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1 Flexural strike-slip basins

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9

10 **ABSTRACT**

11 Strike-slip faults are classically associated with pull-apart basins where continental crust is thinned
12 between two laterally offset fault segments. Here we propose a subsidence mechanism to explain
13 the formation of a new type of basin where no substantial segment offset or syn-strike-slip thinning
14 is observed. Such “flexural strike-slip basins” form due to a sediment load creating
15 accommodation space by bending the lithosphere. We use a two-way coupling between the
16 geodynamic code ASPECT and surface processes code FastScape to show that flexural strike-slip
17 basins emerge if sediment is deposited on thin lithosphere close to a strike-slip fault. These
18 conditions were met at the Andaman Basin Central Fault, where seismic reflection data provide
19 evidence of a laterally extensive flexural basin with a depocenter located parallel to the strike-slip
20 fault trace.

21

22 **MOTIVATION**

23 Near plate boundaries, sedimentary basins usually form by filling accommodation space
24 created via two tectonic processes (Allen & Allen, 2013): 1) isostatic subsidence due to
25 lithospheric stretching or cooling, which particularly governs rift basin formation at divergent
26 boundaries and 2) lithospheric flexure, which is closely associated with foreland basins in
27 convergent settings. So far, basins at transform plate boundaries are thought to be related to the
28 first process, by generating pull-apart basins.

29 Pull-apart basins form between laterally offset strike-slip fault segments (Mann et al.,
30 1983). During strike-slip motion, the area between the offset faults is extended and rapid basement
31 subsidence is generated due to abrupt crustal thinning (van Wijk et al., 2017). Pull-apart basins
32 lengthen over time, but as they are confined within the overstep, they form as long thin basins with
33 a depocenter that is bounded by the strike- or oblique-slip segments (Seeber et al., 2004). While
34 there are many pull-apart basin examples (e.g., Dead Sea Basin, Garfunkel & Ben-Avraham, 1996;
35 Death Valley Basin, SERPA et al., 1988), there has not been much discussion on other types of
36 strike-slip basins.

37 Flexural basins form when an overlying load deflects the lithosphere. This is the case
38 during mountain building, wherein the orogenic load creates accommodation space for sediment
39 infill, increasing sedimentation. However under certain conditions basement subsidence may be a
40 consequence, and not a cause, of enhanced sedimentation (Morley & Westaway, 2006).
41 Previously, the process of sedimentation-induced lower crustal flow and accommodation space
42 creation has been suggested where deep sedimentary basins form in areas featuring very weak
43 lower crust (Clift et al., 2015), such as in the fans of the Red (Clift and Sun, 2006) and Pearl Rivers
44 (Dong et al., 2020), both located at the northern continental margin of the South China Sea.

45 Generalizing conditions for sedimentation-induced creation of accommodation space, we
46 infer that it requires 1) an easily deformable tectonic environment, and 2) focused sedimentation,
47 both of which can occur in regions of prior tectonic subsidence. Considering that strike-slip faults
48 have been suggested to represent highly weakened plate boundaries (Zoback et al., 1987; Provost
49 and Houston, 2003), and that transform continental margins often follow a phase of thinning
50 (Jourdon et al., 2021), we formulate the key hypothesis of this study: sedimentary loading near
51 strike-slip faults in subsided regions may further generate basement subsidence and
52 accommodation space. This positive feedback between focused sedimentation and isostatic
53 flexural subsidence leads to the creation of a previously unrecognized type of basin that we term
54 “flexural strike-slip basin”. We test our hypothesis by two means: 1) numerical forward modeling
55 of a strike-slip system subjected to asymmetric sedimentation, and 2) interpreting seismic
56 reflection data from the East Andaman Basin in the Andaman Sea.

57 **GEOLOGICAL SETTING OF THE ANDAMAN SEA**

58 During the Cenozoic, the Andaman sea formed as a transtensional backarc basin due to the
59 coupling of India to western Myanmar (Curry, 2005). Multiple strike-slip faults exist in the
60 region, one being the Sagaing Fault (SF, Fig. 1), an active (18 mm/yr, Maurin et al., 2010; Vigny
61 et al., 2003) fault in the north of the region that connects south-westward to the Andaman spreading
62 center (Curry, 2005). However, south of the SF is another, inactive, strike-slip fault: the Andaman
63 Basin Central Fault (ABCF; Morley, 2016, 2017; Mahattanachai et al., 2021).

