

The Formation of Seaward-dipping Reflectors in
Volcanic Margins: Insights from High-resolution
Visco-elasto-plastic Geodynamic Models with
Extrusive Surface Processes

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Abstract

Seismic reflection data from volcanic margins reveal thick packages of seaward-dipping reflectors (SDRs) that are commonly interpreted as buried subaerial lava flows. This study presents the first systematic investigation of SDR formation using high-resolution visco-elasto-plastic geodynamic models that incorporate strain weakening and couple melt generation to a surface-processes model that simulates sediment transport and extrusive lava-flow emplacement using a cellular automata approach. The models show that ridgeward-dipping, downward-concave lava-flow packages, which become SDRs once a margin is flooded, naturally result from syn-kinematic subaerial basaltic lava-flow emplacement at an active spreading center above anomalously hot asthenosphere. Near-horizontal lava-flow packages are emplaced and bisected by extension and frictional-plastic deformation at the ridge axis into conjugate ridgeward-dipping packages with dips that increase in the down-dip direction. Lava-flow packages are subsequently modified by additional plastic deformation and rotation driven by progressive burial by younger flows, continued spreading, lithospheric density changes associated with thermal cooling and partial-melt solidification, and volcanic-load-induced ductile thinning of the gabbroic magmatic crust. The models identify several extensional mechanisms capable of producing SDR-like geometries, including normal faulting within volcanic edifices above the ridge axis, syn-kinematic emplacement of subaerial lava above listric normal faults that detach into ductile gabbroic crust, and syn-kinematic lava emplacement above triangular horsts formed within newly accreted gabbroic crust at the ridge axis. The dominant mechanism, together with the detailed geometry of lava-flow packages and associated magmatic crustal structures, is sensitive to the presence or absence of melt damage above partially molten asthenospheric domains, the magnitude of the asthenospheric thermal anomaly, the spreading rate, the gabbroic crustal rheology, and the distribution and timing of volcanic loads that drive ductile deformation of the magmatic crust. The results further demonstrate that reproducing the key geometric features of conceptual SDR models based on seismic reflection data requires a probabilistic melt-damage

weakening mechanism above zones of melt focusing. These features include symmetric, downward-concave lava-flow packages with relatively small dip variations in their up-dip sections, progressively increasing dips in the down-dip direction, and relatively low-relief gabbroic basement. This melt-damage mechanism approximates the rheological weakening effects of channelized melt networks and dike injection within the lithosphere. The models also show that volcanic-load-induced ductile deformation of the gabbroic crust can modify the geometry of dipping lava-flow packages and promote asymmetric accretion of gabbroic crust.

Keywords: Visco-elasto-plastic geodynamic modeling, Lithospheric extension, Magmatic margins, Seaward-dipping reflectors, Lava-flow modeling

1. Introduction

Seismic reflection data from extensional margins in the North and South Atlantic commonly image thick, downward-concave packages of high-amplitude dipping reflectors that diverge and steepen oceanward [1, 2, 3, 4, 5, 6, 7, 8, 9, 10, 11] (Figures D.1 and D.2). Similar reflector geometries have been documented along offshore margins of the Gulf of America, Central Atlantic Ocean, and western India [3, 4, 12]. These seaward-dipping reflectors (SDRs) are widely interpreted as buried subaerial lava-flow sequences emplaced on gabbroic magmatic crust derived from the intrusive fraction of mantle-derived melt [7, 8, 9, 11, 12]. Reconstructions of SDR packages across conjugate margins indicate that some SDRs may have formed from the subaerial emplacement of lava over an active spreading center, resulting in the symmetric flow of basaltic lava away from the ridge axis, and the separation of originally continuous lava-flow sequences into conjugate segments [6, 8, 9, 13] (Figure D.3a). The widespread occurrence of SDRs, together with crustal seismic velocities consistent with thick mafic crust, supports classification of many SDR-bearing margins as volcanic passive margins, where relatively thin crust consists largely, or entirely, of newly accreted intrusive and volcanic material rather than extended pre-existing continental crust [2, 4, 14, 11].

SDR packages are commonly associated with up-dip reflectors that have dips of less than 10° and dip variations of less than 5° over distances of 5–20 km (Figures D.1a–d and D.2a, c, d). In the down-dip portions of SDR packages, reflector curvature generally becomes more strongly downward concave, with dips increasing by as much as 10 – 20° over distances of 5–10 km (Figures D.1b, c, and D.2a, c, d). Maximum reflector dips in these down-dip portions are typically less than 35° (Figures D.1 and D.2). However, because the down-dip parts of SDR packages are commonly poorly imaged, larger maximum dips cannot be ruled out [9] (Figure D.1a, b, c).

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Although many published conceptual and numerical models depict SDRs as relatively simple reflector packages with progressively increasing downward concavity (Figure D.3), observed SDR geometries are commonly more complex. Examples include reflectors that become upward concave down dip (Figure D.1a, d), broad intervals of relatively flat reflectors (Figure D.1d), abrupt dip changes that produce kinked reflector geometries (Figures D.1c and D.2d), and dipping reflectors with low curvature and dip changes of less than 2° over distances of 10–20 km (Figure D.1d). Across volcanic margins, the maximum thickness of SDR packages ranges from approximately 5 to 17.5 km (Figures D.1 and D.2).

Although some SDRs are interpreted to be lava flows that erupted on extended continental crust [9, 13], the crust beneath many SDR packages is commonly interpreted as being composed primarily of gabbroic magmatic material [7, 9, 13, 11]. This interpretation is supported by seismic refraction data from the offshore margins of eastern North America and Namibia, which indicate seismic velocities consistent with mafic intrusive crust [3, 2, 4]. The thickness of interpreted gabbroic magmatic crust ranges from approximately 3–5 km to as much as 20 km (Figures D.1 and D.2). The relief of the magmatic basement surface may be less than 1 km, producing an apparently smooth basement geometry. In other cases, however, basement relief may reach 5–8 km, resulting in more rugose geometries (Figures D.1 and D.2). In many SDR-bearing margins, limited seismic imaging quality along the basement surface makes it difficult to robustly constrain basement geometry.

SDR packages offshore Brazil and Argentina were classified by [9] into two main categories. Type I SDRs are associated with tectonically controlled, proximal, fault-bounded wedges. These packages have relatively planar reflector geometries that downlap onto inclined basement interpreted to record fault-controlled relief [9] (Figure D.3a). The Type I SDRs analyzed by [9] have reflector lengths of 1–51 km, total lava-flow thicknesses of less than 10 km, and reflector dips ranging from 2° to 28° , with a mean dip of 14° [9].

Type II SDRs, in contrast, are interpreted to form on actively spreading magmatic crust and are characterized by diverging reflector packages that become increasingly downward concave in the down-dip direction [9] (Figure D.3a). These reflectors downlap onto subhorizontal, seismically opaque basement with low relief. The Type II SDRs described by [9] have reflector lengths of 1–91 km, total thicknesses of less than 15 km, and reflector dips ranging from 0° to 30° , with a mean dip of 8° [9]. Type II SDRs generally steepen down dip and may display upward convexity near the point where reflectors downlap onto subhorizontal basement [9] (Figures D.1 and D.2).

Despite their widespread nature, the origins of SDRs and associated basement structures remain debated with proposed mechanisms including (1) syn-kinematic extrusion of lava flows on extended continental crust [4], (2) progressive rotation of subaerial lava flows due to volcanic loading and spreading of magmatic crust [8], and (3) syn-kinematic emplacement of lava flows on mobile gabbroic basement [7, 15]. Assessing the plausibility of these mechanisms has implications for paleogeographic reconstructions and models of thermal evolution of the lithosphere and asthenosphere. Geodynamic models that couple

mantle melting with lava-flow emplacement can be used to test the plausibility of these mechanisms by constraining processes and structural evolution with conservation laws that use boundary and initial conditions consistent with geologic and geophysical observations.

Recent geodynamic modeling studies have made substantial progress toward the level of geological realism needed to compare forward-modeling results with observations from extensional margins. Key advances include: (1) increasing numerical resolution to resolve localized shear zones [16, 17]; (2) implementing polyphase boundary conditions consistent with plate reconstructions [16]; (3) coupling lithospheric deformation with erosion and sediment transport [18, 19]; (4) incorporating isostatically consistent sea level [20]; and (5) including mantle melting processes [21, 22, 23, 24]. However, modeling magmatic crust formation and lava-flow emplacement requires extending these approaches by developing methods that couple long-term geodynamic deformation, which occurs over millions of years, with extrusive surface processes that operate over timescales ranging from days to tens of thousands of years.

Previous models of SDR formation have been either conceptual [9] (Figure D.3a) or based on numerical calculations that use a half-spreading-center framework [25, 26] (Figure D.3b). These numerical models neither incorporate realistic visco-elasto-plastic rheology with strain weakening nor fully couple thermo-mechanical evolution, magmatic processes, and lava-flow emplacement.

Here, the formation of dipping subaerial lava flows and magmatic crust is investigated using high-resolution visco-elasto-plastic marker-in-cell geodynamic models that include strain weakening, melt generation, instantaneous melt transport and emplacement, erosion and sediment transport, and a new cellular-automata-based model for extrusive lava-flow emplacement (Figure D.4a). This lava-flow model accounts for subaerial versus submarine emplacement conditions, magma availability for extrusion, and the inter-eruption period associated with large subaerial volcanic eruptions. A further focus of this study is to determine how lithospheric rheological parameters influence the geometry of buried lava-flow packages and magmatic crust. The models of [26] explored the weakening effects of dike-fluid injection in visco-elasto-plastic models by reducing viscosity at the ridge axis. We build on this approach by introducing a probabilistic melt-damage weakening model that approximates the weakening effects of both fluid-filled dikes and channelized melt networks above zones of melt focusing in the partially molten mantle (Figure D.4a). This melt-damage weakening model is spatially and temporally coupled to active melting processes and exerts an important control on the evolution and geometry of buried lava-flow packages and magmatic crust.

2. Methods

The numerical models presented in this work were implemented using Earth-Box.jl (v0.1.6), a Julia package for multiphase visco-elasto-plastic marker-in-cell geodynamic modeling with melt generation, melt intrusion, melt extrusion, frictional-plastic melt damage, lava-flow modeling and marine and terrestrial

sediment transport with compaction [27, 28]. EarthBox.jl discretizes the Stokes-continuity and heat transport equations on a staggered grid using conservative finite differences and free-surface stabilization [29, 30]. Advective processes are modeled using a 4th-order Runge-Kutta scheme that includes both grid and subgrid changes in temperature and deviatoric stress [29].

The fundamental aspects of the marker-in-cell approach for solving incompressible flows with a free-surface used by EarthBox.jl were first published by [31]. This approach was later modified for visco-elasto-plastic problems by [32, 33, 29]. EarthBox.jl uses a visco-elasto-plastic thermo-mechanical algorithm that closely follows the algorithm described by [29] but with a node-based plasticity approach from [30] to improve the convergence of the non-linear visco-elasto-plastic Stokes-continuity solver [Appendix A](#).

The following sections provide a detailed description of methods required to generate the numerical models presented in this work including specific formulations of the governing equations, sediment transport modeling with compaction, melt generation, extraction and transport, lava-flow modeling and rheological models for viscous creep and frictional-plastic failure. The reader is referred to the EarthBox.jl documentation for additional details [27].

2.1. Governing Equations

2.1.1. Conservation of Momentum and Mass

Velocity and pressure are obtained by solving the conservation of momentum and mass equations for a slow, highly viscous incompressible visco-elasto-plastic fluid:

$$\frac{\partial \sigma'_{xx}}{\partial x} + \frac{\partial \sigma'_{xy}}{\partial y} - \frac{\partial P}{\partial x} = -\rho g_x \quad (1)$$

$$-\frac{\partial \sigma'_{xx}}{\partial y} + \frac{\partial \sigma'_{xy}}{\partial x} - \frac{\partial P}{\partial y} = -\rho g_y \quad (2)$$

$$f_{ve} = \frac{\eta_{vp}}{\mu \Delta t + \eta_{vp}} \quad (3)$$

$$\sigma'_{xx} = 2\eta_{vp}\epsilon'_{xx}(1 - f_{ve}) + \sigma'^{co}_{xx}f_{ve} \quad (4)$$

$$\sigma'_{xy} = 2\eta_{vp}\epsilon'_{xy}(1 - f_{ve}) + \sigma'^{co}_{xy}f_{ve} \quad (5)$$

$$\epsilon'_{xx} = \frac{\partial v_x}{\partial x}, \quad \epsilon'_{xy} = \frac{1}{2} \left(\frac{\partial v_x}{\partial y} + \frac{\partial v_y}{\partial x} \right) \quad (6)$$

$$\frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} = 0 \quad (7)$$

where σ'_{xx} is the deviatoric normal stress, σ'^{co}_{xx} is the corrected deviatoric normal stress from the previous time step, σ'_{xy} is the deviatoric shear stress, σ'^{co}_{xy}

is the corrected deviatoric shear stress from the previous time step, P is the pressure, ρ is the density, g_x and g_y are the gravitational accelerations, η_{vp} is the visco-plastic viscosity, μ is the shear modulus, Δt is the computational time step, $\dot{\epsilon}'_{xx}$ is the normal deviatoric strain rate, $\dot{\epsilon}'_{xy}$ is the deviatoric shear strain rate, f_{ve} is the visco-elastic factor and v_x and v_y are the velocity components. Equations (1) and (2) are the x- and y-components of the Stokes equations, respectively. It is noted that equation (2) utilizes the relationships $\sigma'_{yy} = -\sigma'_{xx}$ and $\sigma'_{yx} = \sigma'_{xy}$ to reduce computational storage requirements. The Boussinesq approximation is used to account for the effects of temperature and pressure on density in the gravitation terms in (1) and (2) while maintaining the assumption of incompressibility from (7):

$$\rho = \rho_{ref} (1 - \alpha(T - T_{ref})) (1 + \beta(P - P_{ref})) \quad (8)$$

where ρ_{ref} is the reference density, α is the thermal expansion coefficient, T_{ref} is the reference temperature (298.15 K), β is the compressibility coefficient, and P_{ref} is the reference pressure (10^5 Pa). Gravitational acceleration in the x-direction (g_x) is set equal to zero, and gravitational acceleration in the y-direction (g_y) is set equal to 9.8 m/s^2 . The visco-plastic viscosity term η_{vp} is given by:

$$\eta_{vp} = \begin{cases} \eta_{creep} & \text{for } \sigma_{II,elastic} \leq \sigma_{yield}, \text{ and} \\ \mu \Delta t \frac{\sigma_{yield}}{\sigma_{II,elastic} - \sigma_{yield}} & \text{for } \sigma_{II,elastic} > \sigma_{yield} \end{cases} \quad (9)$$

where η_{creep} is the viscosity associated with creep mechanisms, σ_{yield} is the yield stress and $\sigma_{II,elastic}$ is the second invariant for purely elastic stress buildup [30] as given by

$$\sigma_{II,elastic} = \frac{\mu \Delta t + \eta_{creep}}{\eta_{creep}} \sigma_{II} \quad (10)$$

where σ_{II} is the second invariant of the stress tensor given by

$$\sigma_{II} = \sqrt{(\sigma'_{xy})^2 + (\sigma'_{xx})^2}. \quad (11)$$

Visco-plastic viscosity η_{vp} is limited by a minimum value of $10^{18} \text{ Pa}\cdot\text{s}$ and a maximum value of $10^{24} \text{ Pa}\cdot\text{s}$. The yield stress σ_{yield} is given by the following equations:

$$\sigma_{yield} = \begin{cases} \sigma_c \cos \theta + \sin \theta (P - P_f) & \text{for } P \geq P_f \\ \sigma_c \cos(\theta) & \text{for } P < P_f \end{cases} \quad (12)$$

where σ_c is the cohesion, θ is the friction angle, and P_f is the fluid pressure. Yield stress is limited by a minimum value of 1 MPa.

The corrected deviatoric stresses from previous time steps, σ'^{co}_{xx} and σ'^{co}_{xy} , which appear in equations (4) and (5), take into account rotation using the following equations:

$$\sigma'_{xx}{}^{co} = \sigma'_{xx} (\cos(\omega\Delta t)^2 - \sin(\omega\Delta t)^2) - \sigma'_{xy} \sin(2\omega\Delta t) \quad (13)$$

$$\sigma'_{xy}{}^{co} = \sigma'_{xx} \sin(2\omega\Delta t) + \sigma'_{xy} \cos(2\omega\Delta t) \quad (14)$$

$$\omega = \frac{1}{2} \left(\frac{\partial v_y}{\partial x} - \frac{\partial v_x}{\partial y} \right) \quad (15)$$

where ω is the rotation rate or spin of the material [29].

A free surface is implemented in the model using a sticky-rock interface with stabilization to avoid the so-called drunken sailor effect. This is achieved using the approach described by [30] whereby advection-related density changes are incorporated into (2) to better approximate density at the end of the time step. The modified y-Stokes equation is given by:

$$-\frac{\partial \sigma'_{xx}}{\partial y} + \frac{\partial \sigma'_{xy}}{\partial x} - \frac{\partial P}{\partial y} - g_y \Delta t \left(v_x \frac{\partial \rho}{\partial x} + v_y \frac{\partial \rho}{\partial y} \right) = -\rho g_y \quad (16)$$

The algorithm described in [Appendix A](#) is used to update the deviatoric stress tensor components on markers and solve equations (1), (2), and (7) using the non-linear visco-elasto-plastic constitutive law described in equations (4), (5) and (9).

2.1.2. Conservation of Energy

Temperature is obtained by solving the conservation of energy equation as given by:

$$\rho C_p \frac{DT}{Dt} = \frac{\partial}{\partial x} \left(k \frac{\partial T}{\partial x} \right) + \frac{\partial}{\partial y} \left(k \frac{\partial T}{\partial y} \right) + H_{rad} + H_{shear} + H_{adi} + H_{melt} \quad (17)$$

where T is the temperature, C_p is the specific heat capacity, k is the thermal conductivity, H_{rad} is the radiogenic heat production term, H_{shear} is the shear heating term, H_{adi} is the adiabatic heating or cooling term, H_{melt} is a melt-processes term associated with the latent heat of melting and crystallization, and $\frac{DT}{Dt}$ is the substantive temperature derivative defined as:

$$\frac{DT}{Dt} = \left(\frac{\partial T}{\partial t} + v_x \frac{\partial T}{\partial x} + v_y \frac{\partial T}{\partial y} \right) \quad (18)$$

The thermal conductivity term k in (17) is described with the following temperature-dependent model:

$$k = 358(1.0227k_{20^\circ C} - 1.882)(1.0/T_K - 0.00068) + 1.84 \quad (19)$$

where $k_{20^\circ C}$ is the thermal conductivity of the rock at 20°C, T_K is temperature in Kelvin [34]. The heat capacity term C_p in (17) is described with the following

temperature-dependent model:

$$C_p = C_{p20} \left(0.953 + 2.29 \cdot 10^{-3} T_{\circ C} - 2.835 \cdot 10^{-6} (T_{\circ C})^2 + 1.191 \cdot 10^{-9} (T_{\circ C})^3 \right) \quad (20)$$

where C_{p20} is the heat capacity of the rock at 20°C and $T_{\circ C}$ is the temperature in °C [34].

The shear heating term H_{shear} is defined using the following equation:

$$H_{shear} = \frac{\sigma'_{xx}{}^2}{\eta_{vp}} + \frac{\sigma'_{xy}{}^2}{\eta_{vp}} \quad (21)$$

where σ'_{xx} and σ'_{xy} are the deviatoric components of the stress tensor and η_{vp} is the visco-plastic effective viscosity. The adiabatic heating term H_{adi} is defined using the following equation:

$$H_{adi} = c_{adi} \frac{DP}{Dt} \quad (22)$$

where $\frac{DP}{Dt}$ is the substantive pressure derivative and c_{adi} is the adiabatic coefficient given by:

$$c_{adi} = \alpha T \quad (23)$$

where α is the thermal expansion coefficient. The substantive pressure derivative $\frac{DP}{Dt}$ is defined as:

$$\frac{DP}{Dt} = \left(\frac{dP}{dx} v_x + \frac{dP}{dy} v_y \right). \quad (24)$$

The melt-processes term H_{melt} is divided into two components:

$$H_{melt} = H_{melt,adi} + H_{melt,T} \quad (25)$$

where $H_{melt,adi}$ is the melting-related adiabatic component of the latent heat of melting associated with changes in pressure within partially molten domains and $H_{melt,T}$ is the temperature-dependent component. The adiabatic component of the latent heat of melting and the temperature-dependent components are given by:

$$H_{melt,adi} = -\rho L \frac{\partial M}{\partial P} \frac{DP}{Dt} \quad (26)$$

$$H_{melt,T} = -\rho L \frac{\partial M}{\partial T} \frac{DT}{Dt} \quad (27)$$

where L is the latent heat of melting, M is the melt fraction, and $\frac{DT}{Dt}$ is the substantive temperature derivative. The latent heat of melting term L is set equal to 400,000 J/kg for gabbroic and ultramafic rocks [35].

Equation (27) is included in an effective heat capacity term in the energy equation (17) to account for the temperature-dependent latent heat of melting and crystallization. The effective heat capacity term is given by:

$$C_p^{eff} = C_p + L \frac{\partial M}{\partial T} \quad (28)$$

The partial derivative $\frac{\partial M}{\partial T}$ in (28) is calculated using a finite difference approximation as follows:

$$\frac{\partial M}{\partial T} = \frac{M_{T+\Delta T, P} - M_{T-\Delta T, P}}{2\Delta T} \quad (29)$$

where $\Delta T = 1K$ and $M_{T+\Delta T, P}$ and $M_{T-\Delta T, P}$ are the melt fractions at the current pressure P and temperatures $T + \Delta T$ and $T - \Delta T$, respectively.

Similarly, equation (26) is included in an effective adiabatic coefficient term in equation (22) to account for the adiabatic component of the latent heat of melting and crystallization. The effective adiabatic coefficient term is given by:

$$c_{adi}^{eff} = c_{adi} - \rho L \frac{\partial M}{\partial P} \quad (30)$$

The partial derivative $\frac{\partial M}{\partial P}$ in (30) is calculated using a finite difference approximation as follows:

$$\frac{\partial M}{\partial P} = \frac{M_{T, P+\Delta P} - M_{T, P-\Delta P}}{2\Delta P} \quad (31)$$

where $\Delta P = 1000Pa$ and $M_{T, P+\Delta P}$ and $M_{T, P-\Delta P}$ are the melt fractions at the current temperature T and pressures $P + \Delta P$ and $P - \Delta P$, respectively.

The substantive pressure derivative that appears in (22) and (26) is approximated using lithostatic pressure gradients [30] as described in the following equation:

$$\left(\frac{dP}{dx} v_x + \frac{dP}{dy} v_y \right) = \rho g_x v_x + \rho g_y v_y. \quad (32)$$

The algorithm described in Appendix B is used to solve equation (17) where the conductive component of equation (17) is obtained by solving the purely conductive formulation of equation (17) on Eulerian grid nodes and the advective component of equation (17) is solved by advecting markers using a 4th-order Runge-Kutta scheme and correcting for subgrid changes in temperature [29].

2.2. Topography Advection, Sediment Transport and Compaction

Sediment transport is modeled on an Eulerian topography grid $(x_{topo, i}, y_{topo, i})$ where i ranges from 1 to N_{topo} , N_{topo} is the number of topography nodes and $y_{topo, i}$ is the y-coordinate of the topography at the i -th node. The spacing of the topography grid in the x-direction is denoted by Δx_t , which remains constant throughout the model domain. Prior to applying the sediment transport model, topography grid nodes are advected in the current velocity field using a 4th-order Runge-Kutta scheme and then interpolated to the original topography grid to obtain new y-coordinates of topography that track the interface between rock and the sticky-air/water layer.

Sediment transport is modeled using a diffusion equation that assumes density does not differ between sediment and bedrock and that the cohesion of bedrock is negligible so as not to limit sediment supply at the surface. When

applied along the topography marker chain, the sediment transport equation takes the form of the following equation:

$$\frac{\partial y_{topo}}{\partial t} = \frac{\partial}{\partial x} \left(\kappa_s \frac{\partial y_{topo}}{\partial x} \right) + R_{pelagic} \quad (33)$$

where y_{topo} is the y-coordinate of topography in the 2D model domain, κ_s is the sediment transport diffusivity, and $R_{pelagic}$ is a sediment source term that accounts for pelagic sedimentation and sediment transport in and out of the plane of the model domain. The approach described by [18] is used to define κ_s so that it includes the effects of subaerial slope diffusion, subaerial fluvial concentrative diffusional transport associated with river systems within a given drainage basin and water-depth dependent submarine slope diffusion that accounts for wave and tidal effects at shallow water depths:

$$\kappa_{s,i} = \begin{cases} \kappa_{sm} \exp\left(-\frac{W}{\lambda_{sm}}\right) & \text{if } W_i > 0 \\ \kappa_{sa} + \kappa_{fluvial,i} & \text{otherwise} \end{cases} \quad (34)$$

where W_i is water depth in m , κ_{sm} is the submarine slope diffusivity in m^2/s , λ_{sm} is the submarine diffusivity water-depth decay term in m , κ_{sa} is the subaerial slope diffusivity in m^2/s and $\kappa_{fluvial}$ is the diffusivity associated with fluvial transport processes in m^2/s . The fluvial diffusivity $\kappa_{fluvial,i}$ is described by the following equation:

$$\kappa_{fluvial,i} = R_{precip} C_{sa} D_{stream,i} \quad (35)$$

where R_{precip} is the precipitation rate in m/s , C_{sa} is the subaerial transport coefficient, and $D_{stream,i}$ is the downstream distance in m between drainage divides to the left and right of node i . The downstream distance $D_{stream,i}$ essentially scales fluvial diffusivity with the size of the drainage basin. A 1D conservative finite difference method is used to solve the sediment transport equation (33) with a time step of Δt_{sed} . See Table D.1 for the sediment transport model parameters used in this work.

The sediment transport equation (33) does not take into account the effects of sediment compaction and assumes that sediment density is equal to bedrock density. This assumption is equivalent to assuming that the sediment is deposited in a state of maximum compaction and that the substrate upon which new sediment is deposited does not compact. A compaction correction is applied to the topography solution at each time step of the sediment transport solver to account for deposition of sediment in a less compacted state and the compaction of pre-existing sediment:

$$y_{topo,i} = y_{topo,trans,i} - \left[\left(H_{sed,i}^f + \Delta H_{sed,i}^f \right) - \left(H_{sed,i}^o + \Delta H_{sed,i}^o \right) \right] \quad (36)$$

where $y_{topo,trans,i}$ is the topography obtained by solving the sediment transport equation (33) at node i , $H_{sed,i}^o$ is the initial sediment thickness prior to the deposition of new sediment, $\Delta H_{sed,i}^o$ is the newly deposited sediment thickness

in the zero-porosity maximum compaction state, $\Delta H_{sed,i}^f$ is the newly deposited sediment thickness in a de-compacted state and $H_{sed,i}^f$ is the initial sediment thickness compacted to a depth of $\Delta H_{sed,i}^f$. The parameter $\Delta H_{sed,i}^o$ is related to the new and old topography solutions obtained from solving (33) as follows:

$$\Delta H_{sed,i}^o = \max \left[(y_{topo,i}^o - y_{topo,trans,i}), 0 \right]. \quad (37)$$

The newly deposited decompacted sediment thickness $\Delta H_{sed,i}^f$ is calculated by de-compacting $\Delta H_{sed,i}^o$ from an assumed depth below mudline of 12 km to 0 km using the conservation of mass and assuming that the relationship between porosity and depth in the sediment column can be described by the following equation:

$$\phi = \phi_o \exp \left(-\frac{y_{sm,max}}{\lambda_{comp}} \right) \quad (38)$$

where ϕ is the sediment porosity, ϕ_o is the initial sediment porosity at the mudline, $y_{sm,max}$ is the maximum depth below the seafloor encountered by a particle during burial and λ_{comp} is the compaction decay length [36]. A similar approach is used to obtain $H_{sed,i}^f$ whereby equation (38) is used to compact $H_{sed,i}^o$ to a depth below mudline of $\Delta H_{sed,i}^f$. With each time step of the sediment transport solver a correction is also applied to the location of sediment and sticky air/water markers whereby pre-existing sediment and sticky air/water markers are advected vertically using the compaction displacement field. Interactive tests and benchmarks of this approach can be accessed through the EarthBox.jl API [27].

The marker composition field is updated to account for erosion and sedimentation processes after the sediment transport solver and compaction correction have been applied to update the y-coordinate of topography $y_{topo,i}$. The updated topography is first interpolated at the x-coordinate x_m of marker m to determine the elevation of the topography at the marker $y_{topo,m}$. If the y-coordinate of the marker y_m is less than $y_{topo,m}$ and the marker is lithological, the marker is transformed to either a sticky-air or sticky-water marker depending on whether the marker is above or below sea level to account for erosion. If the marker is sticky-air or sticky-water and y_m is greater than $y_{topo,m}$, the marker is transformed to sediment.

For a given marker m , sediment density ρ_m , thermal conductivity k_m and heat capacity $C_{p,m}$ are updated to account for water-filled porosity as described in equation (38) using the following equations:

$$\begin{aligned} \rho_m &= (1 - \phi_m) \rho_{matrix,m} + \phi_m \rho_{water}, \\ k_m &= (1 - \phi_m) k_{matrix,m} + \phi_m k_{water} \\ C_{p,m} &= (1 - \phi_m) C_{p,matrix,m} + \phi_m C_{p,water} \end{aligned} \quad (39)$$

where $\rho_{matrix,m}$ is the density of the rock matrix, $k_{matrix,m}$ is the thermal conductivity of the rock matrix, $C_{p,matrix,m}$ is the heat capacity of the rock matrix, ρ_{water} is the density of water, k_{water} is the thermal conductivity of water and $C_{p,water}$ is the heat capacity of water.

2.3. Melt Generation, Transport and Emplacement

The models presented in this work use a linearized equilibrium melt fraction model based on the following equation:

$$M_m = \frac{T_m - T_{solidus,m}}{T_{liquidus,m} - T_{solidus,m}} \quad (40)$$

where m is the index of the marker, M_m is the melt fraction for the marker, T_m is the temperature of the marker, $T_{solidus,m}$ is the solidus temperature of the marker, and $T_{liquidus,m}$ is the liquidus temperature of the marker [22].

The solidus temperature of markers with ultramafic composition is defined for anhydrous peridotite by the following equation:

$$T_{solidus}^{mantle} = \begin{cases} -5.1(P_{GPa})^2 + 132.9P_{GPa} + 1085.7 & \text{for } P_{GPa} \leq 8 \text{ GPa} \\ T_{solidus,K03}^{8GPa} + \frac{(T_{solidus20} - T_{solidus,K03}^{8GPa})}{12} (P_{GPa} - 8) & \text{for } P_{GPa} > 8 \text{ GPa} \end{cases} \quad (41)$$

where P_{GPa} is the pressure in GPa, $T_{solidus,K03}^{8GPa}$ is the solidus temperature in K at 8 GPa according to the model of [37], and $T_{solidus20}$ is the solidus temperature in K at 20 GPa set equal to 2250 K . Anhydrous peridotite is used as water is quickly removed from the system during the early stages of melting. The liquidus temperature for ultramafic mantle markers is defined by the following equation for anhydrous peridotite:

$$T_{liquidus}^{mantle} = \begin{cases} -2(P_{GPa})^2 + 45.0P_{GPa} + 1780.0 & \text{for } P_{GPa} \leq 8 \text{ GPa} \\ T_{liquidus,K03}^{8GPa} + \frac{(T_{liquidus20} - T_{liquidus,K03}^{8GPa})}{12} (P_{GPa} - 8) & \text{for } P_{GPa} > 8 \text{ GPa} \end{cases} \quad (42)$$

where $T_{liquidus,K03}^{8GPa}$ is the liquidus temperature in K at 8 GPa according to the model of [37], and $T_{liquidus20}$ is the liquidus temperature in K at 20 GPa set equal to 2600 K .

This study employs two models to determine the solidus and liquidus temperatures for gabbro: (1) a standard gabbro model representing the composition of primary melt products derived from the mantle, and (2) a model for gabbroic materials that have experienced fractional crystallization during emplacement. The solidus and liquidus temperatures for markers that have normal gabbro composition are defined by the following equations from [29]:

$$\begin{aligned} T_{solidus}^{gabbro} &= 1327.0 + 91P_{GPa}, \\ T_{liquidus}^{gabbro} &= 1423 + 105P_{GPa} \end{aligned} \quad (43)$$

The solidus and liquidus temperatures for fractionated gabbro are defined by the following equations:

$$\begin{aligned} T_{solidus}^{frac} &= T_{solidus}^{gabbro} + \Delta T_{solidus}^{frac} \\ T_{liquidus}^{frac} &= T_{liquidus}^{gabbro} + \Delta T_{liquidus}^{frac} \end{aligned} \quad (44)$$

where $\Delta T_{solidus}^{frac}$ and $\Delta T_{liquidus}^{frac}$ are the temperature differences between the solidus and liquidus temperatures for normal gabbro and fractionated gabbro. We use the following values for fractionated shifts in solidus and liquidus temperatures based on the work of [38]: $\Delta T_{solidus}^{frac} = 100 \text{ K}$ and $\Delta T_{liquidus}^{frac} = 300 \text{ K}$.

Melt is assumed to migrate instantaneously to the top of the partially molten mantle domain and then flow horizontally toward local high points (Figure D.4a). This assumption is justified by laboratory measurements of permeability and melt flow rates in olivine aggregates that have been used to infer melt transport times across the entire asthenosphere of around 1000 years [39]. At the local maximum of the partially molten domain melt is divided into two fractions: one that is immediately emplaced at the base of the crust via dikes and channelized networks and another that is transported to the surface (Figure D.4a). This transport and emplacement model is similar to the model described by [40, 41] where melt from focusing zones in the partially molten mantle is transported to the crustal domain at the spreading axis via a melt extraction zone composed of dike swarms. This simplified model of melt transport and emplacement is implemented independently within each melt drainage basin bounded by local minima along the top of the partially molten domain.

The extractable melt fraction for a marker m is defined by the following equation:

$$M_{extractable,m} = M_m - M_{extracted,m}^o \quad (45)$$

where M_m is the total melt fraction for the marker as described in (40) and $M_{extracted,m}^o$ is the melt fraction already extracted. The extracted melt fraction is updated using:

$$M_{extracted,m} = M_{extracted,m}^o + M_{extractable,m} \text{ if } M_m > 0 \text{ and } M_{extractable,m} > 0 \quad (46)$$

Marker density ρ_m is impacted by partial melting by depleting the rock matrix and through the presence of melt in the pore space. The impact of depletion is modeled using the approach of [20] as described in the following equation:

$$\rho_m = \begin{cases} \rho_m^o - 3.8M_{extracted,m}100 & \text{if } M_{extracted,m} < 0.075 \\ \rho_m^o - (24.84 + 0.488M_{extracted,m}100) & \text{if } M_{extracted,m} \geq 0.075 \end{cases} \quad (47)$$

where ρ_m^o is the density of the marker prior to the update step. The effect of melt in pore space on density is modeled using the following equation:

$$\begin{aligned} \rho_{melt} &= \rho_{ref,melt} + f_{eos}P_{Pa,m}, \\ \rho_m &= \rho_m^o(1 - M_{extractable,m}) + \rho_{melt}M_{extractable,m} \end{aligned} \quad (48)$$

where ρ_m^o is the density of the marker prior to this update step, $\rho_{ref,melt}$ is the reference density of the melt set equal to $2750 \frac{kg}{m^3}$, $P_{Pa,m}$ is the pressure in Pa and f_{eos} is the linearized derivative of melt density with respect to pressure equal to $8.97 \cdot 10^{-8} \frac{kg}{m^3 Pa}$ derived from the Birch-Murnaghan equation-of-state for anhydrous basaltic melt using $K_{to} = 20.8$ and $K_{tp} = 4.6$ [42, 43].

The total number of magma markers in drainage basin k if all extractable melt were consolidated into a single body is calculated as:

$$N_{magma,tot,k} = f_{extraction} \left(\sum_{m=0}^{N_{marker}} M_{extractable,m} + R_{melt,k}^o \right) \quad (49)$$

subject to the conditions:

$$M_m > 0, \quad M_{extractable,m} > 0 \quad \text{and} \quad x_{divide,k} < x_m < x_{divide(k+1)} \quad (50)$$

where $f_{extraction}$ is the extraction efficiency, $R_{melt,k}^o$ is the residual melt from the previous time step, $x_{divide,k}$ and $x_{divide(k+1)}$ are the x-coordinates of the left and right drainage divides for drainage basin k , respectively, and x_m is the x-coordinate of the marker. For the experiments presented in this work $f_{extraction} = 0.99$ to account for a small amount of melt retained in the mantle during transport.

The number of magma markers available for emplacement or extrusion in drainage basin k for a given time step is:

$$N_{magma,k} = \text{int}(N_{magma,tot,k}) \quad (51)$$

and the residual component used in the next time step is updated as:

$$R_{melt,k} = N_{magma,tot,k} - N_{magma,k}. \quad (52)$$

The total extractable magma volume in the drainage basin k is then:

$$V_{magma,k} = N_{magma,k} \Delta x_m \Delta y_m \quad (53)$$

where Δx_m and Δy_m are the average widths of the markers in the x and y directions, respectively.

The height of a new column of crust formed if $V_{magma,k}$ is emplaced as a dike within a spreading plate is:

$$H_{mc,k} = \frac{V_{magma,k}}{v_{ext} \Delta t} \quad (54)$$

where v_{ext} is the full extension velocity and Δt is the model time step. This characteristic magmatic crust height, $H_{mc,k}$, is used as a proxy for the efficiency of melt transport to the surface.

