaluate lithospheric thinning processes. The motivation for the modeling is the observation that if the lithosphere is initially more than 200 km thick – typical of many cratons – then thinning by at least a factor of two is required to allow decompression melting of an ascending plume, assuming low volatile content and high excess temperature (potential temperature >1500 °C). Fully dynamic models were applied to investigate the extent and rate of lithosphere thinning assuming an initial structure representative of the Wyoming craton. We find that thermal erosion by plume impingement alone appears incapable of providing the required lithospheric thinning. Alternative models (e.g., low-angle Laramide subduction, lithospheric delamination) also conflict with geochemical evidence that SRPY basalts contain a dominant contribution of old, isotopically evolved mantle material – presumably derived from subcontinental lithospheric mantle (SCLM). We conclude that SCLM is likely to be preserved, that the thick SCLM lid prevents substantial melting of rising plume material (tomographically imaged), and that SRPY basalts are predominantly derived by melting of lithospheric mantle.

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1. Introduction

The relative contributions of lithospheric vs. sublithospheric processes to the formation of basaltic magmas are a fundamental and unresolved issue for the Snake River Plain–Yellowstone (SRPY) province. It is commonly accepted that the magmatism in this area is related to migration of the North American plate over the Yellowstone plume (Leeman, 1982, 1989; Pierce and Morgan, 1992). Partial melting

of a mantle plume impinging the continental lithosphere has been invoked to account for onset (ca. 16 Ma) of voluminous flood basalt volcanism associated with the Columbia River Plateau (cf. Camp et al., 2003; Camp and Ross, 2004). The fact that the majority of the Columbia River basalt (CRB) lavas erupted in a short time (<1.5 Ma) suggests that melting of a large plume head was involved. By analogy, younger, age-transgressive (15 to 0 Ma) rhyolite–basalt volcanism of the SRPY has been implicitly linked to continued migration of North America over the tail of this plume. This view is supported by recent seismic tomographic imaging beneath Yellowstone National Park (Yuan and Dueker, 2005; Waite et al., 2006). These studies reveal a low velocity pipe, extending through a locally thinned transition zone to about 500 km depth, that is interpreted as a hot plume. Other

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evidence in favor of a mantle plume includes: the distribution of SRP silicic volcanism (generally progressively younger to the east [but see below]; Armstrong et al., 1975; Pierce and Morgan, 1992), high $^3\text{He}/^4\text{He}$ ratios in SRP basalts attributed to an anomalous mantle source (Craig et al., 1978; Graham et al., 2009-this volume), excess mantle temperature (ca. 55–200 °C or lower; Schutt and Dueker, 2008; Leeman et al., this volume) beneath the eastern SRP and Yellowstone, and high regional heat flow (Blackwell et al., 1991).

An important aspect of the current position of the hotspot is its location entirely beneath cratonic North America. The Snake River Plain is bounded on the west by the western Idaho suture zone that marks a structural transition between thick cratonic lithosphere to the east and thinner oceanic lithosphere of accreted terranes to the west. This boundary also coincides with a marked transition in Sr–Nd–Pb isotopic compositions of Mesozoic to Neogene igneous rocks that reflects a lateral change in the respective magma sources having oceanic character to the west and cratonic characteristics to the east (Armstrong et al., 1977; Leeman et al., 1992; Hanan et al., 2008). SRPY eruptive centers show a two-stage history beginning with an initial phase (2–4 Myr) that produced voluminous high-silica rhyolite lavas followed by a later phase of predominantly basaltic activity (Bonnichsen et al., 2008; Leeman et al., this volume). The rhyolites are metaluminous and have distinctive Sr and Nd isotopic compositions compared with similar rocks in central Nevada and eastern Oregon (Leeman et al., 1992; Brueske et al., 2008; Streck and Gruner, 2008) that implicate partial melting of the underlying cratonic crust in response to intrusion of massive volumes of basaltic magma (Bonnichsen et al., 2008; Leeman et al., 2008). In other words, the thick continental crust of the cratonic plate acts as a density filter that masks the true magnitude of basaltic magmatism. Despite this complication, cumulative SRPY basaltic magma production is probably at least an order of magnitude less than that associated with the Columbia River flood basalt event and it was distributed over a longer period of time.

Although it is logical to associate SRPY magmatism directly with the imaged mantle plume under Yellowstone (Yuan and Dueker, 2005; Waite et al., 2006), it has yet to be demonstrated that the primitive basaltic magmas are indeed derived from such a source. In fact, several lines of evidence seem to contradict this interpretation. First, the pattern of SRPY magmatism differs in important ways from that associated with typical oceanic hotspots. Instead of a regular monotonic age progression, there are notable anomalies in the ages of silicic volcanism with distance from Yellowstone. For example, very broad distribution of silicic centers and vents between 10 and 11 Ma attests to near coeval intrusion of basaltic magmas into the crust over a broad distance exceeding 400 km (Bonnichsen et al., 2008). Second, the chemistry of SRPY basalts suggests that they represent melts of sources distinct from most oceanic hotspots. In particular, Sr–Nd–Pb isotopic and related trace element data in primitive basalts are suggestive of significant melt contributions from ancient lithospheric mantle (Doe et al., 1982; Hildreth et al., 1991; Hanan et al., 2008). At the least, such evidence implies that the subcontinental lithospheric mantle remains at least partly intact beneath this region. Finally, assuming the ascending plume to be “dry” peridotite, melting cannot begin until this material ascends within roughly 100 km of the surface (cf. Leeman et al., 2008; this volume). In other words, if the overriding lithospheric plate exceeds this thickness, rising plume mantle cannot ascend to shallow enough depths for melting to occur unless or until the lid is sufficiently thinned.

Thus, the capacity of the plume to produce significant melt volumes hinges on whether or not the cratonic lithosphere was thinned sufficiently and on a time scale (no more than a few Myr) consistent with migration of SRPY magmatism. By comparison, thinner lithosphere (<100 km) associated with accreted oceanic terrains under the Columbia Plateau ca. 16 Ma would present little or no impediment to melting of ascending plume mantle.

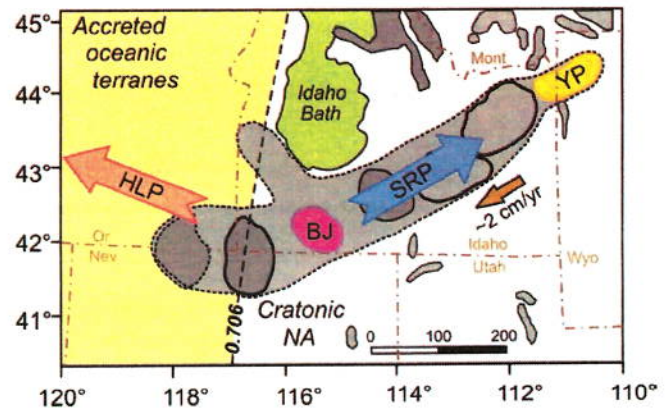


Fig. 1. Map view showing the Yellowstone hot spot track (gray shading) with approximate loci of major silicic eruptive centers. Broad arrows show generalized age progression for onset of magmatism over time. For the Snake River Plain (SRP), earliest magmatism occurred ca. 15 Ma near the Oregon–Idaho–Nevada state juncture, then migrated to Yellowstone (YP) by ca. 2 Ma. In the southeastern Oregon High Lava Plains (HLP), silicic magmatism migrated to the northwest following the outpouring of Columbia River flood basalts (CRBs) ca. 16 Ma. These two regions are underlain by dramatically different lithosphere architectures separated by a narrow suture zone that is defined by the $^{87}\text{Sr}/^{86}\text{Sr} = 0.706$ isopleth for igneous rocks (dashed line; cf. Leeman et al., 1992). To the east is Archean basement of the Wyoming craton (outcrops in dark gray) modified along its western margin by emplacement of the late Cretaceous–Early Tertiary Idaho Batholith. To the west is a collage of accreted oceanic terranes of Paleozoic–Mesozoic age. The proportion of erupted basalt: rhyolite increases sharply west of the suture zone, whereas the overall volume production of post-CRB volcanism is significantly greater to the east.

Although the exact thickness of the pre-plume mantle lithosphere is unclear, the Proterozoic sedimentary record suggests the Archean lithosphere thinned during rifting (e.g. Karlstrom and Humphreys, 1998). However, geochemical evidence supports the existence of a mixed Proterozoic and Archean lithosphere beneath much of this region today (Mueller and Frost, 2006; Hanan et al., 2008), and seismic tomography of the region shows high mantle velocities east of the Sr 0.706 line (cf. Fig. 1), suggesting a possible geochemical mantle boundary here (Roth et al., 2008).

Assuming the Wyoming craton lithosphere had a fairly uniform thickness, the unthinned depth of the lid can be estimated from mantle xenoliths erupted in the Great Falls Tectonic Zone, near the northern margin of the craton (Carlson, 1994). Here, xenoliths sampled from depths shallower than 150 km have cratonic equilibration temperatures and some yield Archean model ages, whereas those from greater depths have younger ages and higher equilibration temperatures. Based on combined gravity and heat flow data, Artemieva (2006) suggests lithospheric thicknesses of 125–150 km in the region. Seismic tomography also suggests comparable lithospheric thicknesses to the east of Yellowstone (Schutt et al., 2008; Stachnik et al., 2008).

In this paper we use numeric models to investigate the impact of a thermo-chemical plume on the lithosphere adjacent to the suture zone and implications regarding contributions of lithosphere vs. upwelling sub-lithospheric mantle in forming SRPY basaltic magmas. Specifically, we attempt to identify conditions under which sufficient lithospheric thinning can plausibly occur such that an ascending plume could produce the SRPY basaltic melt flux. In the models considered in this study, the impacts of several parameters on the thickness of the lithosphere are assessed; these include plume head size, plume excess temperature, plume density anomaly, initial lithospheric thickness, lithospheric viscosity, and lithospheric horizontal velocity (plate motion). Our models are purposely simple as our goal is to determine under what conditions the Yellowstone plume could begin to melt, given a range of starting

lithosphere thickness (100–200 km). This is determined as a range of conditions for which ascending plume mantle is able to cross the dry peridotite solidus.

2. Geodynamic models of plume–lithosphere interaction

A classic mantle plume is an upwelling of high temperature material, with a typical morphology of a large diameter head followed by a relatively narrow tail. As noted earlier, a plume head scenario seems most appropriate for onset of magmatism associated with the Columbia Plateau. In contrast, seismic tomography and relatively smaller magma production associated with the SRPY province are more consistent with a plume tail scenario. Over the last decade or so, several geodynamic studies have linked mantle plumes with flood basalt volcanism. The impact of a plume on the overlying lithosphere has been considered previously (cf. Ribe and Christiansen, 1999; Sleep, 2006; Manglik and Christensen, 2006; Burov et al., 2007). Watson and McKenzie (1991) numerically modeled the impact of a large plume upwelling beneath a stationary plate to explain melt production and geochemical characteristics of Hawaiian basaltic magmatism. Ribe and Christiansen (1999) used 3-D Cartesian models incorporating plate motion to study Hawaiian plume–lithosphere interaction. Other studies focused on the interaction of a plume head with a rifting lithosphere (Leich et al., 1998). Recently, Burov et al. (2007) introduced realistic rheology of the continental lithosphere to study the mantle plume impact and its consequences on tectonic evolution.