64 The Andaman Sea’s transtensional motion led to subsidence, causing the area to act as a
65 sediment trap. Fault trends suggest the region near the ABCF experienced ENE-WSW extension
66 in the Oligocene that shifted to NNW-SSE strike-slip motion during the early to mid-Miocene
67 (lasting ~5 Myr; Morley, 2017). The ABCF formed in a previous necking zone of hyperextended

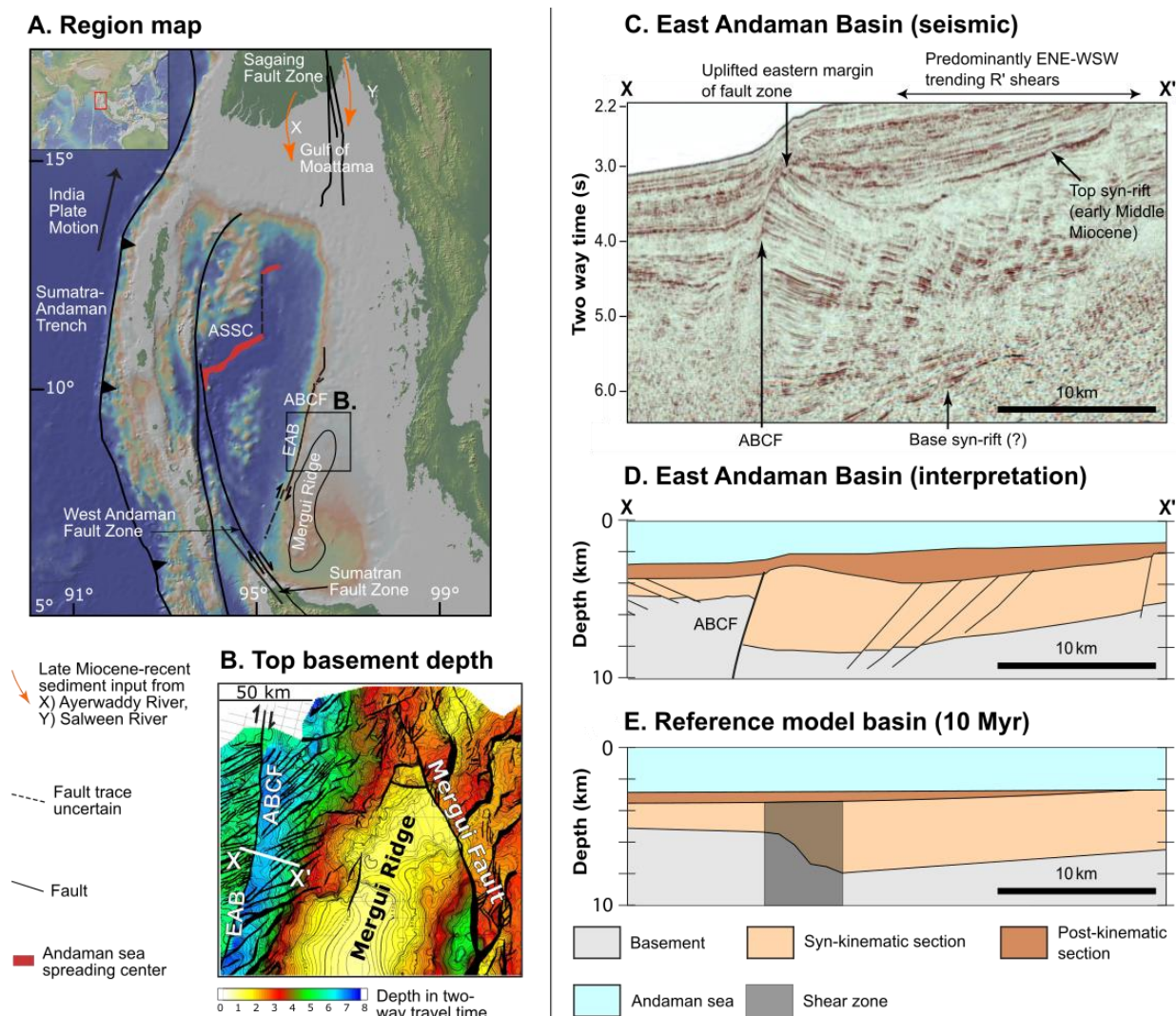


Figure 1. A. Andaman Sea Map. ASSC: Andaman Sea Spreading Center. ABCF: Andaman Basin Central Fault. EAB: East Andaman Basin. B. Depth to the top basement in two-way travel time. C. Seismic data of the EAB. D. Depth interpretation of C. E. Modeled basin. Post-kinematic sediment is found by applying the additional basement subsidence from 5 to 10 Myr to the topography at 5 Myr.

68 continental crust (7-10 km thick; Morley, 2017; Mahattanachai et al., 2021), likely because mantle
 69 heterogeneities influence fault location (Phillips et al., 2021). During strike-slip motion, the
 70 easterly Mergui Ridge (MR) was partially subaerial and acted as an asymmetric clastic sediment
 71 source for the East Andaman Basin (EAB) located along the ABCF (Mahattanachai et al., 2021).

72 The geometry of the EAB in relation to the ABCF is described in detail by Mahattanachai
 73 et al. (2021), who concluded that the long (>200 km), deep (>4 km), westward-thickening basin

74 on the east side of the sub-vertical ABCF did not fit classic extensional, or pull-apart basin
75 characteristics.

76 **MODEL SETUP AND EVOLUTION**

77 We reproduce the key aspects of the EAB region, namely that of thin lithosphere and