The number of volcanic markers that are available for extrusion on the surface as lava flows above drainage basin k is given by:

$$N_{volcanic,k} = \text{int}(f_v N_{magma,k}) \quad (55)$$

where f_v is the extrusion efficiency factor. We assume that the efficiency of melt transport to the surface correlates with the characteristic magmatic crust

height $H_{mc,k}$ as described by the following equation:

$$f_v = \begin{cases} f_{v,min} & \text{for } H_{mc,k} \leq H_{mc,min} \text{ and} \\ f_{v,max} & \text{for } H_{mc,k} \geq H_{mc,max} \text{ and} \\ f_{v,min} + \frac{(H_{mc,k} - H_{mc,min})}{\Delta H_{mc}} \Delta f_v & \text{for } H_{mc,min} < H_{mc,k} < H_{mc,max} \end{cases} \quad (56)$$

where Δf_v is equal to $f_{v,max} - f_{v,min}$ and ΔH_{mc} is equal to $H_{mc,max} - H_{mc,min}$. For this work, $f_{v,min} = 0.06$, $H_{mc,min} = 6,000$ m and $H_{mc,max} = 7,500$ m. Alternative scenarios for parameter $f_{v,max}$ are explored ranging from 0.3 to 0.6, which are consistent with the range of ratios of volcanics to intrusives interpreted from seismic reflection data (Figures D.1 and D.2). The total volume of volcanic markers $V_{vol,k}$ extruded on the surface for a given time step is:

$$V_{vol,k} = V_{vol,k}^o + N_{volcanic,k} \Delta x_m \Delta y_m. \quad (57)$$

where $V_{vol,k}^o$ is the volume of material available for extrusion from previous model time steps within the current inter-eruption period. Inter-eruption periods ranging from 50,000 to 400,000 years are explored for the experiments presented in this work, which encompass the range of observed periods of major pulses of subaerial basaltic lava from the Deccan Traps [44]. The number of magma markers $N_{magma,k}$ can be updated to account for volcanic material as follows:

$$N_{magma,k} = N_{magma,k} - N_{volcanic,k}. \quad (58)$$

Therefore, the volume of magma markers that are emplaced at the base of the crust is:

$$V_{magma,k} = N_{magma,k} \Delta x_m \Delta y_m. \quad (59)$$

Similar to [24], the models presented in this work approximate dike injection by replacing shallow ultramafic material with a volume of gabbroic magma equal to the current volume of magma extracted from the mantle. The approach of [24] defines a single dike for each injection event with a width controlled by extensional velocity and a height controlled by the volume of extracted gabbroic magma. Given the complex nature of dike injection and the difficulty associated with accurately predicting the location, number and size of dikes associated with each injection event, a simpler approach is used in the current work where shallow ultramafic markers within an injection zone are probabilistically selected and replaced with an equivalent volume of gabbroic magma until all extracted magma is emplaced (Figure D.4a).

Gabbroic magma is injected at the base of the Moho within an injection zone of width $W_{inj,k}$ that is centered on the local maximum of the partially molten domain of the mantle (Figure D.4a). The width of the magma injection zone $W_{inj,k}$ is calculated using the following equation:

$$W_{inj,k} = \begin{cases} \frac{V_{magma,k}}{H_{inj,limit}} & \text{for } H_{inj} > H_{inj,limit} \\ W_{inj}^{char} & \text{for } H_{inj} \leq H_{inj,limit} \end{cases} \quad (60)$$

where W_{inj}^{char} is the characteristic injection width, $H_{inj,limit}$ is the maximum injection height limit, and H_{inj} is the injection height defined by:

$$H_{inj} = \frac{V_{magma,k}}{W_{inj}^{char}}. \quad (61)$$

The models presented in this work used $W_{inj}^{char} = 10$ km and $H_{inj,limit} = 1$ km. Equation (60) is used to spread out dike injection and ensure that integrated injection height during an emplacement event does not become geologically unrealistic during the emplacement of large volumes of gabbroic magma when mantle potential temperature is high.

The injection zone is divided into subdomains that are selected using a normal probability distribution centered on the local maximum of the partially molten domain of the mantle (Figure D.4a). The injection of gabbroic magma is implemented by replacing the shallowest ultramafic marker below the Moho within a selected subdomain with a marker that has the composition of gabbroic magma and an emplacement temperature of 1200°C. At the beginning of each time step, gabbroic magma markers that are within 2 km of the Moho are converted into fractionated gabbroic magma markers. This conversion accounts for the fractional crystallization that occurs during the emplacement of magma in the crust, including processes such as dike and sill injection and the flow of crystal mush in the so-called gabbro glacier [38].

Markers undergo solidification based on their updated melt fraction M_m . The solidification process occurs as follows:

1. Solidification is modeled by transforming purely molten markers that have cooled below the liquidus temperature to a solid state.
2. Solid markers with temperatures between the solidus and liquidus temperatures are transformed to a partially molten state.
3. Partially molten markers that have cooled below the solidus are transformed to a solid state.

If the melt fraction M_m from (40) falls below the extracted melt fraction $M_{extracted,m}$ from (45) (i.e., $M_{extractable,m} < 0$) the marker is transformed into a refractory state. See Table D.2 for the melt extraction and emplacement parameters used in this work.

Figure D.4b shows a comparison of the total thickness of oceanic crust produced by this melting and extraction model for different full spreading rates against global observations from [45] and previous modeling results from [46]. Magmatic crustal thicknesses were calculated from models that use the model setup, boundary conditions and rheology described in this work with a thermal anomaly ΔT set to zero. Magmatic crustal thickness is calculated using equation (54) after steady-state magma production is reached.

2.4. Melt Extrusion and Lava Flow

Volcanic extrusion and lava-flow emplacement for a given eruption time step are modeled using a cellular automata approach [47, 48, 49] where the volume of magma available for extrusion in a melt drainage basin is divided into N_{flow} flow events that extrude from a probabilistic eruption location x_e . Flow events are temporally separated by summing the volume of all melt extracted during model time steps within an inter-eruption period Δt_{erupt} . The inter-eruption period Δt_{erupt} of subaerial flows in volcanic margins is not well constrained due to the lack of well penetrations. However, U-Pb geochronology from the Deccan Traps suggests that major pulses of subaerial basaltic lava are separated by intervals ranging from 100,000 to 300,000 years [44].

At the beginning of each flow event, the current topography grid (x_{topo}, y_{topo}) is decimated to a uniform grid (x'_{topo}, y'_{topo}) with spacing $\Delta x'_{topo}$ to improve the computational performance of the cellular automata calculation. Then a decimated flow thickness grid for the flow event, (x'_{topo}, H'_{flow}) , is initialized to zero thickness. Each flow is then divided into N_{pulse} pulses that laterally flow from the eruption location x_e to adjacent cells on the decimated topography grid.

The volume of volcanic material per flow V_{flow} is calculated using the following equation:

$$V_{flow} = \frac{V_{vol,k}}{N_{flow}} \quad (62)$$

where N_{flow} is given by:

$$N_{flow} = \begin{cases} \text{int} \left(\frac{V_{vol,k}}{V_{flow}^{char}} \right) & \text{for } V_{vol,k} > V_{flow}^{char} \\ 1 & \text{otherwise} \end{cases} \quad (63)$$

where $V_{vol,k}$ is the total volume of volcanic material available for extrusion from the drainage basin k calculated during the melt extraction steps that occurred within the current inter-eruption period Δt_{erupt} , and V_{flow}^{char} is the characteristic volume of volcanic material per flow given by:

$$V_{flow}^{char} = \begin{cases} L_{sa} H_{res,sa} & \text{for } y_{e,fc} \leq y_{sl} \text{ and} \\ L_{sm} H_{res,sm} & \text{for } y_{e,fc} > y_{sl} \end{cases} \quad (64)$$

where L_{sa} is the characteristic subaerial flow length, $H_{res,sa}$ is the subaerial residual flow thickness, L_{sm} is the characteristic submarine flow length, and $H_{res,sm}$ is the submarine residual flow thickness, $y_{e,fc}$ is the forecasted y-coordinate of the eruption, and y_{sl} is the y-coordinate of sea level. For the experiments presented in this work lava flow occurs in a subaerial environment by shifting base level into the lithosphere.

The forecasted y-coordinate of the eruption $y_{e,fc}$ in equation (64) is determined by interpolating the y-coordinate from the decimated topography grid (x'_{topo}, y'_{topo}) at the forecasted eruption location $x_{e,fc}$ which is calculated using the following equation:

$$x_{e,fc} = x_{e,min} + \frac{W_e}{2} \quad (65)$$

where $x_{e,min}$ is the minimum x-coordinate of the eruption location given by:

$$x_{e,min} = x'_{shallow,pm} - \frac{W_e}{2} \quad (66)$$

where $x'_{shallow,pm}$ is the average x-coordinate of the shallowest partially molten mantle marker within drainage basin k calculated during melt extraction, and W_e is the width of the eruption zone set equal to 2.5 km.

Each flow is erupted on the surface at eruption location x_e calculated as a normally distributed random variable as described by the following equation:

$$x_e \sim \mathcal{N}(x_{e,min} + 0.5W_e, (0.25W_e)^2) \quad (67)$$

subject to the constraint:

$$x_{e,min} \leq x_e \leq x_{e,min} + W_e. \quad (68)$$

If x_e falls outside this range, it is recalculated.

Each flow is erupted in a series of pulses where the number of pulses is:

$$N_{pulse} = \max \left[\text{int} \left(\frac{V_{flow}}{(\Delta x'_{topo} H_{res})} \right), 1 \right] \quad (69)$$

where $\Delta x'_{topo}$ is the grid spacing of a decimated topography grid used to perform cellular automata calculations, and H_{res} is the residual flow thickness given by:

$$H_{res} = \begin{cases} H_{res,sa} & \text{for } y_e \leq y_{sl} \text{ and} \\ H_{res,sm} & \text{for } y_e > y_{sl} \end{cases} \quad (70)$$

The initial thickness of each pulse is calculated using the following equation:

$$H_{pulse} = \frac{V_{flow}}{N_{pulse} \Delta x'_{topo}} \quad (71)$$

A critical step in the cellular automata calculation is to sort the indices of the decimated topography grid based on the distance from the eruption location x_e with grid index given by:

$$i_e = \text{int} \left(\frac{x_e}{\Delta x'_{topo}} \right). \quad (72)$$

For a given pulse of volcanic material, the cellular automata calculation is initialized as:

$$H'_{flow(i_e)} = H'^o_{flow(i_e)} + H_{pulse}, \quad (73)$$

where $H'^o_{flow(i_e)}$ is the thickness of the flow at index i_e prior to the current pulse. For a given lava pulse, flow thickness H'_{flow} is updated for each cell by radiating outward from index i_e . Basaltic lava material flows from cells with higher relative elevation to adjacent cells with the lower relative elevation until

the integrated thickness of the lava flow H'_{flow} equals the residual thickness H_{res} after which flow ceases to occur. During the iterative update procedure for a given pulse of the lava, the y-coordinates of topography, $y''_{topo(i)}$, $y''_{topo(i-1)}$, and $y''_{topo(i+1)}$ are related to current flow thickness $H'_{flow(i)}$ as follows:

$$\begin{aligned} y''_{topo(i)} &= y_{topo(i)}' - H'_{flow(i)} \\ y''_{topo(i-1)} &= y_{topo(i-1)}' - H'_{flow(i-1)} \\ y''_{topo(i+1)} &= y_{topo(i+1)}' - H'_{flow(i+1)} \end{aligned} \quad (74)$$

and elevation differences that drive flow from cell i to adjacent cells are given by:

$$\begin{aligned} \Delta E'_{left(i)} &= \max(0, y''_{topo(i-1)} - y''_{topo(i)}) \\ \Delta E'_{right(i)} &= \max(0, y''_{topo(i+1)} - y''_{topo(i)}). \end{aligned} \quad (75)$$

where positive values of $\Delta E'_{left(i)}$ and $\Delta E'_{right(i)}$ indicate that flow will occur from cell i to adjacent cells $i-1$ and $i+1$, respectively, if current flow thickness $H'_{flow(i)}$ is greater than the residual thickness H_{res} . The total thickness of lava available to flow to adjacent cells is given by:

$$\Delta H'_{out,total(i)} = \frac{2}{3} \min \left[\max \left(0, H'_{flow(i)} - H_{res} \right), \Delta E'_{limit(i)} \right] \quad (76)$$

where $\Delta H'_{out,total(i)}$ is the total thickness of lava available to flow from cell i to adjacent cells, and $\Delta E'_{limit(i)}$ is given by:

$$\Delta E'_{limit(i)} = \begin{cases} \left(\Delta E'_{left(i)} + \Delta E'_{right(i)} \right) / 2 & \text{if } \Delta E'_{left(i)} > 0 \text{ and } \Delta E'_{right(i)} > 0 \\ \Delta E'_{right(i)} & \text{else if } \Delta E'_{right(i)} > 0 \\ \Delta E'_{left(i)} & \text{else } \Delta E'_{left(i)} > 0. \end{cases} \quad (77)$$

Equation (76) facilitates convergence of the cellular automata iterations by limiting the amount of inter-cellular flow and ensuring that flow does not occur if flow thickness is less than or equal to the residual flow thickness H_{res} .

The total outflow thickness $\Delta H'_{out,total(i)}$ is partitioned to adjacent cells using the following equations:

$$\begin{aligned} \Delta H'_{out,left(i)} &= \min \left(\Delta H'_{out,total(i)} \frac{\Delta E'_{left(i)}}{\left(|\Delta E'_{left(i)}| + |\Delta E'_{right(i)}| \right)}, \Delta E'_{left(i)} \right) \\ \Delta H'_{out,right(i)} &= \min \left(\Delta H'_{out,total(i)} \frac{\Delta E'_{right(i)}}{\left(|\Delta E'_{left(i)}| + |\Delta E'_{right(i)}| \right)}, \Delta E'_{right(i)} \right). \end{aligned} \quad (78)$$

where outflow thickness is limited by elevation differences between cells. The thickness of the flow is updated for cell i and adjacent cells using the following equations:

$$\begin{aligned} H'_{flow(i-1)} &= H'_{flow(i-1)} + \Delta H'_{out,left(i)} \\ H'_{flow(i)} &= H'_{flow(i)} - \Delta H'_{out,left(i)} - \Delta H'_{out,right(i)} \\ H'_{flow(i+1)} &= H'_{flow(i+1)} + \Delta H'_{out,right(i)} \end{aligned} \quad (79)$$

where $H'_{flow(i-1)}$, $H'_{flow(i)}$, and $H'_{flow(i+1)}$ are the thickness of the flow prior to the current cellular automata iteration. For each iteration of the cellular automata calculation, equations (79) are applied to each cell in the decimated topography grid. These iterations are repeated until the maximum thickness difference between iterations falls below a tolerance of 10^{-4} .

After the thickness of the flow on the decimated grid, $H'_{flow,i}$, is updated for each pulse, flow thickness is interpolated and added to the original un-decimated topography grid producing an updated flow thickness $H_{flow,i}$. The topography grid at the end of a flow event is then updated as follows:

$$y_{topo,i} = y_{topo,i}^o - H_{flow,i} \quad (80)$$

where $y_{topo,i}^o$ is the y-coordinate of the topography from the prior flow event. Equation (80) ensures that lava flow calculations are consistent with a dynamically evolving landscape impacted by the flow of volcanic material.

The thickness of lava $H_{flow,i}$ produced during a given flow event is added to $H_{flow,tot,i}$, which is the integrated thickness of all flow events associated with a thermo-mechanical model time step. The total flow thickness $H_{flow,tot,i}$ is used to transform sticky markers to volcanic material if y_m is greater than $y_{topo,m}$, which is the interpolated y-coordinate of the updated topography grid $y_{topo,i}$ at x-coordinate x_m . For cases where topography has been updated for both sedimentation and extrusive flow, sticky markers are transformed to volcanic sediment if y_m is greater than $y_{topo,m} + H_{flow,tot,m}$ and to volcanic markers if y_m is less than $y_{topo,m} + H_{flow,tot,m}$ where $H_{flow,tot,m}$ is the total thickness of lava flows interpolated at the x-coordinate of the marker x_m . This approach assumes that sediment is deposited prior to the eruption of volcanic material. The current volume of volcanic material available for extrusion, $V_{vol,k}$, is reset to zero at the end of each eruption event. See Table D.2 for the extrusion and lava-flow parameters used in this work.

2.5. Composite Viscous Creep

A composite rheology is used to model solid-state creep including diffusion, dislocation, and Peierls creep. The composite effective strain rate invariant $\dot{\epsilon}_{eff}$ is given by:

$$\dot{\epsilon}_{eff} = \max(\dot{\epsilon}_{dis}, \dot{\epsilon}_{pei}) + \dot{\epsilon}_{dif}. \quad (81)$$

where $\dot{\epsilon}_{dis}$, $\dot{\epsilon}_{pei}$, and $\dot{\epsilon}_{dif}$ are the strain rate invariants for dislocation, Peierls, and diffusion creep, respectively. The effective creep viscosity $\eta_{creep,m}$ is then

given by:

$$\eta_{creep,m} = \frac{\sigma'_{II(m)}}{2\dot{\epsilon}_{eff}} \quad (82)$$

where $\sigma'_{II(m)}$ is the second invariant of the stress tensor in MPa .

The strain rate invariant for diffusion creep $\dot{\epsilon}_{dif}$ is calculated using the following equation:

$$\dot{\epsilon}_{dif} = A_{dif} \exp \left[-\frac{(E_{dif(m)}10^3 + V_{dif(m)}P_m10^{-6})}{R_{gas}T_m} \right] (\sigma'_{II(m)}10^{-6}) \quad (83)$$

where $A_{dif(m)}$ is the diffusion creep pre-exponential factor in $1/s/MPa$, $E_{dif(m)}$ is the diffusion creep activation energy in kJ/mol , $V_{dif(m)}$ is the diffusion creep activation volume in $J/MPa/mol$, R_{gas} is the ideal gas constant in $J/mol/K$, T_m is the marker temperature in K , and P_m is the marker pressure in Pa . The strain rate invariant for dislocation creep $\dot{\epsilon}_{dis}$ is given by:

$$\dot{\epsilon}_{dis} = A_{dis} \exp \left[-\frac{(E_{dis(m)}10^3 + V_{dis(m)}P_m10^{-6})}{R_{gas}T_m} \right] (\sigma'_{II(m)}10^{-6})^{n_{dis}} \quad (84)$$

where A_{dis} is the dislocation creep pre-exponential factor in $1/s/(MPa)^n$, E_{dis} is the dislocation creep activation energy in kJ/mol , $V_{dis(m)}$ is the dislocation creep activation volume in $J/MPa/mol$ and n_{dis} is the stress exponent. Finally, the strain rate invariant for Peierls creep $\dot{\epsilon}_{pei}$ is calculated using the following equation:

$$\dot{\epsilon}_{pei} = A_{pei}(\sigma'_{II(m)}10^{-6})^2 \exp \left[-\frac{(E_{pei(m)}10^3 + V_{pei(m)}P_m10^{-6})}{R_{gas}T_m} \left(1 - \left(\frac{\sigma'_{II(m)}10^{-6}}{\sigma_{pei(m)}} \right)^{m_{1pei}} \right)^{m_{2pei}} \right] \quad (85)$$

where $A_{pei(m)}$ is the Peierls creep pre-exponential factor in $1/s/(MPa)^2$, $E_{pei(m)}$ is the Peierls creep activation energy in kJ/mol , $V_{pei(m)}$ is the Peierls creep activation volume in $J/MPa/mol$, $\sigma_{pei(m)}$ is the Peierls stress in MPa and m_{1pei} and m_{2pei} are the Peierls stress exponents. See Table D.3 for the creep parameters used in this work.

The visco-elastic nature of the rheology implemented in this work and the power-law and exponential nature of the creep mechanisms introduce a non-linear time-dependent relationship between stress and effective viscosity. Similar to [29], a bisection algorithm (Appendix D) is used to solve for the effective viscosity consistent with a visco-elastic stress forecast for a given marker m

calculated using the following equations:

$$\begin{aligned}
f_{ve,m} &= \frac{\eta_{creep,m}}{\mu_m \Delta t + \eta_{creep,m}}, \\
\sigma'_{xx,fc,m} &= 2\eta_{creep,m} \dot{\epsilon}'_{xx,m} R_{sr,m} (1 - f_{ve,m}) + \sigma'_{xx,m} f_{ve,m}, \\
\sigma'_{xy,fc,m} &= 2\eta_{creep,m} \dot{\epsilon}'_{xy,m} R_{sr,m} (1 - f_{ve,m}) + \sigma'_{xy,m} f_{ve,m}, \\
\sigma'_{II,fc,m} &= \max \left(\sqrt{(\sigma'_{xx,fc,m})^2 + (\sigma'_{xy,fc,m})^2}, \sigma_{min} \right)
\end{aligned} \tag{86}$$

where σ_{min} is a user-defined minimum stress value, f_{ve} is the visco-elastic factor from equation (3), μ_m is the shear modulus, Δt is the time step and $\dot{\epsilon}'_{xx,m}$ and $\dot{\epsilon}'_{xy,m}$ are the deviatoric strain rate components interpolated from strain rate defined on the staggered grid as described by equation (A.6), and $R_{sr,m}$ is the strain rate ratio given by the following equation:

$$R_{sr,m} = \frac{\dot{\epsilon}'_{II,stress,m}}{\dot{\epsilon}'_{II,velocity,m}} \tag{87}$$

where $\dot{\epsilon}'_{II,velocity,m}$ is the strain rate invariant for marker m calculated using the following equation:

$$\dot{\epsilon}'_{II,velocity,m} = \sqrt{(\dot{\epsilon}'_{xx,m})^2 + (\dot{\epsilon}'_{xy,m})^2}. \tag{88}$$

and $\dot{\epsilon}'_{II,stress,m}$ is the strain rate invariant for marker m calculated using nodal deviatoric stress and deviatoric stress changes interpolated from the staggered grid as follows:

$$\dot{\epsilon}'_{II,stress,m} = \sqrt{\left(\frac{\sigma'_{xx,m_2}}{2\eta_{vp,m}} + \frac{\Delta\sigma'_{xx,m}}{2\Delta t \mu_m} \right)^2 + \left(\frac{\sigma'_{xy,m_2}}{2\eta_{vp,m}} + \frac{\Delta\sigma'_{xy,m}}{2\Delta t \mu_m} \right)^2} \tag{89}$$

where σ'_{xx,m_2} and σ'_{xy,m_2} are the forecasted deviatoric stress components at markers interpolated from forecasted deviatoric grid stress $\sigma'_{xx(i,j)_{p2}}$ and $\sigma'_{xy(i,j)_{b2}}$ from equation (A.7) and $\Delta\sigma'_{xx,m}$ and $\Delta\sigma'_{xy,m}$ are the grid stress changes at markers interpolated from $\Delta\sigma'_{xx(i,j)_p}$ and $\Delta\sigma'_{xy(i,j)_b}$ from equation (A.1). The use of $R_{sr,m}$ in equation (86) makes the visco-elastic stress forecast dependent on nodal stress and reduces numerical diffusion in shear zones [29].

With each time step, the effective visco-plastic viscosity $\eta_{vp,m}$ is initialized with the effective creep viscosity $\eta_{creep,m}$ before the effects of plastic failure and partial melt are taken into account using equation (9) and the following equation, respectively:

$$\begin{aligned}
\eta_{vp,m} &= \eta_{vp,m}^o / \exp(f_\alpha M_{extractable,m}) \quad \text{for } M_{extractable,m} < 0.3 \quad \text{and} \\
\eta_{vp,m} &= \eta_{melt} \quad \text{for } M_{extractable,m} \geq 0.3
\end{aligned} \tag{90}$$

where $\eta_{vp,m}^o$ is the previous visco-plastic viscosity initialized as $\eta_{creep,m}$, f_α is a user-defined parameter that controls the sensitivity of the visco-plastic viscosity to melt fraction and η_{melt} is the viscosity of melt at $M_{extractable,m} = 1$. The parameter f_α is set to 35 in this work.

2.6. Strain Weakening and Melt Damage

Experimental work shows that the yield stress in equation (12) decreases as phyllosilicate minerals and foliation form and the pressure-solution deformation mechanism becomes dominant as grain size is reduced in shear zones [50]. This weakening of frictional-plastic strength is modeled by reducing the cohesion and friction angle with increasing plastic strain as described by the following equations:

$$\begin{aligned}\sigma_{c,m} &= \sigma_{c,m}^o + \frac{(\sigma_{c,m}^f - \sigma_{c,m}^o)}{(\epsilon^f - \epsilon^o)}(\epsilon_{plastic,m} - \epsilon^o) \\ \theta_m &= \theta_m^o + \frac{(\theta_m^f - \theta_m^o)}{(\epsilon^f - \epsilon^o)}(\epsilon_{plastic,m} - \epsilon^o)\end{aligned}\quad (91)$$

where m is the index of the marker, $\epsilon_{plastic,m}$ is the plastic strain of the marker, ϵ^o and ϵ^f are initial and final reference plastic strain, $\sigma_{c,m}^o$ and $\sigma_{c,m}^f$ are initial and final reference cohesion, θ_m^f is the final reference friction angle, and θ_m^o is the initial reference friction angle that is randomized at the beginning of each time step using the following equation from [16]:

$$\theta_m^o = \theta_m'^o + (0.5 - r)f_{random} \quad (92)$$

where $\theta_m'^o$ is the initial reference friction angle of the material associated with marker m , r is a random number in the range [0,1], and f_{random} is a randomization factor controlling the magnitude of the perturbation and set equal to 10 for the experiments of this work. The randomization of friction angles described in equation (92) is used to model the evolving variability of frictional-plastic strength due to fluid-rock interactions [16]. The final reference friction angle θ_m^f is defined as:

$$\theta_m^f = \frac{\theta_m'^f}{\theta_m'^o} \theta_m^o \quad (93)$$

where $\theta_m'^f$ is the unmodified final reference friction angle of the material associated with marker m . Marker plastic strain $\epsilon_{plastic,m}$ in equation (91) is calculated using the following equations:

$$\begin{aligned}\sigma'_{II,m} &= \sqrt{(\sigma'_{xx,m})^2 + (\sigma'_{xy,m})^2}, \\ \dot{\epsilon}_{plastic,m} &= \frac{1}{2}\sigma'_{II,m} \left(\frac{1}{\eta_{vp,m}} - \frac{1}{\eta_{creep,m}} \right), \\ \epsilon_{plastic,m} &= \epsilon_{plastic,m}^o + (\dot{\epsilon}_{plastic,m}\Delta t - \dot{\epsilon}_{healing}\Delta t)\end{aligned}\quad (94)$$

where $\epsilon_{plastic,m}^o$ is the previous marker plastic strain, $\dot{\epsilon}_{plastic,m}$ is the plastic strain rate, $\sigma'_{II,m}$ is the second invariant of the stress tensor, and $\dot{\epsilon}_{healing}$ is the plastic healing rate. See Table D.4 for the strain weakening parameters used in this work.

As melt is transported through the melt-extraction zone in the lithosphere [40, 41] it can weaken the bulk frictional-plastic yield strength by forming pressurized dike swarms that may be weaker than the surrounding rock prior to

solidification and by forming reactive channelized networks. In this work weakening associated with melt transport is assumed to occur in the melt-damage zone located within the melt-extraction zone and above local maxima of the partially molten domain in the mantle (Figure D.4a). The maximum probability of melt damage for a given marker m is assumed to scale with the amount of melt being generated in a melt drainage basin k , which is defined by the characteristic magmatic crust height $H_{mc,k}$ calculated during melt extraction (see equation (54)).

The weakening from damage associated with melt transport is modeled by reducing cohesion, friction angle and flow viscosity using a melt damage factor $f_{damage,m}$ as described by the following equations:

$$\begin{aligned}\sigma_{c,m} &= \frac{\sigma_{c,m}}{f_{damage,m}}, \\ \theta_m &= \frac{\theta_m}{f_{damage,m}}, \\ \eta_{creep,m} &= \frac{\eta_{creep,m}}{f_{damage,m}}.\end{aligned}\tag{95}$$

The melt damage factor $f_{damage,m}$ is defined in a binary manner where $f_{damage,m}$ is set equal to 1 outside of the melt-damage zone and set equal to either 1 or a constant value greater than 1 based on a probability distribution within the melt-damage zone. A melt-damage zone is defined for a given melt drainage basin k by the x-coordinate limits x_{damage}^{min} and x_{damage}^{max} and y-coordinate limit y_{damage}^{max} defined as follows:

$$\begin{aligned}x_{damage}^{min} &= x'_{shallow,pm,k} - \frac{W_{damage}}{2}, \\ x_{damage}^{max} &= x'_{shallow,pm,k} + \frac{W_{damage}}{2}, \\ y_{damage}^{max} &= y'_{shallow,pm,k}\end{aligned}\tag{96}$$

where $x'_{shallow,pm,k}$ is the average x-coordinate of the shallowest partially molten mantle marker and $y'_{shallow,pm,k}$ is the average y-coordinate of the shallowest partially molten mantle marker calculated during melt extraction, and W_{damage} is the width of the damage zone set equal to 5 km for the experiments presented in this work. The melt-damage model is applied to each marker m in the melt-damage zone excluding sticky-air and sticky-water markers.

The melt damage factor $f_{damage,m}$ for a given marker m is calculated using the following equation:

$$f_{damage,m} = \begin{cases} f_{damage,max} & \text{if } r < p_{damage,m} \\ 1.0 & \text{if } r \geq p_{damage,m} \end{cases}\tag{97}$$

where $f_{damage,max}$ is the maximum melt damage factor set equal to values ranging from 1 to 10 for the experiments of this work, and r is a random number

between 0 and 1. The damage probability $p_{damage,m}$ is calculated using the following equation:

$$p_{damage,m} = \begin{cases} 0.0 & \text{if } x_m < x_{damage}^{min} \text{ or } x_m > x_{damage}^{max} \\ \left[\frac{p_{damage}^{central}}{2 \cos\left(\frac{(x_m - x'_{shallow,pm,k})}{W_{damage}} 2\pi\right) + 1} \right] & \text{if } x_{damage}^{min} \leq x_m \leq x_{damage}^{max} \end{cases} \quad (98)$$

where the central damage zone damage probability $p_{damage}^{central}$ is given by:

$$p_{damage}^{central} = \begin{cases} 0.0 & \text{if } H_{mc,k} \leq H_{mc,min} \\ p_{damage}^{max} & \text{if } H_{mc,k} \geq H_{mc,max} \\ \frac{p_{damage}^{inter}}{H_{mc,inter} - H_{mc,min}} \Delta H_1 & \text{if } H_{mc,min} < H_{mc,k} \leq H_{mc,inter} \\ p_{damage}^{inter} + \frac{p_{damage}^{max} - p_{damage}^{inter}}{H_{mc,max} - H_{mc,inter}} \Delta H_2 & \text{if } H_{mc,inter} < H_{mc,k} < H_{mc,max} \end{cases} \quad (99)$$

where $\Delta H_1 = (H_{mc,k} - H_{mc,min})$, $\Delta H_2 = (H_{mc,k} - H_{mc,inter})$, $H_{mc,k}$ is characteristic magmatic crust height for drainage basin k calculated using equation (54), $H_{mc,min}$ is the minimum reference characteristic magmatic crust height, $H_{mc,max}$ is the maximum reference characteristic magmatic crust height, $H_{mc,inter}$ is the intermediate reference characteristic magmatic crust height, p_{damage}^{max} is the maximum damage probability, and p_{damage}^{inter} is the intermediate damage probability. See Table D.5 for the melt damage parameters used in this work.

To account for the effect of migrating melt in the melt-damage zone on marker density, the following equation is used:

$$\rho_m = \rho_m^o(1 - f_{dike}) + \rho_{melt}f_{dike} \quad (100)$$

where ρ_m^o is the density of the marker prior to this update step, ρ_{melt} is the density of the melt that takes pressure into account using the approach described in equation (48), and f_{dike} is the fraction of the marker occupied by migrating melt associated with dike injection and channelized melt networks. Equation (100) is used to model density changes associated with dike injection similar to the approach described by [26] (Figure D.3b).

3. Model Setup

The model domain has dimensions 500 km by 160 km and is initialized with a 10 km thick sticky-air layer, a 22 km thick felsic upper continental crust layer approximated as wet quartzite, a 10 km thick mafic lower continental crust layer approximated as wet anorthite and a 93 km thick ultramafic continental mantle lithosphere layer superimposed on an ultramafic asthenosphere layer (Figure D.5). Similar to [16], lateral strong zones where strain weakening is disabled are used to localize deformation in the central part of the model domain. See Table D.6 for the material properties associated with each layer.

A uniform grid spacing of 200 m is used within a 200 km wide and 50 km thick central region of the model domain (Figure D.5). The average lateral grid spacing to the left and right of the central region is 2 km , and lateral grid spacing is increased in a linear manner from the edge of the high-resolution central region to the edge of the model domain using the approach described by [29]. The vertical resolution of the model below the high-resolution central region is 2 km and is increased in a linear manner from the edge of the high-resolution central region to the bottom of the model domain using a similar approach. The average initial marker spacing is 25 m , and the initial location of markers is randomly perturbed relative to a uniformly spaced grid.

The initial thermal structure within the continental lithosphere is defined using a semi-analytical solution of the steady-state 1D heat conduction equation for a model with constant temperature boundary conditions at the top and bottom of a three-layer lithosphere composed of upper felsic continental crust, lower mafic continental crust and continental mantle lithosphere. The approach used to solve the 1D steady-state heat conduction equation follows the approach described by [51] but modified for a constant temperature bottom boundary condition instead of constant heat flow. Temperature at the top of the lithosphere is set to 0°C, and the temperature at the base of the lithosphere T_{lith} is set equal to 1345°C. For the initial geotherm, constant thermal conductivity is assumed within each layer of the lithosphere with values of 2.25 $W/m/K$, 2.0 $W/m/K$ and 2.0 $W/m/K$ for the upper felsic continental crust, lower mafic continental crust and continental mantle lithosphere, respectively. Radiogenic heat production for each lithospheric layer is defined using the values from Table D.6. The initial thermal structure within the asthenosphere is calculated using an adiabatic gradient of $\gamma = 0.4^\circ C/km$. Figure D.5b shows the calculated temperature profile used to initialize the model. In order to ensure localization occurs in a similar region across all model cases, the initial temperature is perturbed in the central region of the model domain by elevating the depth of the thermal lithosphere-asthenosphere boundary by an amplitude of 5 km in the center of the model domain (Figure D.5a). The perturbation decays in a symmetric linear manner in the x-direction to zero over a distance of 5 km relative to the center of the model domain.

The temperature boundary condition along the top of the model domain is set to a constant temperature of 0°C, and the side boundaries have a zero heat flux boundary condition. The temperature boundary condition at the base of the model is defined using the following equation:

$$T_{base} = T_{lith} + \gamma L_{asthenosphere} + \Delta T \quad (101)$$

where T_{lith} is the temperature at the base of the continental lithosphere, γ is the adiabatic gradient in $^\circ C/km$, $L_{asthenosphere}$ is the thickness of the asthenosphere in km , and ΔT is the thermal anomaly associated with a plume that rapidly ascends through the asthenosphere and spreads out laterally beneath the continental lithosphere.

Velocity boundary conditions along the side boundaries are set to outflow boundary conditions equal to $0.5v_{ext}$ where v_{ext} is the full extensional velocity

of either 1 *cm/yr* or 2 *cm/yr*. This range of full extension velocities is typical of the values inferred from plate reconstruction during the early phases of extension and breakup and SDR formation [14]. Prescribed inflow velocities along the top and bottom boundaries are defined so that the net inflow/outflow volume is zero over the entire model domain and the net inflow/outflow volume is zero within the sticky-air layer. The prescribed inflow velocity along the top boundary v_y^{top} is calculated as:

$$v_y^{top} = \frac{v_{ext}(H_{sticky,l} + H_{sticky,r})}{x_{size}} \quad (102)$$

where $H_{sticky,l}$ and $H_{sticky,r}$ are the thicknesses of the sticky layer on the left and right sides of the model domain, and x_{size} is the width of the model domain. The prescribed inflow velocity along the bottom boundary v_y^{bot} is calculated as:

$$v_y^{bot} = v_y^{top} - 2v_{ext}\frac{y_{size}}{x_{size}} \quad (103)$$

where y_{size} is the height of the model domain (Figure D.5). Inflow velocities v_y^{top} and v_y^{bot} are recalculated at each time step to account for changes in the sticky layer thickness due to crustal thickness changes and isostatic adjustment.

Global sea level is defined at the top of a reference lithospheric column with properties identical to those of the initial lithospheric configuration. Changes in global sea level are calculated from an isostatic balance between this reference column and the laterally averaged pressure at the base of the model domain. To simulate subaerial conditions associated with a restricted basin, a local sea level is prescribed by shifting the global sea level downward by 10 km.

4. Results

A systematic exploration of parameter space was conducted using 44 high-resolution model runs to assess the primary controls on magmatic crust formation, structural evolution and the geometry of dipping lava-flow packages (Table D.7) (Figures D.6 – D.18). The reference model is Case 0, which includes melt damage and uses the following parameter values: $v_{ext} = 2$ *cm/yr*, $\Delta T = 100^\circ\text{C}$, $f_{v,max} = 0.50$, $L_{sa} = 40,000$ m, and $\Delta t_{erupt} = 200,000$ yr (Figure D.6). Table D.7 summarizes the model parameters and key results for all model cases. Animations associated with all model cases in Table D.7 are archived at [52] including compositional animations for all cases and animations of visco-plastic viscosity and density for selected cases.

