

Constraining Lithospheric Delamination on Venus with Flexural Topography and Trench Geometry

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Abstract

Rollback-style lithospheric recycling has been proposed for a subset of Venusian coronae and other tectonic features based on the presence of trench-outer-rise systems, but the mechanism and conditions required to initiate recycling remain poorly understood. Peel-back delamination (PBD) is a proposed mechanism in which dense lithospheric mantle decouples and peels away from a thin layer of overlying crust, resulting in slab rollback. PBD has only been modeled in 2D, and therefore its 3D surface expression is unknown. Here, we present the first three-dimensional models of peel-back delamination on Venus and compare their predicted trench geometry and flexural topography with observations from Artemis Corona, Atahensik Corona, and Derceto Plateau. PBD initiation is facilitated by increasing lithosphere thickness which enhances the negative buoyancy of the lithospheric mantle and drives decoupling from the weak, buoyant crust. Once initiated, rift geometry, lithosphere structure around the rift, and maximum yield strength control the evolution of trench geometry throughout PBD. Elastic thickness (T_e) provides an independent constraint on lithospheric thickness, increasing with thermal lithospheric thickness. Comparisons with observations indicate that models of Artemis and Atahensik Corona best satisfy both trench geometry and flexural topography constraints, whereas the Derceto Plateau model reproduces trench geometry but not flexural topography. At Artemis Corona, along-trench variation in T_e may reflect differences in subsurface slab morphology and plate bending. T_e estimates from the southeastern trench at Artemis Corona favor a thermal lithospheric thickness of approximately 200–250 km for a 1700 K mantle.

1. INTRODUCTION

Although Venus lacks Earth-like plate tectonics, lithospheric recycling may still occur regionally. Key evidence for recycling was first identified from a series of trench-outer-rise structures associated with a subset of coronae (quasi-circular tectono-volcanic features) and a few other tectonic features (Sandwell & Schubert 1992a; Schubert & Sandwell 1992b; Schubert & Sandwell 1995). These structures exhibit flexural topography characteristic of plate bending and have been interpreted as evidence for rollback-style lithospheric recycling in the absence of plate-driven convergence. In this scenario, lithosphere sinks into the mantle while the trench migrates radially outward, producing arcuate trenches with flexural topography comparable to subduction systems on Earth. Although more than 700 coronae have been identified on Venus (Gülcher et al. 2025), only a small subset have been proposed as rollback lithospheric recycling sites based on their trench-outer rise morphology. These sites span a broad range of surface expressions, including differences in overall size, trench geometry, flexural topography, topographic relief, and geologic setting (Fig. 1). The origin of this diversity remains poorly understood, though it may reflect differences in lithospheric structure, evolutionary stage, and/or initiation mechanism.

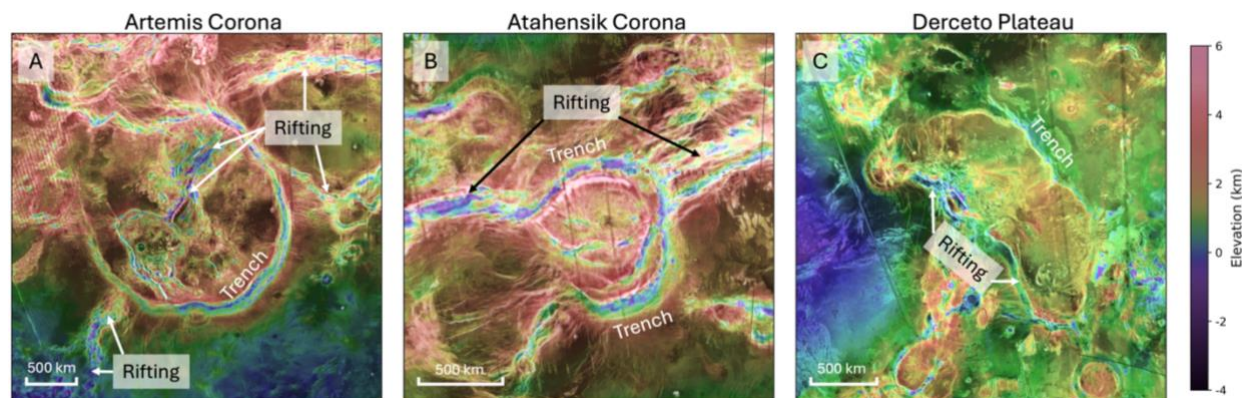


Figure 1: Proposed lithospheric recycling sites on Venus exhibit diverse morphologies. (A) Artemis Corona, an anomalously large quasi-circular trench-outer rise system. (B) Atahensik Corona, a moderately large, nearly circular feature with dual trench segments to the north and south. (C) Derceto Plateau, a non-corona recycling candidate with an elongated trench geometry. Magellan Cycle 3 Mosaic Image Data Record (C3-MDIR) synthetic aperture radar (SAR) imagery (Ford et al. 1993) is overlain by the Magellan Global Topographic Data Record (GTDR) (Rappaport et al. 1999), shown in color.

Two mechanisms have been proposed to explain the initiation of rollback-style recycling on Venus: plume-induced subduction and peel-back delamination (PBD). In PBD, dense, rift-adjacent lithospheric mantle detaches and peels away from the surface through a zone of weakened, low-viscosity crust, allowing rollback-style removal of lithospheric mantle while leaving a thin layer of buoyant upper crust at the surface (Adams et al. 2022, 2023). A similar style of coherent lithospheric peeling has also been proposed for continental lithosphere on Earth (Bird, 1979; Göğüş & Ueda, 2018) and the Archean Earth (Chowdhury et al. 2020). In Venus models, the onset of PBD is facilitated by plume-lithosphere interactions at pre-existing rift zones, consistent with the close association between rifting and many proposed recycling sites (Fig. 1). Two-dimensional models show that PBD only initiates when the lithosphere is sufficiently thick and negatively buoyant to destabilize (Adams et al. 2022, 2023). In plume-induced subduction, a buoyant mantle upwelling impinges and loads the lithosphere, leading to plate bending and outward trench migration (Sandwell & Schubert 1992a; Schubert & Sandwell 1995; Davaille et al. 2017; Gülcher et al. 2020). Laboratory models of plume-induced subduction demonstrated how a symmetric plume could lead to lithospheric tearing, creating a partial arc of subduction consisting of straight segments separated by tears (Davaille et al. 2017). Numerical models of plume-induced subduction have successfully reproduced a range of the most common corona morphologies under comparatively warm, thin lithospheric conditions, where plume emplacement is accompanied by magmatic weakening (Gülcher et al. 2020). The thick lithosphere required in models of PBD and the warm, thin, and magmatically weakened conditions in numerical models of plume-induced subduction suggest that variations in lithospheric thickness and strength may influence both the diversity of coronae on Venus and the viability of different coronae-formation mechanisms.

Flexural topography associated with trench-outer-rise systems provides an observational constraint on lithospheric structure at proposed recycling sites through estimates of elastic thickness (T_e), which provide a measure of lithospheric flexural rigidity and response to loading. Estimates of T_e have been widely used to investigate the thermal and mechanical structure of lithosphere on Venus, including regional variations in lithosphere thickness and heat flow (Johnson & Sandwell 1994; Schubert & Sandwell 1995; O'Rourke & Smrekar 2018, Smrekar et al. 2023; 2025). While most coronae occur on relatively low- T_e lithosphere, proposed recycling sites like Artemis Corona

and Atahensik Corona include some of the largest T_e values reported on Venus (Smrekar et al. 2023; 2025). Their high inferred flexural rigidity is reflected in high-amplitude, long-wavelength trench-outer rise topography that has not yet been quantitatively reproduced in models at Artemis- or Atahensik-like scales. However, T_e estimates also vary considerably within the subset of proposed recycling sites with features like Derceto Plateau having some of the smallest T_e estimates (Schubert & Sandwell 1995), indicating that rollback-style lithospheric recycling may operate across a broad range of lithospheric conditions.

Flexural topography provides one of the few observational constraints on the mechanical properties of Venus' lithosphere. However, previous studies have primarily focused on demonstrating the viability of recycling and reproducing first-order morphological characteristics. While planform feature geometry provides an important constraint on the size and spatial organization of plate bending in recycling systems, it does not uniquely constrain the thermal and rheological structure of the lithosphere. In contrast, flexural topography provides a more direct constraint on lithospheric structure but contains limited information about the three-dimensional evolution of plate bending and slab rollback. Simultaneously reproducing both the flexural topography and trench geometry of specific recycling sites provides a more robust test of recycling mechanisms than either constraint alone.

A key question to understanding lithospheric recycling on Venus is whether the proposed mechanisms can reproduce the observed combinations of trench geometry and flexural topography at any of the proposed recycling sites. Existing studies of PBD have been limited to 2D models, leaving the 3D surface expression largely unknown. Here, we extend PBD models to three dimensions under Venus-like conditions and compare the resulting trench geometry and flexural response with observations from three proposed recycling sites with significant diversity in their topographic and flexural characteristics: Artemis Corona, Atahensik Corona, and Derceto Plateau. By varying lithospheric thickness, lithospheric strength, rift geometry, and mantle temperature, we assess the viability of PBD under a range of conditions and investigate whether variations in rift geometry and lithospheric structure can explain the observed diversity in the surface expressions of proposed recycling sites.

2. METHODS

2.1 Model Setup

We compute a suite of thermo-mechanical models of mantle convection in three dimensions. Making the Boussinesq approximation with a temperature- and viscoplastic, depth-dependent rheology, the equations are solved with a finite-volume method using the code, StagYY, for planetary interiors (Tackley 2008). Simulations are conducted in a spherical-patch domain discretized on a $256 \times 512 \times 256$ grid, with vertical mesh refinement near the surface yielding a minimum radial spacing of 6.9 km and horizontal spacing of 19.0 km. A 152 km-thick low-density “sticky-air” layer overlays the mantle and approximates a free surface allowing the prediction of dynamic topography. Lateral boundaries are reflecting and the bottom boundary is free slip. The model domain extends from the surface to the core-mantle boundary. Unless otherwise stated, model details follow Adams et al. (2022, 2023).

The 3D model setup extends the two-dimensional setup of Adams et al. (2023) to investigate plume-induced PBD initiated at a rift zone. A small plume head (radius $R = 200$ km) is prescribed 700 km beneath a 250 km-wide lateral lithospheric discontinuity representing a simplified chasma rift. A half-space cooling thermal profile is used to prescribe lithospheric structure with a prescribed mantle temperature, T_{mantle} . We consider two endmember configurations: (1) a symmetric case with laterally uniform lithospheric thickness (T_L), and (2) an asymmetric case in which the lithosphere is progressively thinned on one side of the rift to a minimum thickness of 30 km and thickens to T_L over a lateral distance of 2,110 km (Fig. 2). These configurations promote double- and single-sided PBD initiation, respectively, by controlling the distribution of negative buoyancy across the rift. The symmetric case represents uniform lithosphere surrounding the rift, whereas the asymmetric case represents conditions in which insufficient negative buoyancy on one side of the rift (resulting from either locally thinned lithosphere or thickened, positively buoyant crust) prevents the initiation of PBD (Adams et al. 2022, 2023).

These idealized configurations are related to observed Venusian rift systems by approximating differences in rift geometry and lithospheric structure using simplified endmembers. Rifted regions surrounding Artemis and Atahensik Coronae are characterized by branching, discontinuous deformation, whereas Derceto Plateau is adjacent to a continuous, ~1800 km-long chasma (Fig. 1).

We represent these endmembers using short ($L_{\text{rift}} = 300$ km) and long ($L_{\text{rift}} = 900$ km) rift segments, corresponding to full lengths of 600 and 1800 km when mirrored across the equatorial reflecting boundary. Asymmetric lithosphere is used to represent cases where one side of the rift is sufficiently thick and negatively buoyant to destabilize (Artemis- and Derceto-like), while symmetric lithosphere represents cases where both sides can destabilize and produce dual-trench morphologies (Atahensik-like).

Viscosity follows a temperature-dependent Arrhenius diffusion-creep law ($E_a = 240$ KJ/mol) and is limited by a Drucker-Prager yield criterion. The brittle stress is given by $\sigma_{y,\text{brittle}} = C + \mu P$, where C is cohesion, P is lithostatic pressure, and μ is the friction coefficient. The effective yield stress is limited by a prescribed maximum yield stress, $\sigma_{\text{yield}} = \min(\sigma_{y,\text{brittle}}, \sigma_{\text{max}})$ and the effective viscosity is $\eta_{\text{eff}} = \min\left(\eta, \frac{\sigma_{\text{yield}}}{2\dot{\epsilon}_{II}}\right)$, where $\dot{\epsilon}_{II}$ is the second invariant of the strain rate tensor. In the mantle, yielding is defined by a cohesion $C_0 = 10$ MPa and friction coefficient $\mu = 0.25$ with an imposed maximum yield stress that caps the strength at depth. The surface is overlain by a 30 km-thick layer of buoyant basaltic crust that is uniformly weak ($C_0 = 10$ MPa, $\mu=0$), providing a simplified representation of a weak plagioclase rheology (Ranalli 1995) commonly used for Venus (Tian et al., 2023; Gülcher et al. 2020). This weak endmember formulation promotes crust-mantle decoupling and facilitates peel-back delamination under warm Venusian conditions. The models do not include melting or crust production.