These and similar studies clearly suggest that efficiency of thermal and mechanical erosion of the lower lithosphere increases with volume and “thermal strength” (defined here in terms of combined effects of excess temperature, size and buoyancy) of the plume; erosional efficiency decreases as velocity of the upper plate increases (owing to decreased thermal coupling) and as lithosphere age and thickness increase (upper plate is cooler). The present paper specifically attempts to define the conditions that might plausibly lead to sufficient thinning of cratonic lithosphere to enable rising plume material to undergo partial melting and at a scale adequate to produce the inferred magma flux for the SRPY province. Our focus differs from most previous studies in that we are interested in the influence that the ‘mechanical and chemical’ lithosphere exerts on ascending plumes.

3. Numerical model

3.1. Algorithm

Mantle convection is governed by the coupling between fluid flow and energy transport. The calculations are performed in a 3-D cut through a sphere on a non-deforming grid, by solving the conservation equations of mass, momentum and energy under the Truncated Anelastic Liquid Approximation which includes dissipative heating in the energy equation (Tan et al., 2006). The equations are:

$$(\rho u_i)_{,i} = 0 \quad (1)$$

$$-P_{,i} + \left(\eta \left(u_{i,j} + u_{j,i} - \frac{2}{3} u_{k,k} \delta_{ij} \right) \right)_{,i} + \delta \rho g \delta_{ir} = 0 \quad (2)$$

$$\rho c_p (T_t + u_i T_{i,t}) = \rho c_p k T_{,ii} + \rho \alpha g u_r (T + T_0) + \phi + \rho (Q_{L,t} + u_i Q_{L,i}) + \rho H \quad (3)$$

where, u_i is the velocity, P is the dynamic pressure, Ra is the Rayleigh number, η is the viscosity, δ_{ij} is the Kronecker delta tensor, $\delta \rho$ is the density anomaly, T is the temperature, T_0 is the temperature at surface, $\delta \rho$ is the density anomaly, c_p is the heat capacity, k is the thermal diffusivity, α is the thermal expansivity, ϕ

is the viscous dissipation, Q_L is the latent heat, and H is the heat production rate. The expression $X_{,y}$ represents the derivative of X with respect to y , where i and j are spatial indices, r is the radial direction, and t is time.

The density anomaly is:

$$\delta \rho = -\alpha \bar{\rho} (T - \bar{T}_a) + \delta \rho_{ch} C \quad (4)$$

where: $\bar{\rho}$ is the radial profile of density, $\bar{T}_a$ is the radial profile of adiabatic temperature, $\delta \rho_{ch}$ is the density difference between the compositions and C is the composition.

The viscosity formulation is the following:

$$\eta(r, T) = e_{ij}^{\frac{1-n}{2}} f(r)^{\frac{1}{2}} e^{\frac{1}{2} \left(\frac{c_1}{c_2 - T} - \frac{c_1}{c_2 - T_m} \right)} \quad (5)$$

where $f(r)$ is a non-dimensional value that can differ by layer. The dimensional viscosity in each layer is obtained by multiplying the non-dimensional value with the reference value, $\eta_0 = 10^{21}$ Pa s. $T_m = 1$ is the non-dimensional temperature of ambient mantle, n is the power-law exponent ($n=3$), e_{ij} is the 2nd invariant of strain rate tensor and c_1 and c_2 are constants (Manea and Gurnis, 2007). A non-dimensional viscosity variation of 10^4 or 0.01–100 is used across the domain, as described in the individual model runs.

3.2. Model setup and boundary conditions

Using the finite element package *CitcomS.py Version 3.0.1* (Tan et al., 2006) available from the Computational Infrastructure for Geodynamics (CIG) (<http://geodynamics.org>), the computations are performed within a 3D spherical domain (θ, φ, r) , where θ is latitude, φ is longitude and r is radius (Fig. 2). The inner radius corresponds to a depth of 1900 km, the outer radius the surface of the Earth. The span in latitude and longitude is 57° . This domain is evenly divided into 129 elements in the radial dimension and 257×257 elements in areal (i.e., latitude and longitude) dimension, corresponding to a $15 \times 25 \times 25$ km resolution.

The boundary conditions are as follows. The top boundary has either a free slip or an imposed velocity boundary condition corresponding to the movement of the North America plate in a hot-spot reference frame (~ 2 cm/yr). The top and bottom boundaries are assumed to be isothermal. The bottom is free slip, and the sides are reflecting. The initial thermal structure is described by a dual age thermal boundary layer (cf. Fig. 3) with old lithosphere as constrained by mantle xenolith data and young lithosphere constrained from a global thermal model for continental lithosphere (Model TC1 of Artemieva, 2006). The mantle potential temperature is considered to be 1450°C (upper limit for SRP basalts (Leeman et al., 2008)). This background mantle is divided into four layers: lithosphere (variable thickness), upper mantle (variable thickness), transition zone (410–670 km) and part of lower mantle (670–1300 km). The lithosphere has a variable thickness, ranging from 100 km for young lithosphere to 100–200 km for the craton. The model calculations start with an initial thermo-chemical mantle plume placed at 1650–1750 km depth (depending on the initial plume diameter: 300–500 km), an average temperature excess $\Delta T_{\text{plume}} = 150$ – 300°C and a density deficit $\Delta \rho_{\text{plume}} = 30$ – 50 kg/m^3 . These parameters strongly influence plume ascent velocity, and lower values result in unreasonably slow ascent. The plume geometry used in the modeling is obtained by integrating the initial model until the plume head arrives at a depth of 660 km. We track the time-space evolution of the plume by a tracer technique, where the initial spherical plume is filled up with 1 million particles (or tracers). Also, a thin layer 50 km thick located at the base of the

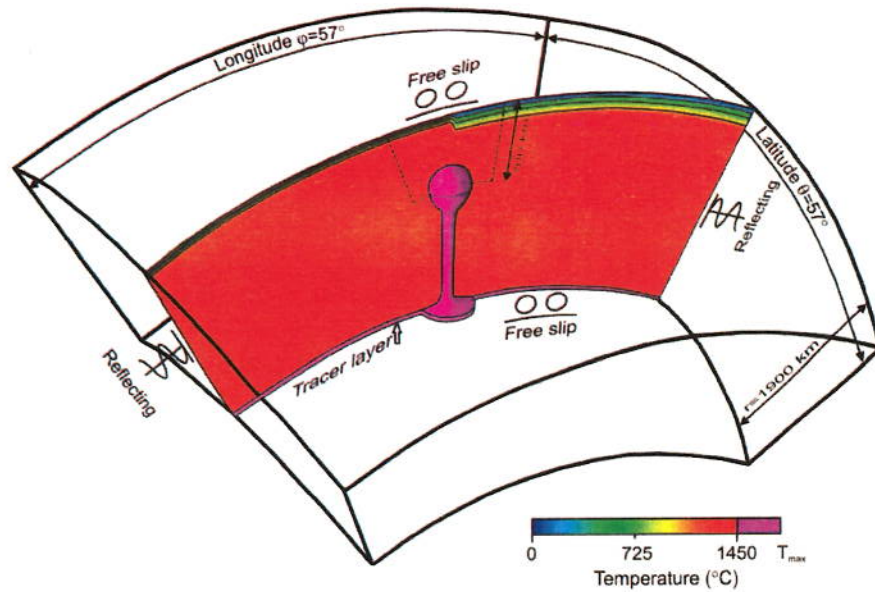


Fig. 2. Model setup and boundary conditions. The 3D box size is $57^\circ \times 57^\circ \times 1900$ km. The lithospheric thickness varies from 100 to 150–200 km. The initial plume-head geometry varies from 300 to 500 km diameter and is filled with tracers (1 million). Also an initial 50 km thick tracer layer covers the bottom of the box. The top boundary condition is either free-slip or kinematic (2 cm/yr). All lateral boundaries are reflecting. The bottom face has a zero velocity imposed. The modeling results are shown in subsequent figures for the 700 km deep dashed box.

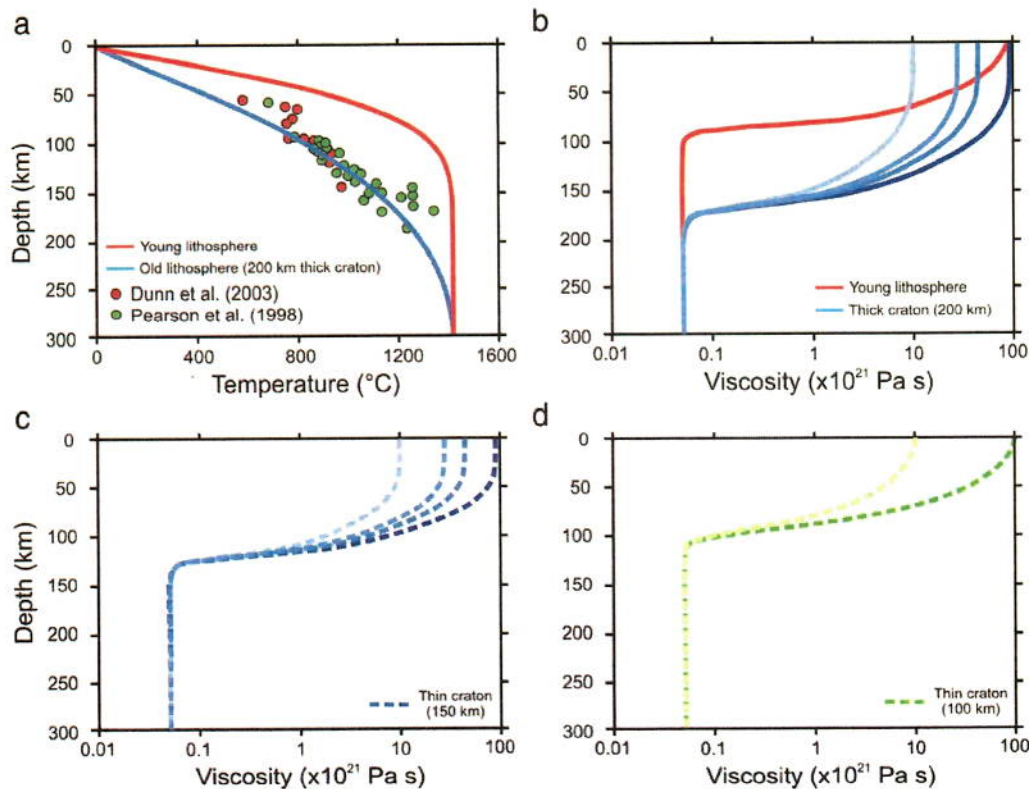


Fig. 3. (a): Geothermal profiles across the craton (blue curve) and young lithosphere (red curve). The craton geotherm is constrained using xenolith data from Dunn et al. (2003) and Pearson et al. (1998). The young geotherm is consistent with model TC1 of Artemieva (2006). Other panels show viscosity vs. depth relations for different model scenarios: (b) Initial viscosity profiles through the young lithosphere (red) and 200 km thick craton (blue tones); dark blue curve stands for strong lithosphere (reference viscosity 10^{23} Pa s); light blue tones are for a weaker lithosphere where the viscosity is lowered by up to an order of magnitude (reference viscosity 10^{22} Pa s); (c) the same as (b) but for a thinner lithosphere (150 km); (d) the same as (c) but for a 100 km thick lithosphere. The upper mantle viscosity is considered 5×10^{19} Pa s from Hirth and Kohlstedt (2003).