78 asymmetric sedimentation, using a visco-plastic 100×8×120 km (X, Y, Z) 3D box model via a
79 two-way coupling of the tectonic code ASPECT (Fig. 2B,C; Kronbichler et al., 2012; Heister et
80 al., 2017; Glerum et al., 2018; Bangerth et al., 2019; supplementary text S1) and the surface
81 processes code FastScape (Braun & Willett, 2013; Yuan et al., 2019a, 2019b; text S2). We assume
82 that a previous extensional event left the region submarine with thinned 40 km thick lithosphere.
83 The model is initialized with 4 km upper crust, 4 km lower crust, 32 km mantle lithosphere, and
84 80 km of asthenosphere (Fig. 2B; Fig. S1). In Fig. 2B, the east boundary is no-slip in all directions,
85 the west boundary is no-slip in the Z direction, 20 mm/yr in Y to induce strike-slip motion, and is
86 given a small (0.2 mm/yr) extensional component in X which helps avoid bending-induced
87 compression but does not affect the presented results (Fig. S2) . The north and south boundaries
88 are periodic to simulate an infinitely long strike-slip fault, and the initial lithostatic pressure at a
89 reference location is prescribed on the bottom boundary to allow for outflow in response to
90 sedimentation. The strike-slip fault forms self-consistently above an initial perturbation of the
91 Lithosphere-Asthenosphere Boundary (10% of lithosphere thickness) in the center of the model
92 that acts as a weak zone for deformation to localize. Accumulated plastic strain weakens the angle
93 of friction over an interval of 0 to 1 from an initial value of 30° to a final value of 7.5°, promoting
94 brittle localization.