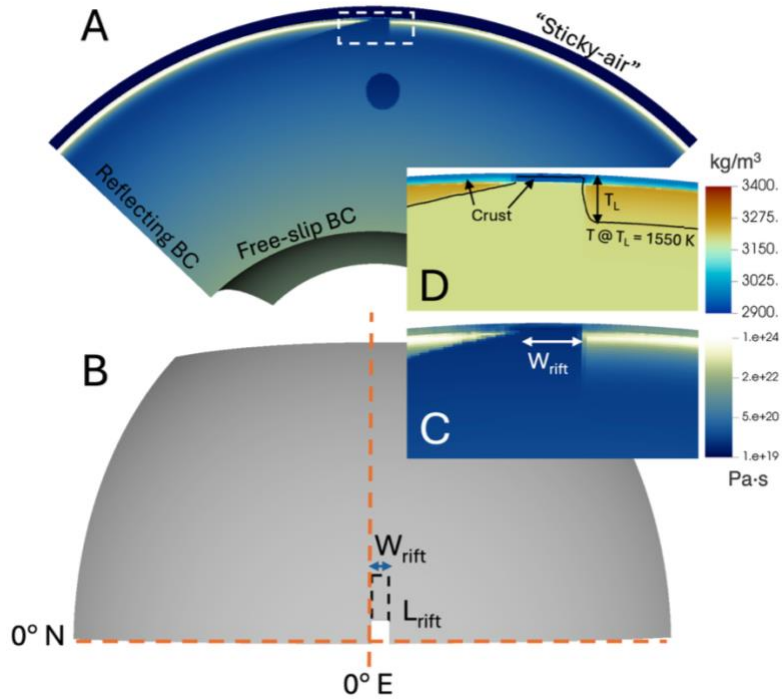


Figure 2: (A) 3D spherical-patch domain showing the initial viscosity field including a prescribed plume (top of plume head at 700 km depth) and a lithospheric rift with width $W_{rift} = 250$ km. (B) Top-down view of lithospheric mantle showing the finite rift geometry, where L_{rift} denotes the half-length of the imposed rift segment ($L_{rift} = 300$ (in white) or 900 km (dashed box)). (C) Zoomed-in view of the asymmetric rift configuration shown in the density field, where thermal lithosphere thickness (T_L) is defined by the 1550 K isotherm. (D) Corresponding viscosity structure for the asymmetric lithosphere case (white box in (A)). Sticky air layer omitted in (B-D) and crust omitted in (B).

2.2 Parameter Space

We explore a range of lithospheric thicknesses $T_L = 100$ –250 km, maximum yield stress $\sigma_{yield} = 60$ –200 MPa, mantle temperatures 1700–1900 K, and rift length $L_{rift} = [300, 900]$ km and symmetry/asymmetry to evaluate their influence on the initiation and surface expression of PBD. These parameters are varied to bracket the transition between stable and unstable regimes and identify a possible minimum lithospheric thickness required to sustain delamination, as previously examined in 2D models (Adams et al. 2022, 2023).

2.3 Phase Transitions

The phase assemblages and densities for Venusian basalt and a harzburgitic mantle are computed using `Perple_X` (Connolly 2009) with the Holland and Powell thermodynamic database (Holland & Powell 2011). Gibbs free energy is minimized over $T = 300\text{--}2000$ K and $P = 0.01\text{--}8$ GPa, spanning the basalt-granulite-eclogite transition under Venus-like conditions. Bulk compositions follow Table 2 of Semprich et al. (2025) with Venus basalt based on Venera/Vega XRF measurements. Because XRF reports total iron, all iron is treated as FeO (fully reduced, dry) consistent with comparable experimental data (Ito & Kennedy 1971) and recent estimates of a reduced Venus interior (Warren & Kite 2023; Weller & Kiefer 2023). The mantle is represented by a moderately depleted harzburgite produced by 15% batch melting (Shaw 1970) of a primitive upper mantle composition (McDonough & Sun 1995). From the `Perple_X` output, we construct a 2D lookup table of crust-mantle density contrasts:

$$\Delta\rho(P,T) = \rho_{\text{crust}}(P,T) - \rho_{\text{mantle}}(P,T) \quad (1)$$

used to compute relative crustal densities. At pressures higher than 5 GPa, the crust-mantle density contrast is constant at $\Delta\rho$ ($P = 5$ GPa, $T = 1700$ K) except for the basalt barrier zone discussed below. Because the incompressible models do not explicitly include adiabatic heating, a depth-dependent temperature correction of 0.5 K km^{-1} is added to the modeled temperature when querying the lookup table, consistent with the mantle adiabatic gradient adopted in previous Venus geodynamic models (Gerya 2014). This correlation affects only the thermodynamic calculation of phase assemblages and densities and the temperature field that evolves within the models remains unmodified. Tracers used to define the crust carry local pressure and temperature as they are advected; they are used to query this table via bilinear interpolation and assign a density contrast $\Delta\rho(P,T)$ value to each crustal tracer.

Mid-mantle phase transitions are implemented following Adams et al. (2022, 2023). The postspinel phase transition (corresponding to the transformation of ringwoodite to bridgmanite and ferropericlasite) is parameterized using a negative Clapeyron slope of -1 MPa/K (Katsura et al. 2003; Fei et al. 2004). Sinking, cold lithosphere undergoes delayed transformation, resulting in buoyant anomalies due to delayed transition beginning at 710 km depth. A basalt barrier, or garnet trap, is included and represents the positive buoyancy of basaltic crust relative to the surrounding mantle at transition zone depths (Davies 2008; Ogawa & Yanagisawa 2014). This effect is

parameterized by maintaining a crust-mantle density contrast of ($\Delta\rho = -130$ kg/m³) across the 710-770 km depth range, thereby resisting slab descent through the transition zone.

2.4 Estimating Elastic Thickness from Flexure

Model topography was calculated from the surface of the crust tracers along profiles oriented perpendicular to the trench. Topographic profiles extended 1000 km radially outward from the trench to capture the forebulge and far-field topography. For each model profile, tracers within $\pm 1^\circ$ of the profile were divided into 8 km bins along the profile, and the maximum tracer radius within each bin was taken to represent the surface elevation. This procedure is used to provide an estimate of the upper surface of the crustal tracer point cloud (Fig. A1). For visualization, model topography shown in figures is represented by a cubic smoothing spline through the binned surface elevation data.

Elastic thickness (T_e) was estimated by fitting a broken elastic plate model to the binned tracer elevation data using the code used in Smrekar et al. (2023) after McNutt (1984). Surface deflection is described by elastic plate bending under an applied load assuming a Cartesian broken plate geometry. The flexural profile is given by:

$$w(x) = -e^{-x/a} \left[c_1 \cos\left(\frac{x}{a}\right) + c_2 \sin\left(\frac{x}{a}\right) \right] + c_3 x + c_4 \quad (2)$$

where x is distance from the trench, $w(x)$ is vertical deflection, c_1 and c_2 control the amplitude and phase of the flexural signal, c_3 accounts for long wavelength tilt, and c_4 is vertical offset. The flexural parameter, α , is related to T_e and flexural rigidity, D by:

$$\alpha = \left(\frac{4D}{\Delta\rho g} \right)^{1/4} \quad (3), \quad D = \frac{ET_e^3}{12(1-\nu^2)} \quad (4)$$

Which gives:

$$T_e = \left[\frac{3\Delta\rho g \alpha^4 (1-\nu^2)}{E} \right]^{1/3} \quad (5)$$

Where $E = 100$ GPa is Young's modulus, $\nu = 0.25$ is Poisson's ratio, $g = 8.87$ m/s² is the Venusian surface gravity, and $\Delta\rho = 3300$ kg/m³ is the density contrast. Model fits were obtained using a Markov Chain Monte Carlo (MCMC) approach implemented with the emcee Python package (Foreman-Mackey et al. 2013). The MCMC returns posterior distributions of α which are converted to T_e using the expressions above. Although the fitted flexural model assumes elastic plate bending, the evolution models do not include elastic

deformation, and the modeled flexural topography develops through viscoplastic bending. Without elasticity, bending stresses in the cold, high-viscosity lithosphere may be overestimated (Patočka et al, 2019); therefore, future work should investigate the effects of elastic deformation on flexural topography. Errors are presented as the 68% confidence interval, however additional error in observed T_e estimates stems from local variations in forebulge topography.

For comparison with model topography, profiles were also taken across trench-outer rise structures in Magellan topography at Artemis Corona, Atahensik Corona, and Derceto Plateau and analyzed using the method described above. To remove long-wavelength regional topography unrelated to flexure, a linear trend was fit to the far-field portion of each full topographic profile, excluding the first 400 km containing the trench and forebulge, and subtracted from the corresponding profile. Observed profiles were then trimmed to exclude portions of the trench influenced by non-flexural processes, only retaining the smooth trench-forebulge signal reflective of plate bending (e.g., Fig. A2). Because T_e is sensitive to trench-forebulge wavelength, modeled and observed profiles were restricted to the portion of the topography interpreted to represent flexural deformation prior to fitting. Modeled profiles were trimmed so that they began in the trench 0.5 km below the far field mean elevation, excluding the deeper trench from the elastic plate fit to better compare to observed trench-outer-rise topography.

3. RESULTS

Model variables and outcomes for 14 computations are summarized (Table 1). Model behavior is separated into two endmembers of plume-rift interactions: (1) stable, stagnant lithosphere, or (2) peel-back delamination. Across the parameter space, lithospheric thickness exerts the primary control on model outcome. All models with $T_L = 100$ km remain stable, whereas all models with thickness $T_L \geq 150$ km undergo PBD under otherwise comparable conditions. Variations in maximum yield stress (60–200 MPa) and mantle temperature (1700–1900 K) influence the timing of PBD and feature geometry (length-to-width ratio), but do not alter the outcome (PBD vs stagnant lid) once this thickness threshold is exceeded. Rift geometry and lithospheric symmetry surrounding the rift zone do not determine whether PBD occurs but instead influence the resulting surface morphology and style of single vs dual slab sinking.

Model #	Target Feature	Thermal Lithosphere Thickness (km)	Half-Space Cooling Age (t_{age}) (Myr)	T_{mantle} (K)	Initial rift half-length (L_{rift}) (km)	Lithosphere symmetry across rift	Maximum yield stress (σ_{yield}) (MPa)	Result	Best fit model time (Myr)	T_e at best-fit model time (km)
1	Artemis	100	110	1700	300	asymmetric	200	stagnant	-	-
2		150	245	1700	300	asymmetric	200	PBD	7.4	27.5
3		150	245	1700	300	asymmetric	100	PBD	5.8	27.7
4		150	245	1700	300	asymmetric	60	PBD	n/a	n/a
5		110	245	1900	300	asymmetric	200	PBD	1.0	24.6
6		200	435	1700	300	asymmetric	200	PBD	6.5	38.2
7		145	435	1900	300	asymmetric	200	PBD	0.8	35.6
8		250	685	1700	300	asymmetric	200	PBD	5.8	60.0
9		220	685	1900	300	asymmetric	200	PBD	0.7	48.8
10	Atahensik	100	110	1700	300	symmetric	200	stagnant	-	-
11		150	245	1700	300	symmetric	200	PBD	4.1	37.4
12		200	435	1700	300	symmetric	200	PBD	3.9	45.0
13	Derceto	75	110	1900	900	asymmetric	200	stagnant	-	-
14		100	110	1700	900	asymmetric	200	stagnant	-	-
15		150	245	1700	900	asymmetric	200	PBD	3.8	34.3

Table 1: Summary of model parameters and outcomes. Best-fit model times and corresponding elastic thickness estimates (T_e) are reported for models that reproduce the observed geometry of Artemis Corona, Atahensik Corona, or Derceto Plateau. T_e values are taken from single profiles through the equatorial plane at latitude 0° perpendicular to the initial rift. n/a indicates a model that underwent PBD but does not provide a geometric match (i.e., comparable size and length to width ratio) to any of the case study sites.