The model parameters varied in this study include: (1) the inter-eruption period Δt_{erupt} (50,000 – 400,000 yr), which controls the integrated volume of lava associated with eruption events and the amount of spreading and initial downward rotation of lava-flow packages prior to burial by subsequent flows (Figures D.8 and D.9); (2) the maximum extrusion factor $f_{v,max}$ (0.30–0.60), which determines the fraction of extracted melt that can reach the surface as subaerial flows (Figures D.10 and D.11); (3) the characteristic length of subaerial flows L_{sa} (20,000–80,000 m), which influences how far subaerial flows can

extend beyond the eruption location and the magnitude of the local volcanic load at the spreading axis (Figures D.12 and D.13); (4) the thermal anomaly ΔT (25–150°C), which controls the total volume of melt produced during the model run (Figures D.14, D.15, and D.16); (5) the full extension velocity v_{ext} (1–2 cm/yr), which controls the rate of passive upwelling of asthenospheric mantle beneath the spreading axis (Figures D.17 and D.18); (6) maximum probability of melt damage p_{damage}^{max} (0.75–1), which controls the continuity of weakening at the ridge axis (Table D.7 Case 16); and (7) the volume fraction of a melt-damaged marker impacted by the density of dike fluid f_{dike} (0–1), which contributes to the density contrast between the ridge axis and newly accreted gabbroic crust (Table D.7 Case 15) (Supplemental Materials Section 1.2 and Figure S3).

4.1. Magmatic Crust Formation with Melt Damage

The base case model with melt damage (Case 0, Table D.7) undergoes an initial phase of crustal stretching whereby small-offset faults produce distributed crustal thinning (Figures D.6a and Supplemental Materials Figure S1a). With increased extension, a phase of more localized crustal thinning occurs that involves faults with larger offsets in the frictional-plastic upper crust and ductile thinning in the lower continental crust (Figure D.6b and Supplemental Materials Figure S1b). The final phase of lithospheric breakup is associated with coupling between the frictional-plastic upper crust and frictional-plastic layers in the lower crust and upper continental mantle lithosphere (Figure D.6c and Supplemental Materials Figure S1c). This final phase of breakup involves localized upwelling of asthenosphere that initializes partial melting (Figure D.6d and Supplemental Materials Figure S1d) and produces a layer of magmatic crust with a relatively small partially molten core of fractionated gabbro. Subaerial lava-flow packages that were erupted during this early phase of magmatic crust formation dip toward the spreading axis (Figure D.6c).

With increased spreading the partially molten domain in the asthenosphere grows larger and a ridge of gabbroic material forms that is cored by partially molten fractionated gabbro (Figure D.6e and Supplemental Materials Figure S1e). The top of the partially molten fractionated gabbro core is approximately denoted by the 1200°C isotherm, which is close to the solidus for fractionated gabbro defined by equations (44) (Figure D.6e and f). The down-dip section of volcanic flow packages undergoes rotation toward the ridge axis during periods between eruption events with increased downward concavity in the down-dip direction due to localized frictional-plastic deformation in the melt-damage zone (e.g., Figure D.6e). Flow packages rotate more than 30° as they spread away from this gabbroic ridge and downlap on a relatively smooth magmatic basement with minor rugosity associated with faulting close to the ridge axis (compare red and green tracers in Figure D.6d-f).

As spreading continues and the magmatic system widens, the partially molten domain in the asthenosphere and the partially molten core of the fractionated ridge reach steady-state volumes, and vertical bands of newly accreted gabbroic

crust undergo ductile deformation and folding due to the load induced by volcanic flows (Figure D.6f) and visco-plastic viscosities that range from $10^{19} Pa \cdot s$ to $10^{21} Pa \cdot s$ in the lower section of the magmatic crust (Supplemental Materials Figure S1d-f). This ductile deformation in the gabbroic magmatic crust leads to magmatic crustal thickness changes of more than 3 km along the flanks of the gabbroic ridge. With continued spreading volcanic flow packages undergo plastic deformation leading to diverging downward-concave geometries that increase in dip in the down-dip direction (Figure D.6f). Maximum dip at the point where the gabbroic ridge flattens out is around 36° , and dips that exceed 50° are observed close to the edge of extended continental crust where flows that formed during the early phases of volcanic extrusion downlap onto magmatic basement.

Using a dry anorthite rheology (Case 43) significantly reduces the amount of ductile deformation in the gabbroic magmatic crust and limits the amount of upward ductile flow of newly accreted gabbroic crust. As a result, the gabbroic ridge is approximately 1 km less thick than the case with wet anorthite rheology (Figure D.6g). The reduced thickness of the gabbroic ridge enhances the early rotation of lava-flow packages during spreading and separation, resulting in kink-like geometries in the down-dip portions of the rotated flow packages (Figure D.6g).

Increasing the subaerial transport coefficient to a high end-member value (Case 13; Table D.7) (Figure D.6h) redistributes material from the initial volcanic edifice and reduces the volcanic load acting on the magmatic crust. This reduction in loading, in turn, decreases deformation within the magmatic crust. Enhanced subaerial transport also increases dip variability in the down-dip portions of volcanic-flow packages, as greater volumes of sediment are delivered to the axial depression generated by spreading during inter-eruption periods.

4.2. Magmatic Crust Formation without Melt Damage

The importance of the probabilistic melt-damage mechanism is demonstrated by comparing Case 0 (Figure D.6a-f), which incorporates melt damage, with Case 19 (Figures D.7 and D.15c) (Table D.7), in which melt damage is disabled. Without melt-damage localization, deformation at the ridge axis becomes more dynamic, alternating between phases in which triangular horst structures form at the ridge axis and phases in which these horsts are bisected by normal faults (Figure D.7 and Supplemental Materials Figure S2). The horst structures are bounded by shear zones that are located at the interface between dipping volcanic flows and newly accreted gabbroic crust that has cooled below $600^\circ C$ and is undergoing frictional-plastic deformation. These shear zones terminate in the ductile deformation regime of accreted gabbroic crust, which is approximately bounded by the $600 - 700^\circ C$ isotherms.

The accretion of gabbroic crust is also more complex than in Case 0, with episodes of asymmetric accretion followed by intervals of more symmetric accretion (Figure D.7). Basement geometry exhibits greater variability than in Case 0. Relatively smooth basement morphology is associated with longer-lived episodes of triangular horst formation at the ridge axis, whereas more rugged

basement geometry, with local relief of approximately 2 km, develops when triangular horsts are bisected by normal faults (Figure D.7).

In Case 19, where melt damage is disabled, the early ridgeward rotation of subaerial lava-flow packages is controlled by syn-kinematic lava-flow emplacement onto triangular horst structures at the ridge axis (Figure D.7). Subsequent spreading and burial of lava-flow packages leads to further ridgeward rotation with maximum dip in the down-dip direction of around 35° to 45° (Figure D.15c). In general, with Case 19, dipping rotated flow packages exhibit divergent downward-concave geometries that onlap smooth magmatic basement and show an increase in dip in the down-dip direction (Figure D.15c). However, in contrast to Case 0, Case 19 shows local regions where lava-flow packages underwent deformation associated with the bisection of triangular horsts by normal faults. This deformation produces locally upward-concave geometries in the down-dip sections of lava-flow packages (Figures D.7 and D.15c).

4.3. *Effect of the Inter-Eruption Period*

Figures D.6f, D.8, and D.9 show the effect of the inter-eruption period on the geometry of lava-flow packages and the magmatic crust structure for cases with and without melt damage. The inter-eruption period controls the amount of time between eruptive events, the volume of lava erupted during each eruption event, and the amount of spreading and extensional displacement of the magmatic crust before the next eruption event.

For the model with melt damage and an inter-eruption period of 50,000 yr, the early evolution is characterized by relatively symmetric lava-flow emplacement and gabbroic crustal accretion (Figure D.8a) [52]. This phase is followed by the development of asymmetric lava-flow deposition associated with listric normal faulting within the volcanic package above the ridge axis. Fault-related subsidence produces an asymmetric depression that preferentially diverts lava toward the left side of the spreading axis. Because the erupted lava volume is relatively small, lava flows are unable to flow over the rift flank on the right side of the ridge axis. This asymmetric lava emplacement leads to strongly asymmetric gabbroic crustal accretion and promotes migration of the ridge axis toward the right side of the spreading axis. On the left side of the spreading axis, lava-flow packages undergo ridgeward rotation, develop kinked down-dip geometries, and downlap onto a relatively smooth magmatic basement. In contrast, the right side of the spreading axis is characterized by a relatively thin lava-flow package with ridgeward-dipping rotated geometries that are primarily controlled by spreading-related deformation and rotation at the ridge axis.

Increasing the inter-eruption period leads to more symmetric lava-flow emplacement and gabbroic crustal accretion, larger fault offset at the ridge axis and smaller increases in dip in the down-dip direction of the flow packages (Figure D.6f and D.8). With longer inter-eruption periods of 400,000 yr, normal fault offset is around 1.5 to 2 km for cases with melt damage (Figure D.8c) and up to 3 to 4 km for cases without melt damage (Figure D.9c). These cases with longer inter-eruption periods produce downward-concave lava-flow packages that downlap onto the footwall side of normal faults that developed in

gabbroic magmatic crust and dip toward extended continental crust (Figures D.8c and D.9c). These structures are geometrically similar to the Type I SDRs described by [9] (Figure D.3a). An important distinction between the model results presented here and the conceptual model of [9] is that Type I geometries are observed to develop within faulted magmatic crust rather than extended continental crust. This result has important implications for stratigraphic age models of some SDR-bearing margins because it suggests that strata within and beneath Type I SDR packages are younger than the solidification age of the underlying gabbroic magmatic crust.

Cases without melt-damage exhibit greater asymmetry in lava-flow package geometry and magmatic crust structure compared to cases with melt damage and produce gabbroic magmatic basement with greater relief than in cases with melt damage (Figure D.8). The relief in the gabbroic basement can reach up to 5 km in cases without melt damage and relatively small inter-eruption periods (Figure D.9a).

4.4. Effect of Maximum Extrusion Factor

Figures D.6f, D.10, and D.11 show the influence of the maximum extrusion factor ($f_{v,max}$) on the geometry of lava-flow packages and the magmatic crust structure for cases with and without melt damage. At low maximum extrusion factors ($f_{v,max} < 0.50$), the gabbroic ridge becomes mobile and thickens to more than 10 km, and the upper boundary of the partially molten domain in the asthenosphere develops small-amplitude undulations (Figure D.10a and b). These undulations cause the melt-damage zone to migrate laterally through time, leading to more complex deformation at the ridge axis even though the melt damage mechanism is active. This instability in melt-damage zone location leads to the development of asymmetry in lava-flow package geometry with dominant flow directions relative to the ridge axis changing through time (Figure D.10a and b). Furthermore, the surface of the magmatic basement exhibits a smooth and uniform morphology with minor relief associated with axial faulting, and the gabbroic magmatic crust exhibits broad asymmetric undulations (Figure D.10a and b).

When extrusion factors exceed 0.4 for cases with melt damage, the ridge axis becomes fixed, and deformation remains stably localized at the ridge axis. This produces lava-flow packages that are approximately symmetric about the ridge axis, as well as symmetric accretion of gabbroic magmatic crust (Figures D.6f and D.10c). In contrast, models without melt damage exhibit greater ridge mobility and more asymmetric accretion of gabbroic crust, even at high extrusion factors (Table D.7) (Figure D.11c). Across all cases, increasing the extrusion factor results in thicker lava-flow packages and higher maximum dips within the downward-concave lava-flow packages (Figures D.6f and D.10).

4.5. Effect of Characteristic Subaerial Flow Length

Increasing the characteristic subaerial flow length (L_{sa}) promotes a more stable ridge location and produces ridgeward-dipping, downward-concave lava-flow packages that steepen in the down-dip direction (Table D.7) (Figures D.6f,

D.12 and D.13). Longer subaerial flow lengths lead to wider, thinner volcanic edifices, which impose smaller local volcanic loads and therefore drive less ductile deformation of the magmatic crust (Figures D.12 and D.13). Longer subaerial flow lengths also produce longer up-dip sections of ridgeward-dipping lava-flow packages characterized by small changes in dip that transition into more steeply dipping down-dip sections that downlap onto magmatic basement with low relief (Figure D.12b,c).

For models with melt damage, increasing the characteristic subaerial flow length is accompanied by a reduction in both the height of the gabbroic ridge and the dip of the ridge flanks (Figure D.12). The greater ridge height produced by shorter subaerial flow lengths reflects larger localized volcanic loads, which enhance upward flow of ductile gabbroic crust at the ridge axis and increase deformation along the magmatic-crust flanks (Figure D.12a). In models without the localizing effects of melt damage, volcanic loads associated with flow lengths shorter than approximately 30 km drive substantial ductile deformation of the magmatic crust. These cases produce a gabbroic magmatic basement composed of multiple ridges, with relief of up to 5 km and minimum crustal thicknesses of approximately 3 km (Figure D.13a).

4.6. Effect of the Plume Thermal Anomaly

The plume thermal anomaly ΔT controls the temperature of the sub-lithospheric mantle and the total volume of melt produced during the model run. Model results for cases with parameters similar to Case 0 (Figure D.6f) but with different thermal anomalies are shown in Figure D.14. A similar suite of models is shown for cases without melt damage in Figure D.15.

During early magmatic crust formation for cases with a relatively low plume thermal anomaly of 25°C the characteristic magmatic crust height is less than the maximum reference value from Table D.2 as described in equation (56) (Figures D.14a and D.15a). This results in an increase in the total thickness of the lava-flow packages over time (Figures D.14a and D.15a). With a larger thermal anomaly of 50°C, the characteristic magmatic crust height exceeds the maximum reference value during the early stages of magmatic crust formation, resulting in a larger total thickness of lava-flow packages (Figures D.14b and D.15b).

With larger thermal anomalies, the height of the gabbroic ridge increases and broad undulations develop in the magmatic crust (Figures D.14c and D.15c). There is also a systematic increase in the maximum dip of downward-concave lava-flow packages with larger thermal anomalies (Figures D.14 and D.15). Figure D.16 shows results at the maximum thermal anomaly of 150°C explored in this work for different assumptions about the length of subaerial flows (L_{sa}) and maximum extrusion factors ($f_{v,max}$). At these high thermal anomalies, a diversity of lava-flow package geometries and magmatic crustal structures are observed depending on extrusion and lava flow parameters including (1) downward-concave lava-flow packages that become upward concave in the down-dip direction (Figure D.16a and c), (2) very thick gabbroic magmatic crust with asymmetric accretion (Figure D.16c), (3) ridgeward-dipping flow packages that

have lengths exceeding 30 km and show very small changes in dip in the down-dip direction (Figure D.16b and c), and (4) volcanic packages with thicknesses that exceed 11 km and dips of downward-concave flow packages that exceed 70° in the down-dip direction (Figure D.16d).

4.7. Magmatic Crust Formation at Ultraslow Spreading Rates

Plate reconstructions of the Central Atlantic that account for onshore syn-rift deformation and magmatic crustal formation suggest that full spreading rates may have been between 1 and 1.5 cm/yr during formation of the SDRs imaged in seismic reflection data from the eastern U.S. margin [14]. To investigate the effects of these ultraslow spreading rates on SDR formation, models were run with a spreading rate of 1 cm/yr using a thermal anomaly of 100°C and 150°C with different assumptions about the inter-eruption period (Δt_{erupt}) and the length of subaerial flows (L_{sa}) (Figures D.17 and D.18).

The reduction in spreading rate from 2 to 1 cm/yr produces lower temperatures within the gabbroic ridge and sub-ridge asthenosphere, reduces ductile deformation of the magmatic crust, decreases the amount of partial melting in the asthenosphere, and results in a smaller, or absent, core of partially molten fractionated gabbro (Figures D.17 and D.18). Consequently, when all other parameters are held fixed, halving the spreading rate reduces the volume of volcanic material erupted during each eruptive event.

For an inter-eruption period of 200,000 yr, a smaller lava volume per event is erupted than in the equivalent case with a full spreading rate of 2 cm/yr (Figures D.6f and D.17a). This reduced eruptive volume leads to kinked down-dip geometries within the lava-flow packages (Figures D.17a and D.18a). Increasing the inter-eruption period by a factor of four, to 400,000 yr, allows larger lava volumes to accumulate during eruption events and produces more downward-concave lava-flow packages characterized by a smoother increase in dip in the down-dip direction (Figures D.17b,c and D.18b,c).

The ultraslow case shown in Figure D.17d (Case 42), which has a thermal anomaly of 100°C , an inter-eruption period of 400,000 yr, and a subaerial flow length of 80 km, produces downward-concave, ridgeward-dipping lava-flow packages that are similar to those in Figure D.12c (Case 9), which uses equivalent parameters but has a higher full spreading rate of 2 cm/yr and an inter-eruption period of 200,000 yr. Cases 9 and 42 also produce dipping lava-flow package geometries that are broadly consistent with the conceptual model presented by [9, 13] for Type II SDRs. These similarities include small dip variations within gently dipping up-dip sections over distances of 20–40 km, followed by a progressive down-dip increase in reflector dip to maximum values of less than 40° (Figure D.3a).

5. Discussion

This study presents the first systematic investigation of SDR formation using high-resolution visco-elasto-plastic geodynamic models that couple mantle

melting, magma transport, and extrusive lava-flow emplacement. Two methodological advances underpin the results. The first is the coupling of long-time-scale lithospheric deformation with a short-time-scale, cellular-automata lava-flow model that represents subaerial and submarine emplacement and episodic eruptions separated by an inter-eruption period. The second is a probabilistic melt-damage weakening model that approximates the effects of melt networks, elevated melt pressures, and dike injection on lithospheric rheology. This weakening model is spatially and temporally coupled to active melting and stabilizes the spreading ridge during magmatic crust formation and produces the symmetric ridgeward-dipping geometries inferred from seismic reflection data [8, 9, 13] (Figure D.3a).

The models presented here show that ridgeward-dipping, downward-concave lava-flow packages, which become SDRs once a margin is flooded, are a natural consequence of syn-kinematic lava emplacement at an active ultraslow to slow spreading center located above anomalously hot asthenosphere (Figure D.19a,b). Near-horizontal lava-flow packages are emplaced and bisected by extension and frictional-plastic deformation at the ridge axis into conjugate ridgeward-dipping packages with dips that increase in the down-dip direction. This process forms an axial depression that is filled by subsequent flows. Lava-flow packages that experienced this initial extension and rotation at the ridge axis are subsequently modified by additional plastic deformation and rotation driven by progressive burial by younger flows, continued spreading, lithospheric density changes associated with thermal cooling and partial-melt solidification, and volcanic-load-induced ductile thinning of the gabbroic magmatic crust.

The initial extension, bisection, and rotation of near-horizontal lava-flow packages at the ridge axis, together with the formation of an axial depression, are important mechanisms for producing ridgeward-dipping, downward-concave lava-flow packages in the models presented here. Axial valleys with 1–2 km of relief are common features of slow to ultraslow spreading ridges [53, 45, 54, 55, 56]. These valleys are commonly associated with normal faulting at the ridge axis, as supported by focal-mechanism data [45]. More broadly, extensional deformation, including detachment faulting, is well documented at slow to ultraslow spreading systems [55]. The work of [53] further showed that axial valleys at many slow-spreading ridges can be explained by flexural models of lithospheric extension that incorporate plastic rheology and variable lithospheric thickness. Thus, the axial depressions and associated ridgeward-dipping lava-flow packages produced in the models presented here are consistent with both observations and mechanical models of slow to ultraslow spreading ridges.

The specific extensional mechanisms that generate ridgeward-dipping packages, as well as the detailed geometry of the lava-flow packages and associated magmatic crustal structures, are sensitive to several model parameters. These include (1) the presence or absence of melt damage, (2) the magnitude of the thermal anomaly in the asthenosphere, (3) the spreading rate, (4) the rheology of the gabbroic magmatic crust, and (5) the distribution and timing of volcanic loads that drive ductile deformation of the magmatic crust (Table D.7; Figures D.6 – D.18). The latter factor includes the inter-eruption period, the maximum

extrusion factor, and the length of subaerial lava flows.

The presence or absence of melt damage exerts an important first-order control on the geometry of subaerial lava-flow packages, the structure of the magmatic crust and the mobility of the gabbroic ridge (Table D.7). For most cases, melt damage stabilizes deformation at the gabbroic ridge axis, promotes more symmetric accretion of gabbroic crust, leads to lower-relief, relatively smooth gabbroic basement, and produces more laterally balanced ridgeward-dipping lava-flow packages (Figures D.6–D.15). The melt-damage mechanism is therefore required to reproduce the symmetric lava-flow package geometries and magmatic crustal structures from conceptual models of SDR formation inferred from seismic reflection data [9, 8] (Figure D.3a).

Several cases produce maximum dips exceeding 40° near the edge of extended continental crust, as in the models of [26] (Figure D.3b). Such dips have not been reported in seismic reflection data, which typically show SDR dips below 40° (Figures D.3a, D.1, and D.2), but they cannot be ruled out because the down-dip portions of SDR packages are commonly poorly imaged [9] and seismic coverage is incomplete. Dips most consistent with the imaged record arise under longer inter-eruption periods, syn-kinematic emplacement during listric normal faulting in magmatic crust, low to moderate extrusion factors, longer subaerial flow lengths, and thermal anomalies below 100°C (Figures D.8, D.10, D.12, D.17, and D.14).

A diversity of magmatic crustal structures and basement relief is observed in the models presented in this work including uniform magmatic crustal thickness with low-relief basement surfaces away from ridge flanks, cases with crustal thickness changes and basement surface relief that exceeds 5 km, highly deformed magmatic basement with high-amplitude ridges, and asymmetric gabbroic accretion sometimes controlled by syn-kinematic emplacement of lava-flow packages during listric normal faulting. More symmetric cases with low-relief basement surfaces often require the presence of melt damage whereas more asymmetric cases with high-relief basement surfaces are often associated with the absence of melt damage. Although many published models of SDR formation are consistent with these cases with melt damage, observed SDR geometries and high-relief basement surfaces from Eastern Greenland (Figure D.2c) appear consistent with some of the cases without melt damage (e.g., Figure D.11a).

The kinematic-dynamic numerical models with a half-spreading configuration presented by [25, 26] use the solidification of a 15 km-thick dike at the ridge axis to drive early SDR rotation (Figure D.3b). This model setup successfully reproduced several key geometric features of Type II SDRs, including downward concavity and divergent reflector packages. However, it did not reproduce the gently dipping up-dip sections characterized by small dip variations that transition into more steeply dipping down-dip sections, as illustrated in the conceptual model shown in Figure D.3a. An additional limitation of the models of [25, 26] is that they assume SDR development occurs within an integrated volcanic and magmatic system whose thickness is comparable to that of undeformed continental crust (Figure D.3b). In contrast, magmatic crustal structures interpreted from seismic reflection data (Figures D.1 and D.2), together

with the visco-elasto-plastic models with strain weakening presented here, suggest that most SDRs develop within integrated volcanic and magmatic systems that are approximately 9–20 km thick, with the volcanic section representing 30–60% of the total thickness. These observations make the assumption of a 15 km-thick dike in [26] (Figure D.3b) difficult to justify for most SDR-bearing systems except for the early evolution of SDRs in the Pelotas Basin (Figure D.1d).

The models presented in this work use a realistic density formulation for the spreading system that accounts for density variations associated with temperature, pressure, and partial melt content in the asthenosphere and gabbroic magmatic crust. To evaluate the potential importance of the dike-solidification mechanism proposed by [25, 26], this density formulation was modified to include the effects of dike fluid in markers probabilistically selected for melt damage (Equation (100)) (Case 15 in Table D.7; Supplemental Materials Section 1.2 and Figure S3). Case 15 shows that the presence of dike fluid within the melt-damage zone produces a density reduction of only approximately 50 kg/m^3 relative to the already hot and partially molten gabbroic ridge. In addition, the maximum vertical extent of the dike-fluid zone is limited to approximately 10 km. Even assuming zero strength at the ridge axis, the predicted isostatic adjustment in ridge elevation is less than 160 m. These results suggest that the presence of dike fluid within the melt-damage zone is unlikely to be a major control on the early rotation of lava-flow packages (Supplemental Materials Section 1.2).

Reproducing the full complexity of the SDRs and magmatic crustal structures from volcanic margins (Figures D.1 and D.2) will likely require more sophisticated models that include (1) the full evolution of plume thermal structure that captures the dissipation of thermal anomalies over time, (2) post-SDR depositional events, (3) more realistic episodic eruption models that account for temporal variations in inter-eruption periods, (4) velocity boundary conditions based on plate reconstructions that account for local and variable extension rates, and (5) models that account for three-dimensional deformation, melt transport, magmatic crust formation and lava-flow emplacement. Despite these limitations, the relatively simple 2D models presented in this work successfully reproduce the first-order SDR characteristics represented in the conceptual models of SDR formation from [9] (Figure D.3a). These characteristics include (1) downward-concave ridgeward-dipping lava-flow packages with up-dip sections that have dip changes ranging from 2 to 5° , (2) down-dip sections that increase in dip in the down-dip direction and reach maximum dips of less than 35° and (3) ridgeward-dipping lava-flow packages that downlap onto the footwalls of landward-dipping faults. These geometries are observed in seismic reflection data from many SDR-bearing margins as shown in Figures D.1 and D.2.

The numerical models presented in this work support the use of SDRs as indicators of subaerial conditions in paleogeographic reconstructions, and suggest that the early basin-forming periods of many global plate reconstructions may need to be revised to account for subaerial conditions and SDR-related spreading under slow to ultraslow spreading rates [14]. Furthermore, the models provide

a quantitative framework for testing hypotheses about the thermal evolution of the deep interior. The numerical experiments presented in this work support the conceptual models of [8] and [9] where SDRs are underlain by large volumes of mafic magmatic crust. This indicates that potential temperatures were sufficiently high for millions of years during the early phases of basin formation to produce large volumes of mafic melt. The preservation of thick symmetric SDR packages at many margins around the globe may indicate that thermal anomalies of 50 to 150°C were present in the sub-continental asthenosphere for several million years during the early basin-forming period.

6. Conclusions

The 44 high-resolution visco-elasto-plastic geodynamic models presented in this work demonstrate that ridgeward-dipping, downward-concave lava-flow packages, which would form SDRs after marine flooding, are a natural consequence of syn-kinematic lava emplacement at an active spreading center above anomalously hot asthenosphere. Extension and frictional-plastic deformation bisect near-horizontal subaerial flows into ridgeward-dipping conjugate packages at the ridge axis where an axial depression forms. Bisected conjugate lava-flow packages are then rotated further by burial by younger flows, continued spreading, density increases associated with thermal cooling and partial-melt solidification, and volcanic-load-induced ductile thinning of the gabbroic crust. This work also shows that the additional density reduction from fluid-filled dikes has little effect on axial topography and ridgeward-dipping lava-flow package geometries.

SDR geometry, magmatic crustal structure, and ridge-axis mobility are governed by the presence or absence of melt damage, the asthenospheric thermal anomaly, the spreading rate, the gabbroic-crust rheology, and the distribution and timing of volcanic loads. The strong sensitivity to the latter, which includes the inter-eruption period, maximum extrusion factor, and subaerial flow length, shows that the details of melt extraction and extrusive emplacement are themselves first-order controls on SDR geometry, not merely surface expressions of deeper processes.

When melt damage is active, the relatively simple 2D models reproduce the first-order characteristics of the conceptual SDR models of [9] that are based on a large volume of seismic reflection data: downward-concave packages with gently varying up-dip sections (2–5°) that steepen down-dip and downlap onto smooth magmatic basement at maximum dips below 35° (Type II SDRs), and packages that downlap onto the footwalls of landward-dipping faults (Type I SDRs). A key implication of this work is that Type I geometries can develop within faulted magmatic crust rather than requiring extended continental crust, with consequences for the stratigraphic age models of some SDR-bearing margins.

Although several models produce maximum dips exceeding 40° near the edge of extended continental crust, dips most consistent with published seismic reflection data (commonly < 40°) arise under longer inter-eruption periods, low to moderate extrusion factors, longer subaerial flow lengths, and thermal anomalies below 100°C. Together these results support the use of SDRs as indicators

of subaerial conditions and suggest that asthenospheric thermal anomalies of 50–150°C persisted for several million years during the early basin-forming period.

Reproducing the full complexity of SDRs at volcanic margins will require extending beyond the 2D framework used here, most importantly to capture the temporal evolution and dissipation of plume-related thermal anomalies, temporally variable eruption histories, plate-reconstruction-based extension rates, and fully three-dimensional deformation, melt transport, and emplacement.

Appendix

Appendix A. Algorithm for Solving the Visco-elasto-plastic Stokes-Continuity Equations

The frictional-plastic failure model used in this work introduces a strong non-linearity where the visco-plastic viscosity $\eta_{vp,m}$ is dependent on deviatoric stress and deviatoric stress is in turn dependent on the visco-plastic viscosity as described in equations (9) and (10). This work uses the Picard iteration method to iteratively solve the non-linear Stokes-continuity equations on the staggered grid. The deviatoric stress tensor is updated on markers that are advected with the velocity field. Updating the deviatoric stress tensor components on markers requires accounting for stress changes from three sources:

1. deviatoric grid stress changes obtained on the Eulerian staggered grid by solving the non-linear Stokes-continuity equation with visco-elasto-plastic stress forecast,
2. deviatoric stress changes occurring at a subgrid scale on individual markers,
3. deviatoric stress changes associated with the rotation of markers during advection.

The following algorithm is used to solve equations (1), (2), and (7) using the visco-elasto-plastic constitutive law described in equations (4) and (5), and to update the deviatoric stress tensor components on markers for the three sources of stress changes.

For each time step t do:

1. Set model time step Δt equal to visco-elastic time step Δt_{ve} .
2. Calculate deviatoric stress components $\sigma'_{xx(i,j)_{p1}}$ and $\sigma'_{xy(i,j)_{b1}}$ and visco-plastic viscosity $\eta_{vp(i,j)_p}$ and $\eta_{vp(i,j)_b}$ on staggered grid by interpolating marker deviatoric stress components $\sigma'_{xx,m}$ and $\sigma'_{xy,m}$ and marker visco-plastic viscosity $\eta_{vp,m}$ using equation (C.1). The subscripts p and b refer to the pressure and basic grids, respectively.

3. Run the Picard iteration loop (Section [Appendix A.0.1](#)) to obtain new solutions for the x-component of velocity $v_{x(i,j)_{vx}}$, the y-component of velocity $v_{y(i,j)_{vy}}$ and pressure $P_{(i,j)_p}$. The subscripts v_x and v_y refer to the x-velocity and y-velocity staggered grids, respectively [29, 27].
4. Interpolate visco-plastic viscosity from basic grid $\eta_{vp(i,j)_b}$ to markers using a harmonic average (Section [Appendix A.0.3](#)).
5. Update model time step Δt ensuring that displacement is less than 50% of the staggered grid spacing based on current velocity solutions $v_{x(i,j)_{vx}}$ and $v_{y(i,j)_{vy}}$ and grid spacings $\Delta x_{(i,j)_{vx}}$ and $\Delta y_{(i,j)_{vy}}$.
6. Forecast deviatoric stresses $\sigma'_{xx(i,j)_{p2}}$ and $\sigma'_{xy(i,j)_{b2}}$ using equation (A.7) with updated deviatoric strain rates $\dot{\epsilon}'_{xx(i,j)_p}$ and $\dot{\epsilon}'_{xy(i,j)_b}$, updated visco-plastic viscosity $\eta_{vp(i,j)_p}$ and $\eta_{vp(i,j)_b}$ and updated model time step Δt .
7. Update grid deviatoric stress changes using the following equations:

$$\begin{aligned}\Delta\sigma'_{xx(i,j)_p} &= \sigma'_{xx(i,j)_{p2}} - \sigma'_{xx(i,j)_{p1}} \\ \Delta\sigma'_{xy(i,j)_b} &= \sigma'_{xy(i,j)_{b2}} - \sigma'_{xy(i,j)_{b1}}\end{aligned}\tag{A.1}$$

8. Update marker deviatoric stress components $\sigma'_{xx,m}$ and $\sigma'_{xy,m}$ for subgrid changes using subgrid stress diffusion (Section [Appendix A.0.4](#)).
9. Advect markers and calculate marker angular velocity ω_m using the 4th-order Runge-Kutta method and angular velocity on the basic grid $\omega_{(i,j)_b}$ given by:

$$\omega_{(i,j)_b} = 0.5 \left(\frac{(v_{y(i,j+1)_{vy}} - v_{y(i,j)_{vy}})}{\Delta x_{vy(j)}} - \frac{(v_{x(i+1,j)_{vx}} - v_{x(i,j)_{vx}})}{\Delta y_{vx(i)}} \right)\tag{A.2}$$

10. Update marker deviatoric stress components $\sigma'_{xx,m}$ and $\sigma'_{xy,m}$ for rotation using the following equations:

$$\begin{aligned}\sigma'_{xx,m} &= \sigma'^o_{xx,m} (\cos(\omega_m \Delta t)^2 - \sin(\omega_m \Delta t)^2) - \sigma'^o_{xy,m} \sin(2\omega_m \Delta t), \\ \sigma'_{xy,m} &= \sigma'^o_{xx,m} \sin(2\omega_m \Delta t) + \sigma'^o_{xy,m} \cos(2\omega_m \Delta t)\end{aligned}\tag{A.3}$$

where $\sigma'^o_{xx,m}$ and $\sigma'^o_{xy,m}$ are the deviatoric marker stress components before advection.

11. Advance model time t by Δt .

End time step loop

Appendix A.0.1. Picard Iteration Loop

While Picard criterion $R_{L2} > \text{Tolerance}$ and $N_{iterations} < N_{max}$ do:

- (1) Calculate the visco-plastic viscosity on pressure grid $\eta_{vp(i,j)_p}$ as a harmonic average of the visco-plastic viscosity $\eta_{vp(i,j)_b}$ on surrounding basic nodes:

$$\eta_{vp(i,j)_p} = \frac{4}{1/\eta_{vp(i,j)_b} + 1/\eta_{vp(i-1,j)_b} + 1/\eta_{vp(i,j-1)_b} + 1/\eta_{vp(i-1,j-1)_b}} \quad (\text{A.4})$$

- (2) Obtain new solution vector S^{new} and updated solutions for x-component of velocity $v_{x(i,j)_{vx}}$, y-component of velocity $v_{y(i,j)_{vy}}$ and pressure $P_{(i,j)_p}$ by solving equations (1), (2), and (7) using current $\eta_{vp(i,j)_b}$ and $\eta_{vp(i,j)_p}$.
- (3) Update the convergence criterion R_{L2} defined as:

$$R_{L2} = \frac{\sqrt{\sum_i (S_{v_{xy},i}^{new} - S_{v_{xy},i}^{old})^2}}{\sqrt{\sum_i (S_{v_{xy},i}^{new})^2}} \quad (\text{A.5})$$

where $S_{v_{xy}}^{old}$ is the old velocity solution vector and $S_{v_{xy}}^{new}$ is the new velocity solution vector.

- (4) Calculate deviatoric strain rate components $\dot{\epsilon}'_{xx(i,j)_p}$ and $\dot{\epsilon}'_{xy(i,j)_b}$ using updated velocity solutions $v_{x(i,j)_{vx}}$ and $v_{y(i,j)_{vy}}$ and the following equation:

$$\begin{aligned} \dot{\epsilon}'_{xx(i,j)_p} &= \frac{1}{2} \left(\frac{v_{x(i+1,j+1)_{vx}} - v_{x(i+1,j)_{vx}}}{\Delta x_{j_b}} - \frac{v_{y(i+1,j+1)_{vy}} - v_{y(i,j+1)_{vy}}}{\Delta y_{i_b}} \right) \quad \text{and} \\ \dot{\epsilon}'_{xy(i,j)_b} &= \frac{1}{2} \left(\frac{v_{x(i+1,j)_{vx}} - v_{x(i,j)_{vx}}}{\Delta y_{i_p}} + \frac{v_{y(i,j+1)_{vy}} - v_{y(i,j)_{vy}}}{\Delta x_{j_p}} \right), \end{aligned} \quad (\text{A.6})$$

- (5) Forecast deviatoric stresses $\sigma'_{xx(i,j)_{p2}}$ and $\sigma'_{xy(i,j)_{b2}}$ using updated deviatoric strain rates $\dot{\epsilon}'_{xx(i,j)_p}$ and $\dot{\epsilon}'_{xy(i,j)_b}$, updated visco-plastic viscosity $\eta_{vp(i,j)_p}$ and $\eta_{vp(i,j)_b}$ with the following equations:

$$\begin{aligned} \sigma'_{xx(i,j)_{p2}} &= 2\eta_{vp(i,j)_p} \dot{\epsilon}'_{xx(i,j)_p} \frac{\mu_{(i,j)_b} \Delta t}{\mu_{(i,j)_b} \Delta t + \eta_{vp(i,j)_p}} \\ &\quad + \sigma'_{xx(i,j)_{p1}} \frac{\eta_{vp(i,j)_p}}{\mu_{(i,j)_b} \Delta t + \eta_{vp(i,j)_p}} \\ \sigma'_{xy(i,j)_{b2}} &= 2\eta_{vp(i,j)_b} \dot{\epsilon}'_{xy(i,j)_b} \frac{\mu_{(i,j)_b} \Delta t}{\mu_{(i,j)_b} \Delta t + \eta_{vp(i,j)_b}} \\ &\quad + \sigma'_{xy(i,j)_{b1}} \frac{\eta_{vp(i,j)_b}}{\mu_{(i,j)_b} \Delta t + \eta_{vp(i,j)_b}} \end{aligned} \quad (\text{A.7})$$

- (6) Update visco-plastic viscosity $\eta_{vp(i,j)_b}$ for yielding and plastic failure flag $\chi_{(i,j)_b}$ on basic grid (Section [Appendix A.0.2](#)).

End Picard iteration loop

It was observed that the node-based plasticity scheme exhibited more stable convergence compared to marker-based methods, an observation consistent with the work of [30]. Furthermore, for the cases tested in this study, convergence criteria could be achieved if the model time step was sufficiently small. For the models presented in this work $R_{L2} = 10^{-3}$ was used as the Picard convergence criterion. See the Supplemental Materials Section 2 for additional details on the convergence behavior of the Picard iteration method used in this work.

