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SUPPLEMENTARY INFORMATION

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A continuous 55-million-year record of transient mantle plume activity beneath Iceland

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Supplementary Table S1 | Constants and variables used in buoyancy flux calculations

Symbol	Description	Value	Unit
B	Buoyancy flux	–	Mg s^{-1}
T_p	Asthenospheric potential temperature	–	°C
t	Travel time, VSR 1	4.8	Myrs
ΔT	Average excess Iceland plume temperature[31,32]	150 ± 50	°C
r	Radial distance from Iceland, VSR 1	700	km
h	Vertical plume head thickness[33]	125 ± 25	km
ρ_m	Density of lithospheric mantle	3.2×10^3	kg m^{-3}
α	Thermal expansion coefficient[34]	3×10^{-5}	$°C^{-1}$
u	Plate half-spreading rate	12.5	mm yr^{-1}
κ	Thermal diffusivity[35]	0.8×10^{-6}	m^2 s^{-1}
g	Gravitational acceleration	9.8	m s^{-2}

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Supplementary Table S2 | Constants and variables used in thermal boundary layer analysis

Symbol	Description	Value	Unit
η	Viscosity		Pa s
E	Activation energy		kJ mol^{-1}
R	Gas constant	8.314	J K^{-1} mol^{-1}
T	Temperature		K
γ	Viscosity scaling factor		
δ_0, δ_1	Boundary layer thickness		km
d	Length scale of convecting layer		km
Ra_i	Rayleigh number of convecting interior		
α	Thermal expansion coefficient, 660 km[34]	2.5 x 10^{-5}	$^{\circ}$C^{-1}
	Thermal expansion coefficient, CMB[34]	1.0 x 10^{-5}	$^{\circ}$C^{-1}
ρ	Density, 660 km[36]	4.0 x 10^{3}	kg m^{-3}
	Density, CMB[36]	5.4 x 10^{3}	kg m^{-3}
κ	Thermal diffusivity, 660 km[35]	0.8 x 10^{-6}	m^{2} s^{-1}
	Thermal diffusivity, CMB[37]	2.0 x 10^{-6}	m^{2} s^{-1}
t	Time		Myrs

Thermal Boundary Scaling

The observed periodicity of VSRs can be used to investigate the likely origin of thermal anomalies within the convecting mantle. Episodic thermal anomalies are probably generated at a thermal boundary layer (e.g. 660-km phase transition, core-mantle boundary[38,39]). We can test if the observed periodicity is consistent with generation at one or other of these boundaries. We assume that mantle material has a Newtonian viscosity, η, where

$$\eta = \eta_0 \exp(E/RT) \tag{3}$$

where η_0 is the viscosity at $0°$C, E is the activation energy, R is the gas constant and T is temperature. We also define a dimensionless parameter, p, where

$$p = \gamma \Delta T. \tag{4}$$

γ is a constant scaling factor and ΔT is the temperature difference within the convecting layer[40]. It follows from Equation (3) that

$$\log \eta = \log \eta_0 + \frac{E}{RT} \tag{5}$$

and the change in viscosity with temperature can be written

$$\frac{-d \log \eta}{dT} = \frac{E}{RT^2} = \gamma. \tag{6}$$

Solomatov[40] showed that if convection initiates at a lower boundary layer with thickness δ_1, then

$$\delta_1 \sim \frac{\delta_0}{\gamma \Delta T} = \frac{1}{p}\delta_0 \tag{7}$$

where δ_0 is the thickness of the upper boundary layer. Convection above the lower boundary layer is reduced to constant viscosity convection with an effective driving temperature scale, given by

$$\Delta T_{rh} \sim \gamma^{-1} = p^{-1} \Delta T \tag{8}$$

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where ΔT_{rh} is the rheological temperature difference. Thus

$$\delta_0 \sim dp^{4/3} Ra_i^{-1/3} \tag{9}$$

where d is the depth of the convecting layer and Ra_i is the Rayleigh number of the convecting interior. Combining Equations (7) and (9), we obtain

$$\delta_1 \sim dp^{1/3} Ra_i^{-1/3} \tag{10}$$

where

$$Ra_i = \frac{\alpha g \rho \Delta T d^3}{\kappa \eta_i}. \tag{11}$$

α is thermal expansivity, g is the acceleration due to gravity, ρ is density, ΔT is the temperature above the thermal boundary layer, κ is thermal diffusivity and η_i is viscosity of the convecting interior. Combining Equations (10) and (11) we obtain

$$\delta_1 = \left(\frac{\kappa \eta_i p}{\alpha g \rho \Delta T}\right)^{1/3}. \tag{12}$$

Using Equation (4), we obtain

$$\delta_1 = \left(\frac{\kappa \eta_i \gamma}{\alpha g \rho}\right)^{1/3}. \tag{13}$$

At the onset of a thermal transient event, growth in thickness of the thermal boundary layer is given by

$$\delta(t) = \sqrt{(\pi \kappa t)} \tag{14}$$

where t is time. Combining Equations (13) and (14) we obtain

$$\eta_i = \frac{(\pi \kappa t)^{3/2} \alpha g \rho}{\kappa \gamma}. \tag{15}$$

Using Equation (6), we can describe viscosity as a function of T and E and then estimate γ for boundary layers located at 660 km depth and at the core-mantle boundary (Figure S1a). Equation (15) is used to test whether VSR periodicities of 4–8 Myrs are compatible with independent viscosity constraints (Figures S1b and S1c).