Table 1

Model parameters held constant.

Parameter		Value
Reference density	ρ_0	3300 kg/m ³
Thermal diffusivity	K	1×10^{-6} m ² /s
Thermal expansion	α	2×10^{-5} 1/K
Earth radius	R_0	6371.137 km
Gravitational acceleration	g	10 m/s ²
Reference viscosity	η_0	1×10^{21} Pa s

model is filled with 1 million tracers (see Fig. 2). This is used to fill with tracers the plume tail during the entire numeric simulation. Tracers' velocities are calculated by interpolating the eight nodal velocities with bilinear shape functions. Tracers are advected using the mid-point method (2nd order accuracy). Parameters held constant are summarized in Table 1.

4. Modeling results

We present the results in terms of the time–space temperature variation of plume and lithosphere, beginning at the time when the plume head first encounters the base of the model lithosphere. We also show the modeling results when we vary the following initial parameters: plume diameter, temperature excess and density contrast, and the initial thickness and strength (viscosity) of the cratonic lithosphere. The testing parameter table is presented in Table 2. We divided the models in two large categories depending on the upper plate boundary condition: free-slip or kinematic (imposed velocity). Within each category we vary the other parameters described above.

4.1. Free-slip upper boundary condition (upper plate stationary with respect to plume)

A first set of models is presented to show the effects of impingement of a plume centered beneath the suture between young accreted lithosphere (oceanic-like) and old cratonic lithosphere, and assuming that the upper plate remains stationary with respect to the ascending plume (a free slip upper surface is used as top boundary condition). Such models are likely to produce maximal warming and deformation of the upper plate. The results are presented as 2D slices through the central part of the 3D model. The plume head begins flattening when it approaches the suture zone. It is then divided into two distinct lobes which spread in the radial direction at a rate similar to the plume ascent rate. In a 2D section through the model, one lobe moves under the young lithosphere and the other lobe shifts under the craton. Significant portions of the lithospheric mantle are thermally eroded by the plume impingement. In order to investigate whether the lithospheric thinning might produce melting or not, we quantify the modeling results by comparing pressure–temperature profiles through the plume to the peridotite dry *solidus* of Hirschmann (2000).

For the first three models (A_1, A_2 and A_3 – see Table 2), the initial plume head diameter and excess temperature are varied while keeping constant the density excess (30 kg/m³) as well as lithosphere thickness (200 km) and strength (10^{23} Pa s). The results show how the mantle part of the lithosphere is thinned and when and where the plume head intercepts the dry *solidus*. For example, in model A_2 (Fig. 4), the plume head (400 km diameter) intersects the *solidus* beneath the young accreted lithosphere at ~120 km depth after ~3 Ma. On the other hand, the thick cratonic lithosphere to the east prevents the plume from melting although there is ~50 km of thinning. For comparison, a weaker plume with diameter of 300 km and excess temperature of 150 °C (Model A_1) fails to produce plume melting.

Table 2

Model runs.

Model ID	Plume head initial diameter (km)	Initial temperature excess (°C)	Density excess (kg/m ³)	Craton thickness (km)	Lithosphere viscosity (Pa s)	Top BC
A_1	300	150	30	200	10^{23}	Free slip
A_2	400	250	30	200	10^{23}	Free slip
A_3	500	300	30	200	10^{23}	Free slip
B_1	300	150	30	150	10^{23}	Free slip
B_2	400	250	30	150	10^{23}	Free slip
B_3	500	300	30	150	10^{23}	Free slip
C_1	300	150	50	200	10^{23}	Free slip
C_2	500	300	50	200	10^{23}	Free slip
D_1	300	150	50	150	10^{23}	Free slip
D_2	400	250	50	150	10^{23}	Free slip
E_1	300	150	50	150	5×10^{22}	Free slip
E_2	400	250	50	150	5×10^{22}	Free slip
F_1	300	150	30	200	5×10^{22}	Free slip
F_2	300	150	30	200	3×10^{22}	Free slip
F_3	300	150	30	200	1×10^{22}	Free slip
G_1	300	150	50	200	5×10^{22}	Free slip
G_2	400	250	50	200	5×10^{22}	Free slip
G_3	500	300	50	200	5×10^{22}	Free slip
H_1	400	250	30	200	3×10^{22}	Free slip
H_2	400	250	50	150	3×10^{22}	Free slip
I_1	300	150	30	100	10^{23}	Free slip
I_2	300	150	50	100	10^{23}	Free slip
J_1	500	300	30	100	10^{23}	Free slip
J_2	500	300	50	100	10^{23}	Free slip
K_1	300	150	30	100	10^{22}	Free slip
K_2	300	150	50	100	10^{22}	Free slip
L_1	500	300	30	100	10^{22}	Free slip
L_2	500	300	50	100	10^{22}	Free slip
AA_1	300	150	30	200	10^{23}	2 cm/yr
AA_2	400	250	30	200	10^{23}	2 cm/yr
AA_3	500	300	30	200	10^{23}	2 cm/yr
BB_1	300	150	30	150	10^{23}	2 cm/yr
BB_2	400	250	30	150	10^{23}	2 cm/yr
BB_3	500	300	30	150	10^{23}	2 cm/yr
CC_1	300	150	50	200	10^{23}	2 cm/yr
CC_2	500	300	50	200	10^{23}	2 cm/yr
DD_1	300	150	50	150	10^{23}	2 cm/yr
DD_2	400	250	50	150	10^{23}	2 cm/yr
EE_1	300	150	50	150	5×10^{22}	2 cm/yr
EE_2	400	250	50	150	5×10^{22}	2 cm/yr
FF_1	300	150	30	200	5×10^{22}	2 cm/yr
FF_2	300	150	30	200	3×10^{22}	2 cm/yr
FF_3	300	150	30	200	1×10^{22}	2 cm/yr
GG_1	300	150	50	200	5×10^{22}	2 cm/yr
GG_2	400	250	50	200	5×10^{22}	2 cm/yr
GG_3	500	300	50	200	5×10^{22}	2 cm/yr
GG_4	500	300	50	150	1×10^{22}	2 cm/yr
HH_1	400	250	30	200	3×10^{22}	2 cm/yr
HH_2	400	250	50	150	3×10^{22}	2 cm/yr
II_1	300	150	30	100	10^{23}	2 cm/yr
II_2	300	150	50	100	10^{23}	2 cm/yr
JJ_1	500	300	30	100	10^{23}	2 cm/yr
JJ_2	500	300	50	100	10^{23}	2 cm/yr
KK_1	300	150	30	100	10^{22}	2 cm/yr
KK_2	300	150	50	100	10^{22}	2 cm/yr
LL_1	500	300	30	100	10^{22}	2 cm/yr
LL_2	500	300	50	100	10^{22}	2 cm/yr

Models with thin cratonic lithosphere (150 km) show similar behavior (Model B_1, B_2 and B_3 – Table 2). Increasing the plume density contrast from 30 kg/m³ to 50 kg/m³ does not change the modeling results significantly (Models C and D – Table 2). However, changing the lithospheric viscosity has a significant effect. Lowering this parameter from 10^{23} Pa·s to 10^{22} Pa·s allows a strong plume (500 km initial diameter, and 300 °C initial temperature excess and 30 kg/m³ density deficit) to thin the cratonic lid by ~100 km and to intersect the dry *solidus* at ~100 km depth in ~13 Ma (Fig. 5; Model F_3). Also, unlike in the previously discussed models, there is an eastward tilt of the plume. This is caused by a strong convection