Appendix A.0.2. Nodal Plasticity Update

For each basic node (i, j) do:

- (1) Calculate the second invariant of deviatoric stress $\sigma'_{II(i,j)_b}$ using updated deviatoric stresses $\sigma'_{xx(i,j)_{b2}}$ and $\sigma'_{xy(i,j)_{b2}}$ and the following equation:

$$\sigma'_{xx,avg(i,j)_{b2}} = \frac{1}{4} \left(\sigma'_{xx(i,j)_{p2}} + \sigma'_{xx(i+1,j)_{p2}} + \sigma'_{xx(i,j+1)_{p2}} + \sigma'_{xx(i+1,j+1)_{p2}} \right),$$

$$\sigma'_{II(i,j)_b} = \sqrt{\left(\left(\sigma'_{xy(i,j)_{b2}} \right)^2 + \left(\sigma'_{xx,avg(i,j)_{b2}} \right)^2 \right)} \quad (\text{A.8})$$

- (2) Compute the second invariant of deviatoric stress for purely elastic stress buildup $\sigma'_{II,elastic(i,j)_b}$ using the following equation:

$$\sigma'_{II,elastic(i,j)_b} = \frac{\mu_{(i,j)_b} \Delta t + \eta_{vp(i,j)_b}}{\eta_{vp(i,j)_b}} \sigma'_{II(i,j)_b} \quad (\text{A.9})$$

where $\mu_{(i,j)_b}$ is the shear modulus at basic nodes and Δt is the model time step.

- (3) Interpolate pressure from staggered pressure grid $P_{(i,j)_p}$ to basic pressure grid $P_{(i,j)_b}$:

$$P_{(i,j)_b} = \frac{1}{4} (P_{(i,j)_p} + P_{(i+1,j)_p} + P_{(i,j+1)_p} + P_{(i+1,j+1)_p}) \quad (\text{A.10})$$

- (4) Compute the Drucker-Prager yield stress on basic grid $\sigma_{yield(i,j)_b}$ using the following equation:

$$\sigma_{yield(i,j)} = \begin{cases} \sigma_{c(i,j)_b} \cos(\theta_{(i,j)_b}) \\ \quad + \sin(\theta_{(i,j)_b}) (P_{(i,j)_b} - P_f) & \text{for } P_{(i,j)_b} \geq P_f \\ \sigma_{c(i,j)_b} \cos(\theta_{(i,j)_b}) & \text{for } P_{(i,j)_b} < P_f \end{cases} \quad (\text{A.11})$$

where $\sigma_{yield(i,j)}$ is the second invariant of the deviatoric stress tensor at basic grid node at yield, $\sigma_{c(i,j)_b}$ is the cohesion, $\theta_{(i,j)_b}$ is the friction angle and P_f is the fluid pressure.

- (5) Update visco-plastic viscosity $\eta_{vp(i,j)_b}$ for plastic failure using the following equation:

$$\eta_{vp(i,j)_b} = \begin{cases} \eta_{creep(i,j)_b} & \text{for } \sigma'_{II,elastic(i,j)_b} \leq \sigma_{yield(i,j)_b} \\ \min(\eta_{yield(i,j)_b}, \eta_{creep(i,j)_b}) & \text{for } \sigma'_{II,elastic(i,j)_b} > \sigma_{yield(i,j)_b} \end{cases} \quad (\text{A.12})$$

where $\eta_{creep(i,j)_b}$ is the composite creep viscosity calculated on markers using equation (82) and $\eta_{yield(i,j)_b}$ is the visco-plastic viscosity at the yield state as given by:

$$\eta_{yield(i,j)_b} = \mu_{(i,j)_b} \Delta t \frac{\sigma_{yield(i,j)_b}}{\sigma'_{II,elastic(i,j)_b} - \sigma_{yield(i,j)_b}} \quad (\text{A.13})$$

- (6) Update the plastic yielding state on basic nodes $\chi_{(i,j)_b}$ using the following equation:

$$\chi_{(i,j)_b} = \begin{cases} 1 & \text{if } \sigma'_{II,elastic(i,j)_b} > \sigma_{yield(i,j)_b} \text{ and } \eta_{yield(i,j)_b} < \eta_{creep(i,j)_b} \\ 0 & \text{otherwise} \end{cases} \quad (\text{A.14})$$

where a value of 1 indicates that the basic node has undergone plastic failure and a value of 0 indicates that the basic node has not undergone plastic failure.

End basic node loop

Appendix A.0.3. Harmonic Viscosity Average for Yielding

For each marker m do:

- (1) Determine if plastic yielding has occurred for any basic node associated with the cell that contains marker m using the plastic failure criterion $\chi_{(i,j)_b}$ defined on the basic grid.
- (2) Initialize the visco-plastic viscosity of the marker $\eta_{vp,m}$ to the creep viscosity $\eta_{creep(m)}$.
- (3) If plastic yielding has occurred on a node associated with marker m calculate the visco-plastic viscosity of the marker $\eta_{vp,m}$ using a harmonic average limited by the flow viscosity $\eta_{creep,m}$ as follows:

$$\eta_{yield,m} = (\chi_{(UL)_b} w_{UL,m} + \chi_{(LL)_b} w_{LL,m} + \chi_{(UR)_b} w_{UR,m} + \chi_{(LR)_b} w_{LR,m}) \left(\frac{\chi_{(UL)_b} w_{UL,m}}{\eta_{vp(UL)_b}} + \frac{\chi_{(LL)_b} w_{LL,m}}{\eta_{vp(LL)_b}} + \frac{\chi_{(UR)_b} w_{UR,m}}{\eta_{vp(UR)_b}} + \frac{\chi_{(LR)_b} w_{LR,m}}{\eta_{vp(LR)_b}} \right)^{-1},$$

$$\eta_{vp,m} = \min(\eta_{yield,m}, \eta_{creep,m}) \quad (\text{A.15})$$

where $UL = (i_{ul,m}, j_{ul,m})$, $LL = (i_{ul,m} + 1, j_{ul,m})$, $UR = (i_{ul,m}, j_{ul,m} + 1)$, and $LR = (i_{ul,m} + 1, j_{ul,m} + 1)$ are the basic nodes indices associated with the cell containing marker m and $w_{UL,m}$, $w_{LL,m}$, $w_{UR,m}$, and $w_{LR,m}$ are the associated node-marker weights calculated using equation (C.3) and $i_{ul,m}$ and $j_{ul,m}$ are the indices of the upper-left basic grid node of the cell containing marker m .

End marker loop

Appendix A.0.4. Subgrid Stress Diffusion Steps

- (1) Calculate initial nodal stresses $\sigma'_{xx,nodal,m}$ and $\sigma'_{xy,nodal,m}$ for each marker m by interpolating $\sigma'_{xx(i,j)_{p1}}$ and $\sigma'_{xy(i,j)_{b1}}$ to markers using equation (C.2).
- (2) Calculate the total subgrid stress differences $\Delta\sigma'_{xx,sgt,m}$ and $\Delta\sigma'_{xy,sgt,m}$ for each marker m using the following equations:

$$\begin{aligned}\Delta\sigma'_{xx,sgt,m} &= \sigma'_{xx,nodal,m} - \sigma'_{xx,m} \\ \Delta\sigma'_{xy,sgt,m} &= \sigma'_{xy,nodal,m} - \sigma'_{xy,m}.\end{aligned}\tag{A.16}$$

- (3) Calculate the relaxed subgrid stress differences $\Delta\sigma'_{xx,sg,m}$ and $\Delta\sigma'_{xy,sg,m}$ using the following equations:

$$\begin{aligned}\Delta\sigma'_{xx,sg,m} &= \Delta\sigma'_{xx,sgt,m} \left(1 - \exp\left(-D_{sg} \frac{\Delta t}{t_{max}}\right)\right) \\ \Delta\sigma'_{xy,sg,m} &= \Delta\sigma'_{xy,sgt,m} \left(1 - \exp\left(-D_{sg} \frac{\Delta t}{t_{max}}\right)\right)\end{aligned}\tag{A.17}$$

where D_{sg} is the subgrid diffusion coefficient and t_{max} is the Maxwell time defined as:

$$t_{max} = \frac{\eta_{vp,m}}{\mu_m}.\tag{A.18}$$

- (4) Calculate the relaxed subgrid stress differences on the staggered grids $\Delta\sigma'_{xx,sg(i,j)_p}$ and $\Delta\sigma'_{xy,sg(i,j)_b}$ by interpolating the relaxed subgrid stress differences from markers $\Delta\sigma'_{xx,sg,m}$ and $\Delta\sigma'_{xy,sg,m}$ to the staggered grid using the bilinear interpolation method described in equation (C.1).
- (5) Calculate the remaining stress changes on staggered grid nodes by removing the interpolated relaxed subgrid stress differences using the following equations:

$$\begin{aligned}\Delta\sigma'_{xx,vepr(i,j)_p} &= \Delta\sigma'_{xx(i,j)_p} - \Delta\sigma'_{xx,sg(i,j)_p}, \\ \Delta\sigma'_{xy,vepr(i,j)_b} &= \Delta\sigma'_{xy(i,j)_b} - \Delta\sigma'_{xy,sg(i,j)_b}\end{aligned}\tag{A.19}$$

- (6) Calculate the remaining visco-elasto-plastic stress changes $\Delta\sigma'_{xx,vepr,m}$ and $\Delta\sigma'_{xy,vepr,m}$ for each marker m by interpolating the remaining stress changes on staggered grid nodes $\Delta\sigma'_{xx,vepr(i,j)_p}$ and $\Delta\sigma'_{xy,vepr(i,j)_b}$ to markers using equation (C.2).

- (7) Update marker deviatoric stress components $\sigma'_{xx,m}$ and $\sigma'_{xy,m}$ using the following equations:

$$\begin{aligned}\sigma'_{xx,m} &= \sigma'^{lo}_{xx,m} + \Delta\sigma'_{xx,sg,m} + \Delta\sigma'_{xx,vepr,m}, \\ \sigma'_{xy,m} &= \sigma'^{lo}_{xy,m} + \Delta\sigma'_{xy,sg,m} + \Delta\sigma'_{xy,vepr,m}\end{aligned}\tag{A.20}$$

where $\sigma'^{lo}_{xx,m}$ and $\sigma'^{lo}_{xy,m}$ are the previous marker deviatoric stress components.

Appendix B. Algorithm for Solving the Heat Advection-Diffusion Equation

The Eulerian-Lagrangian marker-in-cell approach from [29] is used to solve equation (17) whereby markers are advected using a 4th-order Runge-Kutta scheme and marker temperature T_m is updated using the conductive temperature solution on Eulerian grid nodes and temperature changes associated with subgrid thermal diffusion as described in the following steps:

For each time step do:

1. Interpolate current marker temperature T_m to basic grid nodes $T_{(i,j)_b}$ using equation (C.1).
2. Calculate purely conductive temperature solution over time interval Δt on basic grid $T_{solu(i,j)_b}$ by solving equation (17) without advective terms and using current grid temperature $T_{(i,j)_b}$ as the initial condition. An adaptive time-stepping scheme is applied whereby the time step is reduced if the change in temperature exceeds a threshold of 70°C.
3. Calculate the conductive temperature change on the Eulerian grid using the following equation:

$$\Delta T_{(i,j)_b} = T_{solu(i,j)_b} - T_{(i,j)_b}\tag{B.1}$$

4. Update marker temperature T_m for subgrid temperature changes using subgrid thermal diffusion (Section Appendix B.0.1).
5. Advect markers using a 4th-order Runge-Kutta scheme.
6. Advance model time t by Δt .

End time step loop

Appendix B.0.1. Subgrid Thermal Diffusion Steps

- (1) Calculate the initial nodal temperature on markers $T_{nodal,m}$ by interpolating $T_{(i,j)_b}$ back to markers using equation (C.2).
- (2) Calculate the total subgrid temperature difference between interpolated nodal temperatures and current marker temperatures using the following equation:

$$\Delta T_{sgt,m} = T_{nodal,m} - T_m \quad (\text{B.2})$$

- (3) Calculate the relaxed subgrid temperature difference on markers using the following equation:

$$\Delta T_{sg,m} = \Delta T_{sgt,m} \left(1 - \exp \left(-D_{sg} \frac{\Delta t}{t_{sg,m}} \right) \right) \quad (\text{B.3})$$

where D_{sg} is the subgrid diffusion coefficient, Δt is the time step and t_{sg} is the thermal diffusion time scale given by:

$$t_{sg,m} = \frac{\rho_m C_{p,m}}{k_m \left(\frac{2}{\Delta x^2} + \frac{2}{\Delta y^2} \right)} \quad (\text{B.4})$$

where ρ_m is the density of the marker, $C_{p,m}$ is the specific heat capacity of the marker, k_m is the thermal conductivity of the marker, Δx and Δy are the grid spacings in the x and y directions, respectively.

- (4) Calculate the relaxed subgrid temperature difference on Eulerian grid $\Delta T_{sg(i,j)_b}$ by interpolating $\Delta T_{sg,m}$ using equation (C.1).
- (5) Calculate the remaining temperature change $\Delta T_{r(i,j)_b}$ on Eulerian nodes by removing the relaxed subgrid temperature difference using the following equation:

$$\Delta T_{r(i,j)_b} = \Delta T_{(i,j)_b} - \Delta T_{sg(i,j)_b} \quad (\text{B.5})$$

- (6) Interpolate $\Delta T_{r(i,j)_b}$ to markers using equation (C.2) to obtain $\Delta T_{r,m}$, which is the nodal temperature change with subgrid effects removed.
- (7) Update marker temperature T_m taking into account both subgrid effects and the remaining temperature change from the basic grid using the following equation:

$$T_m = T_m + \Delta T_{sg,m} + \Delta T_{r,m} \quad (\text{B.6})$$

Appendix C. Marker-grid Interpolation

Marker-in-cell methods require tensor and scalar parameters to be interpolated back and forth between the Eulerian grid and the Lagrangian markers. We

use the first-order bilinear interpolation scheme similar to [29] for interpolating marker information to Eulerian grid nodes:

$$S_{i,j} = \frac{\sum_m S_m w_{m(i,j)}}{\sum_m w_{m(i,j)}}, \quad w_{m(i,j)} = \left(1 - \frac{\Delta x_m}{\Delta x}\right) \left(1 - \frac{\Delta y_m}{\Delta y}\right) \quad (\text{C.1})$$

where $S(i, j)$ is the interpolated parameter value at the Eulerian grid node (i, j) , S_m is the parameter value at the marker m located within one of four surrounding grid cells, $w_{m(i,j)}$ is the weight of the marker m at the Eulerian grid node (i, j) , Δx_m and Δy_m are the distances between the marker m and the Eulerian grid node (i, j) in the x and y directions, respectively, and Δx and Δy are the grid spacings in the x and y directions, respectively. With our modeling approach the index i corresponds to the vertical y -axis that increases with depth and the index j corresponds to the x -axis. Two types of interpolation are used based on the size of the search radius used in (C.1): (1) inclusive interpolation where markers within the search radius equal to Δx and Δy are used in the interpolation, and (2) exclusive interpolation where markers within the search radius equal to $0.5\Delta x$ and $0.5\Delta y$ are used in the interpolation.

A similar first-order bilinear scheme is used for interpolating parameter values from Eulerian grid nodes to Lagrangian markers:

$$S_m = S_{(i_{ul,m}, j_{ul,m})} w_{UL,m} + S_{(i_{ul,m}, j_{ul,m}+1)} w_{UR,m} + S_{(i_{ul,m}+1, j_{ul,m})} w_{LL,m} + S_{(i_{ul,m}+1, j_{ul,m}+1)} w_{LR,m} \quad (\text{C.2})$$

where S_m is the interpolated parameter value at the Lagrangian marker m , $i_{ul,m}$ is the y -index of the upper left grid node for marker m , $j_{ul,m}$ is the x -index of the upper-left grid node for marker m , $S_{i_{ul,m}, j_{ul,m}}$, $S_{i_{ul,m}, j_{ul,m}+1}$, $S_{i_{ul,m}+1, j_{ul,m}}$, and $S_{i_{ul,m}+1, j_{ul,m}+1}$ are the parameter values at the four surrounding Eulerian grid nodes, and $w_{UL,m}$, $w_{LL,m}$, $w_{UR,m}$, and $w_{LR,m}$ are the weights for the upper-left, lower-left, upper-right, and lower-right grid nodes, respectively. The weights of the four surrounding Eulerian grid nodes for marker m are calculated using the following equations:

$$\begin{aligned} w_{UL,m} &= (1.0 - \Delta x'_{UL})(1.0 - \Delta y'_{UL}) \\ w_{LL,m} &= (1.0 - \Delta x'_{UL})\Delta y'_{UL} \\ w_{UR,m} &= \Delta x'_{UL}(1.0 - \Delta y'_{UL}) \\ w_{LR,m} &= \Delta x'_{UL}\Delta y'_{UL} \end{aligned} \quad (\text{C.3})$$

where $\Delta x'_{UL}$ and $\Delta y'_{UL}$ are the normalized distances in the x and y directions, respectively, between the marker and the upper-left node of the cell containing the marker as defined by the following equations:

$$\begin{aligned} \Delta x'_{UL} &= \frac{(x_m - x_{b,j_{ul,m}})}{\Delta x_{b,j_{ul,m}}} \\ \Delta y'_{UL} &= \frac{(y_m - y_{b,i_{ul,m}})}{\Delta y_{b,i_{ul,m}}} \end{aligned} \quad (\text{C.4})$$

where x_m and y_m are the coordinates of the marker m , $x_{b,j_{ul,m}}$ and $y_{b,i_{ul,m}}$ are the coordinates of the upper-left basic grid node of the cell containing the marker m , and $\Delta x_{b,j_{ul,m}}$ and $\Delta y_{b,i_{ul,m}}$ are the basic grid spacings in the x and y directions, respectively. The upper-left indices $i_{ul,m}$ and $j_{ul,m}$ of the cell containing the marker and weights $w_{UL,m}$, $w_{LL,m}$, $w_{UR,m}$, and $w_{LR,m}$ are pre-computed for each marker using bisection to improve the efficiency of the interpolation process. These pre-computed weights and indices can also be used to optimize (C.1) as described in the following algorithm:

1. Calculate the numerator, denoted with subscript n , and denominator, denoted with subscript d , summations from (C.1) by evaluating weights associated with each marker m :

For Each Marker m Do:

$$\begin{aligned}
S_{avg,d(i_{ul,m},j_{ul,m})} &= S_{avg,d(i_{ul,m},j_{ul,m})} + w_{UL,m} \\
S_{avg,n(i_{ul,m},j_{ul,m})} &= S_{avg,n(i_{ul,m},j_{ul,m})} + S_m w_{UL,m} \\
S_{avg,d(i_{ul,m},j_{ul,m}+1)} &= S_{avg,d(i_{ul,m},j_{ul,m}+1)} + w_{UR,m} \\
S_{avg,n(i_{ul,m},j_{ul,m}+1)} &= S_{avg,n(i_{ul,m},j_{ul,m}+1)} + S_m w_{UR,m} \\
S_{avg,d(i_{ul,m}+1,j_{ul,m})} &= S_{avg,d(i_{ul,m}+1,j_{ul,m})} + w_{LL,m} \\
S_{avg,n(i_{ul,m}+1,j_{ul,m})} &= S_{avg,n(i_{ul,m}+1,j_{ul,m})} + S_m w_{LL,m} \\
S_{avg,d(i_{ul,m}+1,j_{ul,m}+1)} &= S_{avg,d(i_{ul,m}+1,j_{ul,m}+1)} + w_{LR,m} \\
S_{avg,n(i_{ul,m}+1,j_{ul,m}+1)} &= S_{avg,n(i_{ul,m}+1,j_{ul,m}+1)} + S_m w_{LR,m}
\end{aligned} \tag{C.5}$$

End marker Loop

2. Calculate bilinear average values at each grid node (i, j) using the numerator and denominator summations:

$$S_{i,j} = \frac{S_{avg,n}(i,j)}{S_{avg,d}(i,j)} \tag{C.6}$$

Appendix D. Non-Linear Creep Viscosity Update Steps

For each marker m do:

1. Initialize the deviatoric stress invariant $\sigma'_{II,A,m}$ using equation (11) with current marker deviatoric shear stress $\sigma'_{xy,m}$ and deviatoric normal stress $\sigma'_{xx,m}$.
2. Calculate effective creep viscosity $\eta_{creep,m}$ using equation (82) with deviatoric stress invariant $\sigma'_{II,A,m}$ and $\dot{\epsilon}_{eff}$ calculated with current marker temperature T_m , pressure P_m and deviatoric stress invariant $\sigma'_{II,A,m}$.
3. Forecast the deviatoric stress invariant $\sigma'_{II,B,m}$ using equation (86) with current creep viscosity $\eta_{creep,m}$.

While $N_{iterations} < N_{max}$ **and** $\Delta\sigma > 1$ **do**:

- (1) Calculate average stress invariant $\sigma'_{II,avg,m}$ using the following equation:

$$\sigma'_{II,avg,m} = \frac{1}{2} (\sigma'_{II,A,m} + \sigma'_{II,B,m}) \quad (D.1)$$

- (2) Calculate effective creep viscosity $\eta_{creep,m}$ using equation (82) with deviatoric stress invariant $\sigma'_{II,avg,m}$ and ϵ_{eff} calculated with marker temperature T_m , pressure P_m and deviatoric stress invariant $\sigma'_{II,avg,m}$.
- (3) Update the forecasted visco-elastic stress invariant $\sigma'_{II,fc,m}$ using equation (86) with $\eta_{creep,m}$ and current marker deviatoric shear stress $\sigma'_{xy,m}$ and deviatoric normal stress $\sigma'_{xx,m}$.
- (4) Update bisection limits:

if $(\sigma'_{II,B,m} > \sigma'_{II,A,m})$ **and** $(\sigma'_{II,fc,m} > \sigma'_{II,avg,m})$ **or**
 $(\sigma'_{II,B,m} < \sigma'_{II,A,m})$ **and** $(\sigma'_{II,fc,m} < \sigma'_{II,avg,m})$ **then**
 $\sigma'_{II,A,m} = \sigma'_{II,avg,m}$
else
 $\sigma'_{II,B,m} = \sigma'_{II,avg,m}$
end if
- (5) Update the stress change $\Delta\sigma$ using the following equation:

$$\Delta\sigma = \sigma'_{II,B,m} - \sigma'_{II,A,m} \quad (D.2)$$

End while loop

4. Initialize visco-plastic marker viscosity for time step by copying updated marker creep viscosity $\eta_{creep,m}$ to $\eta_{vp(m)}$

End marker loop

Data Availability Statement

A snapshot of the Julia package EarthBox.jl (v0.1.6) used to generate the models from this work and input files for all cases are archived on Zenodo at <https://doi.org/10.5281/zenodo.20785566> [28]. And animations associated with all model cases in Table D.7 are archived on Zenodo at <https://doi.org/10.5281/zenodo.20788288> [52].

The code is hosted on GitHub at <https://github.com/geonomyllabs/EarthBox.jl>.

Declaration of Competing Interest

The author declares that there are no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Tables

Parameter	Symbol	Units	Value
Subaerial Transport coefficient	C_{sa}	-	$10^{-4}, 10^{-2}$
Subaerial slope diffusivity	κ_{sa}	m^2/yr	0.25
Submarine slope diffusivity	κ_{sm}	m^2/yr	100
Submarine diffusion decay depth	λ_{sm}	m	1000
Precipitation rate	R_{precip}	m/yr	1.0
Pelagic sedimentation rate	$R_{pelagic}$	mm/yr	0.0
Initial porosity of sediment at mudline	ϕ_o	-	0.4
Sediment porosity decay depth	λ_{comp}	m	2500

Table D.1: Sediment transport model parameters.

Parameter	Symbol	Units	Value
Extraction efficiency factor	$f_{extraction}$	-	0.99
Characteristic injection width	W_{inj}^{char}	km	10
Injection height limit	$H_{inj,limit}$	m	1000
Minimum extrusion efficiency factor	$f_{v,min}$	-	0.06
Maximum extrusion efficiency factor	$f_{v,max}$	-	0.3 0.4 0.5 0.6
Minimum reference characteristic magmatic crust thickness	$H_{mc,min}$	m	6000
Maximum reference characteristic magmatic crust thickness	$H_{mc,max}$	m	7500
inter-eruption period	Δt_{erupt}	yr	50,000 100,000 200,000 400,000
Characteristic length of submarine flows	L_{sm}	km	2
Characteristic length of subaerial flows	L_{sa}	km	20 40 60 80
Residual thickness of submarine flows	$H_{res,sm}$	m	30
Residual thickness of subaerial flows	$H_{res,sa}$	m	30
Minimum width of eruption domain	$W_{e,min}$	m	2500
Maximum width of eruption domain	$W_{e,max}$	m	2500

Table D.2: Melt extraction, emplacement and extrusion model parameters. Bold font indicates a parameter used in case 0, the base case model.

Parameter	Symbol	Units	Wet Quartzite ^a	Wet Anorthite ^b	Dry Anorthite ^b	Dry Olivine ^{c,d}
Dislocation Creep						
Tensor Invariant Pre-exponential Factor	A_{dis}	$s^{-1}MPa^{-n}$	8.57e-4	3.57e+3	2.26e+13	6.51e+5
Stress Exponent	n_{dis}		4	3	3	3.5
Activation Energy	E_{dis}	kJ/mol	223	345	641	530
Activation Volume	V_{dis}	J/MPa/mol	0	38	24	13
Diffusion Creep						
Tensor Invariant Pre-exponential Factor	A_{dif}	$s^{-1}MPa^{-1}$				1.04e-2
Diffusion Creep Activation Energy	E_{dif}	kJ/mol				375
Diffusion Creep Activation Volume	V_{dif}	J/MPa/mol				6
Peierls Creep						
Tensor Invariant Pre-exponential Factor	A_{pei}	$s^{-m1}MPa^{-m2}$				2.5e+7
Stress Exponent $m1$	$m1_{pei}$					1
Stress Exponent $m2$	$m2_{pei}$					2
Peierls Stress	σ_{pei}	MPa				9100

Table D.3: Viscous flow law parameters. Grain size is assumed to be 6 mm. Sources: a: [57], b: [58], c: [59], d: [60]

Parameter	Symbol	Units	Value
Initial reference friction angle	θ_m^{io}	degrees	30 ^a
Final reference friction angle	θ_m^{if}	degrees	7 ^b
Initial reference cohesion	$\sigma_{c,m}^o$	MPa	20 ^a
Final reference cohesion	$\sigma_{c,m}^f$	MPa	10 ^a
Initial reference plastic strain	ϵ_m^o	-	0
Final reference plastic strain	ϵ_m^f	-	1.0

Table D.4: Plastic failure parameters for markers. Sources: a: [16], b: [50]

Parameter	Symbol	Units	Value
Maximum melt damage factor	$f_{damage,max}$	-	1, 10
Width of melt-damage zone	W_{damage}	m	5000
Intermediate damage probability	p_{damage}^{inter}	-	0.1
Maximum damage probability	p_{damage}^{max}	-	0.75, 1.0
Minimum reference magmatic crust height	$H_{mc,min}$	m	750
Intermediate reference magmatic crust height	$H_{mc,inter}$	m	2000
Maximum reference magmatic crust height	$H_{mc,max}$	m	3000

Table D.5: Melt damage model parameters. Bold font indicates a parameter used in case 0, the base case model.

	Reference Density (kg/m ³)	Thermal Expansivity (1/K)	Compressibility (1/Pa)	Shear Modulus (Pa)	Reference Heat Capacity C_{p20} (J/kg/K)	Reference Thermal Conductivity $k_{20^\circ C}$ (W/m/K)	Radiogenic Heat Production (W/m ³)	Flow Law
Sticky Air ^a	1.0 ^a	0.0	0.0	1.0e+20 ^a	100.0 ^a	200.0 ^a	0.0	Isoviscous: 10 ¹⁸ Pa · s
Clastic Sediment ^b	2710 ^b	3e-05	1.8e-11 ^d	3.6e+10 ^d	890 ^b	2.3 ^b	1.4e-6 ^b	Wet Quartzite
Felsic Continental Crust	2860 ^c	3e-05	1.8e-11 ^d	3.6e+10 ^d	800 ^b	2.5 ^b	1.8e-6 ^b	Wet Quartzite
Mafic Continental Crust	2900 ^c	3e-05	1.6e-11 ^c	4e+10	800 ^b	2.5 ^b	0.5e-6 ^b	Wet Anorthite
Ultramafic Continental Lithosphere	3291 ^d	3e-05	7.9e-12 ^e	6.6e+10 ^e	800 ^b	4 ^b	0	Dry Olivine
Ultramafic Asthenosphere	3311 ^d	3e-05	7.9e-12 ^e	6.6e+10 ^e	800 ^b	4 ^b	0	Dry Olivine
Gabbroic Crust	2900 ^d	3e-05	1.6e-11 ^d	4e+10	800 ^b	2.5 ^b	0	Wet Anorthite
Gabbroic Magma	2750 ^f	3e-05	1.6e-11 ^d	1e+11	800 ^b	2.5 ^b	0	Isoviscous
Fractionated Gabbroic Crust	2900 ^d	3e-05	1.6e-11 ^d	4e+10	800 ^b	2.5 ^b	0	Wet or Dry Anorthite
Fractionated Gabbroic Magma	2750 ^f	3e-05	1.6e-11 ^d	1e+11	800 ^b	2.5 ^b	0	Isoviscous

Table D.6: Material properties. Sources: a: [29], b: [34], c: [20], d: [61], e: [35], f: [42]

Case	v_{erg} (cm/yr)	Magma- Crust Rheology	$f_{damage,max}$	f_0,max	ΔT (°C)	Δt_e (kyr)	L_{sa} (km)	C_{sa}	f_{dike}	Ridge Built Offset	Listric Faulting	Triangular Hess at Ridge	Ridge Mobility	Ridge Deformation Stability	Lava Package Symmetry	Gabbroic Accretion Symmetry	Max. Lava Subsidence Thickness (km)	On-Axis Gabbroic Thickness (km)	Off-Axis Gabbroic Thickness (km)	Max. Dip Top-Lapping Power (°)	Dominant Up-dip Geometry	Dominant Down-dip Geometry	Gabbroic Ductile Flowing Extent
0	2	wet	10	1	100	50	0.5	40	10 ⁻⁴	0	<0.5	Y	N	fixed	S	A-S	7.3	9.7	5.2-6.4	13-53	DC & CD	K	M
1	2	wet	10	1	100	50	0.5	40	10 ⁻⁴	0	<0.5	Y	N	mobile	S	A-S	8.6	8.6	4.2-8.2	33-54	DC & CD	K	H (left) L (right)
2	2	wet	10	1	100	100	0.5	40	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	7.7	9.5	5.7-6.5	21-51	DC & CD	K	M
3	2	wet	10	1	100	400	0.5	40	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	7.4	8.7	4.9-6.9	14-37	DC & CD	K	M
4	2	wet	10	1	25	200	0.5	40	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	3.2	4.4	3.6-4.8	14-28	DC & CD	SDC	L
5	2	wet	10	1	50	200	0.5	40	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	4.7	5.9	3.6-4.8	21-41	DC & CD	SDC	L
6	2	wet	10	1	150	200	0.5	40	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	9.8	15.0	7.4-9.1	12-46	DC & CD	SDC & UC	H
7	2	wet	10	1	100	200	0.5	20	10 ⁻⁴	0	<0.5-1	N	N	mobile	A	A	6.9	10.4	4.8-6.9	11-55	DC & CD	SDC	M to H
8	2	wet	10	1	100	200	0.5	20	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	7.4	8.4	6.2-6.3	11-34	DC & CD	SDC	M
9	2	wet	10	1	100	200	0.5	80	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	7.8	7.6	5.9-6.3	12-28	DC & CD	SDC	L to M
10	2	wet	10	1	100	200	0.3	40	10 ⁻⁴	0	<0.5	N	N	mobile	A-S	A-S	6.4	10.8	8.1-10.0	1-30	DC & CD	SDC & UC	L to M
11	2	wet	10	1	100	200	0.4	40	10 ⁻⁴	0	<0.5	N	N	mobile	A-S	A-S	6.6	10.1	6.9-9.0	8-37	DC & CD	SDC & UC	L to M
12	2	wet	10	1	100	200	0.6	40	10 ⁻⁴	0	0.5	N	N	fixed	S	S	7.7	7.7	4.4-5.6	21-51	DC & CD	SDC	M
13	2	wet	10	1	100	200	0.8	40	10 ⁻²	0	<0.5	N	N	fixed	S	S	9.5	9.5	6.1-6.7	12-58	DC & CD	K	M
14	2	wet	10	1	100	200	0.2	40	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	7.0	7.9	6.1-6.7	22-58	DC & CD	K	L
15	2	wet	10	1	100	200	0.5	40	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	7.7	9.8	4.6-6.3	11-49	DC & CD	SDC	M to H
16	2	wet	10	0.75	100	200	0.5	40	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	7.6	9.8	4.7-5.9	12-50	DC & CD	SDC	M
17	2	wet	1	1	100	50	0.5	40	10 ⁻⁴	0	2	Y	N	mobile	A	A	10.5	8.7	2.9-11.4	2-30	DC & F	SDC	H
18	2	wet	1	1	100	100	0.5	40	10 ⁻⁴	0	2	Y	N	mobile	A	A	10.1	8.0	3.9-8.8	13-31	DC & F	SDC	H (left) L (right)
19	2	wet	1	1	100	200	0.5	40	10 ⁻⁴	0	<0.5	N	Y	fixed	A-S	A-S	8.8	9.5	3.6-7.7	10-39	DC & CD	SDC	H
20	2	wet	1	1	100	200	0.5	20	10 ⁻⁴	0	2	Y	Y	mobile	A-S	A-S	9.2	10.0	2.8-9.3	12-19	DC & F	SDC	H
21	2	wet	1	1	100	200	0.5	40	10 ⁻⁴	0	<0.5	N	Y	fixed	S	S	9.4	10.1	4.9-7.9	18-36	DC	SDC	H
22	2	wet	1	1	100	200	0.5	60	10 ⁻⁴	0	1	N	Y	fixed	S	S	9.1	8.4	4.2-7.7	15-24	DC	SDC	M
23	2	wet	1	1	25	200	0.5	40	10 ⁻⁴	0	2	N	Y	mobile	S	S	4.1	5.3	2.5-5.2	8-35	DC & F	SDC	L
24	2	wet	1	1	50	200	0.5	40	10 ⁻⁴	0	2	N	Y	fixed	S	S	5.5	6.2	1.7-6	9-22	DC & F	SDC	L
25	2	wet	1	1	150	200	0.5	40	10 ⁻⁴	0	3	N	Y	fixed	S	S	11.8	16.1	4.5-11.1	10-34	DC & CD	SDC	H
26	2	wet	1	1	100	200	0.3	40	10 ⁻⁴	0	2	N	Y	fixed	A	A	6.7	10.9	6.8-8.9	9-22	DC & CD	SDC	M
27	2	wet	1	1	100	200	0.4	40	10 ⁻⁴	0	2	N	Y	fixed	S	S	8.3	10.8	6-7.9	10-30	DC & CD	SDC	M to H
28	2	wet	1	1	100	200	0.6	40	10 ⁻⁴	0	3	N	Y	mobile	A	A	9.4	8.8	2.8-7.3	15-39	DC & CD	SDC	M to H
29	2	wet	1	1	100	200	0.5	40	10 ⁻⁴	0	2	N	Y	fixed	S	S	8.4	9.1	4.4-7.6	11-35	DC & CD	SDC	L to M
30	2	wet	1	1	100	400	0.5	40	10 ⁻⁴	0	3	N	Y	fixed	S	S	8.4	9.1	4.4-7.6	11-35	DC & CD	SDC	L to M
31	1	wet	10	1	150	200	0.5	40	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	9.3	12.2	6.0-7.6	22-48	DC & CD	K	M to H
32	1	wet	10	1	100	400	0.5	40	10 ⁻⁴	0	0.5	N	N	fixed	S	S	7.1	7.7	5.5-6.3	22-48	DC & CD	SDC	L
33	1	wet	10	1	100	200	0.5	40	10 ⁻²	0	<0.5	N	N	fixed	S	S	7.1	6.5	5.8-7.2	43-50	DC & CD	K	L
34	1	wet	10	1	150	400	0.5	40	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	10.3	13.6	7.3-9.1	22-37	DC & CD	SDC	H
35	2	wet	10	1	150	200	0.5	60	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	10.5	13.6	7.4-8.7	5-32	DC & CD	SDC & UC	H
36	2	wet	10	1	150	200	0.5	80	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	10.9	12.5	7.2-8.3	8-24	DC & CD	SDC	H
37	2	wet	10	1	150	200	0.3	40	10 ⁻⁴	0	<0.5	N	N	mobile	A-S	A-S	6.9	16.5	10.3-14.3	1-22	DC & CD	SDC & UC	H
38	2	wet	10	1	150	200	0.6	40	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	11.9	13.6	4.1-7.0	20-65	DC & CD	SDC	H
39	1	wet	10	1	150	400	0.5	80	10 ⁻⁴	0	0.5	N	N	fixed	S	S	9.1	10.8	6.4-9.4	18-30	DC & CD	SDC	M
40	1	wet	10	1	150	400	0.5	80	10 ⁻²	0	<0.5	N	N	fixed	S	S	9.9	10.8	8-9.6	50-51	DC & CD	K	L to M
41	1	wet	10	1	100	400	0.5	40	10 ⁻²	0	<0.5	N	N	fixed	S	S	8.3	7.0	6.3-7.0	38-52	DC & CD	K	L to M
42	1	wet	10	1	100	400	0.5	80	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	7.1	7.3	6.1-6.9	19-25	DC & CD	SDC	L
43	2	dry	10	1	100	200	0.5	40	10 ⁻⁴	0	<0.5	N	N	fixed	S	S	7.6	8.3	5.1-7.6	24-53	DC & CD	K	L

Table D.7: Variable input parameters and key results for 44 model cases. Input parameters: v_{ext} , full extension velocity; $f_{damage,max}$, maximum melt-damage factor (1 = no damage); P_{damage}^{max} , maximum melt-damage probability; ΔT , thermal anomaly associated with the mantle plume; Δt_e , inter-eruption period; f_0,max , maximum extrusion-efficiency factor; L_{sa} , characteristic length of subaerial flows; C_{sa} , subaerial transport coefficient; f_{dike} , volume fraction of melt-damaged markers filled with dike fluid. Lava-flow package geometry abbreviations: DC, downward concave; UC, upward concave; SDC, smooth downward concave; CD, constant dip; F, faulted; V, variable; S, symmetric; A, asymmetric. Extent of ductile folding in gabbroic crust: H, high; M, moderate; L, low. Other abbreviations: N, no; Y, yes; (left), left conjugate; (right), right conjugate.