3.1 Time Evolution of Peel-Back Delamination

The time evolution of the reference model (Model #6; Figure 3-4) illustrates the evolution of plume-induced PBD. The initial configuration for a model with $L_{rift} = 300$ km and asymmetric lithospheric thickness (Fig. 3A) illustrates how a buoyant plume rises into the rift, producing localized yielding of the weak crust. Yielded crust forms a low-viscosity crustal zone above and adjacent to the rift (Fig. 4A), and this viscosity contrast across the base of the crust enables lower crust and lithospheric mantle to detach and peel away from a thin layer of crust remaining at the surface. As the lower crust attached to the slab sinks it transitions from buoyant basalt to dense eclogite, resulting in an increase in negative buoyancy that helps sustain slab sinking (Fig. 4E). PBD preferentially initiates on the thicker lithosphere side of the rift, while the thinner side remains stable and coupled to the surface due to plume underplating (Figs. 3B-C).

As the lithosphere detaches and sinks deeper, the slab tip encounters the mid-mantle phase transitions at 710-770 km depth (Figs. 3D-F). A zone of positive buoyancy within the crust and lithospheric mantle results from the basalt barrier (BB) and the deflection of the postspinel phase boundary (PSB), respectively. This buoyant zone laterally deflects the slab tip and inhibits descent. Sinking accelerates once the slab tip passes through the positive buoyancy zone, which increases stresses at the slab hinge causing thinning and tearing of the slab at the surface along the arcuate trench (Fig. 3F).

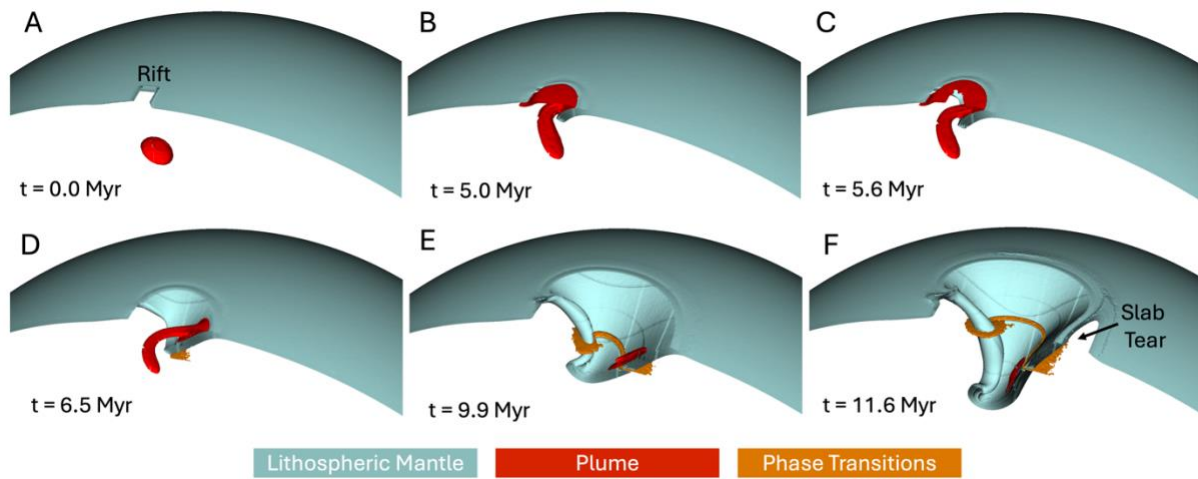


Figure 3: Time evolution of Model #6 ($T_L = 200$ km, $L_{\text{rift}} = 300$ km, asymmetric lithosphere) from the initial condition to 11.6 Myr. Panels show oblique cutaway views through the 3D regional model showing the interior on one side of the equatorial plane. Crust is omitted to highlight the evolution of the plume and lithospheric mantle. Panels show the initial plume beneath the rift (A), plume impingement and lithospheric detachment (B-C), progressive rollback and slab descent (D-E), and development of a slab tear near the surface (F). Orange contours indicate the phase-transition interval (PT; 710-770 km depth). Light blue contours denote delaminating lithospheric mantle ($3225 < \rho < 3310$ kg m⁻³) and red contours denote hot plume material ($T > 1725$ K).

3.2 Surface Evolution

The surface expression of PBD evolution is shown in Fig. 4I-P. Following the initial plume-rift interaction, a low-amplitude, arcuate trench-outer-rise structure forms within the thicker delaminating lithosphere surrounding the rift. As slab sinking continues, the trench expands radially outward and

deepens as the height of the flexural rise increases and the trench-forebulge wavelength decreases. The ends of the trench are coupled to the ends of the rift zone throughout delamination, forming a quasi-circular trench geometry that expands radially outward over time (Fig. 4J-L). Topography is highest over the thinned edge of lithosphere due to the isostatic uplift and underplating of the warm plume and remains at high elevation throughout the model evolution. Though the interior of the delamination area is initially elevated due to plume-induced thermal uplift (Fig. 4I), it undergoes broad subsidence driven by the motion of the sinking slab. Since models lack melting or crust formation, we leave the topography of the interior for future investigation.

Flexural profiles extracted from model topography yield elastic thickness estimates that evolve over time despite no changes in lithospheric structure. Best-fit elastic plate models yield initially high T_e values during the onset of PBD (Fig. 4M), which decrease over ~ 0.6 Myr to a quasi-steady-state value during continued rollback (Fig. 4N-P). This initial decrease reflects an adjustment in the characteristic wavelength of flexure as the plate transitions from initial bending to sustained rollback rather than a change in intrinsic lithospheric strength. Once rollback is established, the trench-forebulge wavelength and inferred T_e remain approximately constant, while the amplitude of the flexural rise progressively increases as the slab sinks (Fig. 4N-P). For a plate of constant flexural rigidity, increasing the magnitude of an applied load increases flexural amplitude without changing the characteristic flexural wavelength (Turcotte & Schubert 2002). Thus, the increasing forebulge amplitude is consistent with increasing slab pull during continued sinking, while the nearly constant wavelength and T_e indicate little change in effective plate rigidity.

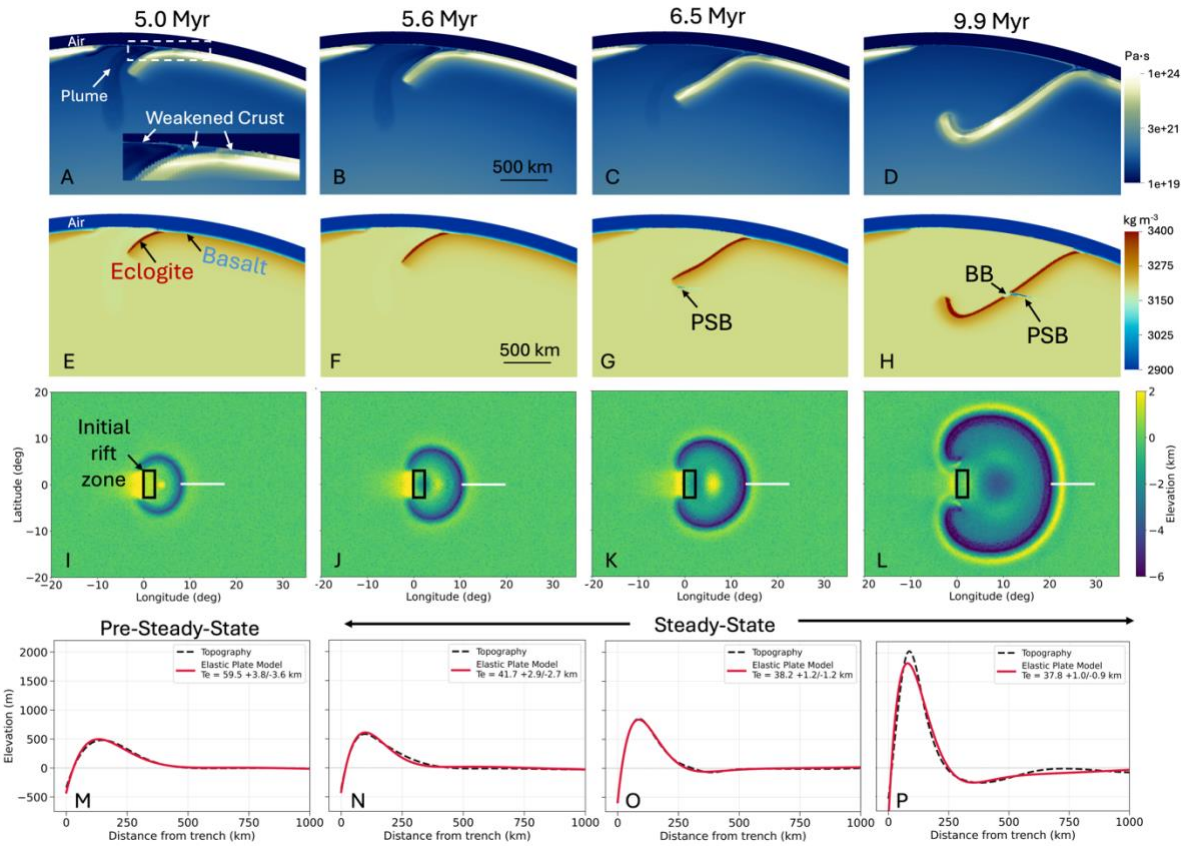


Figure 4: Time evolution of Model #6 at 5.0, 5.6, 6.5, and 9.9 Myr (left to right; corresponding to Fig. 3B-E). Rows show (A-D) viscosity and (E-H) density along a rift-normal, equatorial cross section through the center of the delaminating slab, (I-L) surface elevation calculated from crust tracer data, and (M-P) elevation profiles extracted along 0° latitude (white lines in I-L) with best-fit elastic plate models and estimated elastic thickness (T_e). Surface topography is mirrored across the equatorial plane. The inset in (A) highlights plume-induced weakening of the crust adjacent to the rift. BB and PSB denote the basalt barrier and postspinel phase-boundary deflection, respectively.

3.3 Artemis Corona

The $T_L = 200$ km reference model with a short rift ($L_{\text{rift}} = 300$ km) and asymmetric lithosphere thickness surrounding the rift zone (Model #6; Fig. 2-4) produces a quasi-circular, single trench resembling the size and geometry of Artemis Corona (Fig. 5). The rift-parallel length of Artemis Corona (~ 1900 km) is used to constrain a best-fit model time of 6.5 Myr where the model geometry most closely matches the observed trench geometry. At this time, the

rift-normal distance from the trench to rift edge, together with the model rift dimensions (600 x 250 km), places the model-equivalent initial rift within the observed rifted region in the interior of Artemis Corona (Fig. 5A). This inferred rift location is interpreted as a model-consistent estimate rather than a unique solution, as the initial rift geometry may have been modified by subsequent tectonic extension or volcanism.

Flexural topography provides a complementary constraint on model-data agreement. A reference topographic profile is selected along the southeast margin of Artemis Corona where the trench-outer-rise signal is well defined (blue line in Fig. 5A). The long wavelength downward sloping regional topographic component is removed from the observed profile as described in Section 2.4. A representative topographic profile across the model trench-outer-rise structure at the geometric best-fit model time (red line in Fig. 5B) is compared to the observed reference profile from Magellan topography in Fig. 5C. The observed and modeled topography yield a comparable flexural amplitude, wavelength, and trench-to-forebulge distance, producing elastic thickness estimates of 38.2 km and 37.8 km for model and observations, respectively. The profile in Fig. 5A represents a well-defined example of the observed flexural signal rather than a uniform T_e around the entire trench system. Additionally, flexural morphology differences result in spatial variations in T_e at Artemis Corona that are discussed further in Section 4.5. Model T_e is relatively consistent spatially along the modeled trench but decreases slightly toward the rift; observed profiles 1-5 (labeled in Fig. 5B) yield T_e values of 38.2, 39.2, 37.3, 37.1, and 35.2 km, respectively. At the geometric best-fit model time, the slab tip has reached 700 km depth, and the cooled mantle surrounding the slab becomes positively buoyant due to interactions with the postspinel phase boundary deflection at 710 km depth (Fig. 5D). This stage is within the range of quasi-steady-state bending characterized by little variation in inferred elastic thickness (Fig. 4N-P).