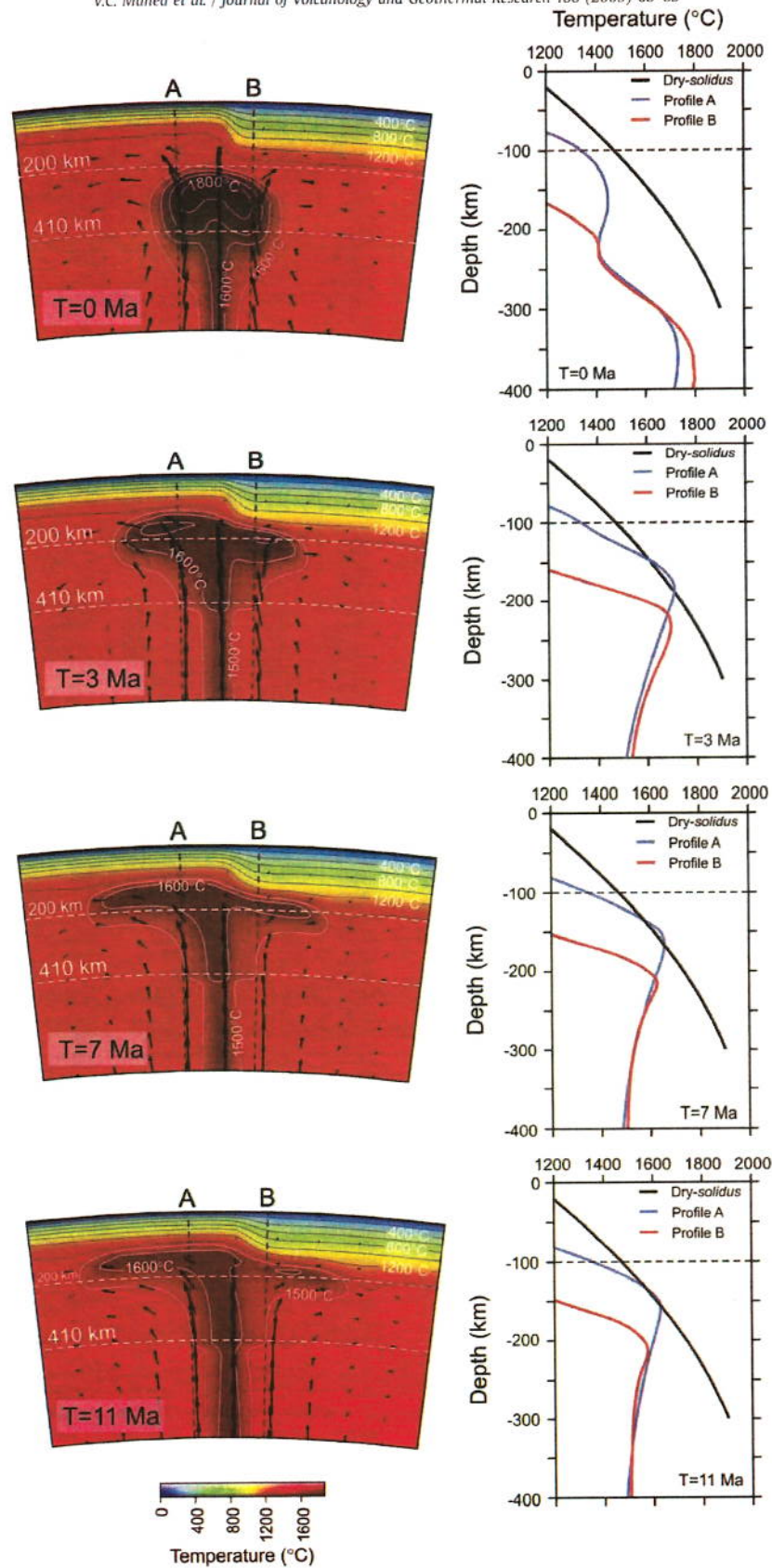


Fig. 4. Time-space evolution of plume–lithosphere impact for Model A.2 (see Table 2). The craton is considered initially 200 km thick and has a reference viscosity of 10^{23} Pa s. The plume head has an initial diameter of 400 km, an initial radial averaged temperature excess of 250 °C and a density deficit of 30 kg/m³. The upper boundary condition is considered free-slip. The right hand panels show P – T cross-sections through the young lithosphere and the craton. After 3 Ma the plume head reaches the dry peridotite solidus beneath the normal lithosphere (cf. blue curve), whereas the craton is not sufficiently thinned to allow melting (cf. red curve).

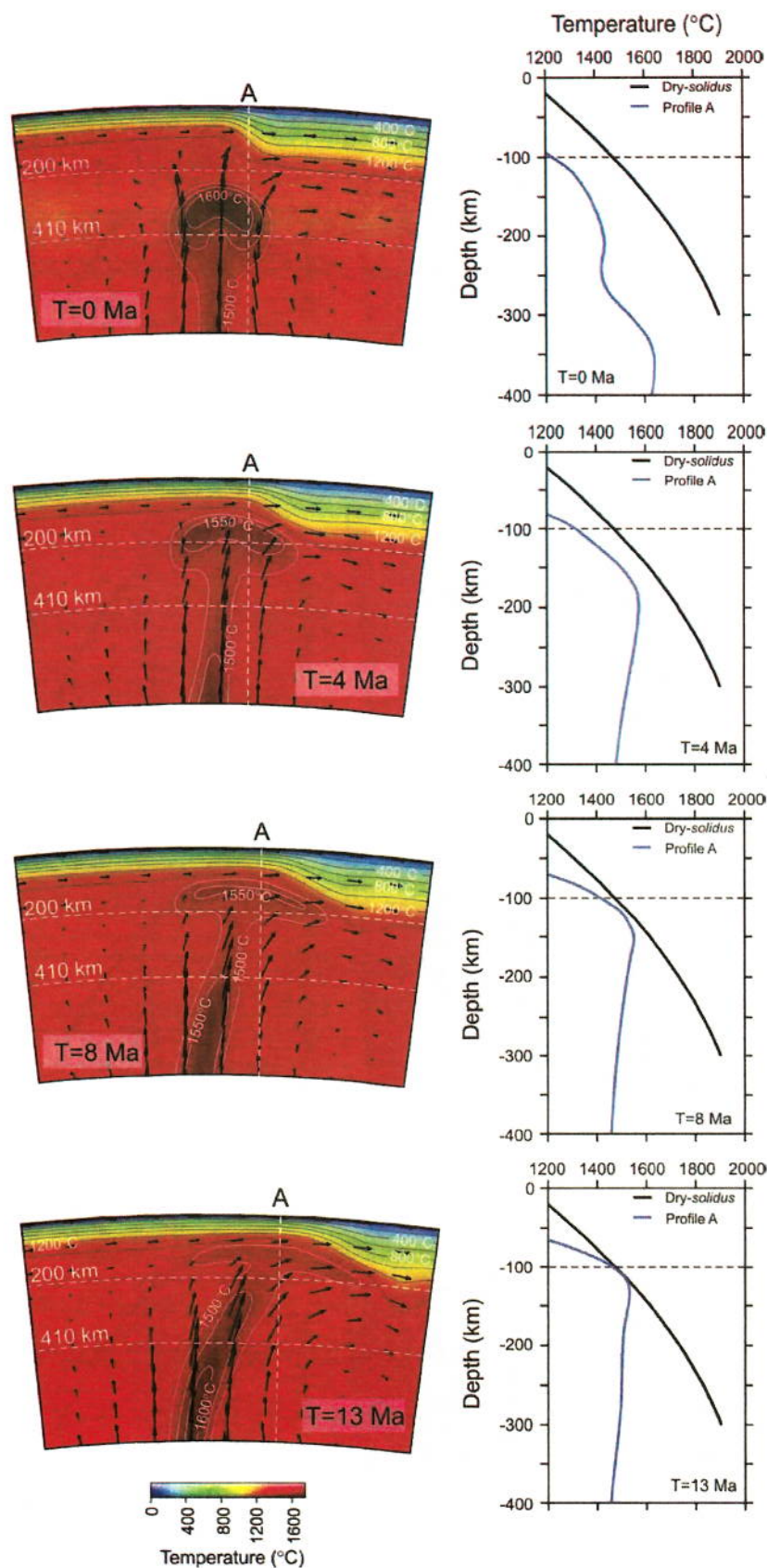


Fig. 5. Time-space evolution of plume-lithosphere impact for Model F.3 (see Table 2). The craton is considered initially 200 km thick and has a low viscosity of 10^{22} Pa s. The plume head has an initial diameter of 300 km, an initial radial averaged temperature excess of 150 °C and a density deficit of 30 kg/m³. Note how the plume tilts significantly toward the craton and after ~13 Ma the thick lithosphere starts to delaminate. In this experiment the craton is sufficiently eroded and the dry solidus is crossed at ~100 km depth.

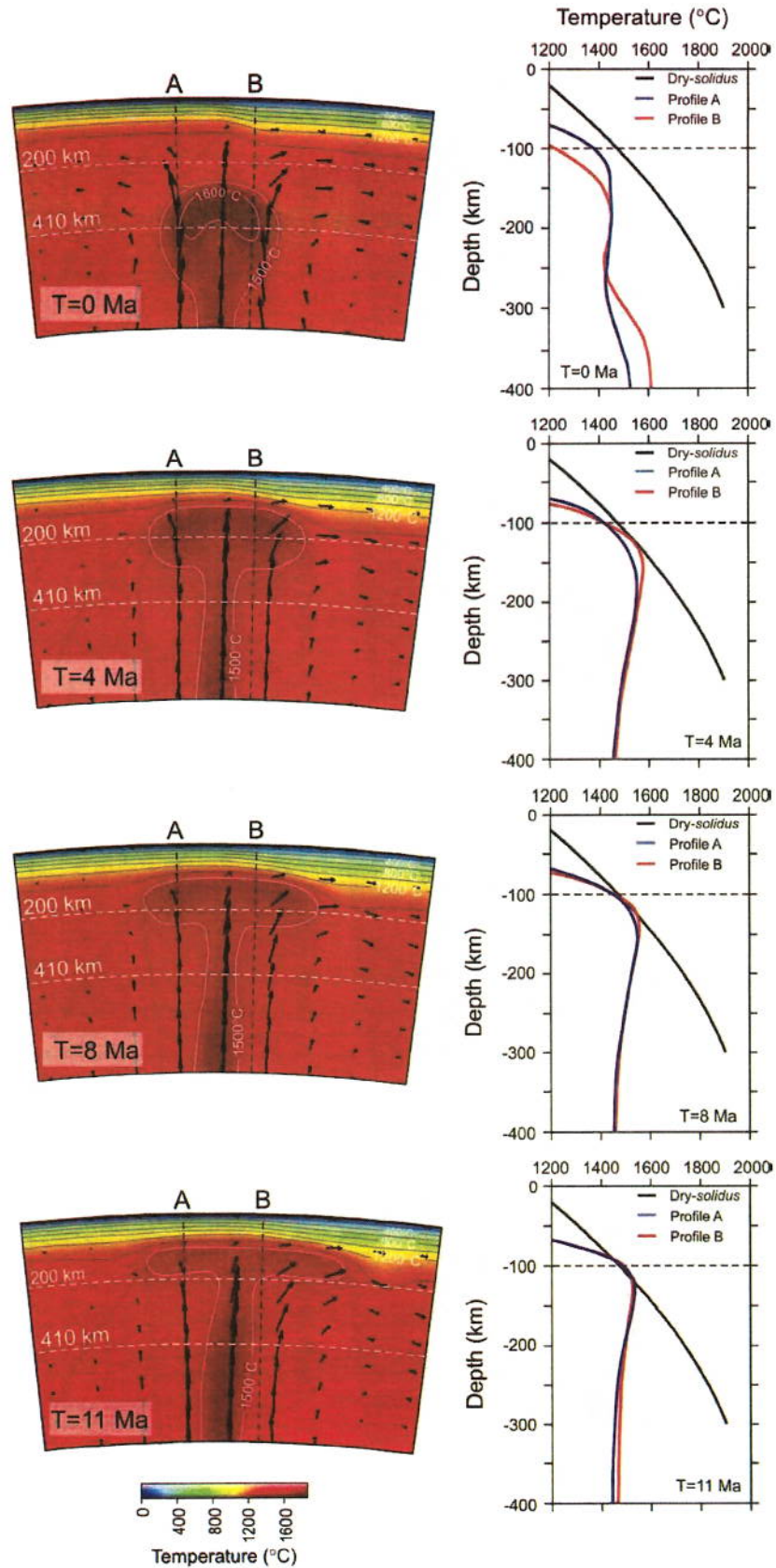


Fig. 6. Time-space evolution of plume-lithosphere impact for Model K.1 (see Table 2). The craton is considered initially 100 km thick and has a low viscosity of 1×10^{22} Pa s. The plume head has an initial diameter of 300 km, an initial radial averaged temperature excess of 150 °C and a density deficit of 30 kg/m³.