Figures

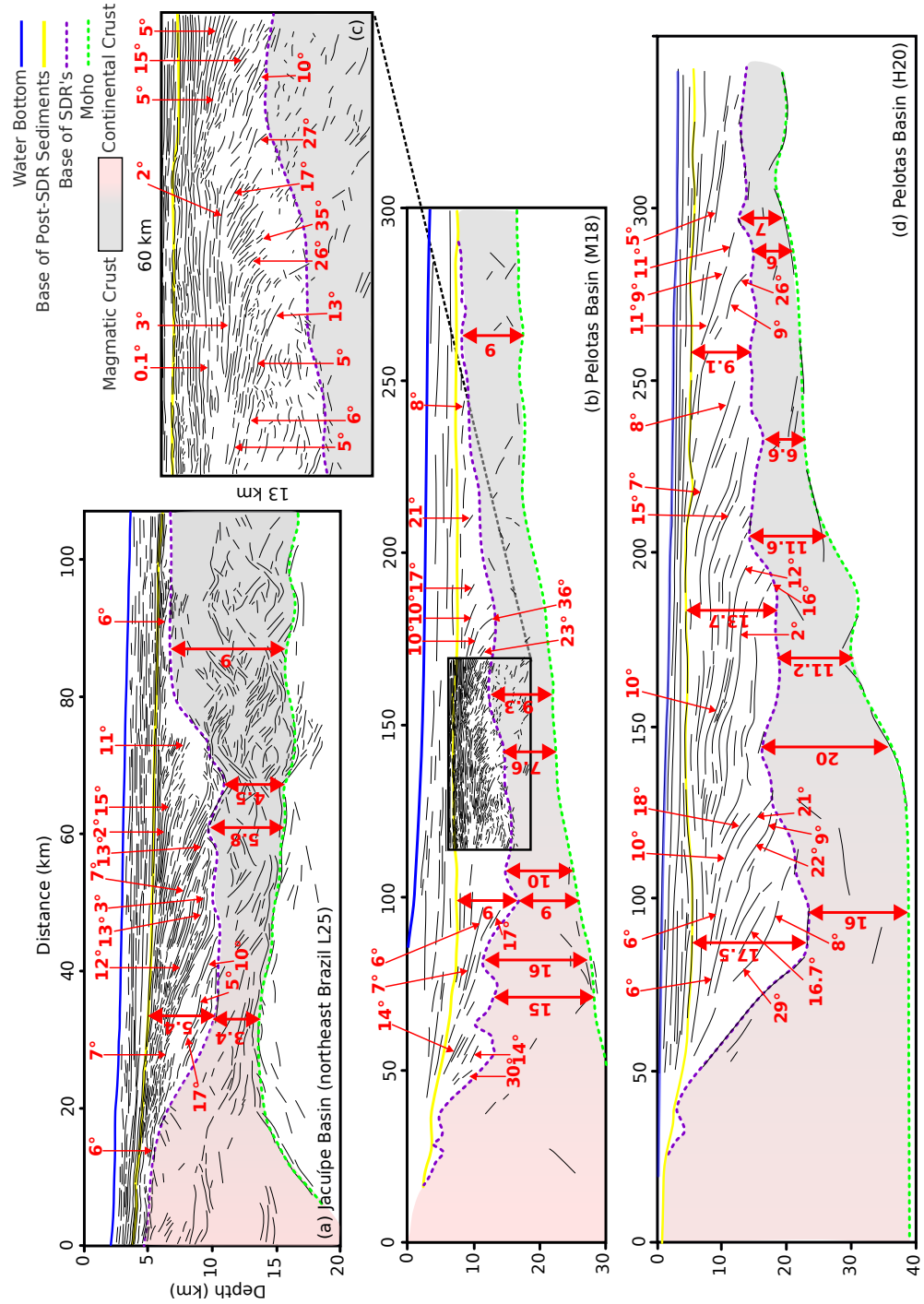


Figure D.1: Seaward-dipping reflector (SDR) geometry from offshore Brazil based on published seismic reflection images from (a) Jacuípe Basin and (b,c,d) Pelotas Basin [12, 9, 10]. Abbreviations shown in seismic cross sections refer to the source of the seismic reflection images and are defined as follows: L25, [12]; M18, [9]; H20, [10]. Thin black lines denote dominant seismic reflectors as interpreted from the seismic reflection images. Interpreted horizons are shown for the water bottom (blue), base of post-SDR sediments (yellow), base of SDRs (purple) and Moho (green). Crust dominated by continental felsic and mafic material is shown in light red, and crust dominated by gabbroic magmatic material is shown in light gray. The gradient between light red and gray indicates uncertainty in the transition between continental and gabbroic crust. Spot measurements of reflector dip in degrees are denoted with thin red arrows. Measurements of magmatic crust and total lava-flow package thicknesses in km are denoted with thick red double arrows.

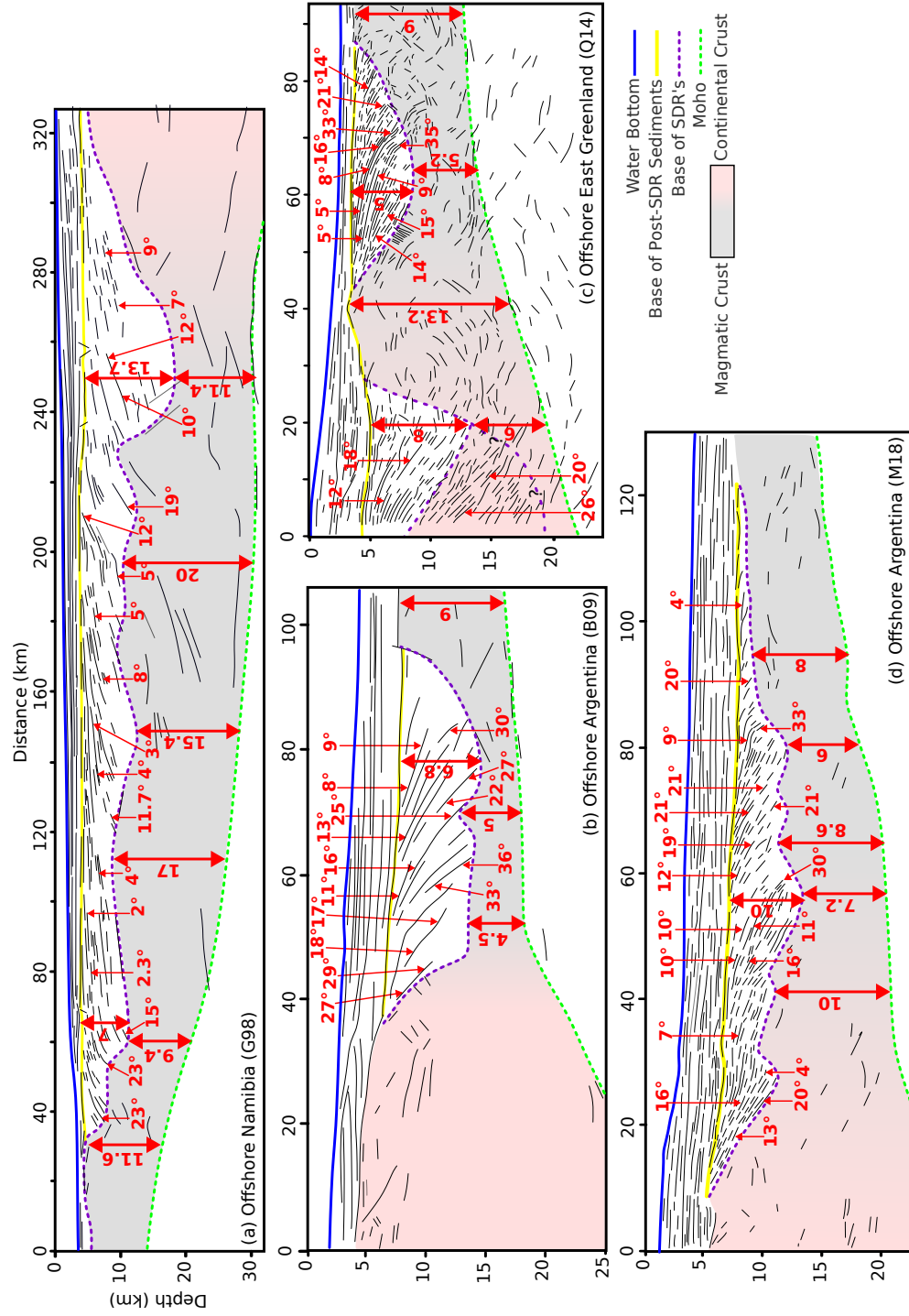


Figure D.2: Seaward-dipping reflector (SDR) geometry from (a) offshore Namibia, (b,c) offshore Argentina, and (d) offshore East Greenland based on published seismic reflection images [1, 5, 7, 9, 10]. Abbreviations shown in seismic cross sections refer to the source of the seismic reflection images and are defined as follows: G98, [1]; B09, [5]; Q14, [7]; M18, [9]; H20, [10]. See Figure D.1 for descriptions of horizons and annotations.

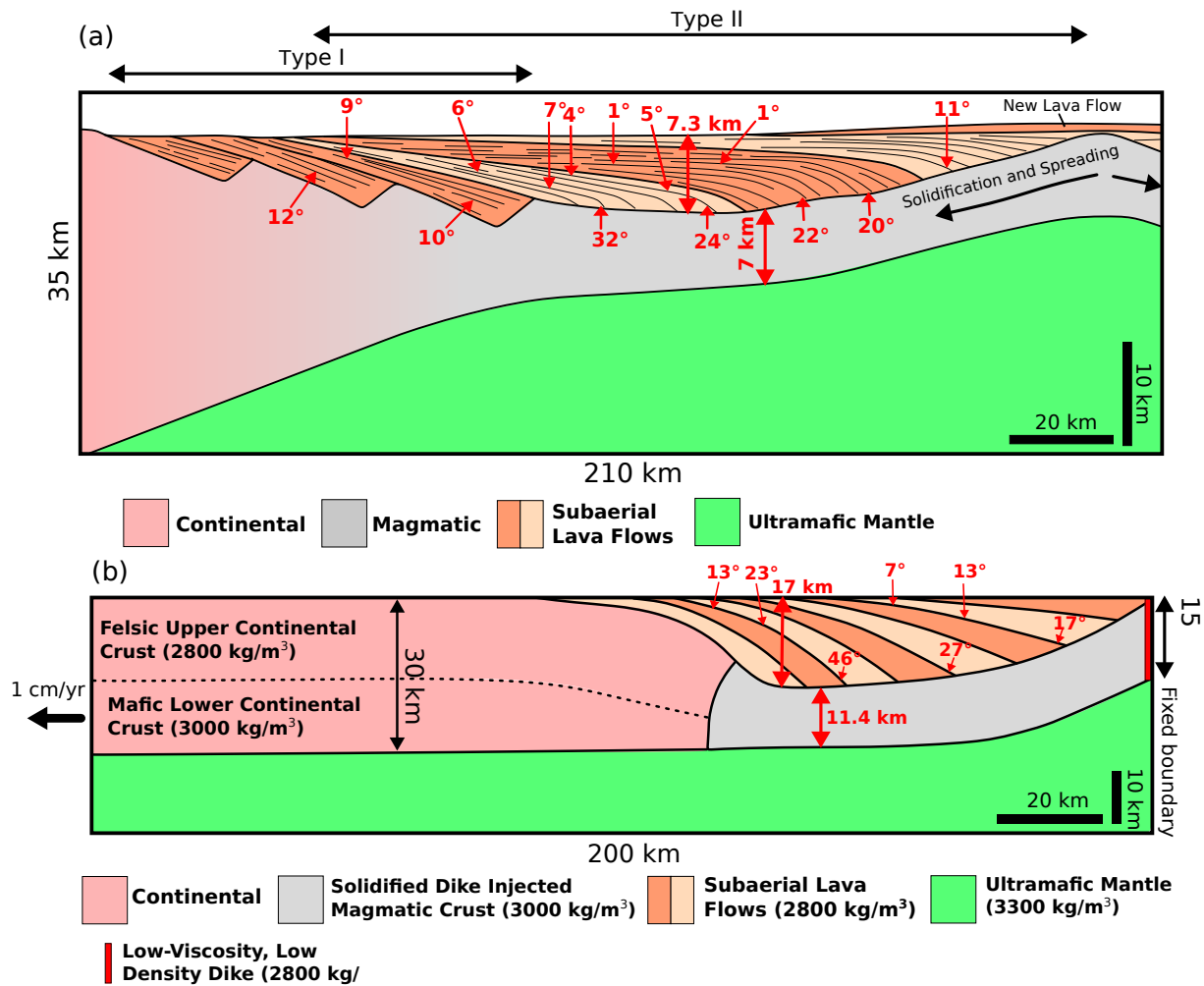
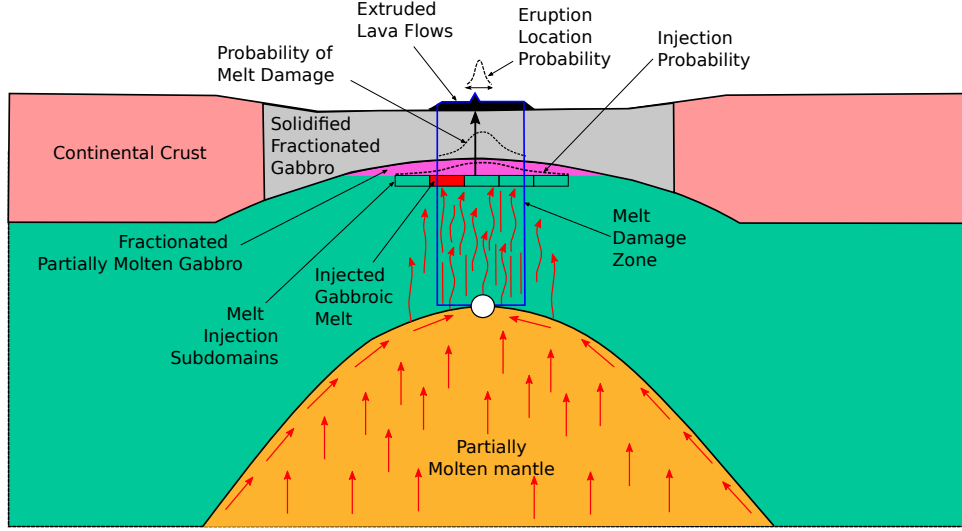


Figure D.3: Published models for SDR formation including (a) a conceptual model based on interpreted seismic reflection data from offshore Brazil and Argentina [9] and (b) a visco-elasto-plastic numerical model without strain weakening from [26] using fixed boundary conditions along the ridge axis and injection of a low-viscosity and low-density dike with a thickness of 15 km at the ridge axis. The Type I and Type II specifications in (a) are based on the classification of [9]. See text for additional details.

(a) Melt Extraction, Intrusion and Extrusion Model



(b) Melt Extraction Model Calibration

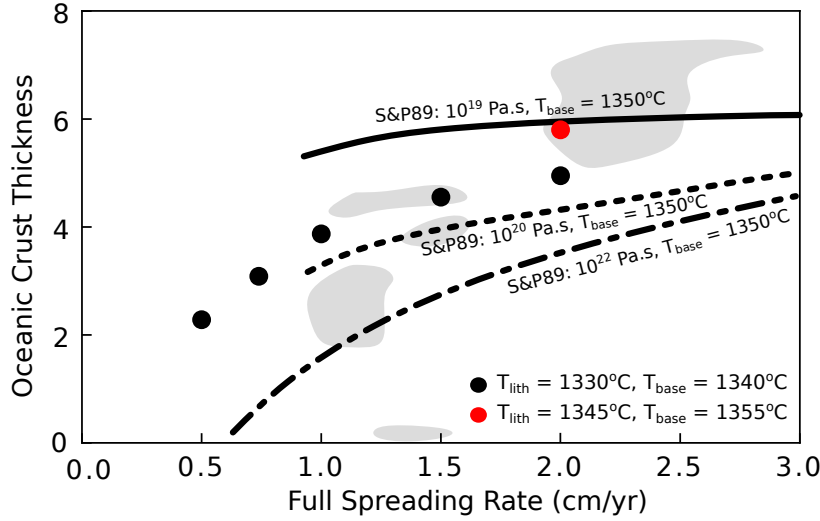


Figure D.4: (a) Schematic illustration of the melt extraction, intrusion and extrusion model used in this work. Melt is assumed to undergo instantaneous vertical migration through the partially molten mantle until a permeability barrier is reached at the top of the partially molten domain. Melt then migrates along the permeability barrier toward the melt focusing point, after which it is transported to the base of the continental crust via channelized networks and dike injection. Gabbroic magma is probabilistically emplaced within melt-injection subdomains at the base of the Moho based on an injection-probability distribution, and fractionated gabbroic melt solidifies to form solidified and partially molten fractionated gabbro. A fraction of the extracted melt is extruded onto the topographic surface as lava flows from probabilistically selected eruption locations and allowed to laterally flow based on a cellular automata model. Melt damage is applied within the melt-damage zone by reducing the friction angle, cohesion and flow viscosity of probabilistically selected markers to approximate the effects of channelized melt networks and dike injection on lithospheric strength. (b) Calibration of the melt extraction model used in this work showing oceanic crust thickness as a function of full spreading rate using the model setup and rheology described in this work. Gray shading indicates the range of global observations from [45]. Black dots show model results for different temperatures at the base of the lithosphere (T_{lith}) using the model setup and boundary conditions described in this work with a thermal anomaly ΔT of 0 °C. Solid, dotted and dashed curves show the model results from [46] for different asthenosphere viscosities with the temperature at the base of the model (T_{base}) set to 1350°C.

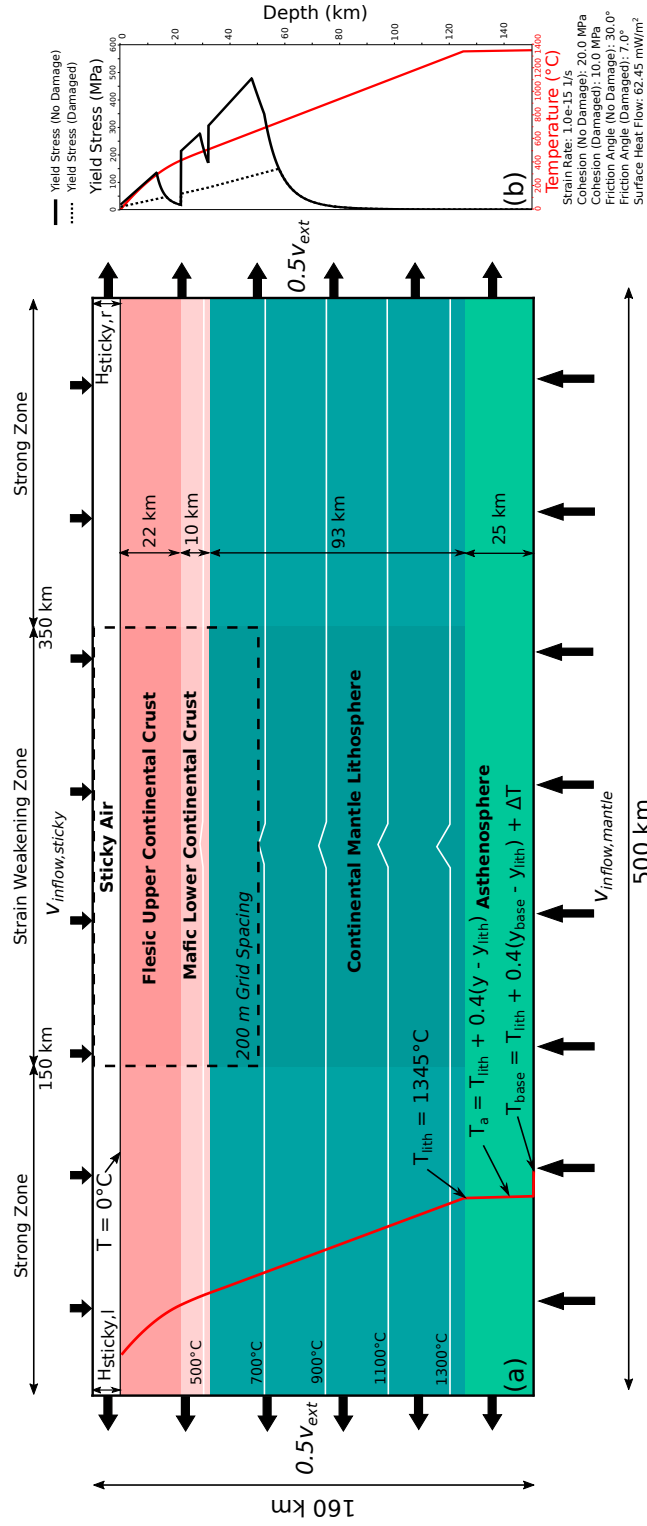


Figure D-5: Model setup showing (a) the initial model geometry and boundary conditions, and (b) the initial thermal structure and yield stress profile. The model domain is 500 km by 160 km and consists of a 10 km thick sticky-air layer overlying a 22 km thick felsic upper continental crust (wet quartzite rheology), a 10 km thick mafic lower continental crust (dry olivine aggregate rheology), a 93 km thick continental mantle lithosphere (dry olivine aggregate rheology) and a 25 km thick asthenosphere layer (dry olivine aggregate rheology). Strain weakening is disabled in lateral strong zones, which are denoted by lighter hues, to localize deformation in the central region of the model domain. The top boundary temperature is fixed at 0° , while the basal temperature is set to $T_{base} = T_{lith} + 0.4(y - y_{lith}) + \Delta T$ where $T_{lith} = 1345^\circ\text{C}$ is the initial temperature at the base of the continental lithosphere and ΔT is a mantle thermal anomaly. A locally refined high-resolution region with a uniform grid spacing of 200 m is denoted by the black dashed lines in panel (a). The initial asthenosphere thermal structure follows an adiabatic gradient of $0.4^\circ\text{C}/\text{km}$. White lines show isotherms from the initial thermal structure that includes a perturbation to ensure deformation localizes in the central part of the model domain. Velocity boundary conditions involve outflow velocities of $0.5v_{ext}$ on the side boundaries and compensating inflow velocities $v_{inflow, sticky}$ and $v_{inflow, mantle}$ on the top and bottom boundaries to maintain volume conservation. Panel (b) shows the temperature profile (red curve) and yield stress profiles for both undamaged material with a friction angle of 30° (solid black line) and damaged material with a friction angle of 7° (dashed black line). See text for additional details.

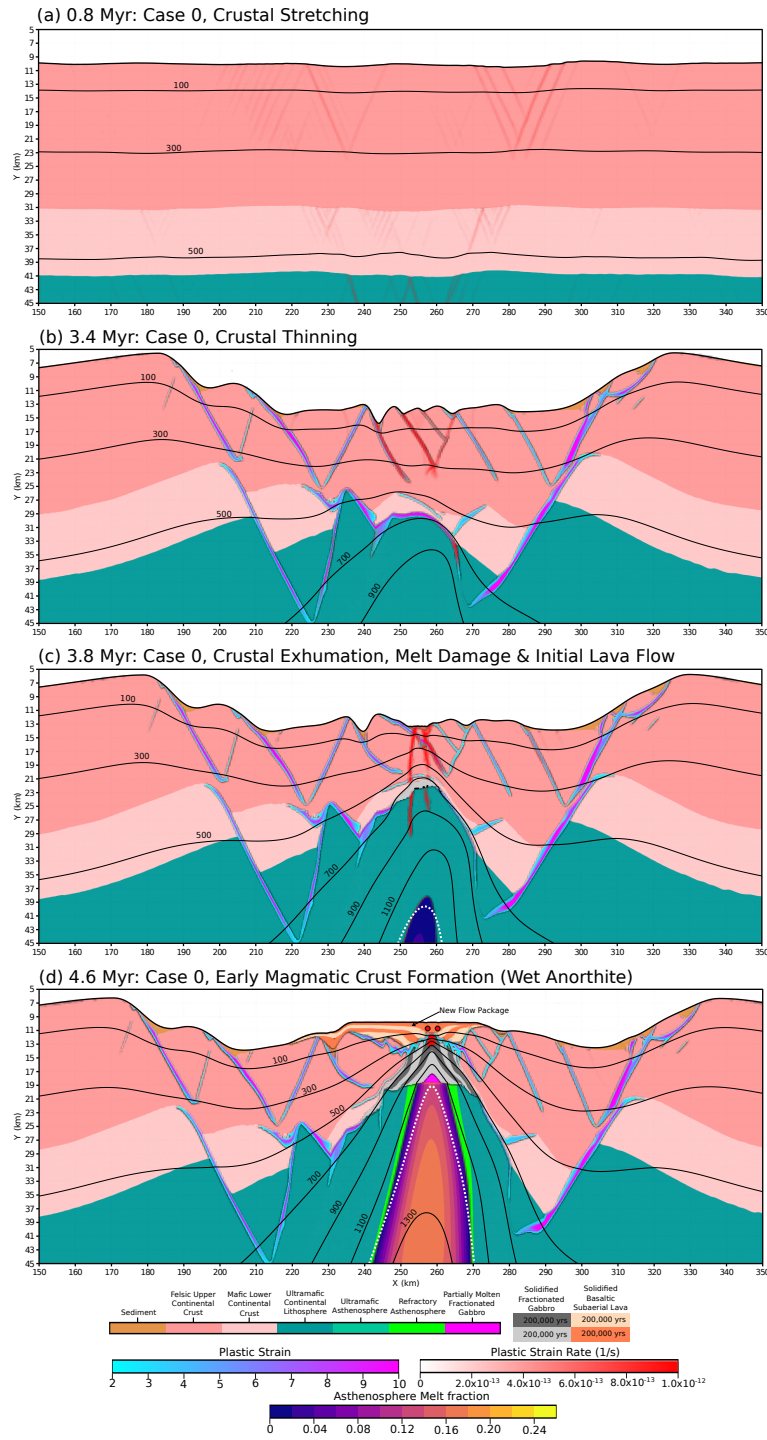


Figure D.6: Panels (a-f) show the evolution of the base-case model (Case 0) with probabilistic melt damage, inter-eruption period of 200,000 yr, maximum extrusion factor of 50%, thermal anomaly of 100 °C, wet anorthite rheology in gabbroic magmatic crust and flow length of 40 km. Panels (g,h) show a time step comparable to (f) but for (g) Case 43 with a dry anorthite rheology in magmatic crust and (h) Case 13 with a two orders of magnitude greater fluvial sediment transport relative to the base case. Markers are colored according to their composition with overlays for asthenosphere melt fraction, plastic strain rate and plastic strain. Decompression melting generates partially molten asthenosphere colored according to the total melt fraction assuming no melt was extracted from the asthenosphere. Partially molten asthenosphere transforms to refractory asthenosphere once sufficient cooling occurs as denoted by light green colors. Melt extraction produces intrusive gabbro bodies represented as dark and light bands of gray to show 200,000 yr solidification intervals. Packages of subaerial flows are represented as dark and light orange bands representing 200,000 yr intervals. Temperature isotherms are denoted with black lines, with the 1200 °C isotherm shown as a dotted white line. Colored circles in (d-f) show tracers in conjugate lava-flow packages. Black lines with central circles in panels (f-h) show measured dip in degrees within lava-flow packages. Vertical double arrows in panels (f-h) show thicknesses in kilometers.

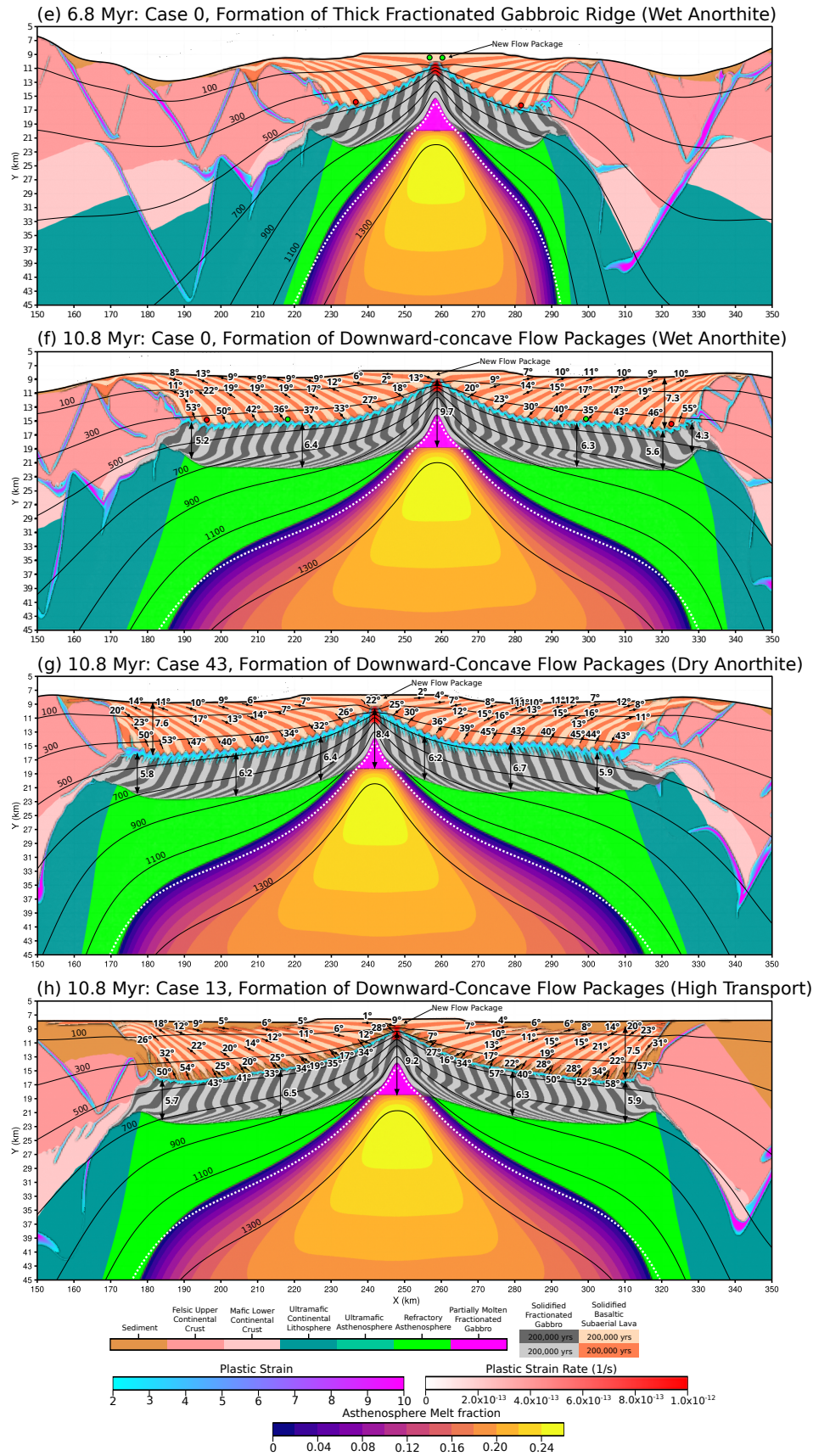


Figure D.6: (continued)

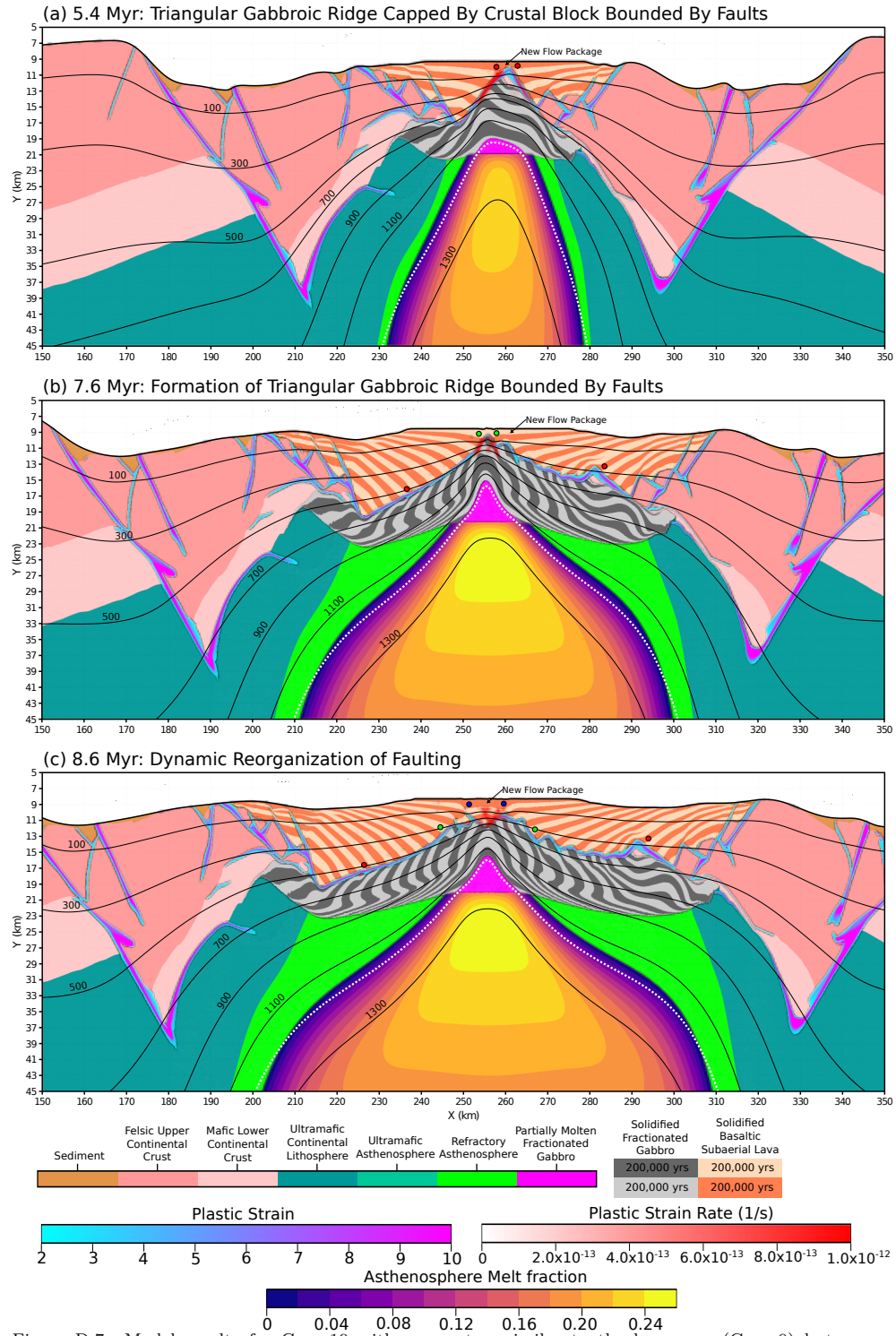


Figure D.7: Model results for Case 19 with parameters similar to the base case (Case 0) but without melt damage. Panels (a-f) show the evolution of the model starting from early magmatic crust formation after lithospheric extension and breakup. Colored circles show tracers in conjugate lava-flow packages. See Figure D.6 and Table D.7 for additional details.