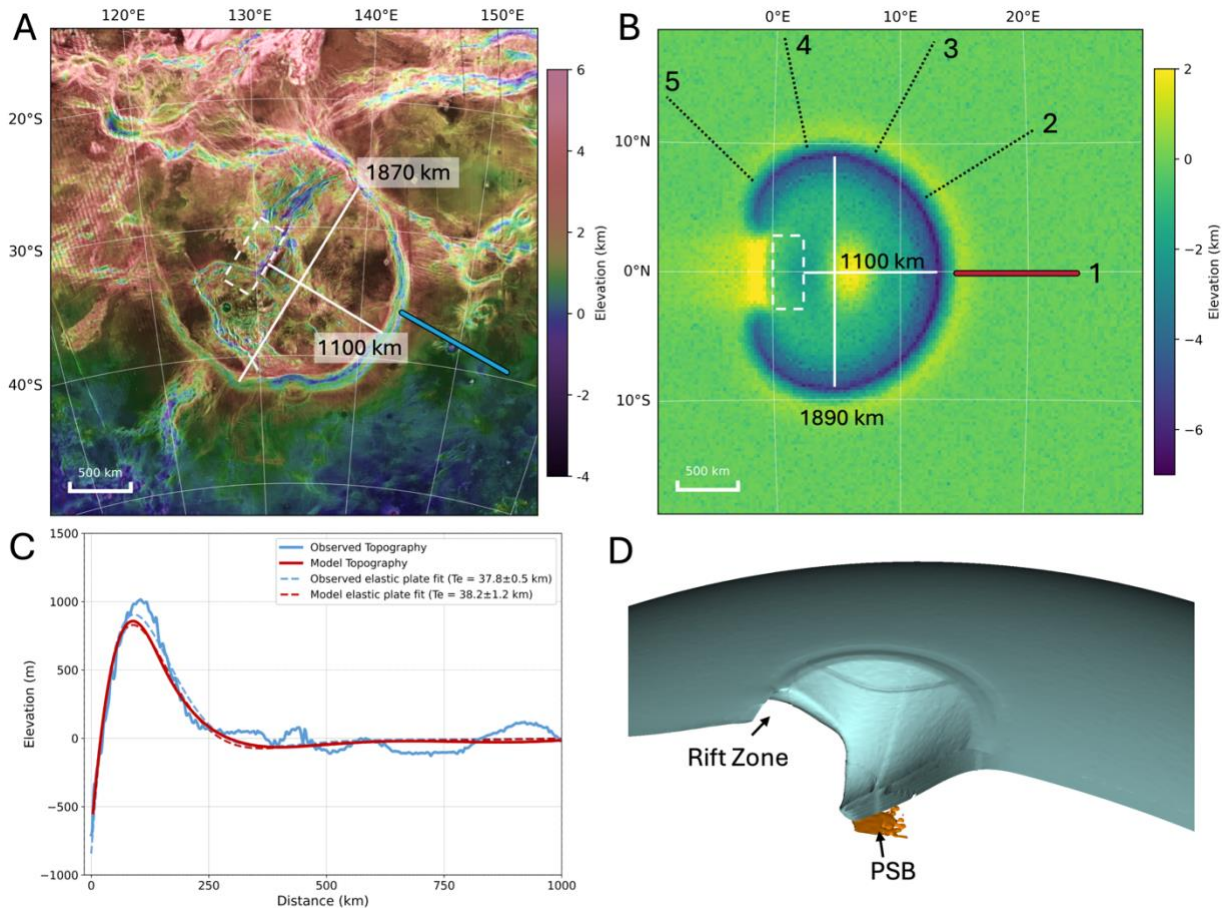


Figure 5: Comparison of Artemis Corona with the geometric best-fit peel-back delamination model with $T_L = 200$ km and $T_{\text{mantle}} = 1700$ K (Model #6; $t = 6.5$ Myr). (A) Observed topography showing the model-inferred initial rift location (dashed box) and the profile used for flexural analysis (blue). (B) Model topography at the best-fit time showing the initial rift geometry (dashed box), reference profile #1 (red), and spatially distributed flexural profiles #2-5 along trench. (C) Observed and modeled topographic profile #1 with best-fit broken-plate flexural models used to estimate elastic thickness ($T_e = 37.8$ km observed; 38.2 km modeled). (D) Three-dimensional view of the modeled lithospheric mantle and the postspinel phase boundary (PSB) at the geometric best-fit time; colors as in Fig. 3.

3.4 Atahensik Corona

The quasi-circular, dual trench structure observed at Atahensik Corona is best matched by a short initial rift ($L_{\text{rift}} = 300$ km) and symmetric lithosphere thickness surrounding the rift zone (Model #12; $T_L = 200$ km). In contrast to the asymmetric lithosphere that produced single-sided delamination resembling Artemis Corona (Fig. 5), the symmetric case produces delamination on both

sides of the rift (Fig. 6). A best-fit model time of 3.9 Myr yields model dimensions of 1025 x 845 km, compared with observed dimensions of 785 x 825 km. The rift-perpendicular width of the model feature closely matches observations, indicating good agreement in the dimension corresponding to the primary direction of trench retreat. The modeled feature is longer along the rift than Atahensik Corona, suggesting that a shorter initial rift may be required to reproduce the observed along-rift trench geometry. The modeled trench system also develops a slight asymmetry, with a greater rift-to-trench distance on one side of the initial rift. Correlating the less-developed side of the model with the northern trench at Atahensik places the model-equivalent initial rift northwest of the feature center, aligned with the SW-NE-trending rifted terrain extending northeast of Atahensik Corona. However, the inferred rift location is not unique and should be regarded as one possible configuration consistent with the observed morphology.

A reference topography profile taken across the northern trench at Atahensik Corona exhibits one of the clearest trench-outer-rise structures (Fig. 6C). Like for Artemis Corona, the long wavelength regional topographic component associated with uplifted rifted topography is removed from the observed profile prior to fitting. The $T_L = 200$ km model reproduces the overall amplitude and wavelength of the observed flexural signal, yielding comparable T_e estimates of 43.7 km and 45.0 km for observed and modeled topography, respectively. At the geometric best-fit time, PBD is still at an early stage, with the slab tip reaching only ~180 km depth. The model therefore remains in the initial low-amplitude, pre-steady-state stage of bending during which the inferred T_e is higher than the quasi-steady-state values reached during continued rollback (Fig. 4M-P). As PBD progresses, the modeled flexural wavelength shortens and T_e decreases despite no change in the lithospheric structure. At a representative later timestep ($t = 8.9$ Myr), the inferred T_e is 38.6 $\pm$ 1.7 km, comparable to the T_e of the reference Artemis model with the same thermal lithosphere thickness $T_L = 200$ km (Fig. 5).

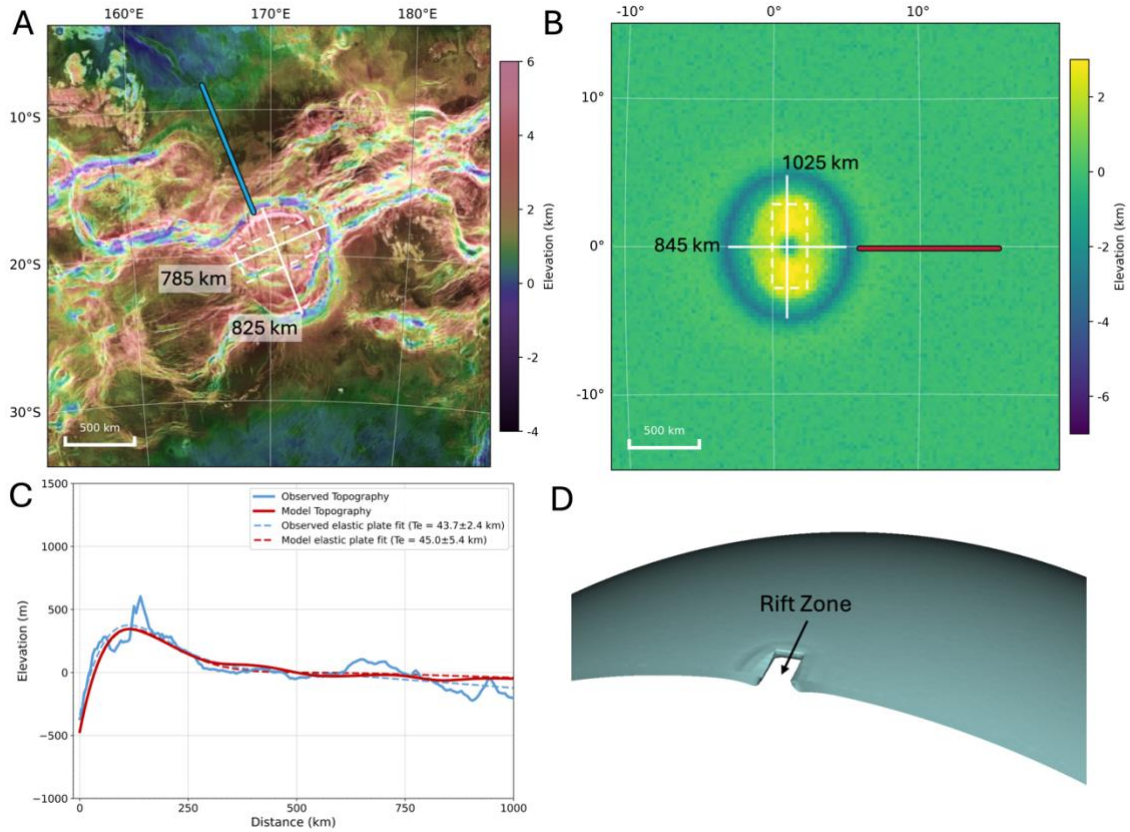


Figure 6: Comparison of Atahensik Corona with the geometric best-fit $T_L = 200$ km PBD model with $T_{\text{mantle}} = 1700$ K (Model #12; $t = 3.9$ Myr). (A) Observed topography showing the model-inferred initial rift location (dashed box) and a flexural profile across the northern trench (blue). (B) Model topography at the best-fit time showing the corresponding initial model rift geometry (dashed box) and a representative model topographic profile (red). (C) Observed and modeled profiles with best-fit broken-plate flexural analyses used to estimate elastic thickness ($T_e = 43.7$ km observed; 45.0 km modeled). (D) Three-dimensional view of the modeled lithospheric mantle at the best-fit time; colors as in Fig. 3.

3.5 Derceto Plateau

The long-rift, asymmetric lithosphere model (Model #15) reproduces the elongated, arcuate trench geometry of Derceto Plateau (Fig. 7). The rift-parallel extent of the modeled feature is determined by the initial rift length and changes little over model evolution. This contrasts with the shorter rift condition which produces quasi-circular trenches for the Artemis and Atahensik Corona models, indicating that the rift-parallel feature length is controlled by the initial rift length. Therefore, the geometric best-fit

model time is selected by matching the modeled rift-perpendicular width of the feature to observations. The observed trench-to-rift width of ~675 km yields a best-fit model time of 3.8 Myr. The model-inferred initial rift (1800 x 250 km) is aligned with Hanghepiwi Chasma, the rift along the southwestern margin of Derceto Plateau (Fig. 7A).

In contrast to Artemis and Atahensik, Derceto Plateau lacks a clear flexural rise. Topographic profiles extracted across the NE edge of the feature (Fig. 7C), corresponding to the trench-outer rise structure identified by Schubert and Sandwell (1995), do not exhibit flexural bulge that is distinct from the background short-wavelength topography. Consequently, a flexural signal which could reliably estimate T_e from the observed topography was not identified. Three sample observed topography profiles are shown in Fig. 7C. In contrast, the model topography shows a clear flexural rise with an estimated $T_e = 34.3$ km at the geometric best-fit model time. Like the Artemis model, the asymmetric lithosphere structure produces PBD only on the thick lithosphere side of the rift (Fig. 7D). At the best-fit time, the slab tip has descended to 375 km depth and represents a pre-steady-state period where T_e is elevated during early stages of PBD. T_e decreases as bending continues and reaches a steady-state value (e.g. $T_e = 27.1 \pm 0.8$ km at $t = 5.1$ Myr), comparable to other models with $T_L = 150$ km (Table 1).