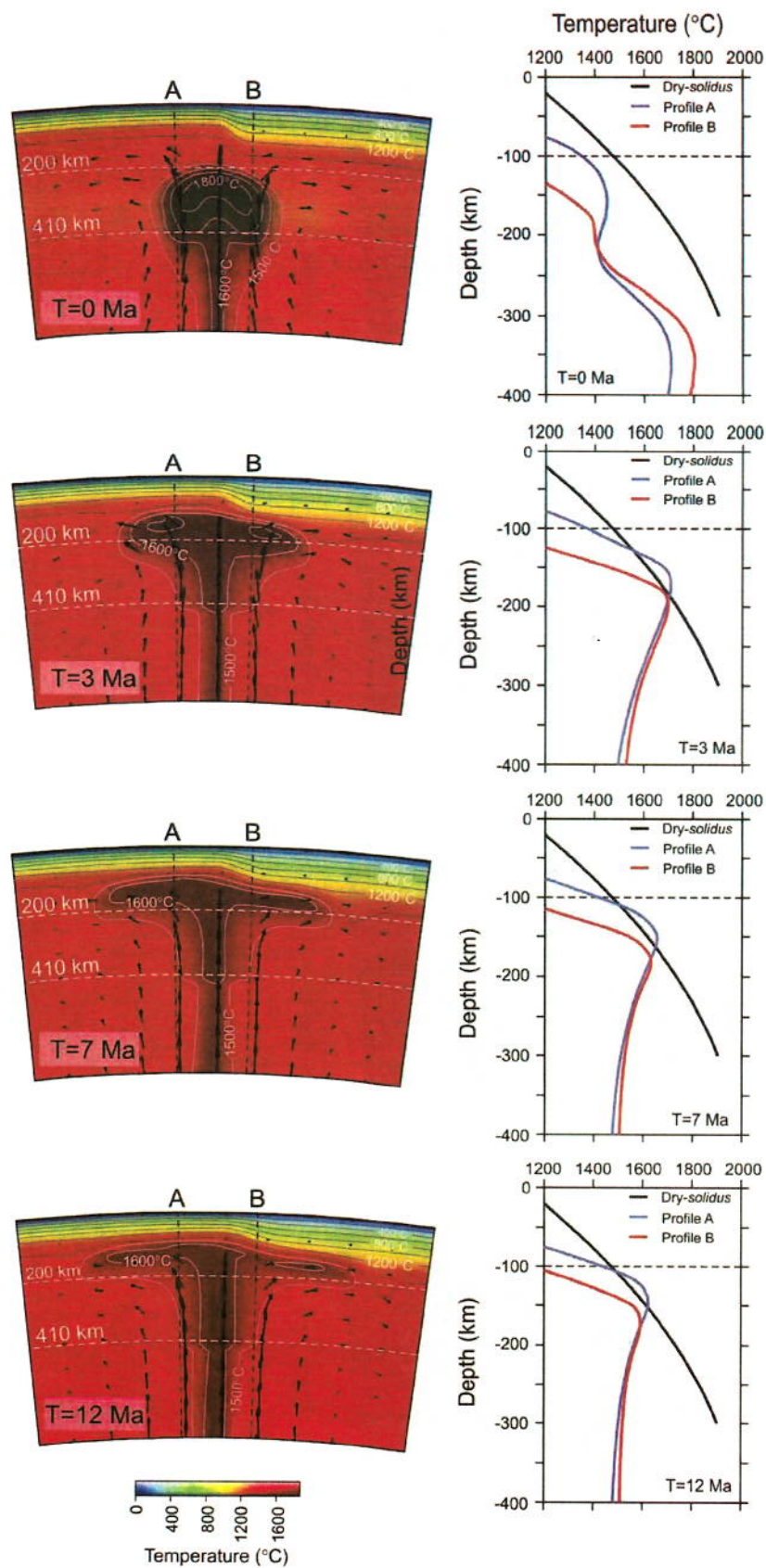


Fig. 7. Time-space evolution of plume-lithosphere impact for Model H.2 (see Table 2). The craton is considered initially 150 km thick and has a low viscosity of 3×10^{22} Pa s. The plume head has an initial diameter of 400 km, an initial radial averaged temperature excess of 250 °C and a density deficit of 50 kg/m³.

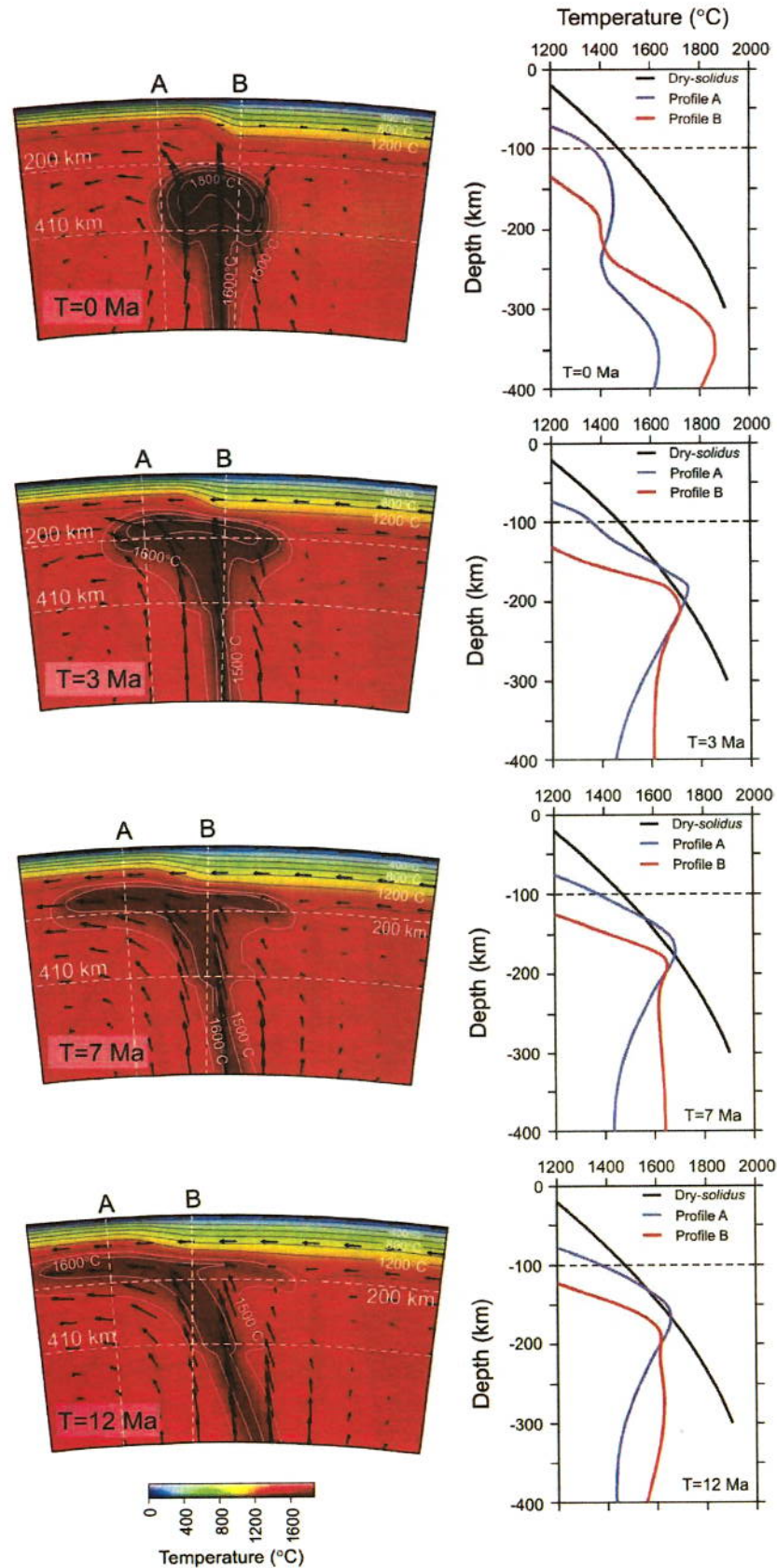


Fig. 8. Time-space evolution of plume-lithosphere impact for Model BB_2 (see Table 2). The craton is considered initially 150 km thick and has a viscosity of 10^{23} Pa s. The plume head has an initial diameter of 400 km, an initial radial averaged temperature excess of 250 °C and a density deficit of 30 kg/m³. In this model the westward movement of the lithosphere at 2 cm/yr deflects the plume head and tail and diminishes the plume efficiency to erode the thick lithosphere compared with the equivalent models with free-slip upper boundary condition.

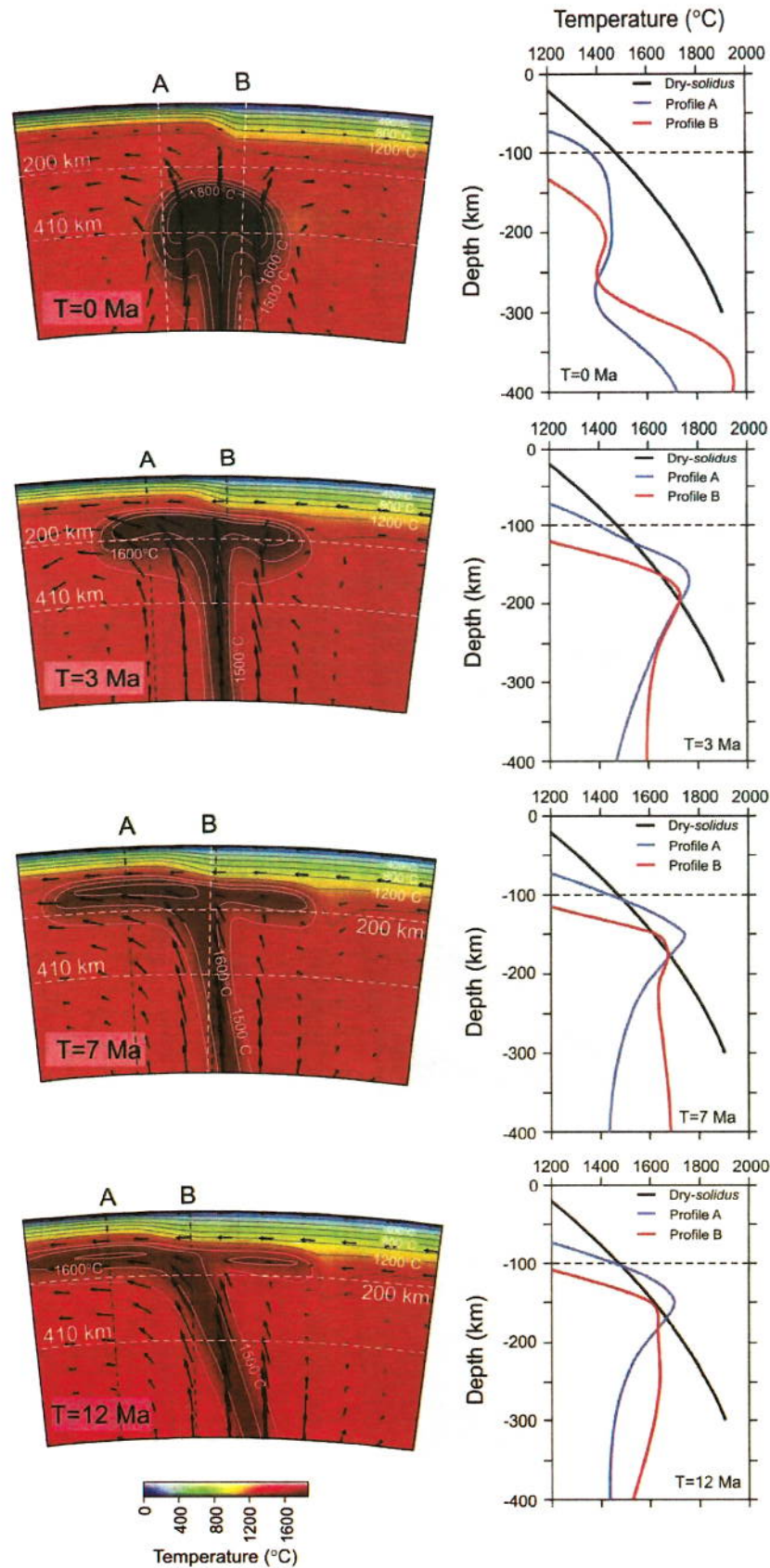


Fig. 9. Time-space evolution of plume-lithosphere impact for Model GG. 4 (see Table 2). The craton is considered initially 150 km thick and has a low viscosity of 10^{22} Pa s. The plume head has an initial diameter of 500 km, an initial radial averaged temperature excess of 300 °C and a density deficit of 50 kg/m³. Even if we push the model parameters beyond the acceptable upper threshold, the plume is not able to thin the thick lithosphere sufficiently to initiate melting at depths near 100 km. However, the high initial plume temperature heats the base of the craton lithosphere and dry solidus is crossed at ~150 km depth.