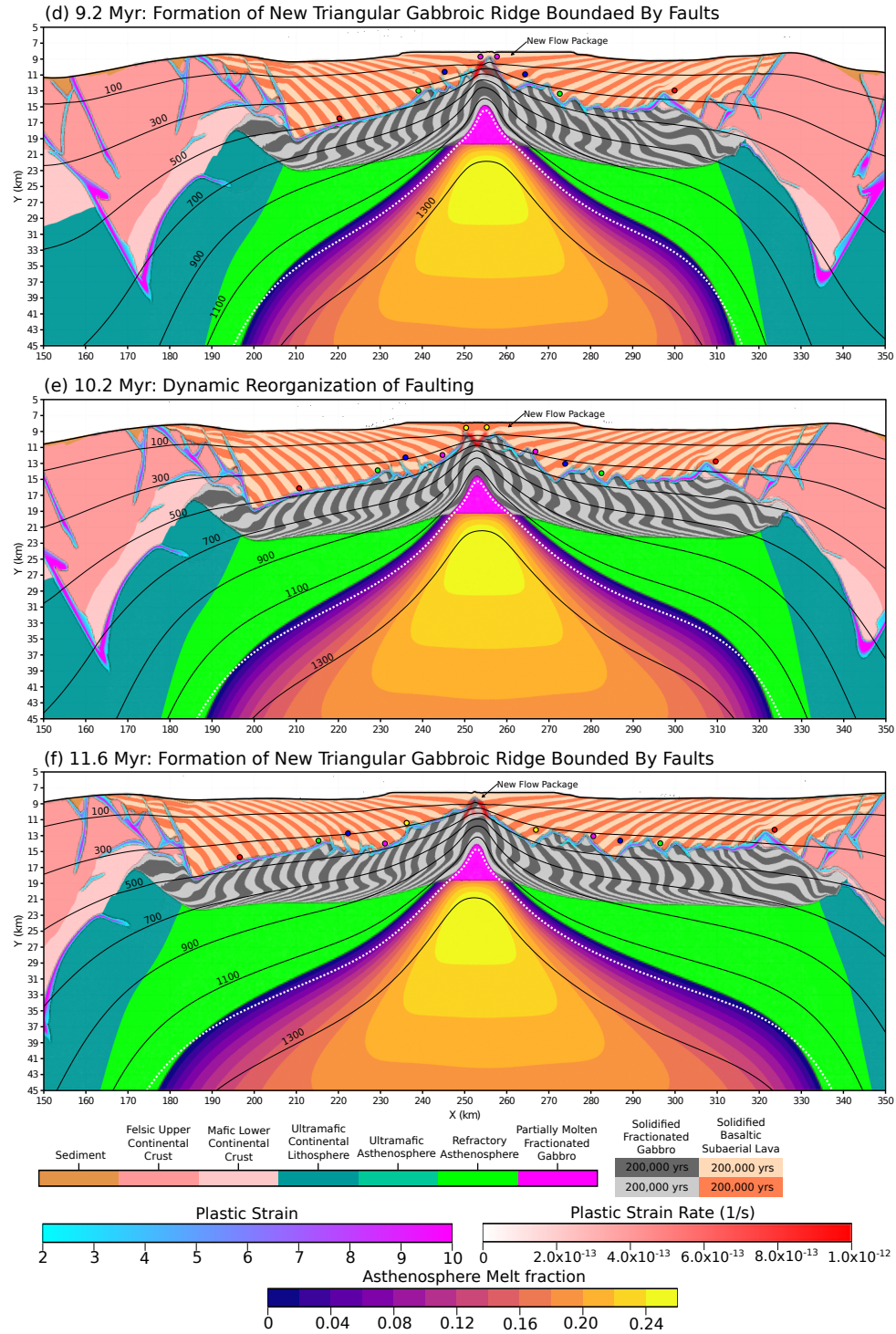


Figure D.7: (continued)

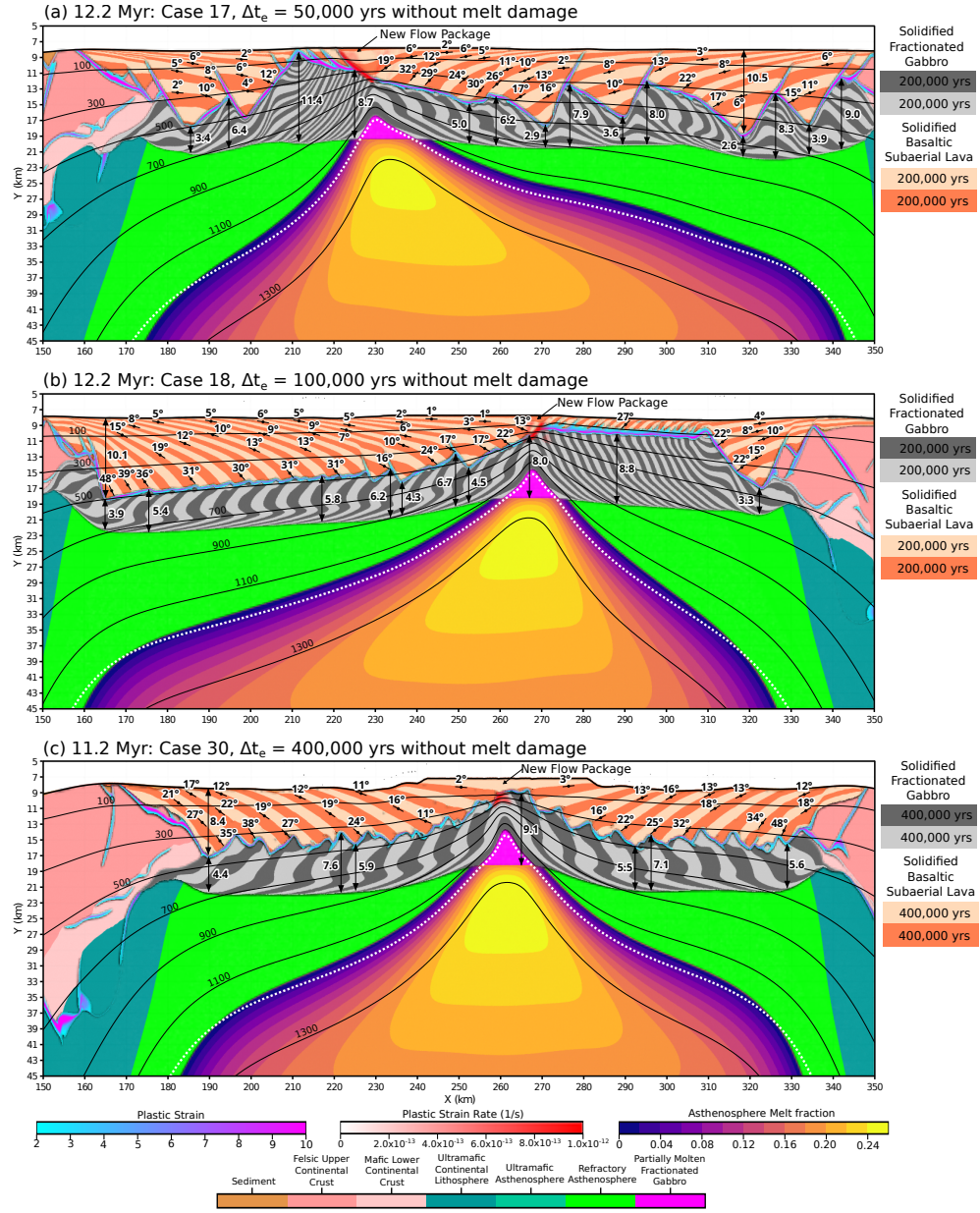


Figure D.9: Panels (a-c) show model results for Cases 17, 18 and 30 with parameters similar to the base case (Case 0) but without melt damage and different inter-eruption periods Δt_{erupt} . See Figure D.6 and Table D.7 for additional details.

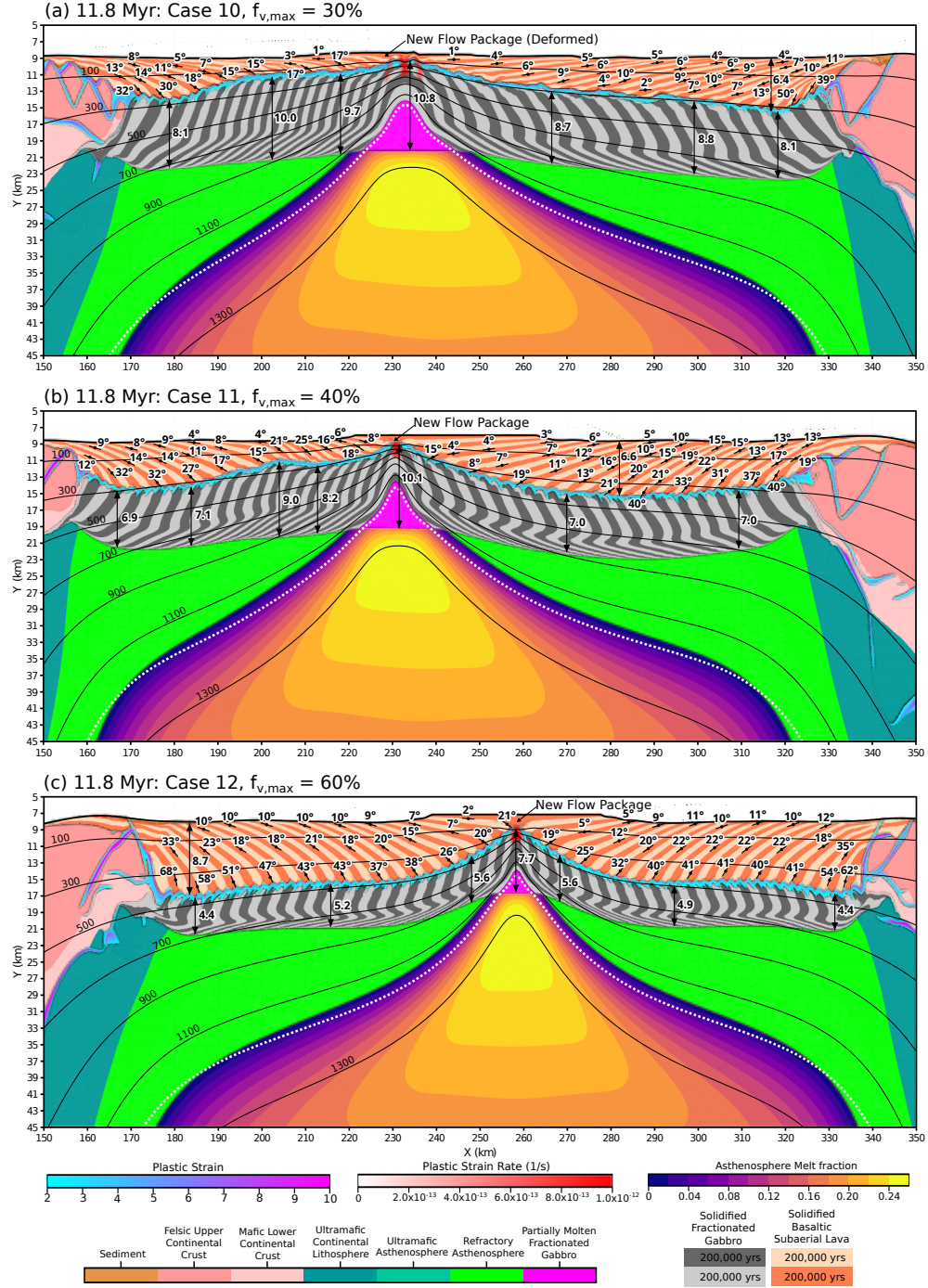


Figure D.10: Panels (a-c) show model results for Cases 10, 11 and 12 with parameters similar to the base case (Case 0) but with different maximum extrusion factors $f_{v,max}$. See Figure D.6 and Table D.7 for additional details.

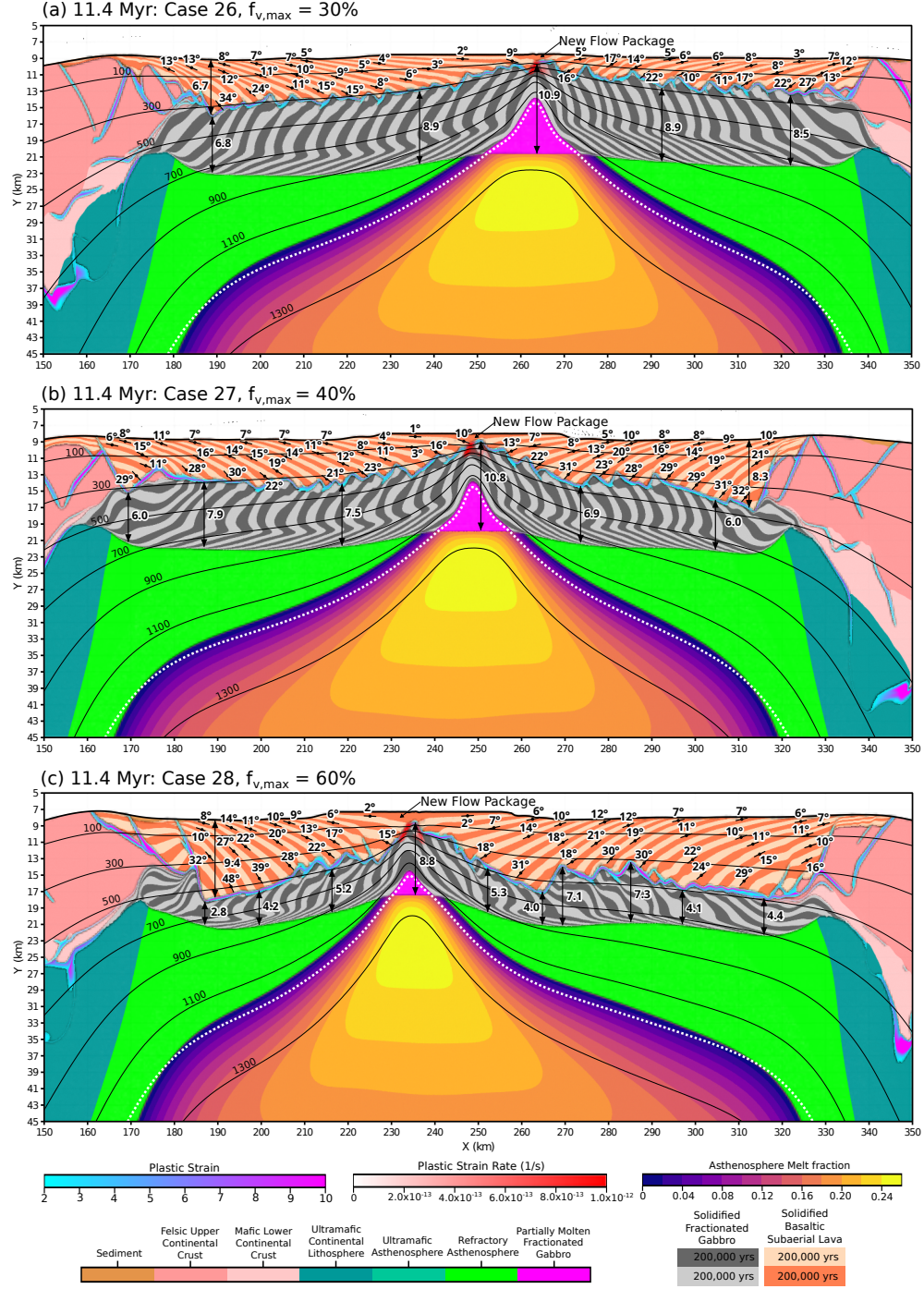


Figure D.11: Panels (a-c) show model results for Cases 26, 27 and 28 with parameters similar to the base case (Case 0) but with different maximum extrusion factors $f_{v,max}$ and no melt damage. See Figure D.6 and Table D.7 for additional details.

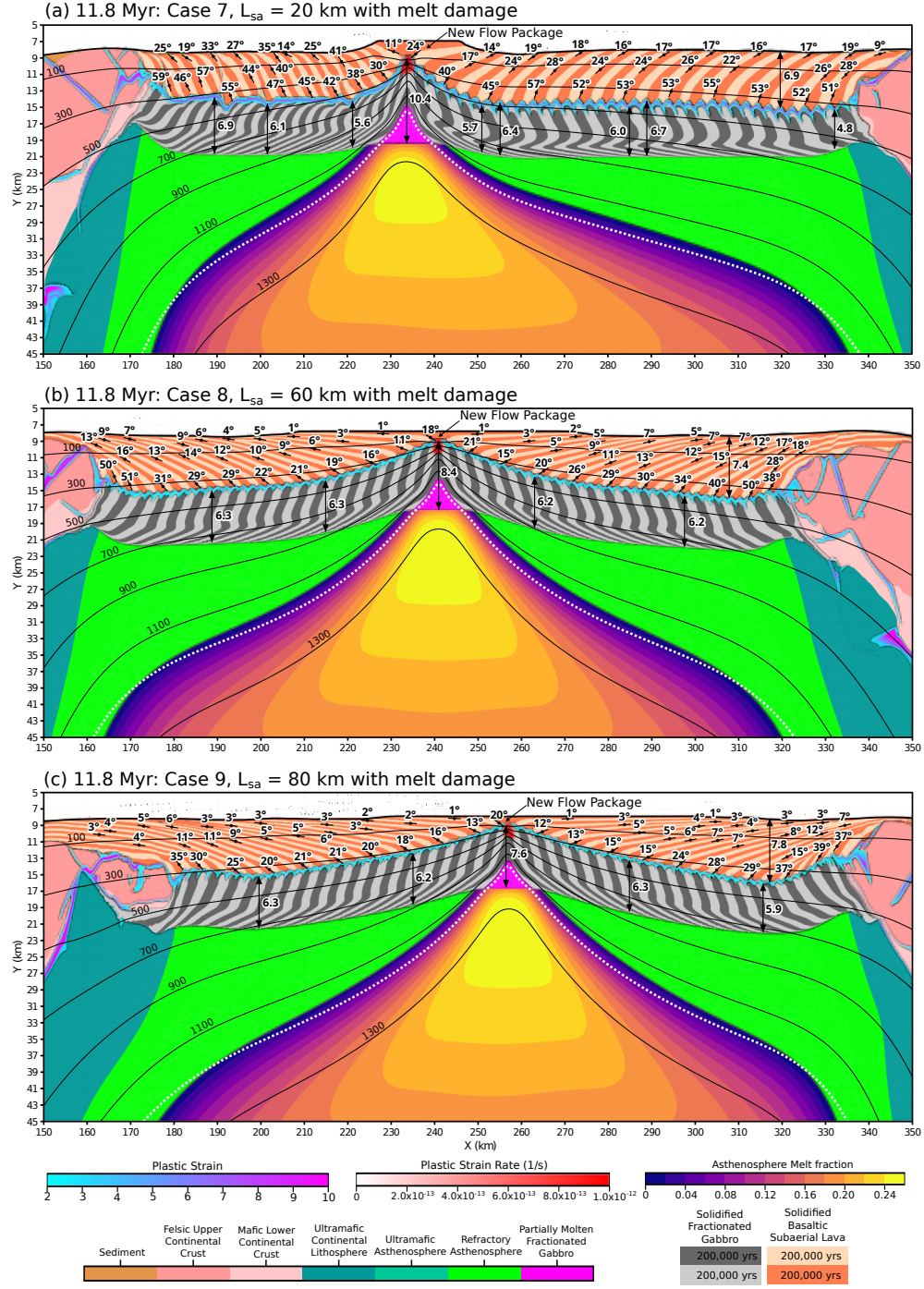


Figure D.12: Panels (a-c) show model results for Cases 7, 8 and 9 with parameters similar to the base case (Case 0) but with different subaerial flow lengths L_{sa} . See Figure D.6 and Table D.7 for additional details.

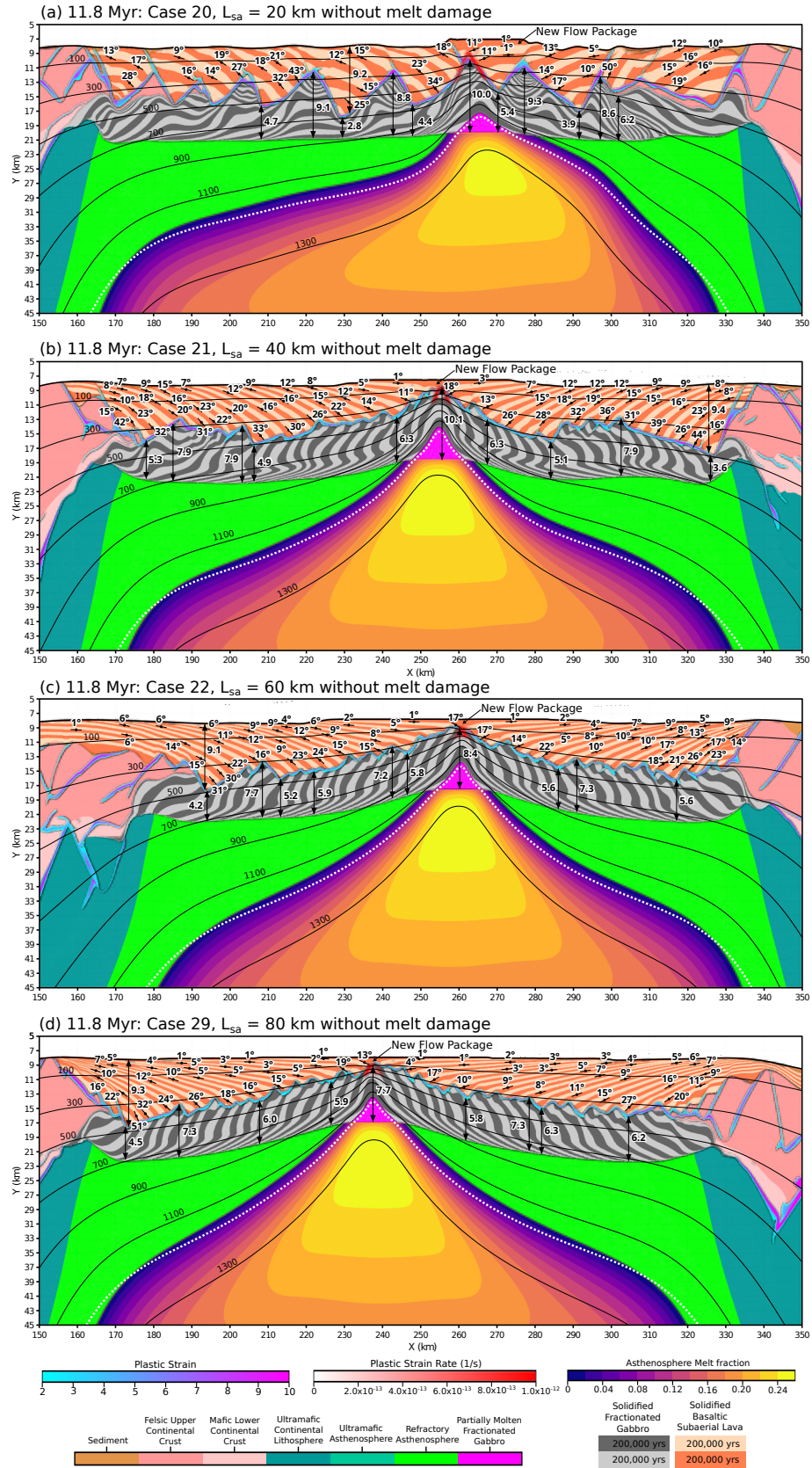


Figure D.13: Panels (a-d) show model results for Cases 20, 21, 22 and 29 with parameters similar to the base case (Case 0) but with different subaerial flow lengths L_{sa} and no melt damage. See Figure D.6 and Table D.7 for additional details.

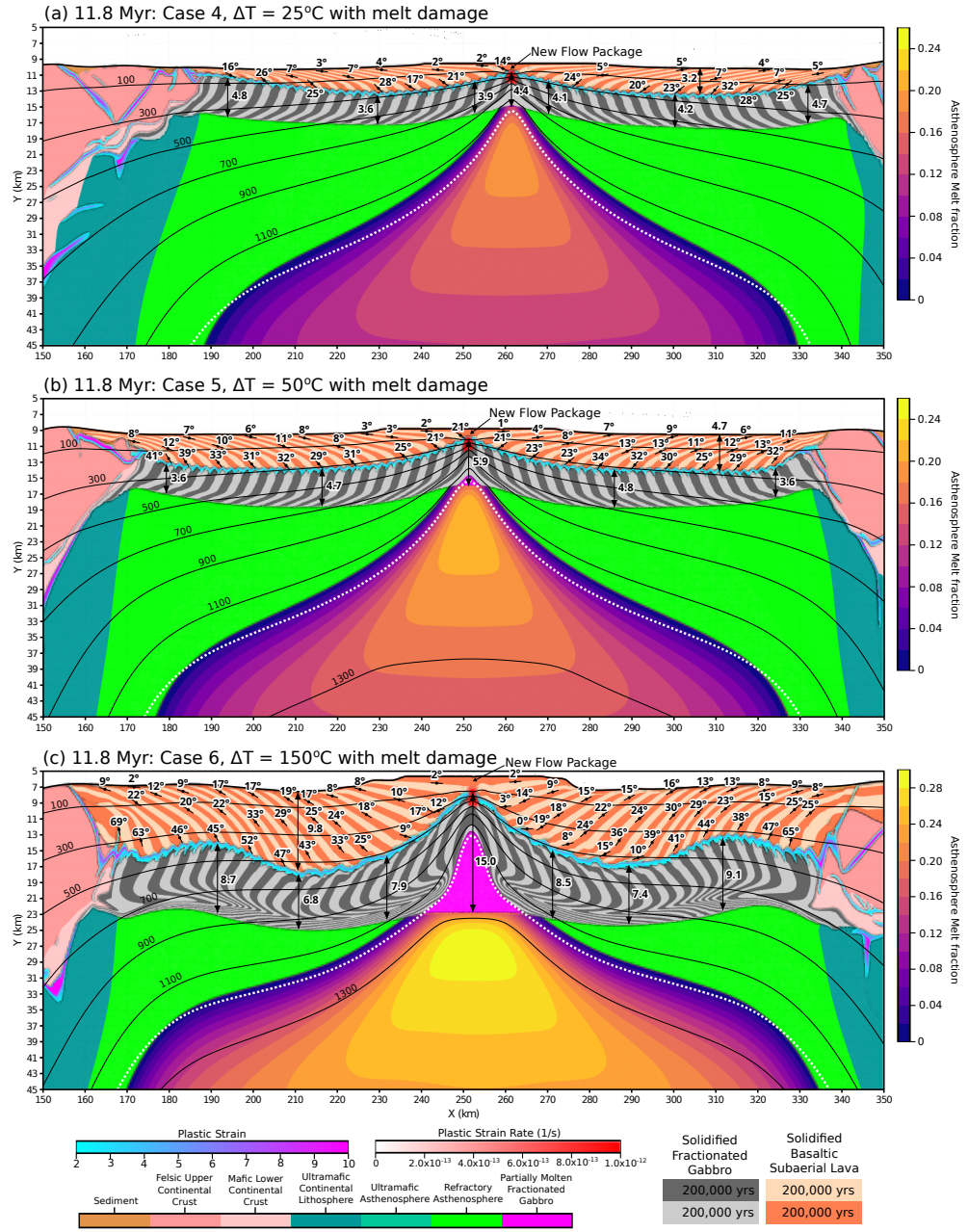


Figure D.14: Panels (a-c) show model results for Cases 4, 5, and 6 with parameters similar to the base case (Case 0) but with different thermal anomalies ΔT . See Figure D.6 and Table D.7 for additional details.

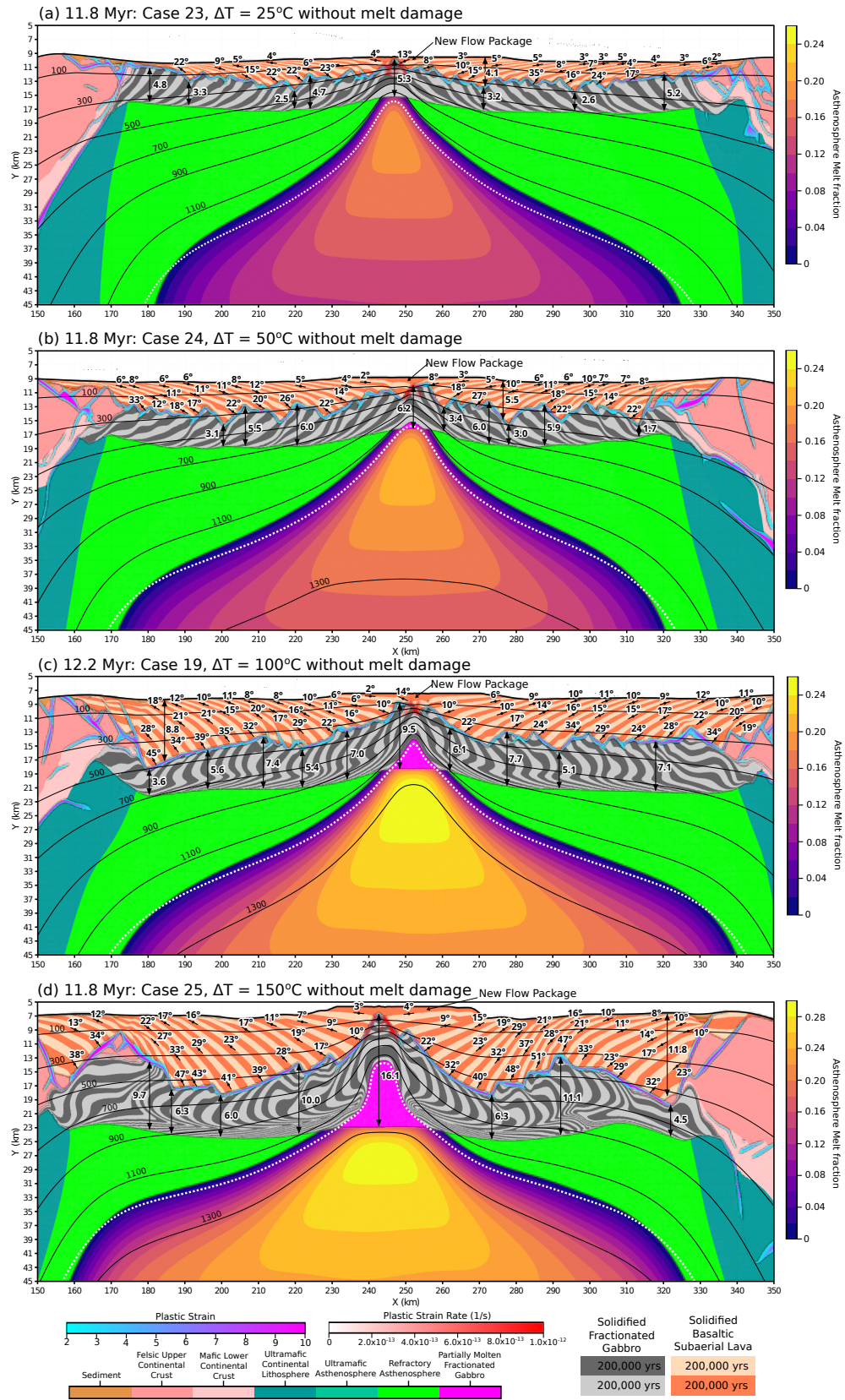


Figure D.15: Panels (a-d) show model results for Cases 23, 24, 19 and 25 with parameters similar to the base case (Case 0) but without melt damage and different thermal anomalies ΔT . See Figure D.6 and Table D.7 for additional details.

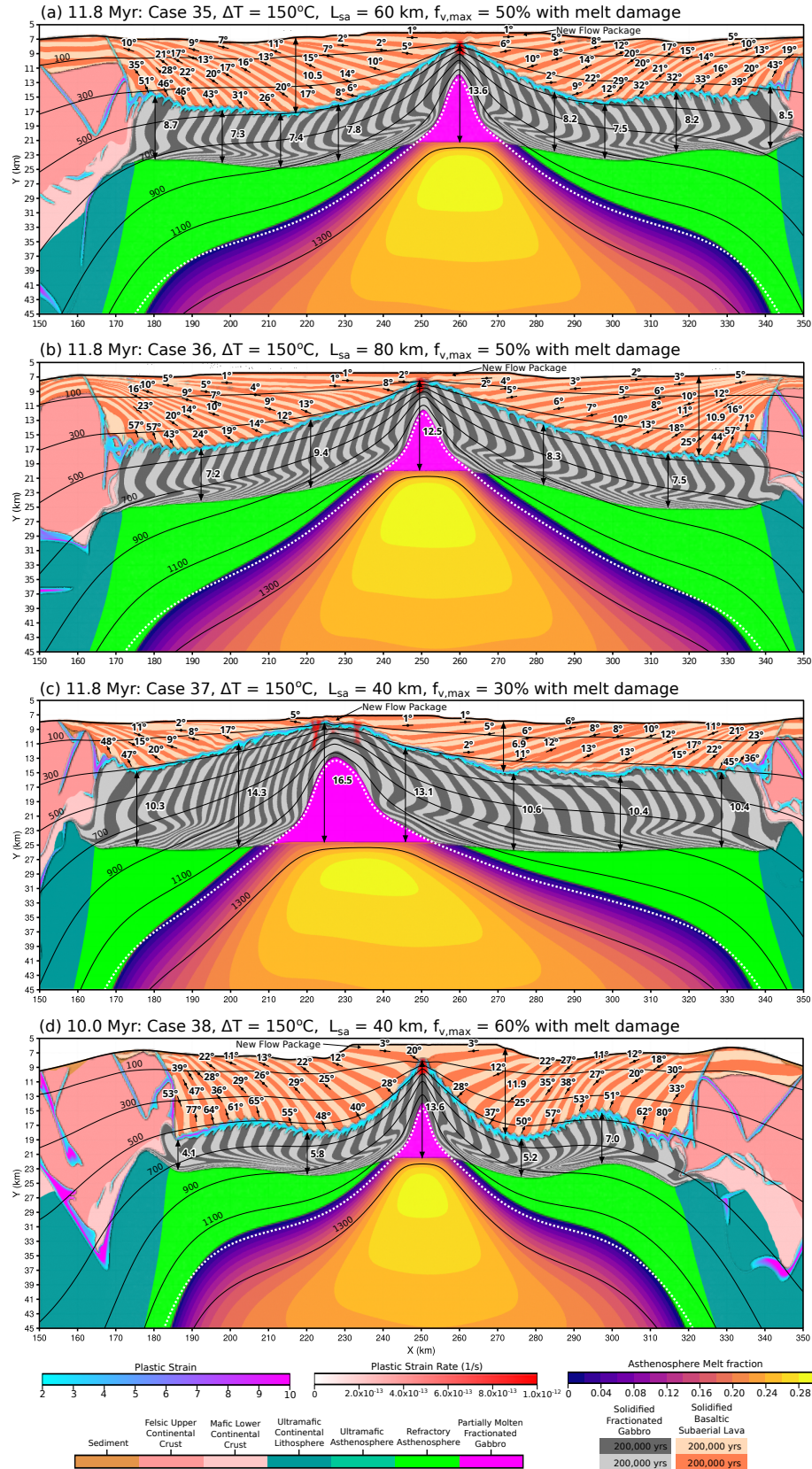


Figure D.16: Panels (a-d) show model results for Cases 35, 36, 37 and 38 with parameters similar to the base case (Case 0) but with a thermal anomaly $\Delta T = 150^\circ\text{C}$, (a-b) different length of subaerial flows L_{sa} , and (c-d) different maximum extrusion factor $f_{v,max}$. See Figure D.6 and Table D.7 for additional details.

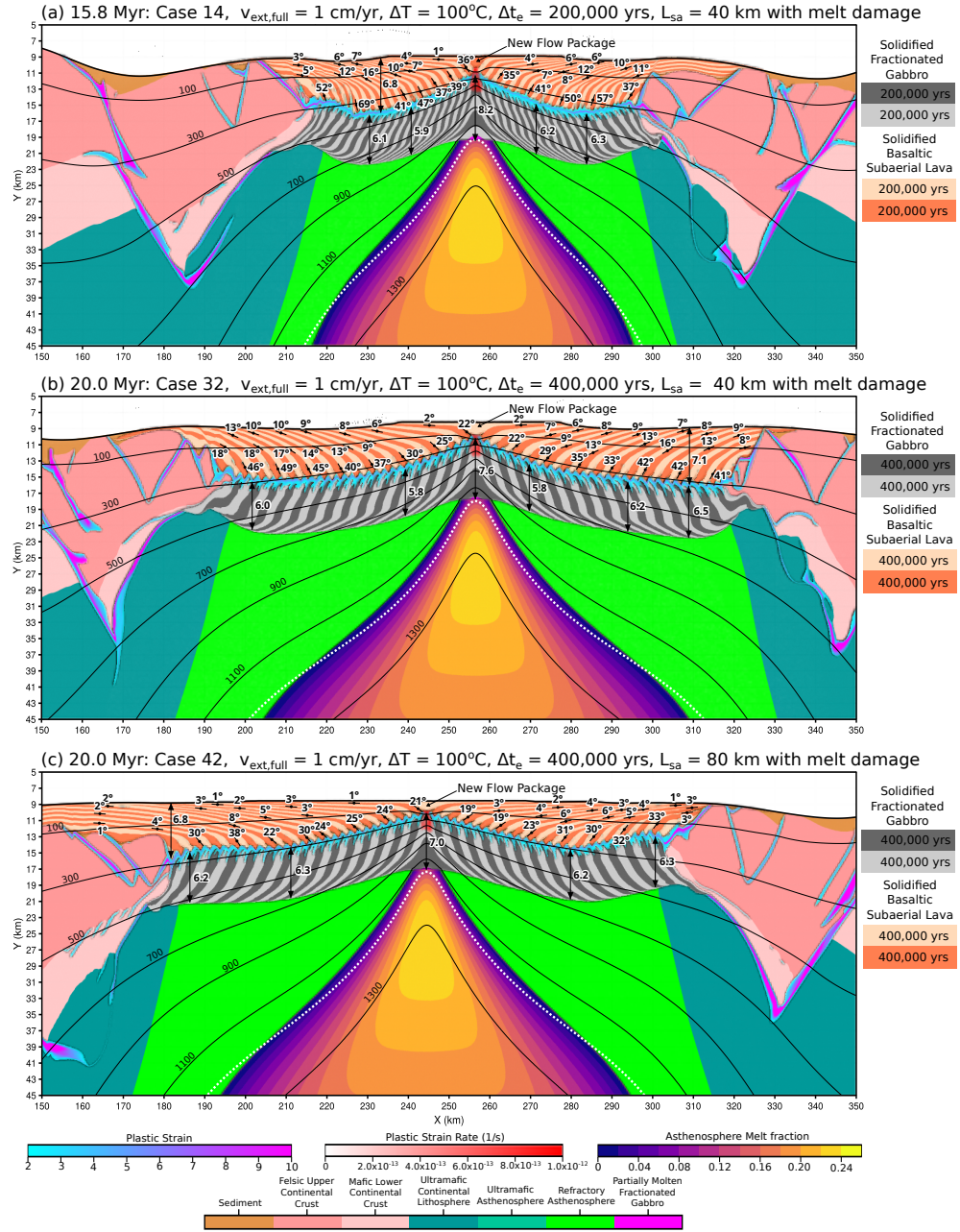


Figure D.17: Panels (a-c) show model results for Cases 14, 32 and 42 with full extensional velocity $v_{\text{ext}} = 1 \text{ cm/yr}$, a thermal anomaly $\Delta T = 100^\circ\text{C}$ and different eruption intervals Δt_{erupt} and subaerial flow lengths L_{sa} . See Figure D.6 and Table D.7 for additional details.