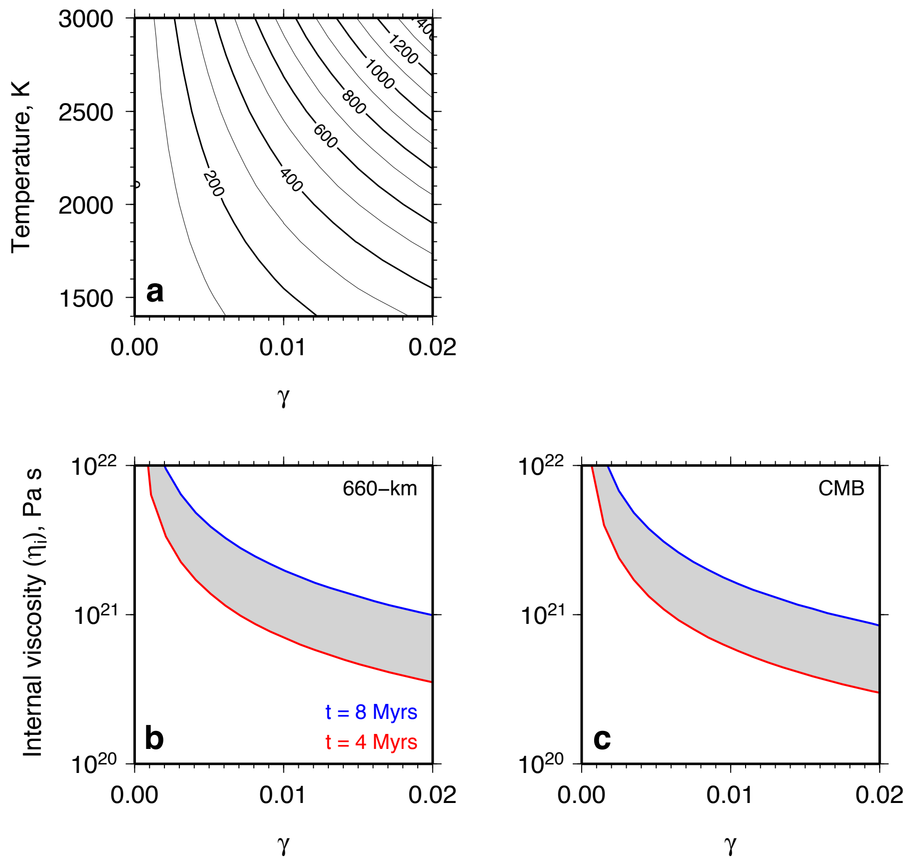

Figure S1: **Boundary layer stability analysis**. **a**, Relationship between temperature and viscosity scaling factor, γ, for range of activation energy values (labelled contours, in kJ mol^{-1}). **b**, Internal mantle viscosity, η_i, plotted as a function of γ using values appropriate for a thermal boundary layer at 660-km discontinuity and VSR periodicities of 4 Myrs (red line) and 8 Myrs (blue line). **c**, Same for core-mantle boundary (CMB).

For an average adiabat of 1600 K, the temperature at the 660-km discontinuity is $\sim$1850 K[41]. Values between 300–500 kJ mol^{-1} represent a sensible range of activation energy estimates at depths of 660 km[42,43]. Therefore γ is 0.011–0.018 (Figure S1a). For VSR periodicities of 4–8 Myrs, this range of γ values requires a viscosity of 10^{20}–10^{21} Pa s at the 660-km discontinuity (Figure S1b). Temperature estimates for the core-mantle boundary can be obtained by extrapolating transition-zone temperatures downward along mantle adiabats, which yields 2500–2800 K[39,44]. In the lower mantle, where deformation by diffusion creep is significant, activation energy is probably 250–350 kJ mol^{-1}. This range yields γ values of 0.004–0.007 (Figure S1a)[43]. In this case, we obtain a viscosity of about 10^{21} Pa s at the core-mantle boundary (Figure S1c). Both results are compatible with estimates of mantle viscosity from laboratory experiments and from glacio-isostatic rebound calculations[39,45,46].

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