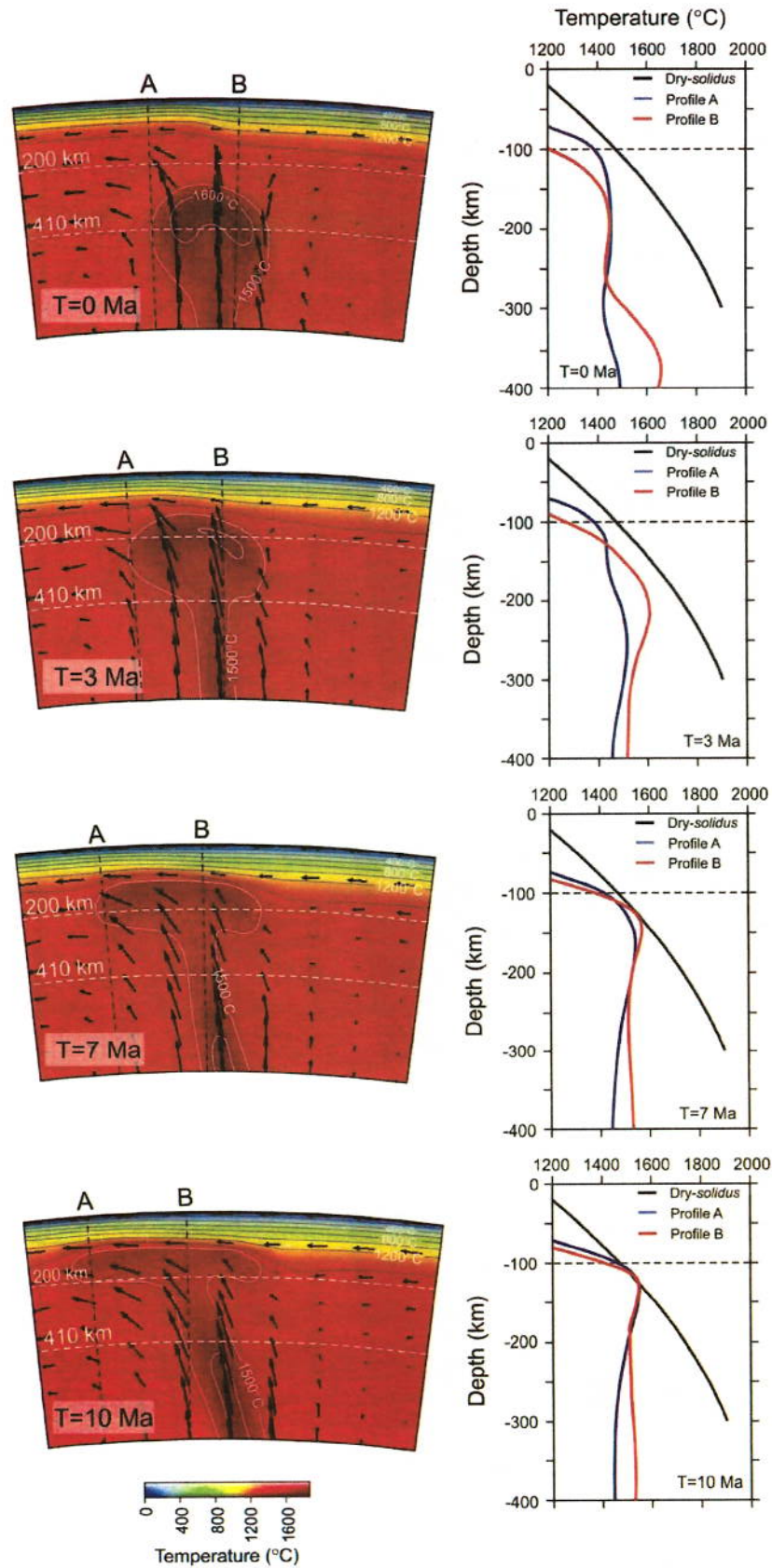


Fig. 10. Time-space evolution of plume-lithosphere impact for Model KK₂ (see Table 2). The craton is considered initially thin (100 km thickness) and has a low viscosity of 10^{22} Pa s. The plume head has an initial diameter of 300 km, an initial radial averaged temperature excess of 150 °C and a density deficit of 50 kg/m³. The plume is able to thin the craton sufficiently to initiate melting at depths of ~125 km depth.

cell beneath the craton created by the interaction between the plume and the weak lithosphere. However, models run with the thinnest cratonic lithosphere (100 km) combined with a small plume diameter (300 km), a temperature excess of only 150 °C, a density deficit of 30 kg/m³ and a weak lithosphere (initial viscosity of 10^{22} Pa s), show that melting is possible after ~8 Ma and at ~100 km depth (Fig. 6; Model K_1).

For thicker (200 km) lithosphere none of the models tested thinned the cratonic lid sufficiently to allow plume melting until viscosity was reduced by a factor of three or more (to $<3 \times 10^{22}$ Pa-s). Interestingly, in all three sets of models (G, H and K series, Table 2; cf. Fig. 7) where temperatures in the plume intersect the mantle solidus, lithospheric convective destabilization can be observed along the eastern lobe of the plume.

4.2. Kinematic upper boundary condition (upper plate moves wrt to plume)

We next present a more realistic set of models in which a plume rises beneath the suture zone, but with the upper plate moving with a velocity (2 cm/yr) representative of migration of North America with respect to a hotspot reference frame (DeMets et al., 1994; Gripp and Gordon, 2002). In the majority of these models (from AA through HH – Table 2) the thermo-chemical plume is unable to significantly erode the old lithosphere (cf. Model BB_2; Fig. 8). This is mainly a consequence of the high efficiency of the upper plate in dissipating thermal energy from the plume. Only if the parameter range is pushed to extreme limits (initial plume diameter of 500 km and 300 °C temperature excess, density deficit of 50 kg/m³, craton thickness less than 150 km and viscosity of 10^{22} Pa-s) can the plume thin the cratonic lid sufficiently to initiate melting of the upwelling mantle (Model GG_4, Fig. 9). With reduction of lithosphere thickness to 100 km, plumes with smaller diameter (300 km) and temperature excess (150 °C) are able to melt the base of the lithosphere at ~125 km depth (Model KK_2, Fig. 10). However, in the context of Yellowstone, these scenarios are unrealistic, since the plume is likely not that hot, nor is there evidence for significant melting below about 100 km (Leeman et al., 2009).

5. Discussion

5.1. General considerations

The work presented here addresses conceptually the effect of a plume impact on a thick overriding continental lithospheric plate, and how this might facilitate partial melting within the ascending plume material. We accept the premise that CRB flood volcanism is logically related to such an event, and some of our models show that it is possible for temperatures in the plume to reach the dry mantle solidus at realistic depths if the lithosphere is initially thin (≤ 100 km) as is likely the case for accreted oceanic terrains that make up the region outboard (west of) cratonic North America. In trying to model the region that underlies most of the SRPY province, we find that it is exceedingly difficult to reach melting conditions within the plume unless one assumes unrealistic model parameters. Our first scenarios, in which the lithosphere is essentially static with respect to the plume axis, produce maximal heating and deformation of the lower part of the lithosphere. Yet, the effects are relatively small unless a plume radius several times wider than the width of the SRPY hotspot track, excess temperature much greater than petrologic estimates or very thin cratonic lithosphere are accepted. If the surface manifestation of SRPY magmatism (swath width ca. 100 km) better approximates the dimensions of mantle upwelling (i.e., a plume–tail scenario is more appropriate), we predict that lithosphere thermal erosion will be inadequate to promote plume melting if the lithosphere thickness prior to plume impact resembled that beneath the undeformed Wyoming craton today (≥ 125 –150 km). More realistic models, in which the upper plate

migrates with respect to the plume axis, worsen the situation in that even greater thermal input is required to thin the lithosphere.

Our models suggest that the only realistic alternatives that might lead to plume melting are that: (a) viscosity of the upper plate is significantly lowered; (b) the lithosphere is thinned by some auxiliary process; or (c) it was already substantially thinned relative to the stable part of the craton prior to plume impact. The last of these is considered least likely because seismic tomography (Schutt et al., 2008) and gravity data (Artemieva, 2006) both support the interpretation that lithospheric thickness exceeds at least 125 km (and could be significantly greater) beneath portions of the Wyoming craton that are unaffected by plume impingement.

It is already well known that the mantle part of the lithosphere is strong and can support surface loads for long geological periods. The viscosity range implied for the continental mantle is in the range of 10^{22} – 10^{23} Pa s. These values are within the acceptable range for continental mantle because they confer the persistence of surface topographic features (i.e. orogenic belts) for several tens of m.y. or longer (Burov and Watts, 2006). A much lower viscosity (10^{19} – 10^{20} Pa s) would produce non-stable orogenic belts that collapse in less than 2 m.y.