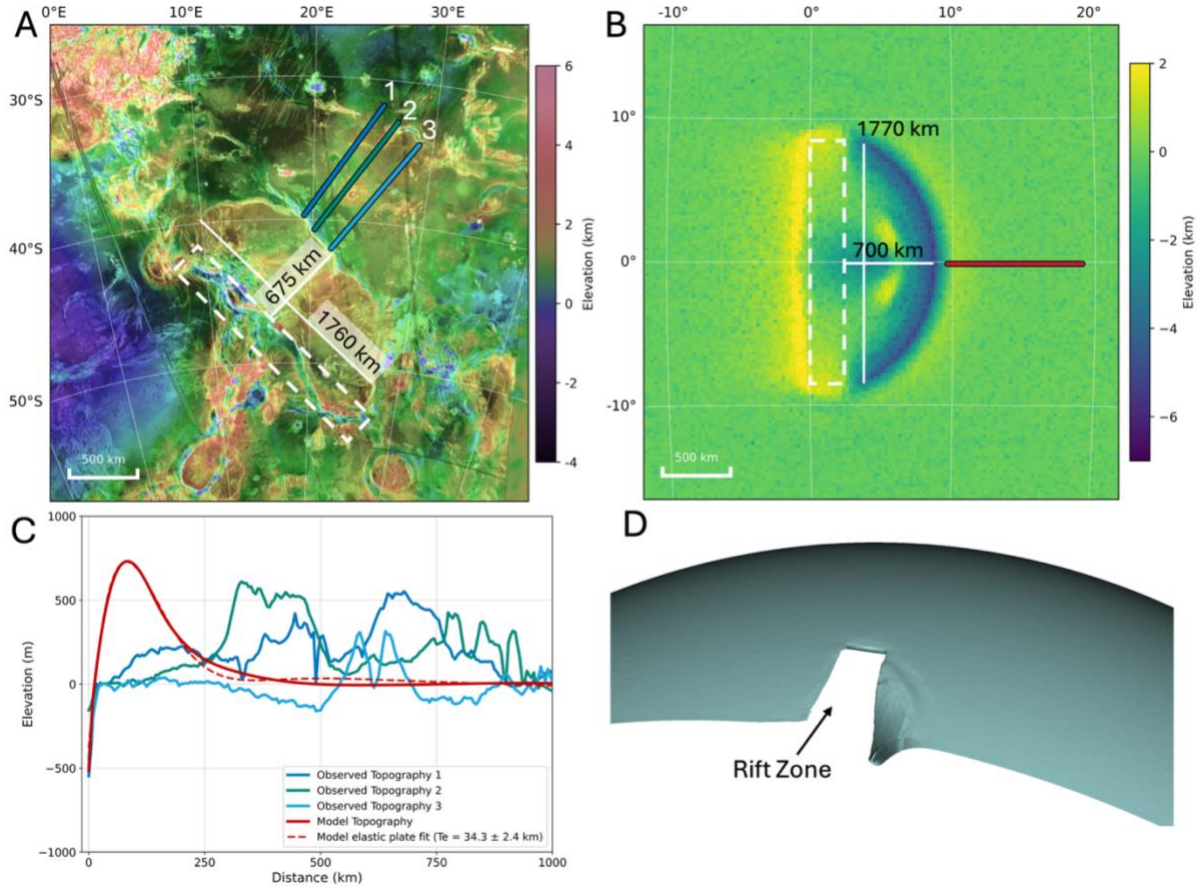


Figure 7: Comparison of Derceto Plateau with the geometric best-fit $T_L = 150$ km PBD model with $T_{\text{mantle}} = 1700$ K (Model #15; $t = 3.8$ Myr). (A) Observed topography showing the model-inferred initial rift location (dashed box) and three reference topographic profiles (green/blue). (B) Model topography at the best-fit time showing the corresponding initial rift geometry (dashed box) and the model topographic profile (red). (C) Observed and modeled topographic profiles. T_e is only estimated from the modeled topography profile, whereas the observed profiles do not exhibit a distinct flexural rise. (D) Three-dimensional view of the modeled lithospheric mantle at the best-fit time; colors as in Fig. 3.

3.6 Effects of Yield Strength and Mantle Temperature

Trench geometry is strongly controlled by rift length and lithospheric symmetry, which were varied to reproduce the observed case study sites. However, lithospheric strength exerts a secondary influence on trench morphology. Reducing the maximum mantle yield strength from 200 MPa to 100 MPa in the Artemis-like model (short rift, asymmetric lithosphere) decreases the width-to-length ratio of the trench and enhances weakening near the slab

hinge, promoting slab breakoff (Fig. 8). At similar trench dimensions, progressively weaker lithosphere exhibits greater extension and thinning of the delaminating slab, allowing the slab to reach greater depths while promoting necking near the hinge. Further reducing the yield strength to 60 MPa produces a shallower and narrower trench system as thinning near the slab hinge reduces slab pull during incipient breakoff. This low-strength model undergoes breakoff before developing the size and geometry observed at Artemis Corona, suggesting that a relatively strong, mechanically coherent slab is required to sustain the broad, arcuate trench geometry characteristic of Artemis. Increasing mantle temperature from 1700 K to 1900 K accelerates PBD, with the model reaching an Artemis-like geometry earlier (1.0 Myr compared with 7.4 Myr; Fig. 8 A,D). For the same half-space cooling plate age, $T_{\text{age}} = 245 \text{ Myr}$, the $T_{\text{mantle}} = 1900 \text{ K}$ model produces a smaller elastic thickness ($T_e = 24.6 \text{ km}$) than the $T_{\text{mantle}} = 1700 \text{ K}$ model ($T_e = 27.5 \text{ km}$).

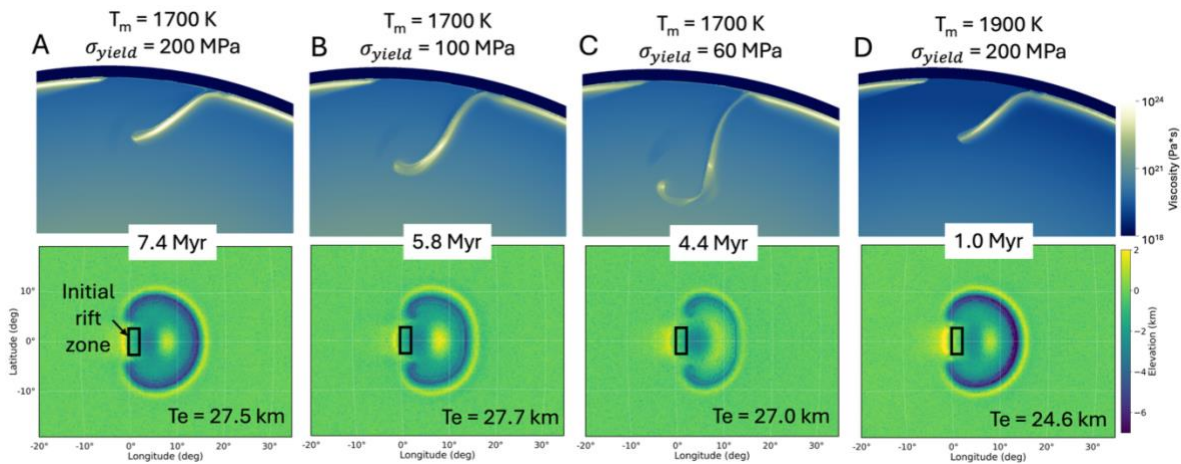


Fig. 8: Effects of mantle yield strength and mantle temperature on PBD relative to (A) the Artemis-like Model #2 with $T_L = 150 \text{ km}$. (A-D) Cross sections show viscosity field (top row) and corresponding planform view (bottom row). Model times are selected which best reproduce the observed size and geometry of Artemis Corona; (C) the 60 MPa model is shown during slab breakoff because it does not reproduce the observed morphology of Artemis Corona or any other case study site.

4. DISCUSSION

4.1 Lithospheric Controls on Peel-Back Delamination Initiation

Whereas the timing and surface expression of PBD are governed by rift geometry, yield strength, and mantle temperature, PBD initiation is primarily

controlled by lithospheric thickness. Lithosphere thickness influences PBD through two complementary effects. First, thicker lithosphere contains a larger volume of cold, dense lithospheric mantle, increasing the gravitational driving force for delamination. Second, lithospheric thickness controls the geothermal gradient, assuming a half-space cooling thermal profile, and therefore the pressure-temperature conditions at which basaltic crust transforms to denser granulite and eclogite assemblages. As the lithosphere cools and thickens, crustal densification occurs at progressively shallower depths, reducing the depth of neutral buoyancy from 81.1 km for a 50 km thermal lithosphere to 48.5 km for a 250 km thermal lithosphere (Fig. 9). Consequently, thicker, cooler lithosphere not only contains a larger volume of dense lithospheric mantle but also causes the crust to contribute to the total negative buoyancy of the lithosphere at shallower depths.

The models exhibit a threshold lithospheric thickness for PBD initiation, consistent with previous 2D studies (Adams et al. 2022, 2023). Models with $T_L \geq 150$ km undergo sustained PBD, whereas models with $T_L = 100$ km remain stable as a stagnant lid (Table 1). This threshold is lower than the minimum $T_L = 200$ km reported by Adams et al. (2023), which prescribed an instantaneous basalt-to-eclogite transition at 70 km depth. Here, the pressure-temperature-dependent basalt stability field produces progressive crustal densification and allows negatively buoyant crust to develop at shallower depths (Fig. 9). The resulting increase in total lithospheric negative buoyancy facilitates PBD initiation in thinner lithosphere than predicted by previous models.

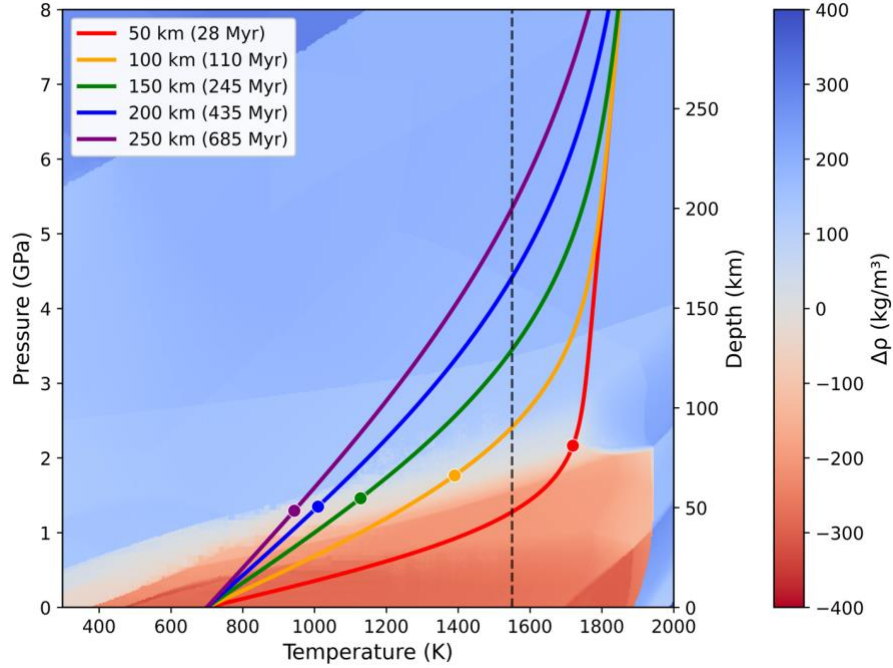


Fig. 9: Density contrast ($\Delta\rho = \rho_{crust} - \rho_{mantle}$) between thermally equilibrated Venusian basaltic crust and a 15% depleted, dry, fully reduced harzburgitic mantle as a function of temperature and pressure, calculated using Perple_X. Pressure as a function of depth is assumed to be the lithostatic pressure with no additional tectonic stresses. Colored curves show half-space cooling geotherms corresponding to various thermal lithosphere thicknesses. Circles mark the depth where the crust becomes neutrally buoyant ($\Delta\rho = 0$) along each geotherm, corresponding to depths of 81.1, 66.1, 54.7, 50.5, and 48.5 km for thermal lithosphere thickness of 50, 100, 150, 200, and 250 km, respectively, under the assumed thermal conditions. The vertical dashed line indicates $T=1550$ K isotherm used to define T_L . Geotherms include an adiabatic temperature gradient of 0.5 K km^{-1} to provide a consistent comparison with the P-T-dependent Perple_X calculations (see section 2.3 for additional details). The geodynamic models, however, are incompressible and therefore do not include adiabatic heating, so the ages corresponding to a given T_L differ slightly from those of the plotted geotherms.

4.2 Crustal Strength and Mechanical Decoupling During PBD

In addition to sufficient negative buoyancy, PBD initiation requires mechanical decoupling of the lithospheric mantle from the overlying buoyant crust. Generally, slab-like peeling delamination is facilitated by the presence of a low-viscosity layer in which deformation localizes (Meissner & Mooney 1998). In continental lithosphere on Earth, the decoupling zone is

commonly considered to form within a weak lower crust (LePourhiet et al. 2006; Göğüş et al. 2011; Krystopowicz & Currie 2013; Göğüş & Ueda 2018). In our models, decoupling occurs within the uniformly weak, 30 km-thick crust ($C_0 = 10$ MPa, $\mu = 0$), which provides a simplified representation of the weak An75 plagioclase rheology of Ranalli (1995) commonly used in geodynamic models of Venus. High levels of intrusive magmatism may also thermally weaken the crust as predicted by plutonic-squishy lid models for Venus (Lourenço et al. 2020). This simplification may overestimate the thickness of the weak decoupling layer by neglecting the stronger upper crust; yet since T_e is primarily controlled by the thicker lithospheric mantle, its effect on modeled T_e is likely limited.

However, to understand the viability of a weak lower crustal layer on Venus, stronger, dry crustal rheologies may be more appropriate given the inferred dry interior of Venus (Grinspoon 1993; Constantinou et al. 2025). Yield strength envelopes calculated using dry plagioclase (Katayama 2021) and dry diabase rheologies (Mackwell 1998) show that the lower crust is stronger for lower geothermal gradients associated with thicker T_L (Fig. 10). This introduces a trade-off in the conditions favorable for PBD: increasing T_L provides greater lithospheric negative buoyancy to drive delamination but the associated strengthening of the lower crust makes mechanical decoupling more difficult. At $T_L = 100$ km, the lower crust is relatively weak, but the lithosphere lacks sufficient negative buoyancy to initiate PBD in the numerical models. At larger T_L , the lithosphere is sufficiently negatively buoyant to delaminate, but the colder and stronger lower crust may inhibit mechanical decoupling. The strength and thickness of the weak lower-crustal layer required for decoupling are not constrained by these models and likely depend on the magnitude of the driving force for delamination, as well as mantle temperature, strain rate, and crustal rheology; we therefore do not define a specific T_L upper limit for PBD.