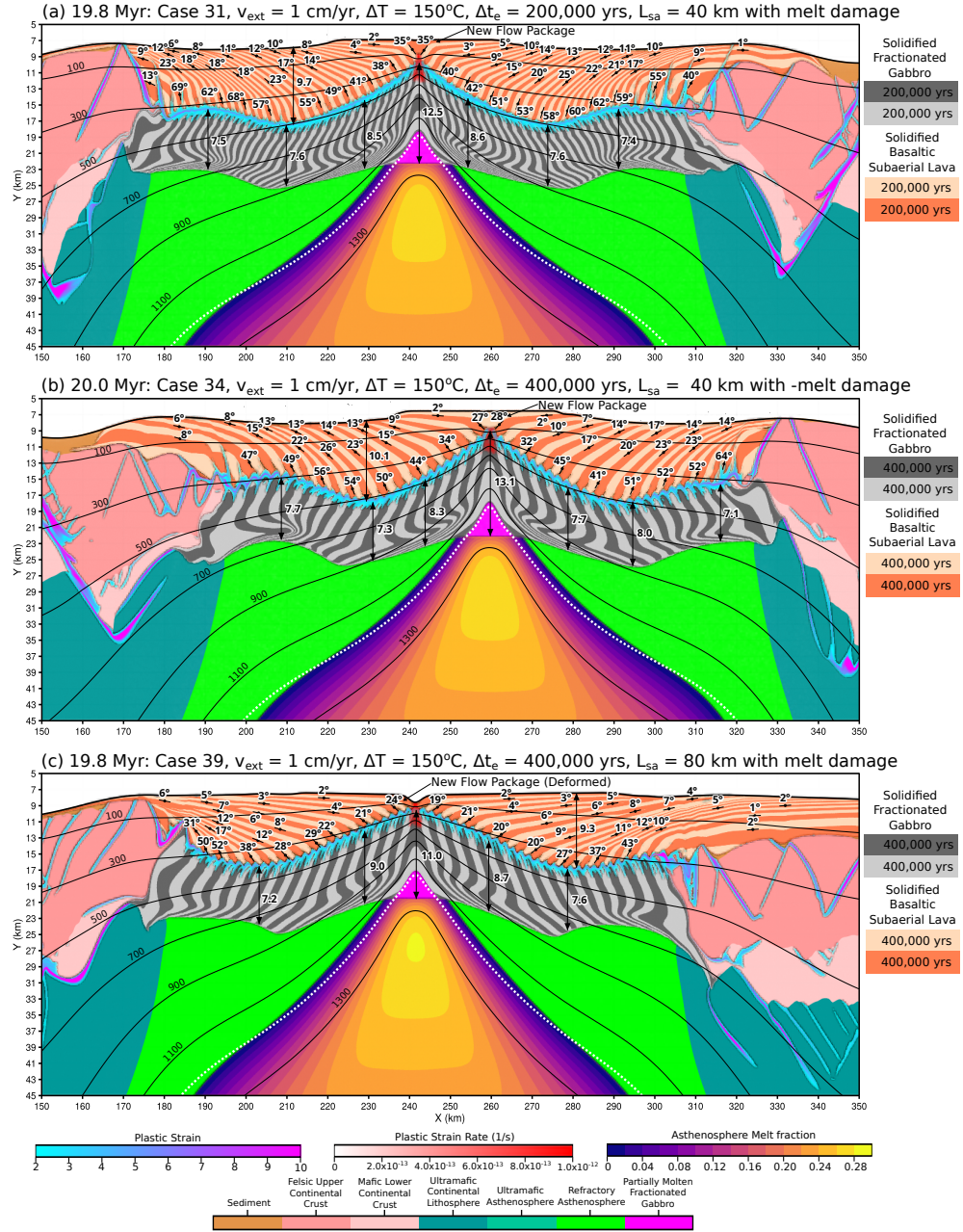
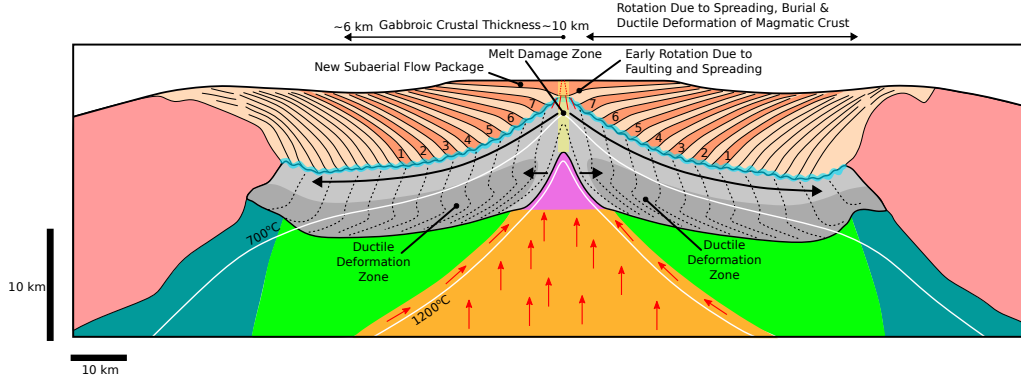
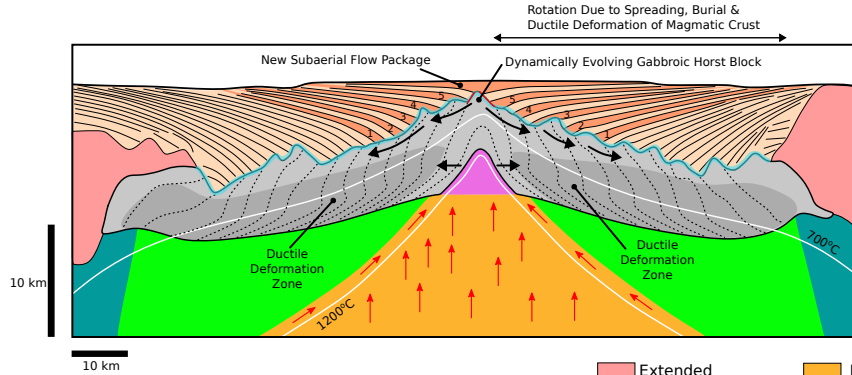


Figure D.18: Panels (a-c) show model results for Cases 31, 34 and 39 with full extensional velocity $v_{\text{ext}} = 1 \text{ cm/yr}$, a thermal anomaly $\Delta T = 150^\circ\text{C}$ and different inter-eruption periods Δt_{erupt} and subaerial flow lengths L_{sa} . See Figure D.6 and Table D.7 for additional details.

(a) Stable Deformation At Ridge, Symmetric Spreading, Gabbroic Crustal Thinning



(b) Dynamic Deformation At Ridge, Symmetric Spreading, Gabbroic Crustal Thinning



(c) Listric Faulting in Magmatic Crust, Asymmetric Spreading, Gabbroic Crustal Thinning

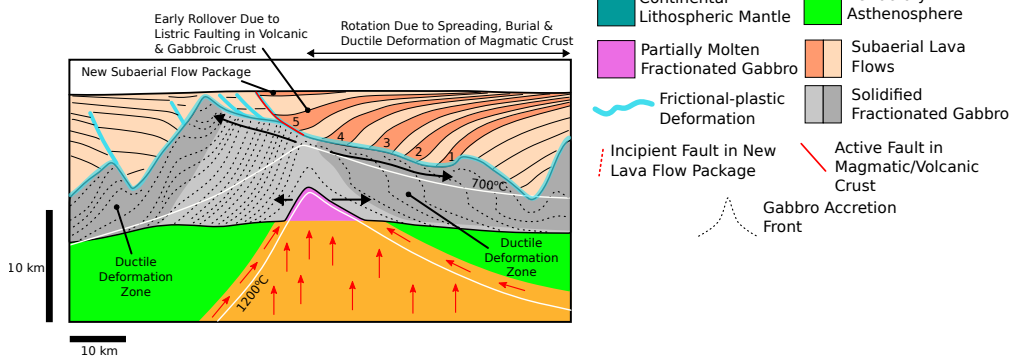


Figure D.19: Schematic illustrations of different thermo-mechanical processes associated with the development of downward-concave ridgeward-dipping lava-flow packages and different gabbroic magmatic crust structures as observed in the numerical models presented in this work. Panel (a) is based on Case 0 and shows a scenario where melt damage leads to stable localized extensional deformation via steep normal faults at the ridge axis, rotation of flow packages with increasing dip in the down-dip direction, and gabbroic basement with low relief. Flow packages spread away from a thick gabbroic ridge cored by partially molten fractionated gabbro and undergo continued rotation due to spreading, burial and gabbroic crustal thinning associated with ductile folding in the lower magmatic crust. The up-dip geometry of lava-flow packages is associated with dip variations that range from a few degrees to less than 10 degrees. Panel (b) is based on Cases 19, 22 and 29 and shows a scenario similar to (a) but without melt damage which leads to dynamic deformation at the ridge and ridgeward-dipping reflectors that form due to a triangular horst at the ridge. Similar to (a) continued rotation of flow packages occurs due to spreading, burial and gabbroic crustal thinning. This scenario is associated with much more relief in the gabbroic basement relative to (a). Panel (c) is based on Cases 17 and 18 and shows the formation of a listric fault within volcanic packages and upper gabbroic basement that detaches in ductile gabbroic crust. Asymmetric ridgeward-dipping lava-flow packages develop due to syn-kinematic infill of lava flows during listric faulting, spreading away from the ridge axis and ductile deformation of gabbroic crust due to volcanic loading. For this scenario, the geometry of lava-flow packages and gabbroic crust accretion is highly asymmetric.

Supplemental Material for The Formation of Seaward-dipping
Reflectors in Volcanic Margins: Insights from High-resolution
Visco-elasto-plastic Geodynamic Models with Extrusive Surface
Processes

Erik Arthur Kneller

1 Supplemental Plots

1.1 Visco-plastic Viscosity

Figures S1 and S2 show plots for visco-plastic viscosity for Case 0 and Case 19, respectively. Output time steps correspond to the time steps shown in Figure 6 and Figure 7 of the main text. Case 0 is the base case with melt damage, a thermal anomaly of 100°C, an inter-eruption period of 200,000 yr, a subaerial flow length of 40 km and a maximum extrusion factor of 50%. Case 19 has parameters identical to Case 0 but melt damage is disabled. See the main text for additional details.

1.2 The Effect of Dike Fluid on Calculated Density and Ridge Elevation

Figure S3 shows plots for density for Case 0 (base case) and Case 15. Case 15 differs from Case 0 in that the density of markers that are probabilistically selected for melt damage is modified to take into account the effects of injected dike fluid. For Case 15, it is assumed that 100% of melt-damaged markers are filled with dike fluid that has a reference density of 2,750 kg/m³. The introduction of dike fluid reduces density at the ridge axis by around 40–50 kg/m³ relative to the calculated density without dike fluid and has a negligible effect on the overall structural evolution of the model.

The maximum effect of a large crustal-scale vertical dike intrusion on elevation Δh can be estimated with the following equation that assumes zero strength of the lithosphere:

$$\Delta h = \frac{t_c \Delta \rho}{\rho_{mantle}} \quad (S1)$$

where t_c is the magmatic crustal thickness, $\Delta \rho$ is the density difference between the magmatic crust and the injected dike fluid and ρ_{mantle} is the density of the mantle. The following parameters that maximize the effect of the dike intrusion on elevation for equation (S1) from Figure S3 are: $t_c = 10$ km, $\Delta \rho = 50$ kg/m³ and $\rho_{mantle} = 3,170$ kg/m³. Substituting these values into equation (S1) gives a maximum elevation difference of less than 160 m, which is negligible for the purposes of this study. This simple calculation is consistent with the results shown in Figure S3, which use a lithospheric rheology with finite strength.

2 Convergence Tests

Figures S4–S11 show the effects of the maximum allowed marker displacement relative to cell size (f_{cell}), initial model time step size (Δt) and Picard convergence tolerance for a case with model parameters similar to the base case from the main paper (Case 0 in Table 7) but with the following changes: (1) a longer characteristic length of subaerial flow of 50 km (vs. 40 km), (2) an initial geotherm that includes the effects of the 100°C temperature anomaly, (3) a transient bottom temperature boundary condition that has the thermal anomaly removed after 1 Myr, (4) no thermal perturbation in the central part of the model and (5) adaptive time stepping that uses the average grid cell size in the x- and y-directions instead of the local grid cell size used in the main text. All convergence tests use the same local refinement approach as the base case (Case 0) that includes a central region where grid spacing is 200 m and initial marker resolution is 25 m. The initial thermal perturbation was not included to test the algorithm’s ability to reach convergence under more challenging conditions relative to the models presented in the main text. Convergence tests are shown for different combinations of f_{cell} , Δt and Picard convergence tolerance during the first 2 Myr of model evolution (Figures S4–S7) and during the latter phases of breakup and magmatic crust formation (Figures S8–S11). Figures S12–S15 show convergence behavior of Picard iterations for solution vectors and the non-linear residual for different numerical parameters.

A key conclusion from these numerical experiments is that for models with an initial model time step size (Δt) less than 10,000 yr and a marker displacement limit (f_{cell}) less than or equal to 0.5, a tolerance of 10^{−3} produces structural evolution broadly consistent with model runs using a tolerance of 10^{−4} and 10^{−5}. The primary difference is that simulations with smaller convergence tolerances localize shear zones approximately 1 Myr earlier. This modest delay in shear zone localization does not affect the overall conclusions from the main text. Therefore, for the purposes of this study, a tolerance of 10^{−3} and a marker displacement limit f_{cell} of 0.5 are considered sufficient to capture the structural evolution of the model when adaptive time stepping based on local grid sizes is used. Under these conditions, model time steps decrease to approximately 1,500 yr to 4,000 yr during the later phases of extension, breakup and magmatic crust formation.

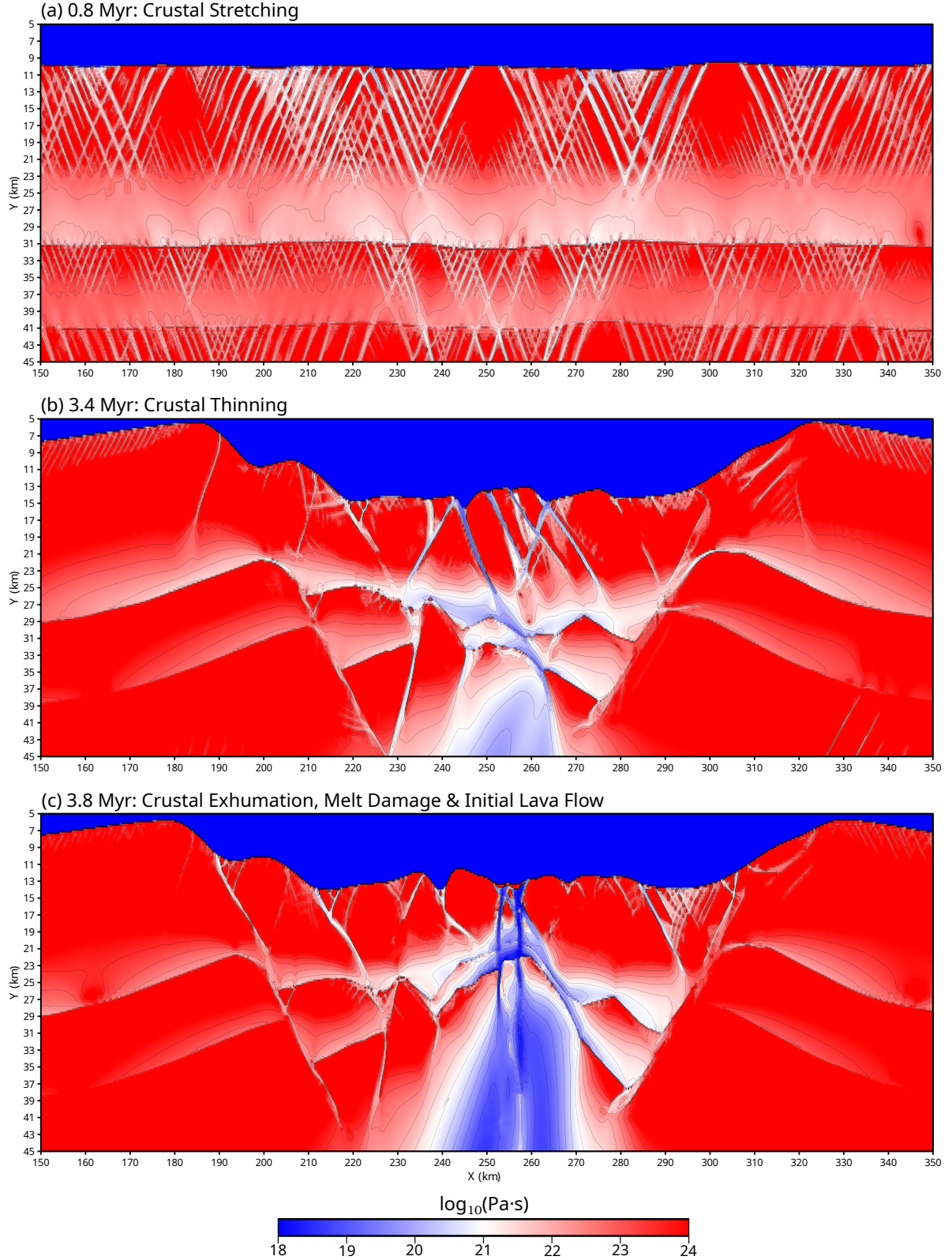


Figure S1: Visco-plastic viscosity for Case 0 (base case) during (a-b) lithospheric extension and (c) early magmatic crust formation and (d-f) continued volcanic extrusion, formation of downward-concave volcanic flow packages that dip toward the ridge axis and ductile deformation in the gabbroic magmatic crust. The vertical zones of reduced viscosity in (c) are associated with a melt-damage zone that is moving laterally, which produces a relict zone of elevated plastic strain rate and reduced visco-plastic viscosity to the left of the current melt damage zone centered above the shallowest point of the partially molten asthenosphere. Compare with Figure 6 of the main text. Contour interval is $0.5 \log_{10}(\text{Pa}\cdot\text{s})$.

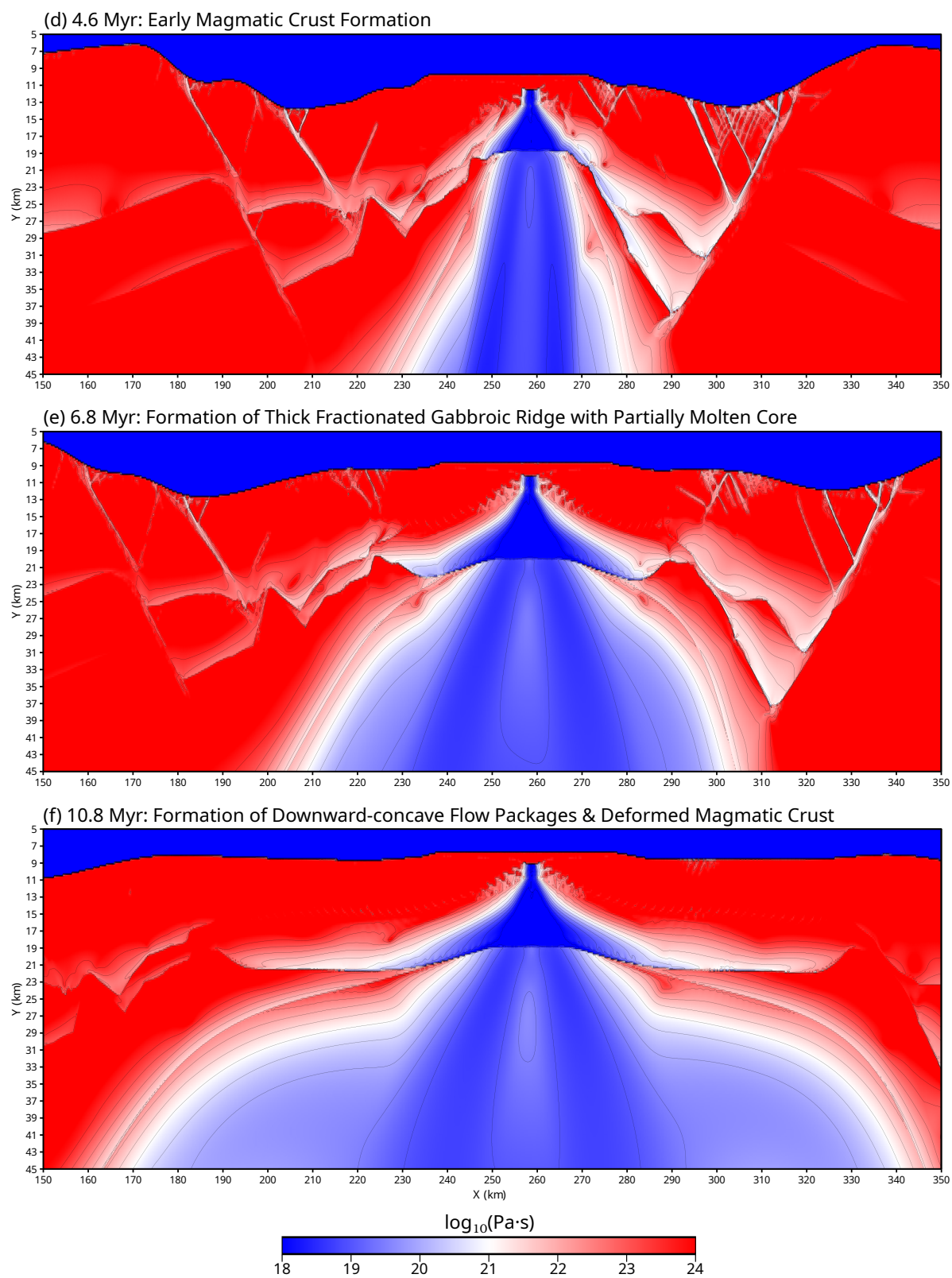


Figure S1: (continued)

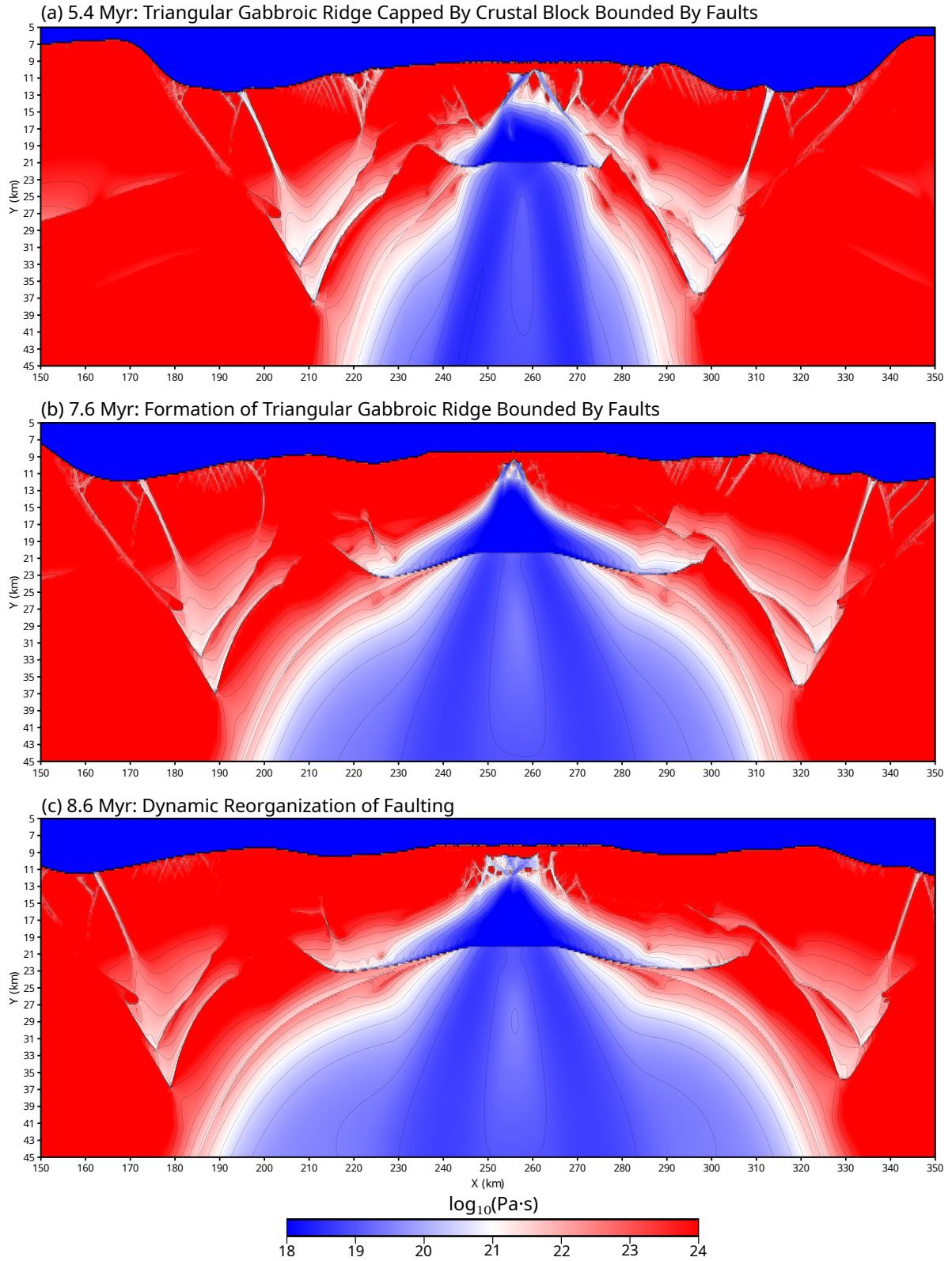
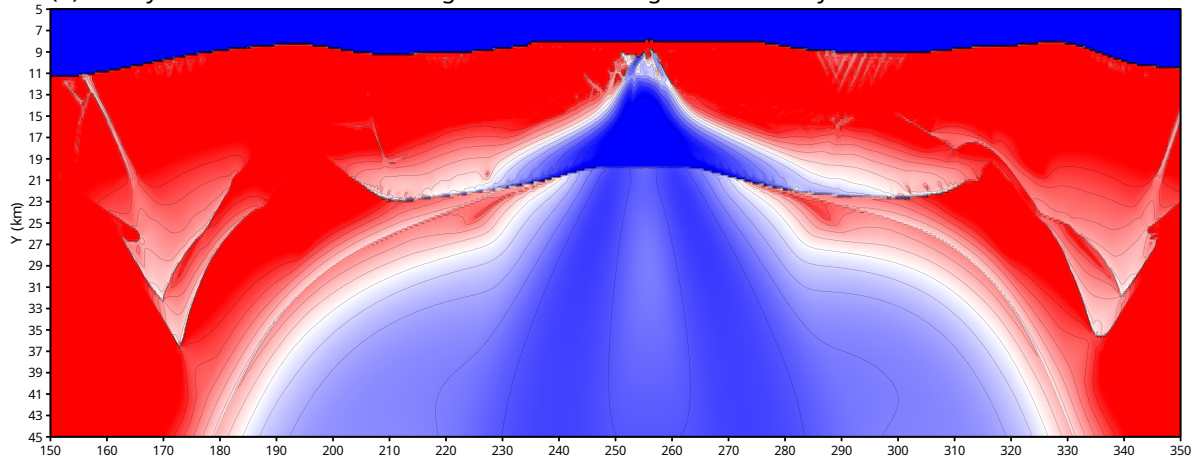
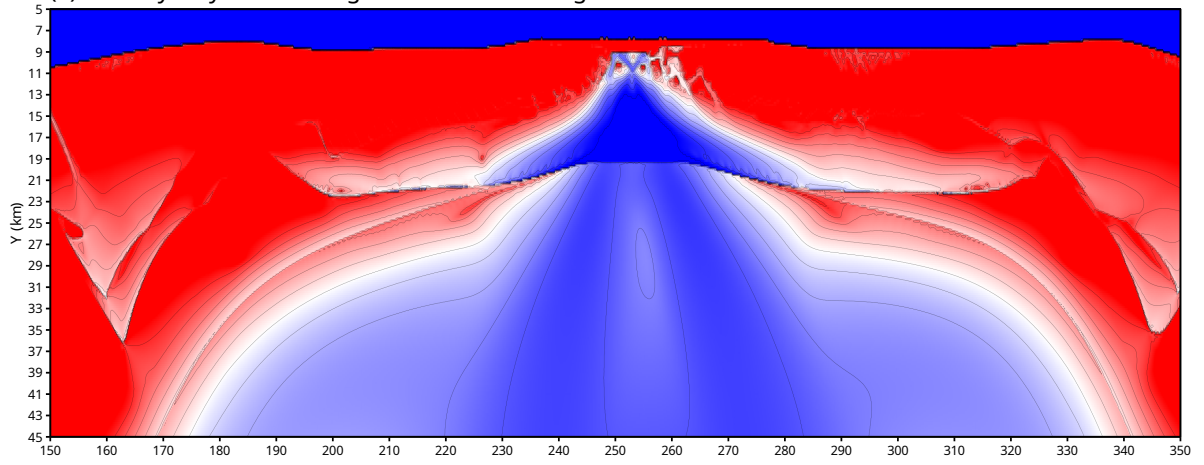


Figure S2: Visco-plastic viscosity for Case 19 (melt damage disabled) during (a-f) continued volcanic extrusion, formation of downward-concave volcanic flow packages that dip toward the ridge axis and ductile deformation in the gabbroic magmatic crust. Case 19 is associated with dynamic deformation at the ridge axis involving the repeated formation and destruction of a triangular horst at the ridge axis that produces downward-concave volcanic flow packages dipping toward the ridge axis. Contour interval is $0.5 \log_{10}(\text{Pa}\cdot\text{s})$.

(d) 9.2 Myr: Formation of New Triangular Gabbroic Ridge Boundaed By Faults



(e) 10.2 Myr: Dynamic Reorganization of Faulting



(f) 11.6 Myr: Formation of New Triangular Gabbroic Ridge Bounded By Faults

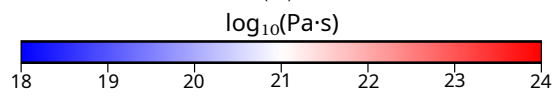
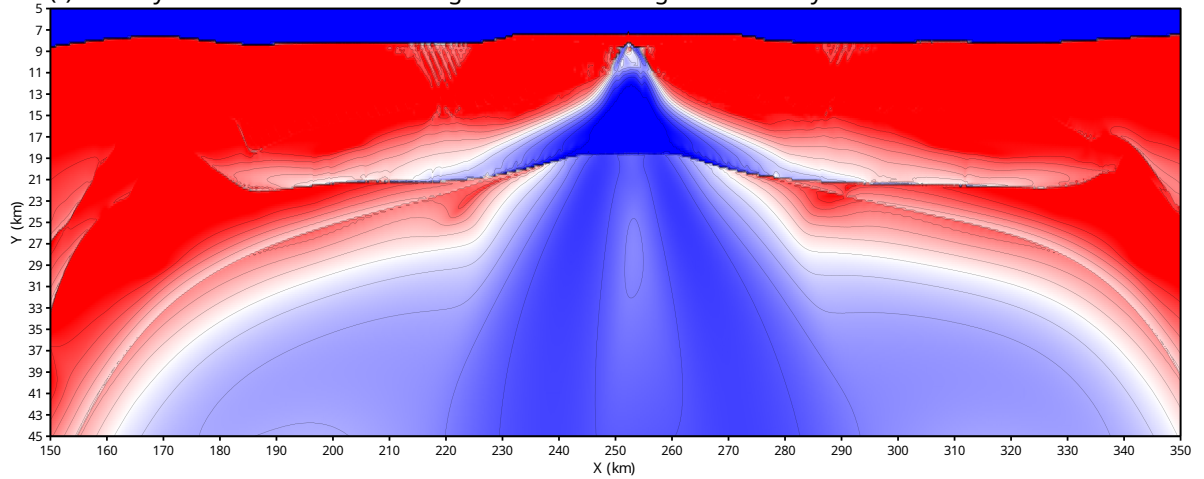


Figure S2: (continued)

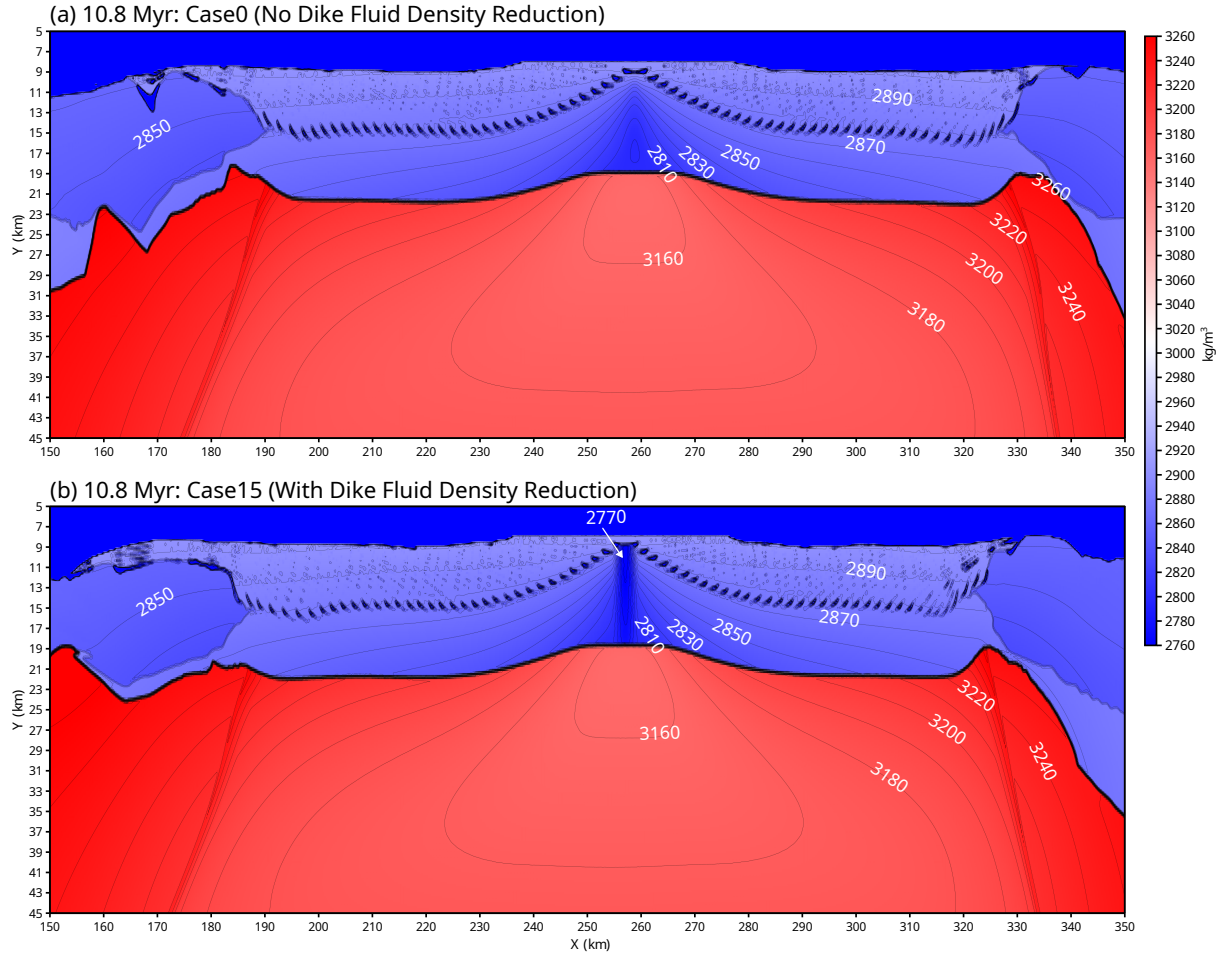


Figure S3: Density plots for (a) Case 0 (base case) and (b) Case 15. Case 15 differs from Case 0 in that the density of markers that are probabilistically selected for melt damage is modified to take into account the effects of injected dike fluid. Including the effects of dike fluid on density has a negligible effect on the overall structural evolution of the model due to the already reduced density associated with thermal expansion and partial melt. See the supplemental and main text for additional details. Contour interval is 10 kg/m^3 .

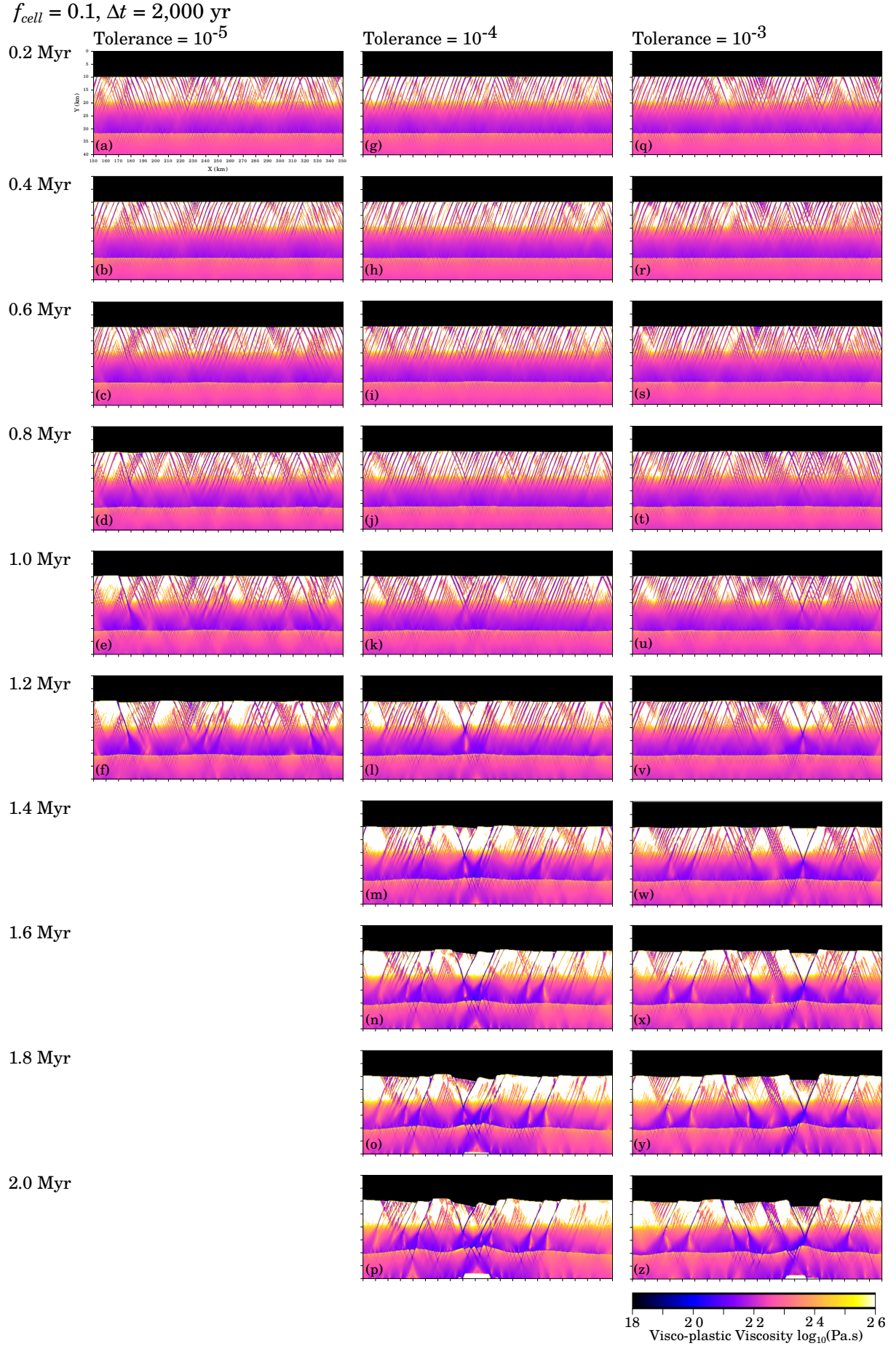


Figure S4: Convergence tests with $f_{cell} = 0.1$ and $\Delta t = 2,000 \text{ yr}$ for cases with a Picard tolerance of (a-f) 10^{-5} , (g-p) 10^{-4} and (q-z) 10^{-3} .