It has been suggested (Tanton-Elkins, 2005, 2007) that regions of prolific magmatism could undergo significant viscosity reduction and/or parts of the lithospheric mantle or crust could become gravitationally unstable due to mafic magmatic inputs. Under favorable conditions, these factors could lead to Rayleigh–Taylor instabilities that might promote destabilization of lower parts of the lithosphere – in which case head-room could be created to allow upwelling and decompression melting of deeper mantle (asthenosphere, plume). Specifically, this scenario has been suggested as one means of triggering flood basalt magmatism in areas underlain by initially thick lithosphere on a time scale of several Myrs. Such a scenario could have contributed to destabilization of at least the westernmost edge of the cratonic margin and initiation of the earliest SRPY magmatism. However, there are several observations that might contradict this scenario. First, insofar as the isotopic compositions of Mesozoic and Cenozoic igneous rocks reflect the architecture of the underlying lithosphere, it is notable that the Sr isotopic boundary (the $^{87}\text{Sr}/^{86}\text{Sr}$ “0.706 line”; Armstrong et al., 1977) defined by Mesozoic granitoids coincides with that defined by Miocene and younger rhyolites and basalts (Leeman et al., 1992; Savov et al., 2009–this volume). Thus, there seems to be no significant lateral dispersion of the lithosphere related to or following mid-Miocene CRB flood volcanism. Second, as noted earlier, a pervasive lithospheric isotopic signature in SRPY volcanic rocks (whether primary source features or inherited by magma–lithosphere interaction) signifies that old lithospheric material remains and could not have been delaminated entirely. A third limitation may be in the timing of magmatism viz a viz the delamination model. Even under favorable conditions, the timescale for delamination is on the order of at least 5 Myr. As discussed above, the earliest SRPY mafic magmatism is assumed to predate the first outpourings of voluminous silicic magmas. In southwestern Idaho, the earliest silicic centers were established at about 16 Ma – essentially contemporaneous with or shortly following CRB magmatism. So, we feel that, as useful as delamination might be in jump-starting SRPY magmatism, we see no compelling reason to invoke such a model for this area.

A more plausible scenario is that SRPY magmatism is somehow related to the regional extensional tectonics associated with northward propagation of the Basin and range province into southern Oregon and Idaho since at least Miocene time. Following the model of Harry and Leeman (1995), parts of the lithosphere can be induced to melt shortly following onset of extension provided that these mantle domains contain fertile lithic components that are near their solidus conditions. However, if the isotopic compositions of SRPY magmas indeed reflect their sources, this fertility must be associated with an ancient (Archean) enrichment process. Of relevance to this paper, we note that contemporaneous extension and

magmatism are likely to weaken the lithosphere (i.e., lower its effective viscosity) in a dramatic fashion. Under these circumstances, impact of upwelling mantle material on lithosphere thickness could be somewhat greater than indicated by the models discussed in this paper. Models of a mantle plume impact combined with an extending lid represent a direction of modeling that we will consider in a future study.

5.2. Overview of model results

In the first parameter test set (free-slip upper boundary condition) the majority of models show insufficient craton thinning to allow melting of the mantle plume (Fig. 11). Even increasing the plume head diameter (from 300 km to 500 km), its initial temperature excess (from 150 °C to 300 °C) or density excess (from 30 kg/m³ to 50 kg/m³) to unrealistically high values fails to thin the cratonic lithosphere sufficiently to induce partial melting of ascending plume material. Some of the models prevent melting even beneath the thinner lithosphere to the west (Model A_1, B_1, C_1, D_1, E_1, F_1, F_2 and

G_1). These models all have a small plume head (300 km) and a modest initial temperature contrast (150 °C). This result might provide a lower limit for the plume initial state. Increasing the plume head diameter to 400 km or more (and temperature contrast to 250 °C) is needed before the young lithosphere is thinned enough to produce partial melting beneath it.

Of all the parameters considered, lithosphere viscosity clearly has the most significant influence on efficiency of thermal erosion and lithosphere thinning. Decreasing the lid viscosity one order of magnitude (from 10²³ Pa s to 10²² Pa s) allows the plume to penetrate deep into the craton and to intersect the dry *solidus* (Model F_3). The thinning continues even after the dispersal of the plume head. In this case, the plume tail (~100 km in diameter) is tilted significantly toward the cratonic lid due to the strong mantle flow created beneath (Fig. 5). Also, the thick lithosphere starts to delaminate and sinks into the mantle about 8 Ma after the plume head impinged the base of the craton. This local flow field might play a role in causing the southeastward tilt of the plume tail (Yuan and Dueker, 2005; Waite et al., 2006). Clearly, lithosphere rheology or heterogeneity can have a

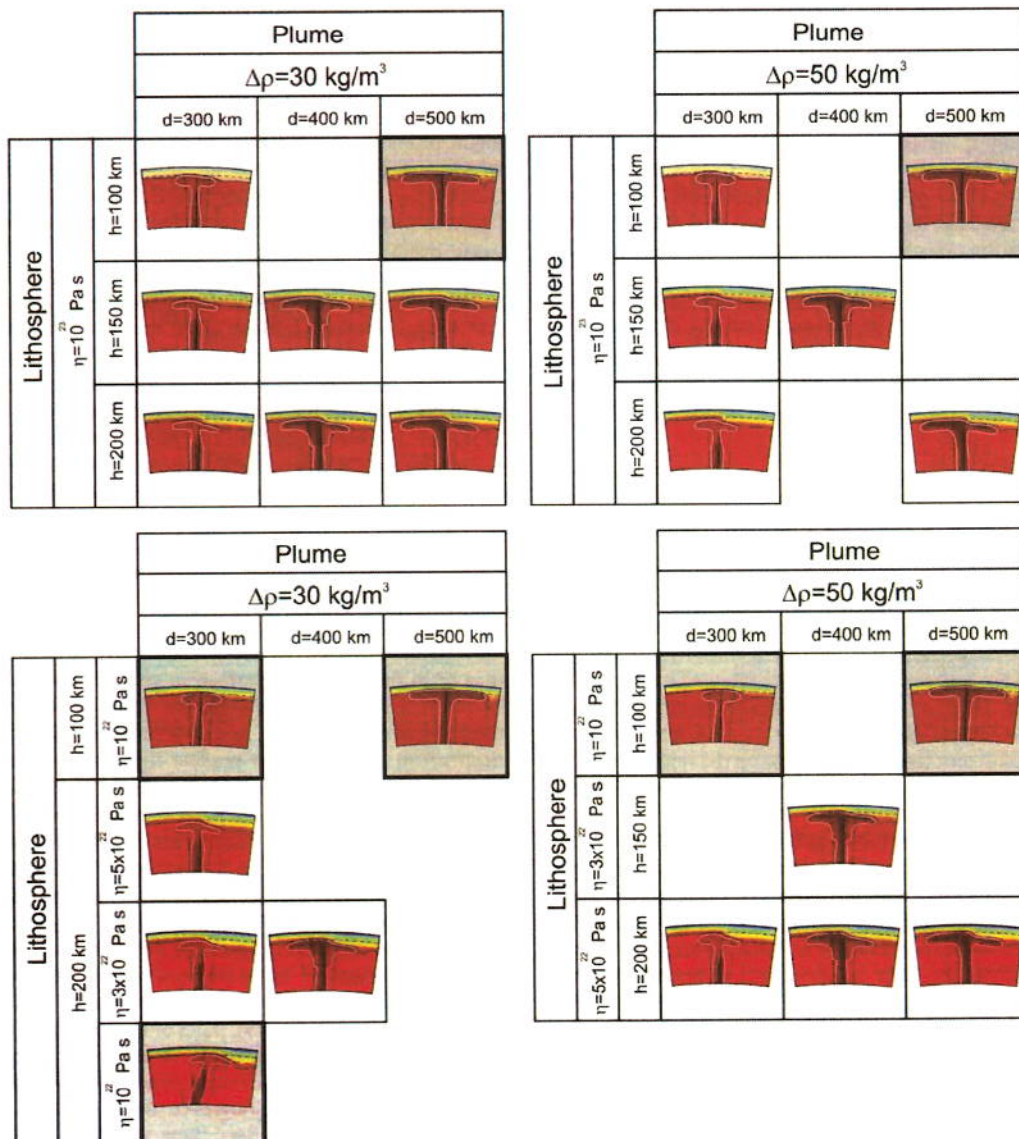


Fig. 11. A summary of the model runs with free-slip upper boundary (i.e., static) condition. All models have been integrated 10 Myr in time. The model which thins the craton sufficiently to produce melting at ~100 km depth is presented in gray background.

significant influence on the pattern of volcanism, albeit in this case the timing is quite slow with respect to apparent migration of SRPY magmatism. Another model (Model H_2) shows the following interesting pattern: an initial plume of 400 km diameter (with $\Delta T = 250^\circ\text{C}$, $\Delta\rho = 50\text{ kg/m}^3$ and $\eta_{\text{LITH}} = 3 \times 10^{22}\text{ Pa s}$) fails to erode the thick lid enough to allow decompression melting of the plume, but it allows the ascending hot plume tail to penetrate deeper and to intersect the solidus at $\sim 150\text{ km}$ after approximately 20 Ma. The last set of model runs correspond to relatively thin craton (100 km thick). For these models, a relatively small (300 km) and cold ($\Delta T = 150^\circ\text{C}$) initial plume might be able to melt beneath a weak lithosphere ($\eta_{\text{LITH}} = 10^{22}\text{ Pa s}$) in less than 8 Ma after the impact (Model K_1, Fig. 6). Although the parameters used in this model are surely unrealistic, it provides an interesting mechanism for the plume–tail interaction with thick lithospheric lids. A summary of the static models (free-slip boundary condition) is presented in Fig. 11. The models that succeed in triggering melting of plume mantle are shown in gray background.

Models with an imposed upper boundary condition that simulates absolute motion of the North American plate in a hotspot reference frame ($v = 2\text{ cm/yr}$) show that plume impingement on

the lithosphere is less efficient in modifying the upper plate than in the static models (Fig. 12). The plume is deflected westward by the upper plate movement and the heat is mostly transferred to the asthenospheric mantle rather than the lithosphere (see the flow distribution in Figs. 8 and 9). Even if lithosphere viscosity is decreased by an order of magnitude, the upper plate movement still has a predominant effect. Again, the solidus is reached at $\sim 150\text{ km}$ depth, but only by using unrealistic plume parameters ($d = 500\text{ km}$, $\Delta T = 300^\circ\text{C}$, $\Delta\rho = 30\text{ kg/m}^3$) the solidus is hit at 150 km depth. This result is inconsistent with melt formation depths (ca. 100 km; Leeman et al., 2008) based on combined tomographic and petrologic constraints. However, a combination of thin (100 km) and weak cratonic lithosphere ($\eta_{\text{LITH}} = 10^{22}\text{ Pa s}$) with a relatively small and cold plume, could induce melting at $\sim 125\text{ km}$ depth, but only $\sim 8\text{ Ma}$ after the plume impact (Model KK_2; Fig. 10).