Crustal thickness provides an additional control on mechanical decoupling because thicker crust extends to greater depths and temperatures, increasing the thickness of the weak ductile lower crustal layer. For example, in a 150 km-thick lithosphere, dry plagioclase and dry diabase rheologies predict that a 15-20 km crust maintains a relatively strong lower crust, whereas a ≥ 30 km-thick crust has a weaker lower crust layer due to warmer temperatures at depth (Fig. 10). Local variations in thermal structure may further modify

these conditions, particularly near rift zones where elevated temperatures could weaken the lower crust relative to the idealized half-space cooling thermal profiles considered here. The most favorable conditions for PBD may therefore combine moderately thick lithosphere ($T_L \geq 150$ km) which provides sufficient negative buoyancy, with thick and warm, rift-adjacent crust that promotes development of a weak lower crustal decoupling layer.

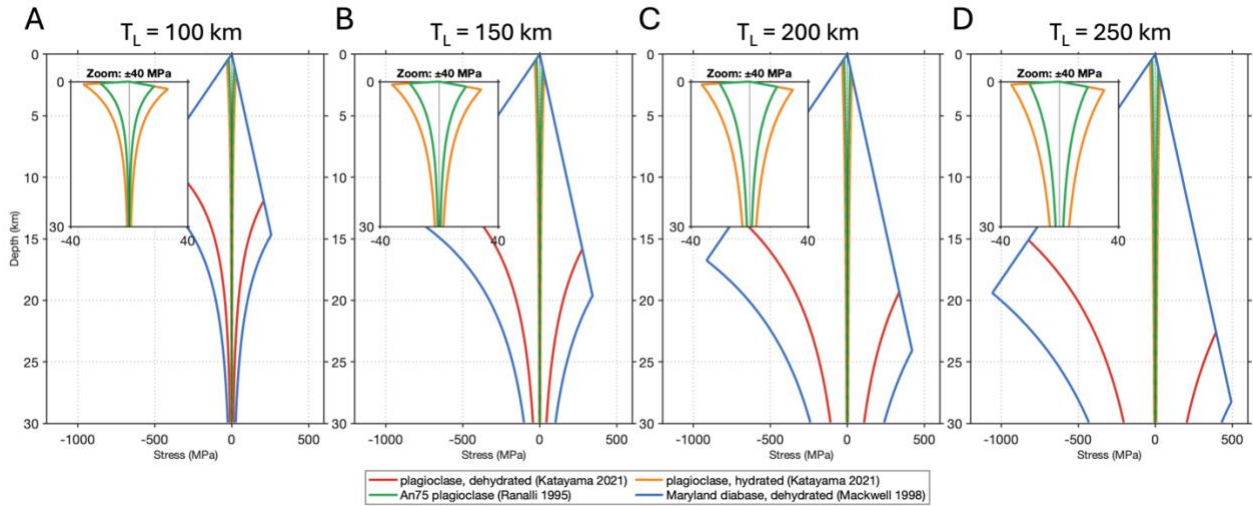


Fig. 10: Crustal yield strength envelopes for thermal lithosphere thicknesses of (A) 100 km, (B) 150 km, (C) 200 km, and (D) 250 km calculated using half-space cooling thermal profiles. Each panel shows strength profiles for a 30 km-thick layer of crust using experimentally derived dislocation creep flow laws proposed for Venus. Calculations assume zero cohesion, $T_{\text{mantle}} = 1700$ K, friction coefficient $\mu = 0.6$, and a strain rate $\dot{\epsilon} = 10^{-17} \text{ s}^{-1}$. Positive stress represents tensile strength, and negative stress represents compressional strength. Insets show zoomed-in hydrated plagioclase and An75 plagioclase flow laws.

4.3 Implications for Candidate Recycling Sites on Venus

Artemis Corona provides strong support for PBD, with agreement between the model and observations in both trench geometry and flexural topography (Fig. 5). The short-rift, asymmetric model reproduces its overall size and quasi-circular trench geometry, while the modeled trench-outer rise structure closely resembles the observed flexural topography in amplitude and wavelength. One difference is that the lateral ends of the modeled trench remain coupled to the ends of the prescribed rift, whereas the observed Artemis trench continues around the corona beyond the inferred rift ends.

This mismatch may reflect the simplified rift geometry or rheological parameters controlling deformation near the slab margins, as well as surface complexities such as nearby rift branches and tesserae. At the geometric best-fit time, the model is in the steady-state phase of PBD characterized by sustained slab rollback and relatively stable T_e . Previous studies have inferred relatively high elastic thickness (Schubert & Sandwell 1995; Anderson & Smrekar 2006; Jiménez-Díaz et al. 2015), consistent with relatively strong lithosphere. Crustal thickness estimates also suggest above-average crust thickness surrounding Artemis Corona (Anderson & Smrekar 2006; James et al. 2013), providing favorable conditions for both sufficient negative buoyancy in the lithosphere and the presence of a weak lower crust necessary for decoupling.

Atahensik Corona provides a second plausible example of PBD. Its characteristic dual-trench morphology and trench-outer-rise structure are reproduced by the model with symmetric lithosphere and a short rift, although the modeled feature is somewhat longer parallel to the initial rift than the observed corona. In the models, the along-rift extent of the trench system is strongly influenced by the prescribed rift length. Atahensik, however, is embedded within the complex, branching Dali-Diana chasmata rift system, and a more realistic initial rift geometry could modify the along-axis extent of PBD relative to the simplified linear rift used here. The observed northern trench at Atahensik preserves a well-defined trench-outer-rise structure that is closely reproduced by the $T_L = 200$ km model. The observed and modeled profiles yield similar elastic thicknesses of 43.7 and 45.0 km, respectively, indicating agreement in both trench geometry and flexural behavior. Atahensik corresponds to a pre-steady-state stage of PBD in the model when T_e is elevated relative to steady state values for a given T_L .

Derceto Plateau presents a more ambiguous case. The asymmetric, long-rift model successfully reproduces its elongated trench morphology, indicating the observed geometry may be consistent with PBD. However, unlike Artemis or Atahensik, the observed topography lacks a distinct flexural outer rise, and previous studies infer much smaller elastic thickness values ($T_e \sim 6$ km; Schubert and Sandwell, 1995) than produced by the model with the minimum lithosphere thickness capable of undergoing PBD (Model #15; $T_L = 150$ km; $T_e = 34.3$ km). Geologic mapping identified lava flows that partially embay and infill portions of the trench (MacLellan et al. 2021), suggesting that the

original flexural topography may not be fully preserved. Trench infilling introduces uncertainty into the interpretation of the present-day flexural signal, however it is unclear if it can account for the large discrepancy between the observed and modeled T_e . Consequently, the flexural observations provide weaker support for PBD than at Artemis or Atahensik Corona.

4.4 Flexural Constraints on Lithospheric Structure at Artemis Corona

Among the candidate PBD sites considered here, Artemis Corona provides the most extensive and well-preserved trench-outer-rise system, allowing flexural topography to be used to further investigate its surrounding thermal lithospheric structure. Across 58 profiles, the median inferred T_e is 47.0 km, comparable to previous T_e estimates of ~35-45 km for Artemis Corona (Sandwell & Schubert, 1992a,b; Schubert & Sandwell, 1995; Smrekar et al. 2023; 2025). However, strong spatial variability in the flexural response is inferred along the trench (Fig. 11a). Profiles along the southern trench segment (profiles 1-19) yield a median T_e of 32.4 km, whereas profiles along the south-eastern segment (profiles 20-58) yield a higher median value of 49.1 km. The T_e estimates appear to define two spatial domains separated near an abrupt change in trench orientation.

Our preferred interpretation of this spatial variation is that it does not reflect a discontinuity in thermal lithospheric thickness. The coincidence between the change in T_e and trench orientation suggests that the two flexural domains may instead reflect differences in slab morphology and deformation state. Slab segmentation, tearing, or partial breakoff could modify slab geometry and enhance plate bending within one segment, promoting localized inelastic yielding and biasing the inferred T_e downward (Garcia et al. 2019). In contrast, the Artemis-like model produces relatively consistent T_e around the trench for a uniform T_L with only a modest decrease toward the rift ends (Fig 5). The larger and more abrupt spatial variation observed at Artemis may therefore reflect localized deformation associated with three-dimensional slab complexity and variations in bending that are not captured by the numerical models. The higher T_e values along the southeastern trench may more closely represent the lithosphere surrounding Artemis, whereas the southern T_e values may be reduced due to enhanced bending and inelastic yielding.

Despite this spatial variability, the range of T_e observed around Artemis can be used to constrain the thermal lithospheric thicknesses compatible with the

flexural observations. For a given mantle temperature, the Artemis-like models show a systematic increase in T_e with increasing T_L (Table 1; Fig. A3), while T_e remains relatively stable during steady-state PBD (Fig. 4). For each T_L , MCMC posterior probability density functions (PDFs) from five flexural profiles distributed along the model trench (profiles 1-5; Fig. 5B) are combined and compared with the observed T_e distributions to place first-order constraints on lithospheric structure (Fig. 11B). Differences in the width of the combined PDFs partly reflect how well T_e is constrained by the modeled flexural morphology. Thinner- T_L models produce higher amplitude, more sharply defined forebulges, whereas thicker- T_L models produce broader, lower-amplitude forebulges for which scatter in the tracer-defined surface introduces greater uncertainty in the flexural fit. Models with $T_L = 150, 200, \text{ and } 250 \text{ km}$ produce combined posterior PDFs with peaks at 26.5, 37.2, and 55.8 km, respectively.

The relationship between T_e and T_L depends on the overall thermal structure of the lithosphere, including surface temperature, mantle temperature, and cooling age. In the 1700 K mantle models, T_e increases systematically with T_L as cooling age increases. At a fixed cooling age, increasing mantle temperature from 1700 to 1900 K reduces both T_L and T_e . A warmer mantle raises temperatures throughout the lithosphere, which shifts the isotherm that defines the base of the thermal lithosphere to shallower depths and thins the portion of the lithosphere that supports bending stresses. If mantle temperatures are higher than the 1700 K reference value assumed here, as suggested in recent models of mantle dynamics on Venus (Kerr et al. 2025), this would produce lower T_e estimates for a given plate age (Fig. 11b). The range of T_e observed at Artemis Corona may be consistent with multiple thermal structures, including different combinations of mantle temperature and thermal age, as well as deviations from the idealized half-space cooling profiles considered here.

Nevertheless, the flexural observations favor relatively thick thermal lithosphere at Artemis Corona within the explored range of thermal conditions (Fig. 11b). The T_e distribution along the southeastern trench, interpreted as more representative of background lithospheric structure, overlaps with models having $T_L \sim 200\text{--}250 \text{ km}$ for 1700 K mantle models. Warmer 1900 K mantle models with a shallower thermal lithosphere predict a lower range of T_e than the 1700 K models, though higher T_e values could be produced with older cooling ages. We interpret $T_L \sim 200\text{--}250\text{+ km}$ for $T_{\text{mantle}} = 1700\text{--}1900 \text{ K}$ as a

plausible range for the lithosphere surrounding Artemis, with uncertainties in subsurface slab morphology and thermal conditions preventing a more precise estimate. This range is independently consistent with the $T_L \geq 150$ km minimum thickness required for PBD initiation in the numerical models. At the upper end of this range ($T_L \geq 250$ km), the negative buoyancy of the lithospheric mantle may be sufficient to drive delamination without requiring a mantle plume (Adams et al. 2022). Thicker lithosphere with greater negative buoyancy also allows smaller plumes to initiate PBD than would be necessary to destabilize thinner lithosphere (Adams et al. 2023), indicating Artemis Corona may not have required a large plume-rift interaction.