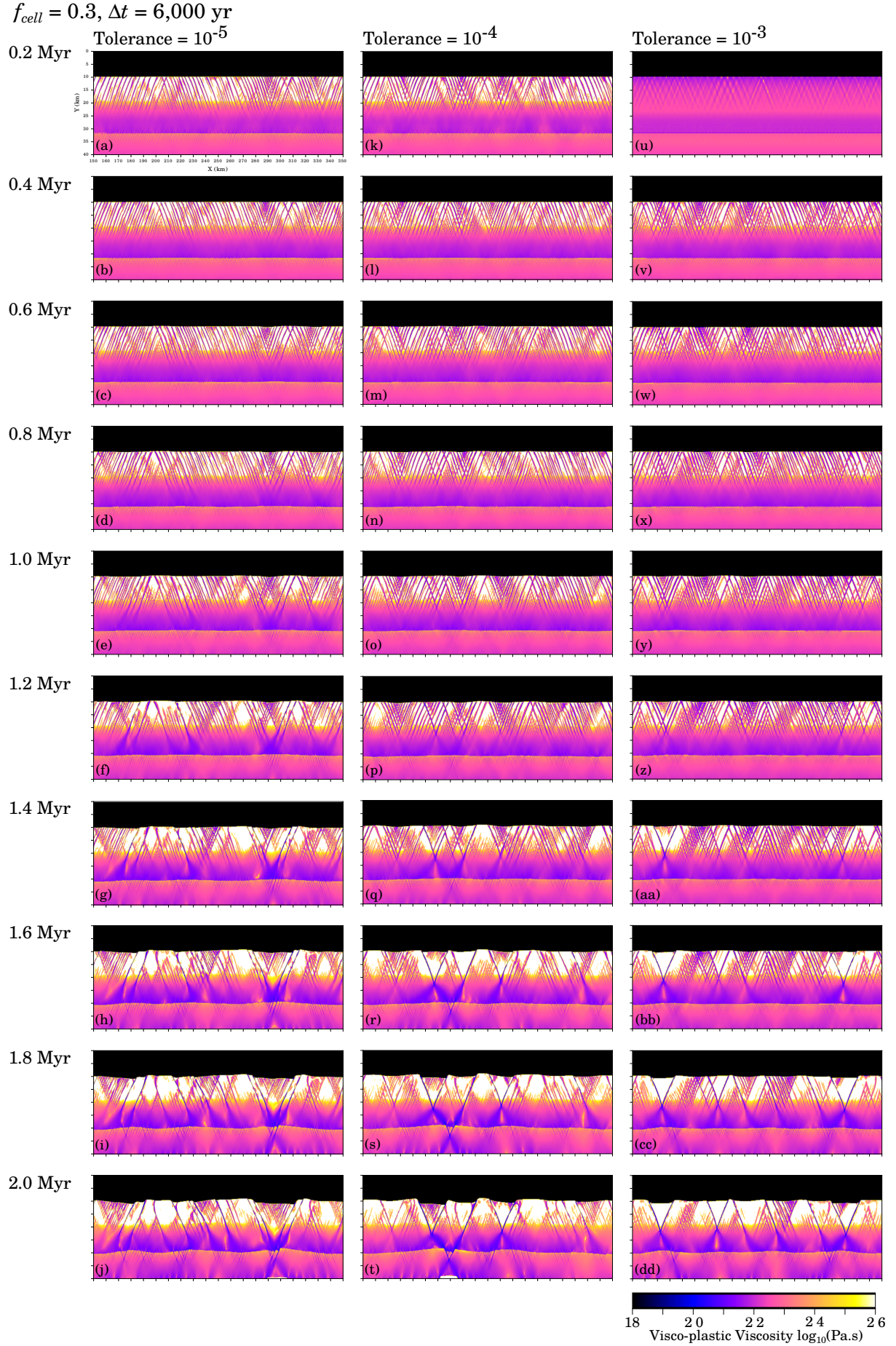


Figure S5: Convergence tests with $f_{cell} = 0.3$ and $\Delta t = 6,000$ yr for cases with a Picard tolerance of (a-j) 10^{-5} , (k-t) 10^{-4} and (u-dd) 10^{-3} .

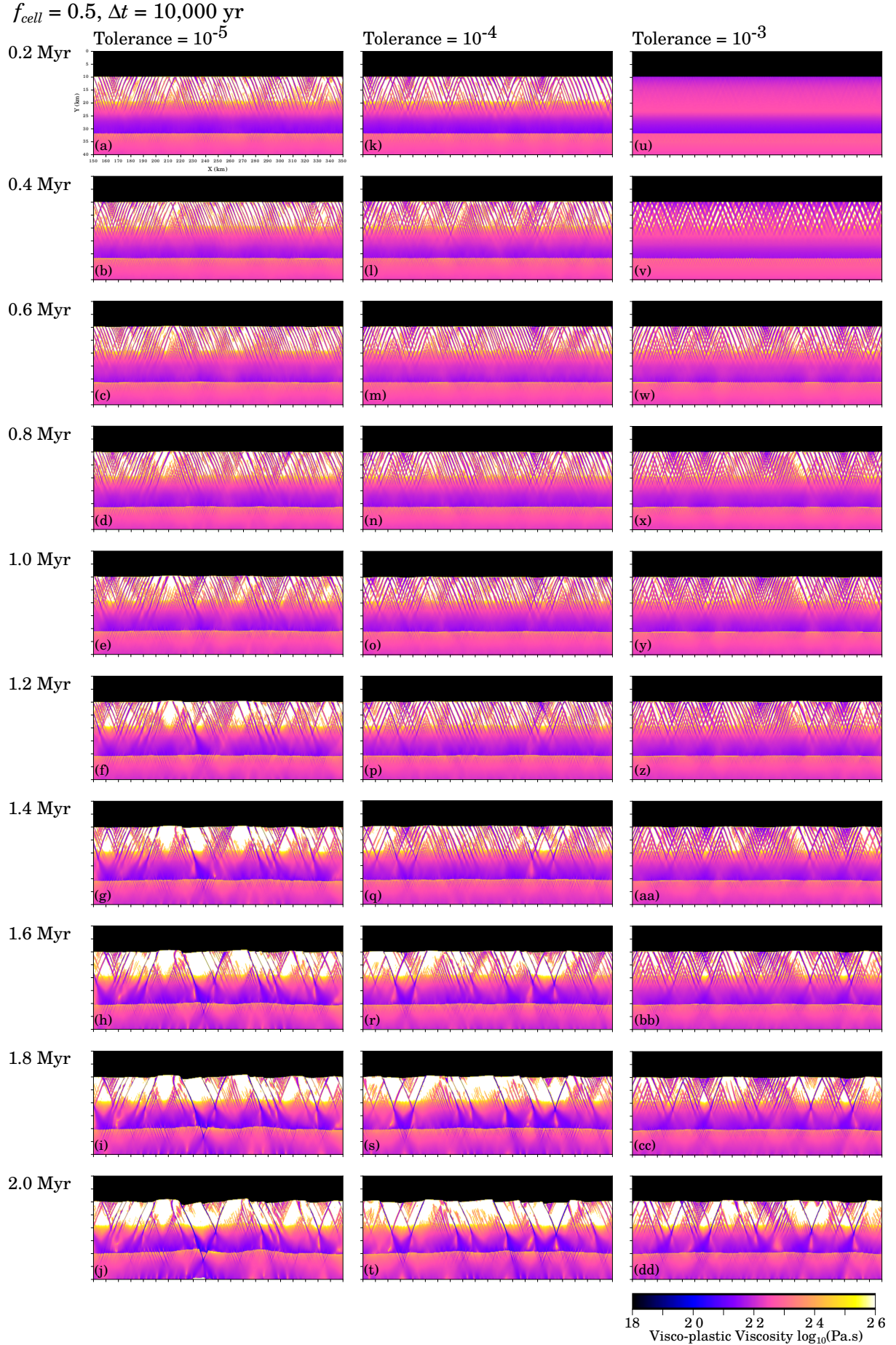


Figure S6: Convergence tests with $f_{cell} = 0.5$ and $\Delta t = 10,000$ yr for cases with a Picard tolerance of (a-j) 10^{-5} , (k-t) 10^{-4} and (u-dd) 10^{-3} .

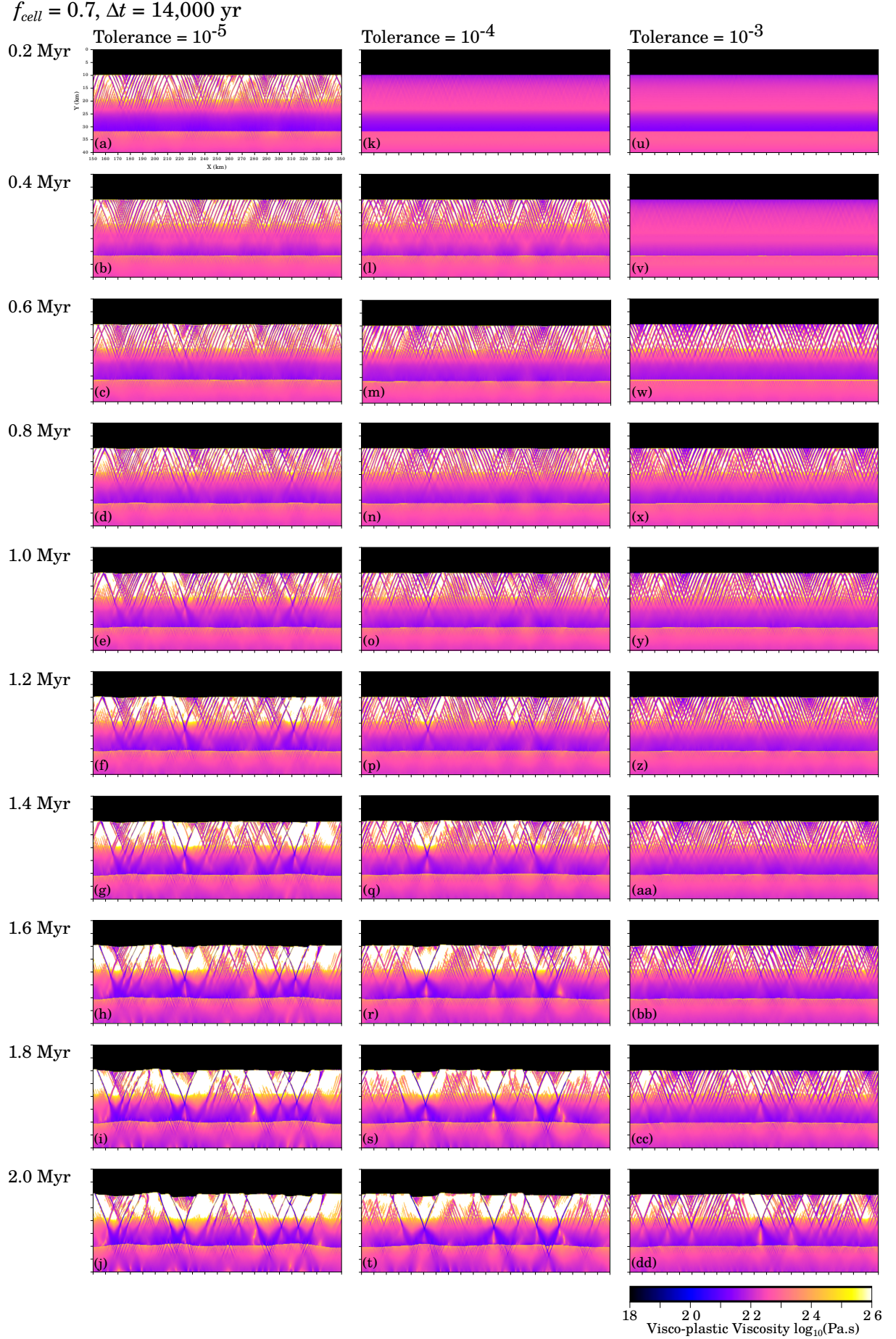


Figure S7: Convergence tests with $f_{cell} = 0.7$ and $\Delta t = 14,000 \text{ yr}$ for cases with a Picard tolerance of (a-j) 10^{-5} , (k-t) 10^{-4} and (u-dd) 10^{-3} .

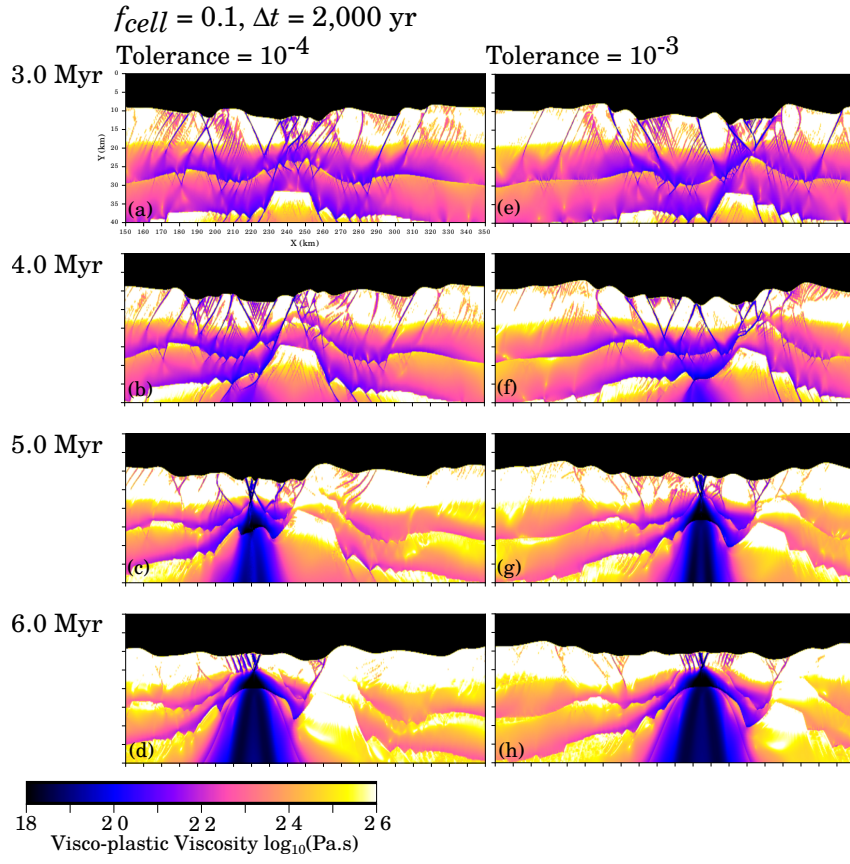


Figure S8: Convergence tests during the later phases of extension and breakup with $f_{cell} = 0.1$, $\Delta t = 2,000 \text{ yr}$, (a-d) Picard tolerance of 10^{-4} and (e-h) Picard tolerance of 10^{-3} .

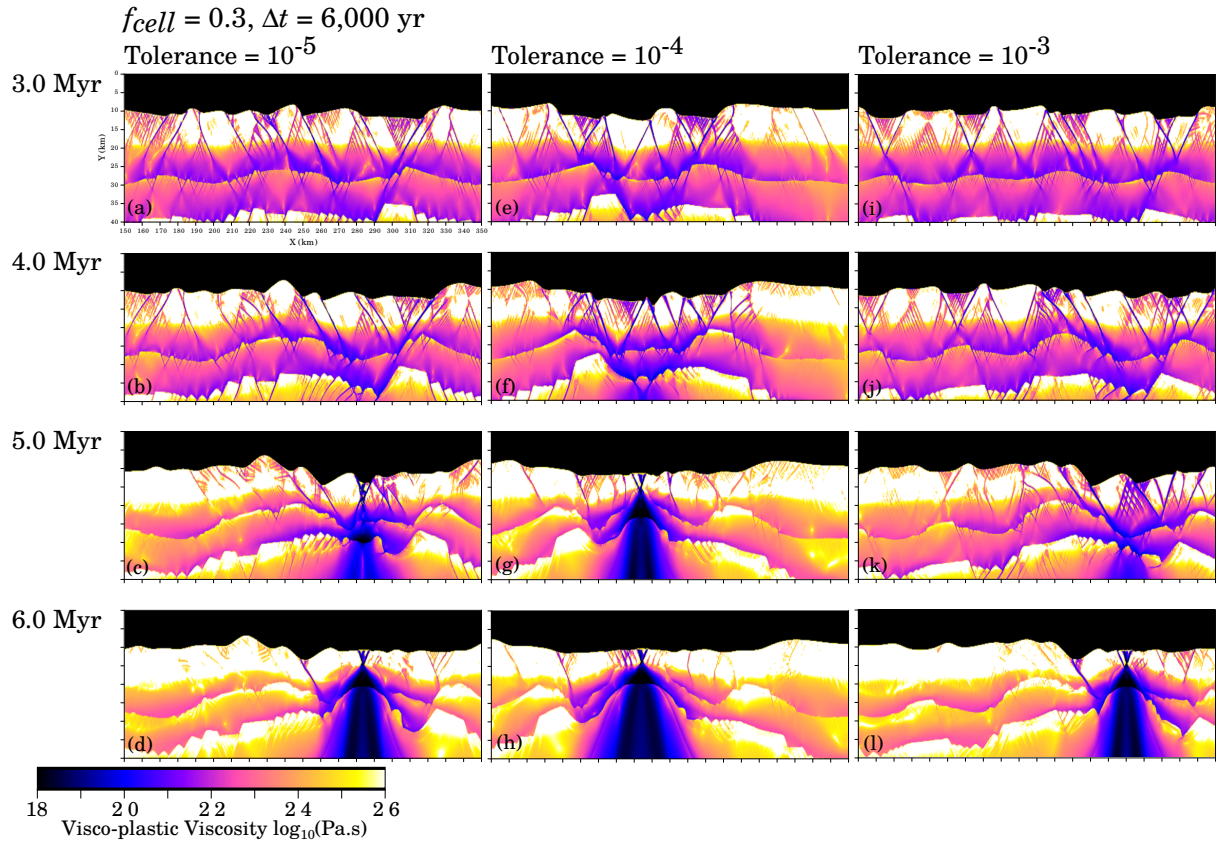


Figure S9: Convergence tests during the later phases of extension and breakup for cases with $f_{cell} = 0.3$ and $\Delta t = 6,000 \text{ yr}$ for cases with a Picard tolerance of (a-d) 10^{-5} , (e-h) 10^{-4} and (i-l) 10^{-3} .

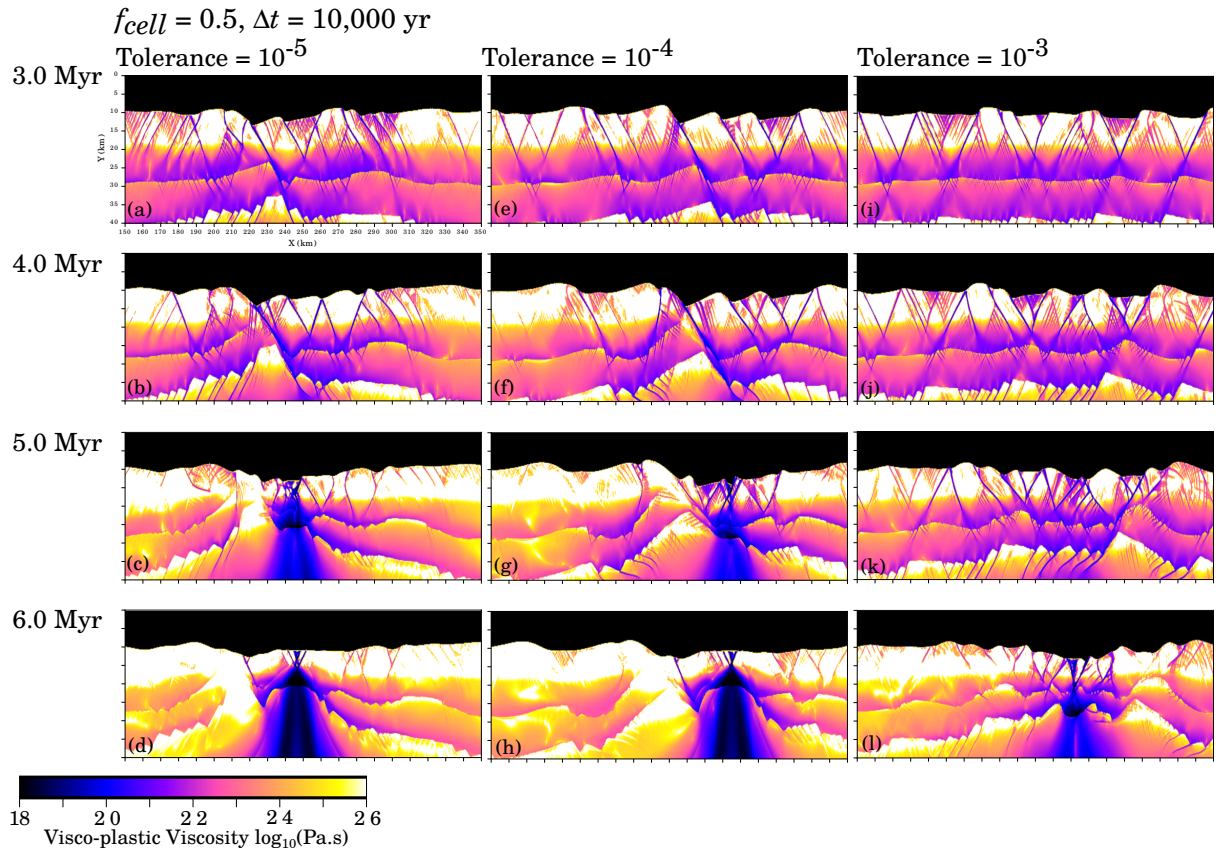


Figure S10: Convergence tests during the later phases of extension and breakup for cases with $f_{cell} = 0.5$ and $\Delta t = 10,000 \text{ yr}$ for cases with a Picard tolerance of (a-d) 10^{-5} , (e-h) 10^{-4} and (i-l) 10^{-3} .

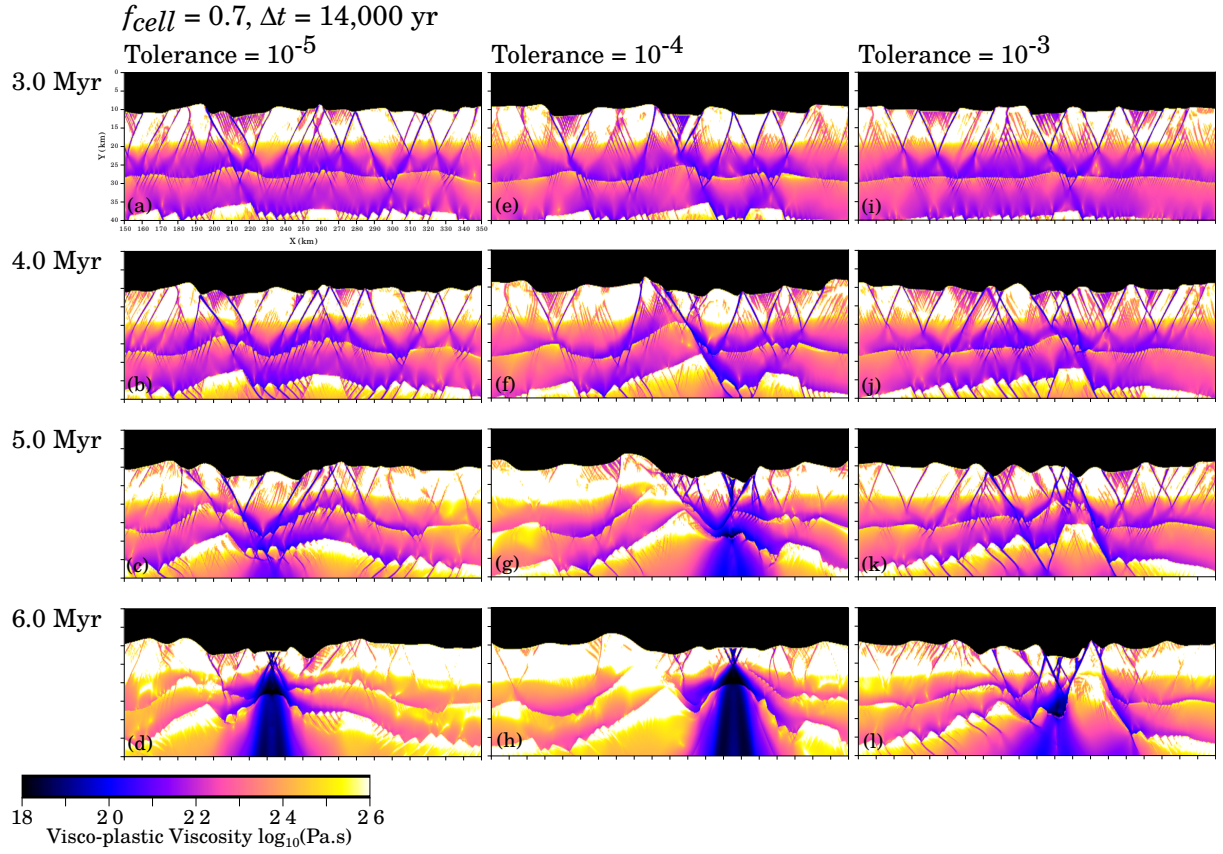


Figure S11: Convergence tests during the later phases of extension and breakup for cases with $f_{cell} = 0.7$ and $\Delta t = 14,000 \text{ yr}$ for cases with a Picard tolerance of (a-d) 10^{-5} , (e-h) 10^{-4} and (i-l) 10^{-3} .

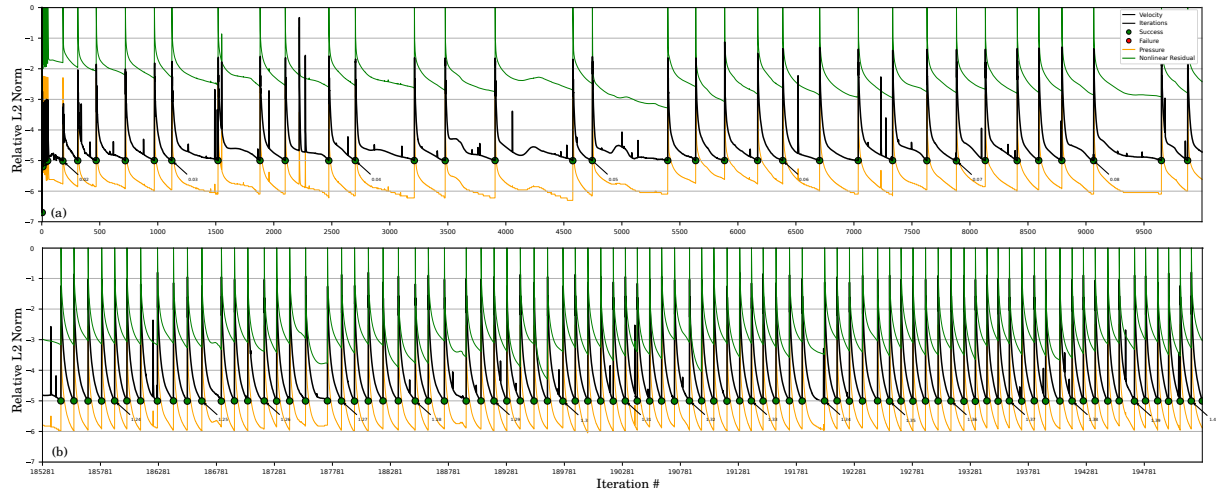


Figure S12: Picard iterations with $f_{cell} = 0.1$, $\Delta t = 2,000 \text{ yr}$, tolerance equal to 10^{-5} and maximum number of iterations equal to 1,000. Convergence of the Picard method is shown at (a) the beginning of extension and (b) at the beginning of shear-zone localization at around 1.2 Myr. These plots correspond to the model shown in Figure S4 (a-f).

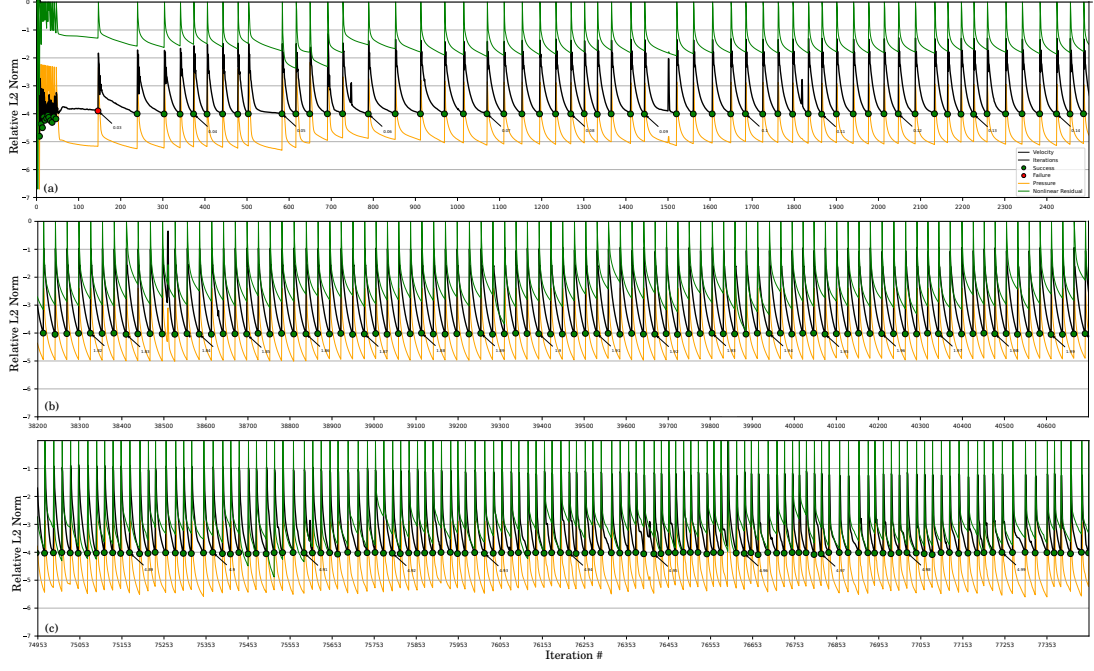


Figure S13: Picard iterations with $f_{cell} = 0.1$ and $\Delta t = 2,000$ yr, tolerance equal to 10^{-4} and maximum number of iterations equal to 100. Convergence of the Picard method is shown at (a) the beginning of extension, (b) after shear zones localize and distinct fault blocks form at around 2 Myr and (c) during onset of breakup and extrusion at around 5 Myr. These plots correspond to the model shown in Figure S4 (g-p) and Figure S8 (a-d).

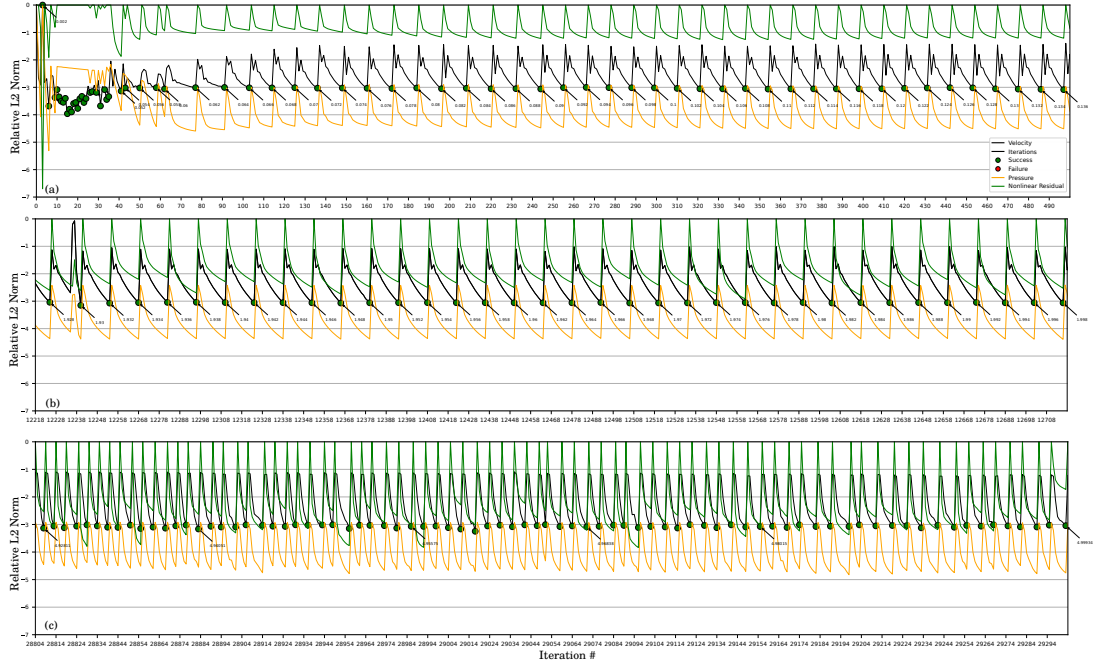


Figure S14: Picard iterations with $f_{cell} = 0.1$ and $\Delta t = 2,000$ yr, tolerance equal to 10^{-3} and maximum number of iterations equal to 100. Convergence of the Picard method is shown at (a) the beginning of extension, (b) after shear zones localize and distinct fault blocks form at around 2 Myr and (c) during onset of breakup and extrusion at around 5 Myr. These plots correspond to the model shown in Figure S4 (q-z) and Figure S8 (e-h).

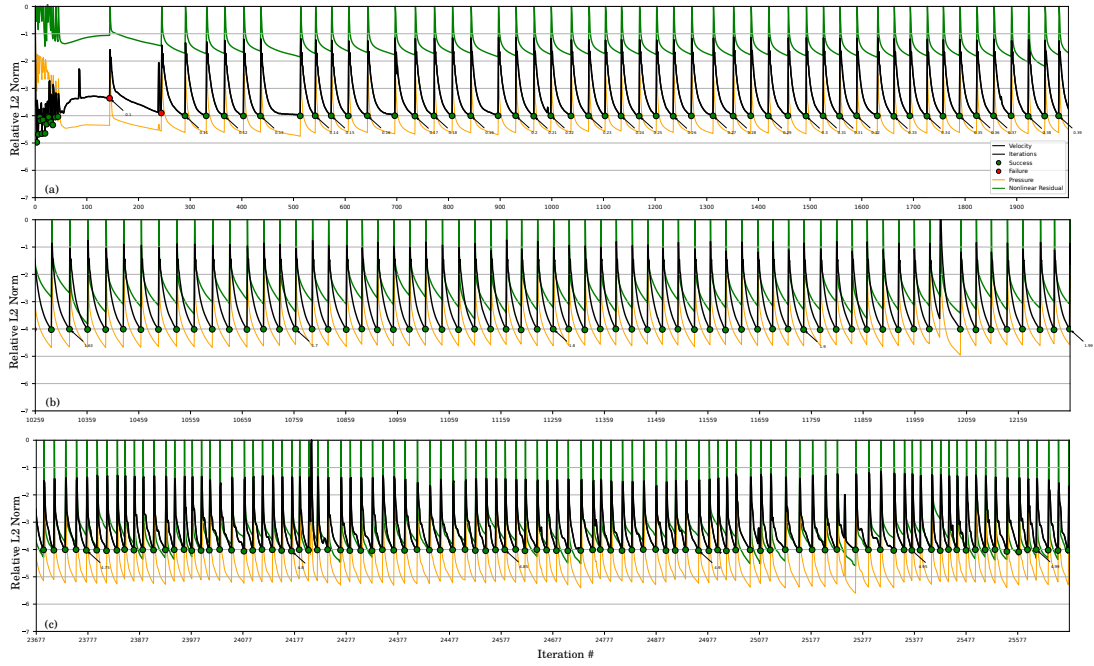


Figure S15: Picard iterations with $f_{cell} = 0.3$ and $\Delta t = 6,000$ yr, tolerance equal to 10^{-4} and maximum number of iterations equal to 100. Convergence of the Picard method is shown at (a) the beginning of extension, (b) after shear zones localize and distinct fault blocks form at around 2 Myr and (c) during onset of breakup and extrusion at around 5 Myr. These plots correspond to the model shown in Figure S5 (k-t) and Figure S9 (e-h).