General outcomes of the modeling exercises are summarized in Fig. 13 where effects on lithosphere thickness are illustrated as a function of time after plume impingement. The viscosity of the lithosphere plays an important role in lithospheric thinning by plume impact. For the same plume parameters, significant thinning

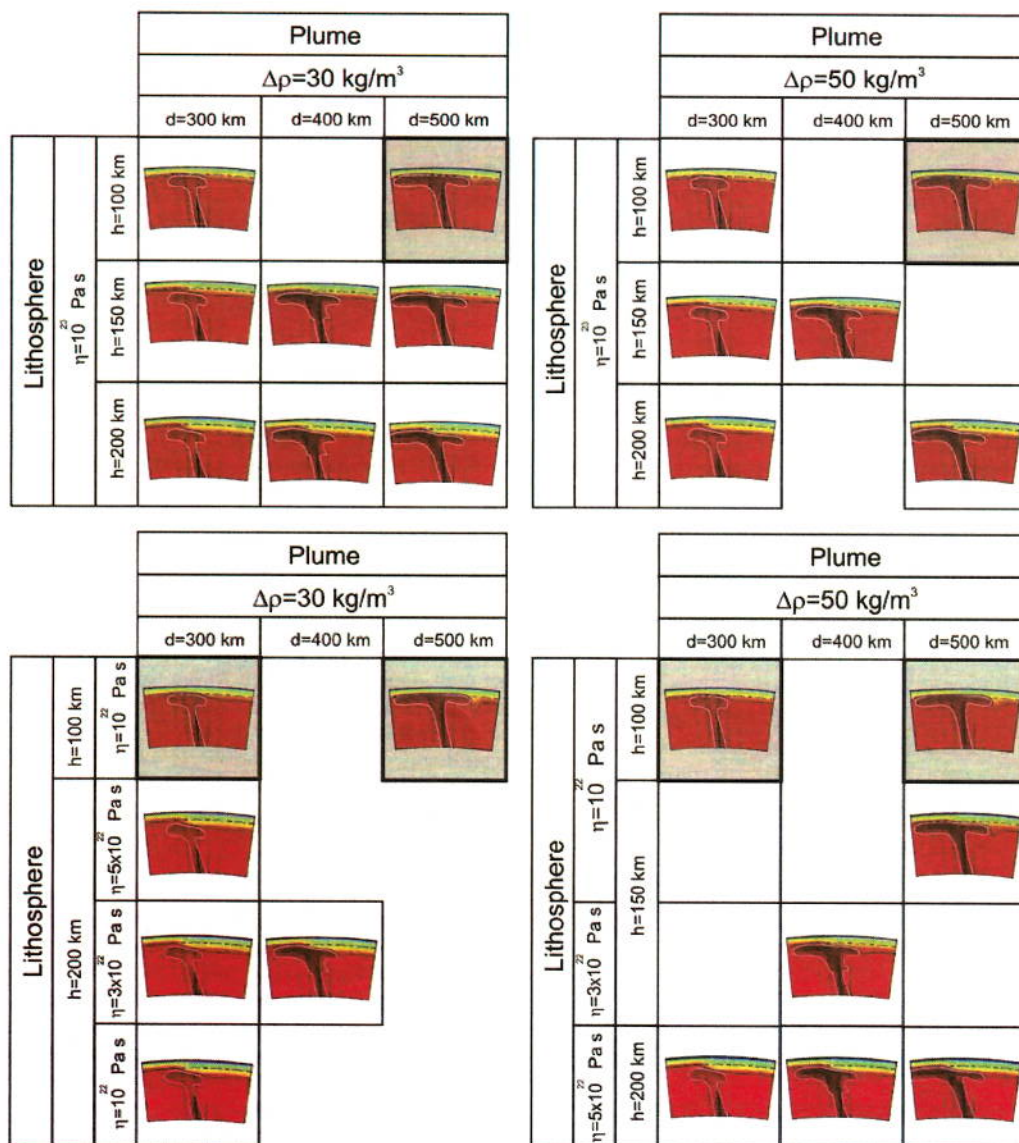


Fig. 12. A summary of the model runs with imposed velocity upper boundary condition. All models have been integrated 10 Myr in time. In these experiments none of the models thin the craton efficiently enough to produce melting at $\sim 100\text{ km}$ depth.

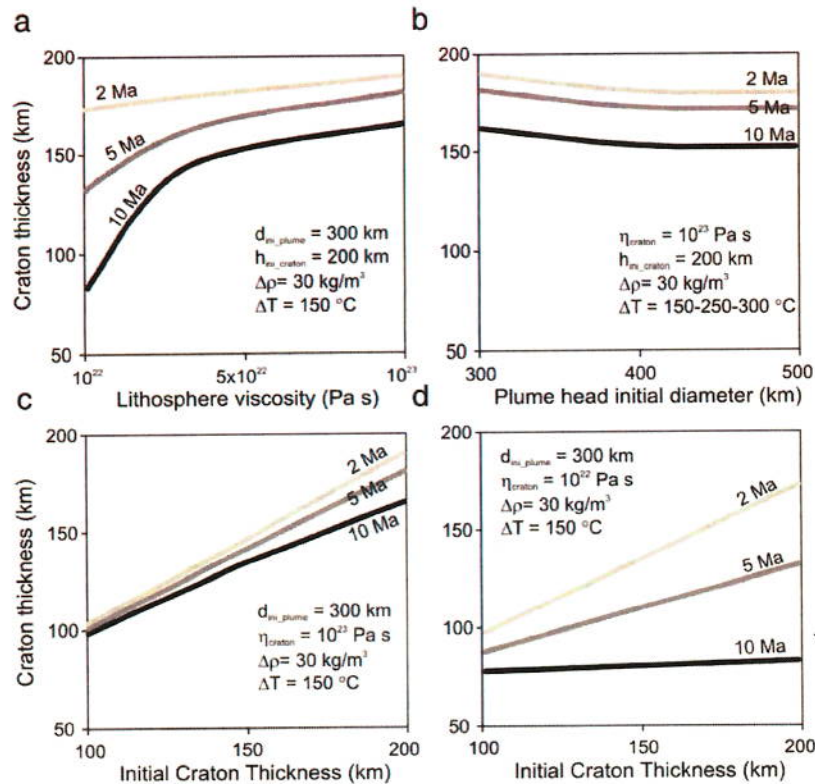


Fig. 13. Craton thickness variation in time as function of a) lithosphere viscosity, b) plume initial diameter and temperature excess, c) and d) initial cratonic thickness. All the data is for models with free slip top boundary condition (static mode) that result in maximum deformation of the lithosphere. Note that an ascending plume with potential temperature ca. 1450 °C is unlikely to melt unless lithosphere thickness is less than 100 km.

occurs for viscosities below 5×10^{22} Pa s (Fig. 13a). For such cases, the magnitude of thinning increases with assumed initial lithosphere thickness – presumably owing to the fact that absolute viscosity decreases with depth. However, we note that this effect depends on the absolute viscosity; it is insignificant for high viscosities (10^{23} Pa s; Fig. 13c), but greatly enhanced for models with low viscosity (10^{22} Pa s; Fig. 13d). Also, for highly viscous lithosphere, the initial plume diameter and temperature excess have less effect. Plume melting beneath either thick or thin litho-

sphere is possible only for models with relatively low lithosphere viscosity (Fig. 14). For these models the depth where dry solidus is crossed is in the range of 115–125 km, but the timing (in the range of 8–12 Ma after plume impingement) seems geologically excessive compared to timing of SRPY magmatism.

5.3. Implications for the SRPY plume track

Our results have several implications for the understanding of the Yellowstone plume–Wyoming craton system, and indeed continental plumes in general. For a typical high-viscosity craton of melt-depleted peridotite with a thermal boundary layer extending to ≥ 150 km, it is very difficult for plume melting to occur. This may explain the paucity of plume magmatic manifestations on continents. This is also consistent with SRPY basalt geochemistry, which shows little evidence for plume-source melting, despite the compelling seismic images of a mantle plume in the region (Yuan and Dueker, 2005). The modeling in this study helps clarify this seeming inconsistency.

Production of SRPY basaltic magmas predominantly from lithospheric sources may require that the Wyoming cratonic lithosphere is relatively fertile compared to typical cratons. There is xenolithic evidence (cf. Carlson and Irving, 1994) for multiple metasomatic “events” in the mantle beneath this region, dating back to at least Archean time in some places. It has also been suggested that parts of the lithospheric mantle beneath the interior western U.S. could have been re-hydrated during Laramide flat subduction (Humphreys et al., 2003). However, we consider this less probable because thermal models of flat slab subduction zones (e.g., Manea et al., 2004) suggest that the flat slab segment would be strongly devolatilized well before reaching distances of cratonic North America. While contributions from such post-Archean processes cannot be ruled out, they cannot

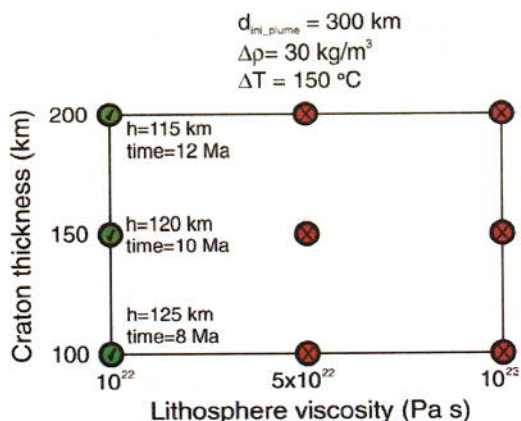


Fig. 14. A summary of parameters (lithospheric viscosity and initial craton thickness) for which lithospheric melting is possible (models with free slip top boundary condition). Red crossed dots denote the models unable to produce melting. Green checked dots show models which predict melting.

easily account for the strong Archean signature in primitive SRPY basalts (Doe et al., 1982; Hanan et al., 2008).

An alternative possibility stems from observation of high velocity lower crust under much of Wyoming. Several authors (Gorman et al., 2002; Schutt et al., 2008; Stachnik et al., 2008) suggest that this feature could represent a basaltic underplate from an ancient (Archean?) melting event. If Archean-aged basalt (now pyroxenite?) veins also exist in the mantle, lithospheric extension (Harry and Leeman, 1995), possibly aided by plume heating from below, could trigger remelting of such veins to produce the SRPY basalts having a lithospheric geochemical signature. Lithospheric delamination could also play a role in enhancing heating from below or in mixing lithospheric material into the underlying mantle [e.g., if cratons are denser than normally assumed (Schutt and Leshere, 2006) or if strength is reduced owing to magmatic processes or shear heating related to deformation (Hartz and Podladchikov, 2008)]. The consequences of such factors remain to be explored with future modeling in conjunction with seismology constraints suggesting that a high velocity lid extends to only ~60 km depth. A fundamental question concerns whether the low velocity material at greater depths is old lithosphere, asthenosphere, or a mixture of the two.

6. Conclusions

Simple models of the effects of a plume impingement on 100–200 km thick cratonic lithosphere shows that it is difficult to thin this lithosphere enough to induce plume melting unless a combination of a very thin (<125 km) and weak ($\eta_{\text{LITH}} = 10^{22}$ Pa s) craton is considered (Fig. 14). Combined with geochemical evidence that SRPY basalts carry radiogenic isotope signatures for involvement of ancient (i.e. lithospheric) mantle, a consistent explanation is that most of the basalts produced by the Yellowstone hotspot are due to melting of reduced solidus lithosphere, perhaps heated from below by a mantle plume.

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