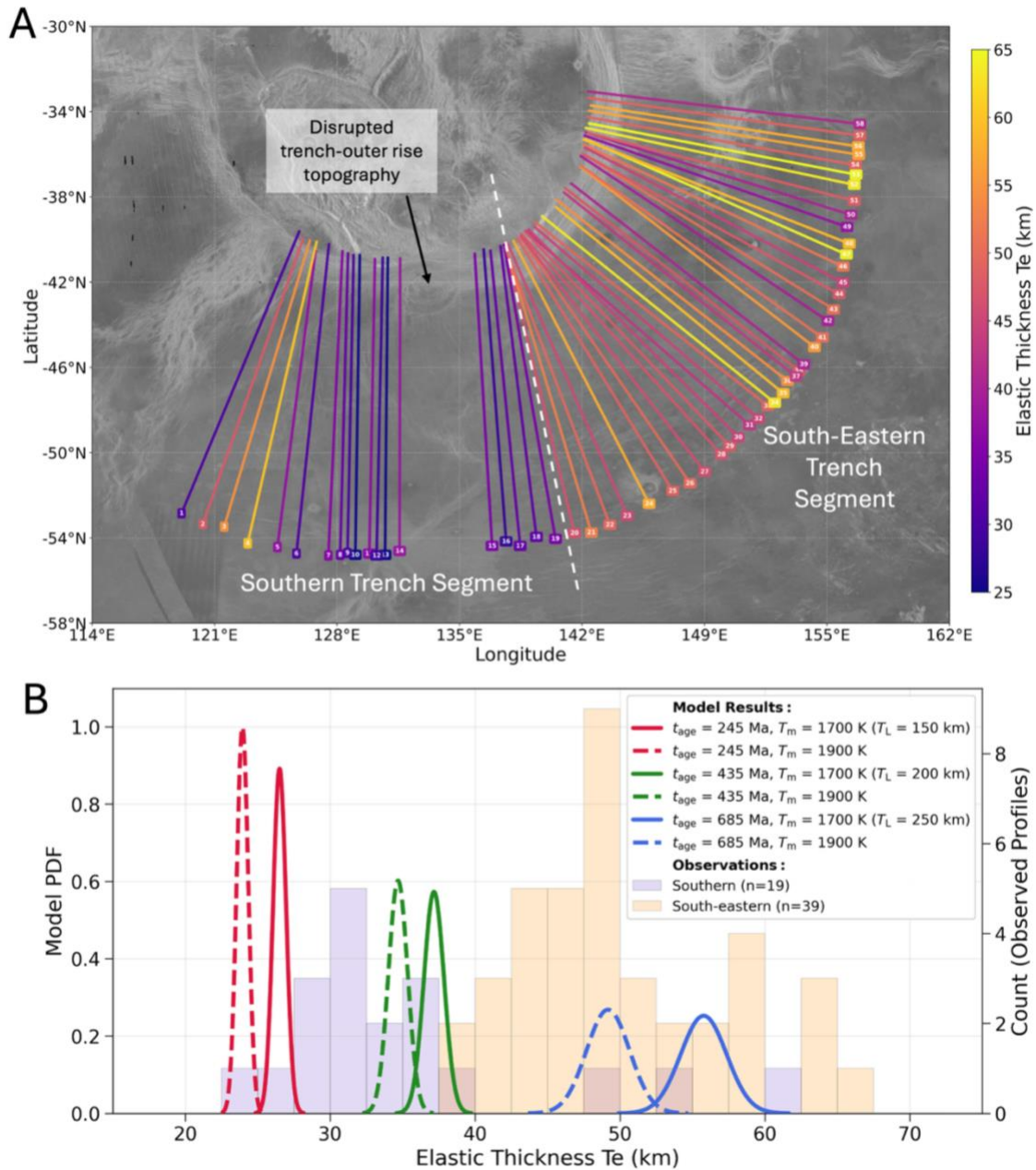


Fig. 11: Spatial variability in observed elastic thickness at Artemis Corona and comparison with modeled elastic thickness for different thermal lithosphere thicknesses and mantle temperatures. (A) T_e estimated from flexural profiles distributed along the Artemis Corona trench. Profiles are numbered sequentially along the trench with T_e values given in Table A1. Large gaps indicate locations where a distinct trench-outer rise profile was not identified. A white dashed line separates the southern trench (profiles 1-19) and south-eastern trench (profiles 20-58) segments. (B) MCMC posterior probability density functions of T_e from Artemis-like PBD models plotted over distributions of observed T_e for the south and south-eastern trench segments

of Artemis Corona. Solid curves show models with $T_{\text{mantle}} = 1700$ K and $T_L = 150$, 200, and 250 km, based on five spatially distributed profiles from each model (locations in Fig. 5B; corresponding topographic profiles in Fig. A4). Dashed curves show models with $T_{\text{mantle}} = 1900$ K. Each 1700 and 1900 K model pair has the same half-space cooling age, t_{age} , but different T_L .

5. CONCLUSIONS

The first three-dimensional models of peel-back delamination (PBD) on Venus are presented which demonstrate that the combined use of trench geometry and flexural topography provides a more rigorous test for evaluating candidate lithospheric recycling sites than either constraint alone. Across the parameter space, lithospheric thickness is the primary control on PBD initiation by governing both the negative buoyancy of the lithosphere and the depth of crustal densification, with models requiring a minimum thermal lithosphere thickness. PBD requires a weak lower crust to mechanically decouple the negatively buoyant lithospheric mantle from the overlying buoyant crust. These conditions are most likely to occur where relatively thick lithosphere coincides with thick and/or warm crust conditions which promote a weak lower crust. Rift geometry and surrounding lithospheric structure control the resulting surface expression, producing both quasi-circular and elongated trench systems.

These results demonstrate that the combined use of trench geometry and flexural topography provides a useful framework for distinguishing candidate PBD sites from other proposed recycling locations. PBD is unlikely to represent a globally ubiquitous mode of lithospheric recycling or corona formation on Venus. Instead, it likely only operates where crustal and lithospheric conditions provide sufficient negative buoyancy and a mechanism to decouple the lithospheric mantle from overlying buoyant crust. This supports the possibility that features such as Artemis Corona and Atahensik Corona represent a tectonically distinct subset of coronae formed in thicker, stronger lithosphere by different processes than the majority of coronae which are thought to develop in warm, thin lithosphere (Smrekar & Stofan 1997; Hoogenboom & Houseman 2006; Gülcher et al. 2020; Smrekar et al. 2023; 2025). PBD may represent one of several tectonic and volcanic processes contributing to the diversity and evolution of coronae and corona-like features on Venus (Cascioli et al., 2025), which may be relevant for understanding lithospheric recycling on early Earth and exoplanets.

ACKNOWLEDGMENTS

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DATA AVAILABILITY

StagYY (Tackley, 2008) is the property of P.J.T. and is available upon request for collaborative studies. Requests can be made to P.J.T (paul.tackley@eaps.ethz.ch). A sample model parameter file and model videos can be found at <https://doi.org/10.5281/zenodo.23780691>.

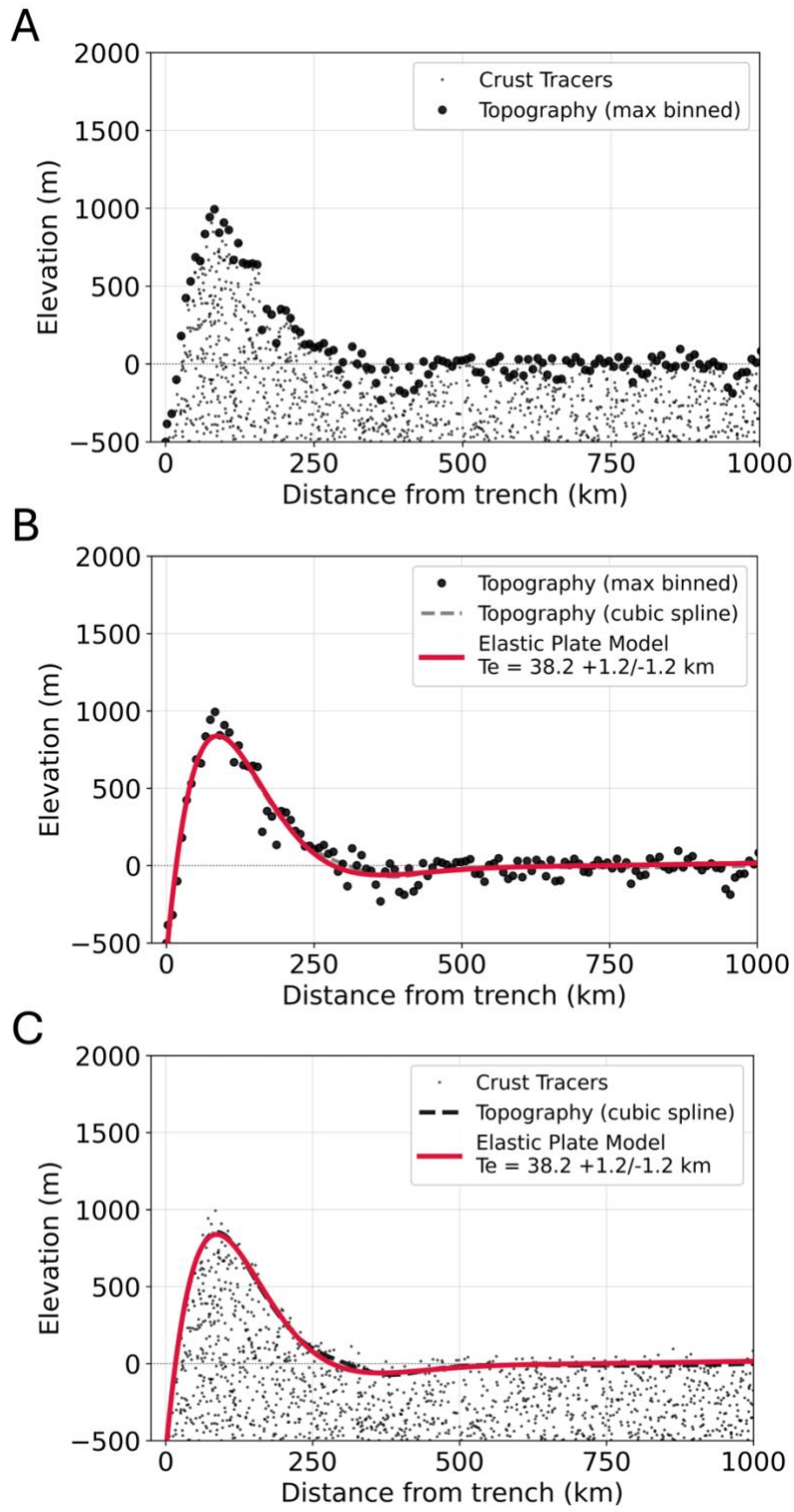


Figure A1: Estimation of model topography and elastic thickness from crustal tracer data. (A) Crustal tracer elevations along a trench-perpendicular profile from a representative model (Model #6 at the geometric best-fit model time, $t = 6.5$ Myr). Crust tracers along the profile are grouped into 8 km distance bins and the maximum elevation in each bin is retained (black circles). (B) The binned topography points are fit with a broken elastic plate model (red) used to calculate elastic thickness, T_e , in addition to a cubic spline fit representing smoothed model topography (dashed line). (C) The spline and elastic model overlain on the full crustal tracer distribution show that the binned surface approximation captures the overall trench-outer-rise flexural profile including the location of the forebulge peak, despite uncertainty in the tracer-defined surface.

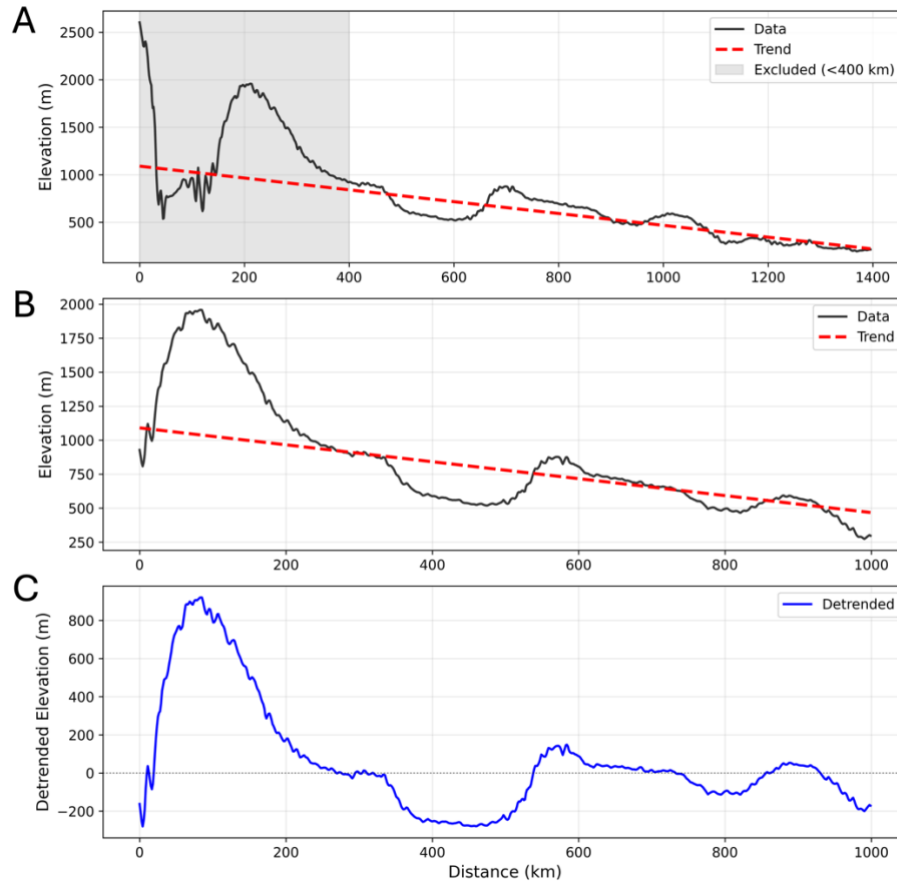


Figure A2: Removal of long-wavelength regional topography and trimming of observed trench-outer-rise profiles. (A) A sample Magellan topographic profile across the Artemis Corona trench-outer rise system. A linear trend is calculated from the far-field portion of the profile beyond 400 km from the trench, excluding the trench and forebulge region, and applied to the entire topographic profile. (B) The profile is trimmed to retain only the smooth trench-forebulge signal and exclude non-flexural topography inside the trench. (C) Detrended profile used for elastic plate fitting obtained by subtracting the linear regional trend from the trimmed profile. Elastic thickness is calculated using the trimmed, detrended profile.

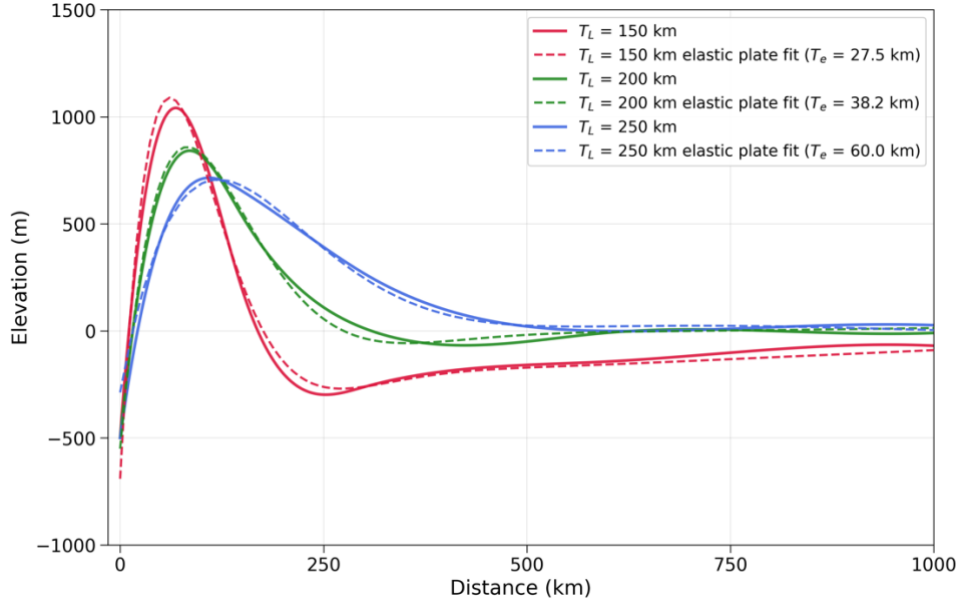


Figure A3: Dependence of modeled flexure on thermal lithosphere thickness in reference Artemis-like models with $T_{\text{mantle}} = 1700$ K. Topographic profiles (solid lines) and corresponding elastic plate fits (dashed lines) for Profile 1 from Models 2, 6, and 8, with $T_L = 150$, 200, and 250 km, respectively. Inferred T_e increases with T_L . Thinner lithosphere produces a higher amplitude, shorter wavelength trench-forebulge signal, whereas thicker lithosphere produces lower amplitude flexure with a broader forebulge.

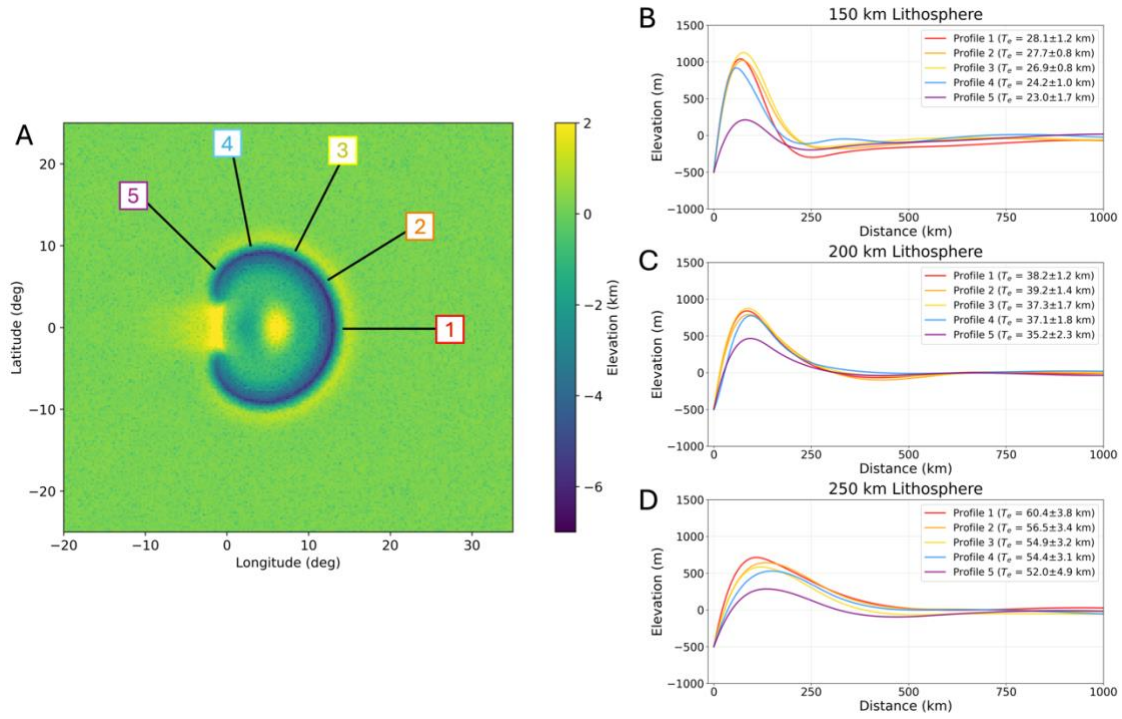


Figure A4: Along-trench variability in modeled elastic thickness. (A) Example Artemis model topography showing the locations of five trench-perpendicular profiles used to calculate the T_e posterior distributions in Figure 11b. (B-D) Elastic plate fits to the five profiles for models with $T_{\text{mantle}} = 1700$ K and $TL = 150, 200, \text{ and } 250$ km, respectively. T_e varies modestly along the trench, and it is typically largest across from the initial rift zone and smallest near the rift ends. Across all models, the amplitude of the flexural rise is lowest near the rift ends.

Profile #	Segment	Start Lat	End Lat	Start Lon	End Lon	Te (km)	Error (km)
1	S	-40.1103	-53.0963	126.4378	119.8612	30.265	0.804233
2	S	-40.4342	-53.5969	126.657	121.0512	47.73545	1.033139
3	S	-40.5225	-53.7147	127.0328	122.2413	54.69716	0.795389
4	S	-40.5814	-54.4803	127.4086	123.5566	60.12101	0.869576
5	S	-40.6109	-54.657	127.8158	125.2164	37.34408	1.384951
6	S	-40.6992	-54.9515	128.0976	126.2812	30.26358	0.844581
7	S	-41.0231	-55.0398	128.8492	128.0663	35.93634	0.707744
8	S	-41.1115	-55.0104	129.1624	128.724	32.40842	0.672907
9	S	-41.1998	-54.922	129.4756	129.1311	29.70637	0.740966
10	S	-41.1998	-55.0104	129.8201	129.5695	23.75761	0.792126
11	S	-41.3765	-54.9515	130.6656	130.3525	36.82979	0.92603
12	S	-41.3176	-55.0398	131.1041	130.7596	28.52172	1.353945
13	S	-41.3176	-55.0104	131.3859	131.2293	25.04097	2.290417
14	S	-41.347	-54.8337	132.0749	132.0436	37.82293	2.217832
15	S	-41.1409	-54.5981	136.2401	137.2109	34.69018	0.908179
16	S	-40.9348	-54.392	136.7412	137.9939	27.82053	0.667403
17	S	-40.9937	-54.5981	137.117	138.7768	32.19508	0.673572
18	S	-40.7875	-54.1859	137.6807	139.685	31.78307	0.346892
19	S	-40.7286	-54.2742	137.8686	140.7498	34.61629	0.245117
20	SE	-40.6992	-54.0092	137.8999	141.8145	46.70075	0.54989
21	SE	-40.817	-53.9797	137.9625	142.7541	52.64232	0.819004
22	SE	-40.9348	-53.6264	137.9939	143.7875	49.62797	0.595971
23	SE	-40.6403	-53.2141	138.0565	144.7584	45.47671	0.432455
24	SE	-40.552	-52.6546	138.401	145.9797	57.70734	0.73385
25	SE	-40.4342	-52.0657	138.6202	147.2951	48.78304	0.429195
26	SE	-40.3458	-51.7123	138.7768	148.2659	49.71425	0.416215
27	SE	-40.3458	-51.1823	138.8394	149.0801	46.23617	0.704556
28	SE	-40.2281	-50.3872	138.8394	150.0197	46.9616	0.554664
29	SE	-40.1103	-50.0044	139.0273	150.4894	47.66116	0.528949
30	SE	-40.0219	-49.5922	139.2465	150.9592	44.7472	0.442951
31	SE	-40.0808	-49.0327	139.09	151.5855	42.88099	0.452409
32	SE	-39.9041	-48.7382	139.591	152.0866	44.89265	0.547493
33	SE	-39.6686	-48.1493	139.591	152.619	49.25283	0.670855
34	SE	-39.4035	-48.002	139.9982	152.9948	63.44209	0.573408
35	SE	-38.9913	-47.5603	140.7811	153.4645	59.39993	0.540822
36	SE	-38.6379	-47.0008	140.7811	153.7151	54.31987	0.671248
37	SE	-38.314	-46.7947	141.2195	154.1848	43.40813	0.60526
38	SE	-38.1079	-46.5297	141.2822	154.3414	47.38726	0.777384
39	SE	-37.9017	-46.2352	141.6266	154.6233	40.5713	0.72946
40	SE	-37.1361	-45.4401	142.159	155.2183	55.86359	0.870441

41	SE	-37.0772	-44.9984	142.159	155.6568	50.75895	0.628477
42	SE	-36.6355	-44.2328	142.159	155.9699	39.80416	0.471228
43	SE	-36.5177	-43.7028	142.253	156.2831	50.56928	0.622466
44	SE	-36.5472	-42.9961	142.2843	156.5963	47.74146	0.611691
45	SE	-36.2527	-42.466	142.4096	156.8155	44.66964	0.71946
46	SE	-36.1055	-41.7298	142.3156	156.8155	51.51172	1.650446
47	SE	-35.8699	-41.1704	142.4409	157.0034	64.47136	2.201672
48	SE	-35.8699	-40.6698	142.3156	157.16	59.19503	1.601865
49	SE	-35.6932	-39.9041	142.3783	157.0347	37.83251	0.529759
50	SE	-35.546	-39.3446	142.4409	157.2539	40.46626	0.576541
51	SE	-35.5754	-38.6968	142.5662	157.4418	48.01917	0.626038
52	SE	-35.3399	-37.9606	142.4409	157.4418	63.68258	0.96076
53	SE	-35.1337	-37.4895	142.6288	157.5358	66.98193	1.124602
54	SE	-34.8393	-37.0478	142.5348	157.5045	49.37323	0.782567
55	SE	-34.5448	-36.5766	142.6288	157.6924	56.98774	1.222449
56	SE	-34.3092	-36.1938	142.7227	157.6611	57.6749	1.70008
57	SE	-33.9559	-35.6932	142.5662	157.755	48.92409	1.193436
58	SE	-33.6614	-35.1632	142.5662	157.755	41.02423	1.598162

1041
1042 Elastic thickness measurements along Artemis Trench. Summary of T_e values
1043 derived from 58 topographic profiles crossing Artemis trench (plotted in main
1044 text Fig. 11a). Profiles are labeled as south (profiles 1-19) and south-
1045 eastern (profiles 20-58) trench segments. Start and end coordinates (latitude
1046 and longitude) are provided for each profile. T_e values were obtained via MCMC
1047 fitting of trimmed and detrended topography using the methods described in
1048 the main text. Errors represent 68% confidence intervals calculated from the
1049 MCMC posterior distribution. The mean and median of the profiles in the
1050 southern trench segment are 35.4 km and 32.4 km, respectively. The mean and
1051 median of the profiles in the south-eastern trench are 51.2 km and 49.1 km,
1052 respectively.

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