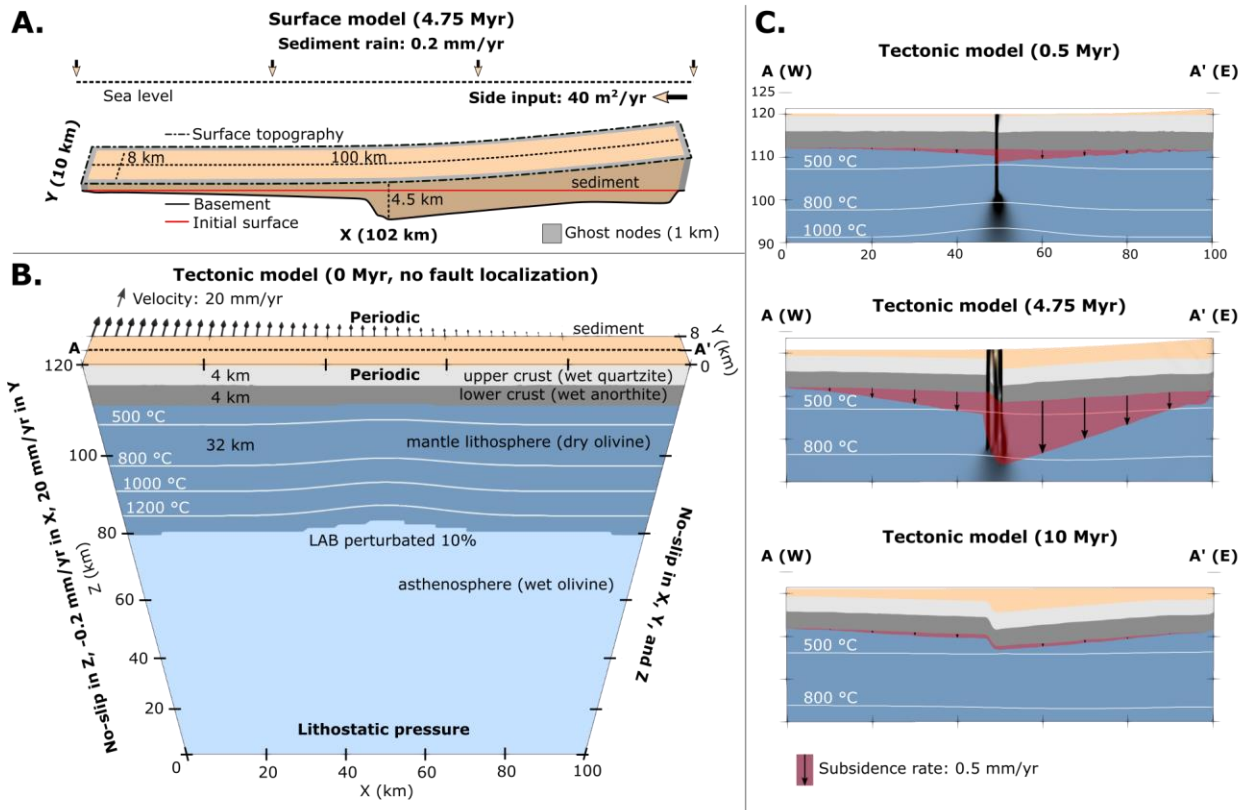


Figure 2. A. Surface processes model at 4.75 Myr and 2× vertical exaggeration. Sediment is the area between the topography (dashed line) and basement (solid black line). Ghost nodes (gray), implemented for periodic advection along Y and to control the sedimentary side input, surround the surface model and do not interact with the tectonic model. B. Initial tectonic setup. Colors represent composition, white isotherms the temperature distribution. Arrows indicate the total velocity magnitude. LAB: Lithosphere-Asthenosphere Boundary. C. Cross-sections of the top 30 km of the tectonic model along A to A' in B showcase the formation of a flexural strike-slip basin in response to sedimentation. Subsidence rate at the Moho is indicated in red. See movies S1 and S2.

95 The surface processes code FastScape is coupled to the top of the tectonic model (text S3).

96 We assume that the model is submarine and transports sediment via diffusion with a coefficient of

97 500 m²/yr, consistent with open marine environments in previous modeling studies (Rouby et al.,

98 2013). Sediment is supplied to the domain in two ways: 1) The entire surface experiences 0.2

99 mm/yr of pelagic/hemipelagic “sediment rain” sedimentation. 2) Ghost nodes (see Fig. 2A) at the

100 east boundary are uplifted each timestep to prescribe a constant sediment flux of 40 m²/yr,

101 mimicking an off-model sediment source, similar to the MR for the EAB.

102 The models are run for 10 Myr, where the first 5 Myr represent the syn-tectonic stage with
103 strike-slip motion and sedimentation. The final 5 Myr constitute the post-tectonic stage with no
104 prescribed motion or sediment supply, although marine sediment transport continues (for setup
105 details, see text S4).

106 **REFERENCE MODEL RESULTS**

107 In the reference model, strain localizes on a vertical fault near the model center (~0.5 Myr;
108 Fig. 2C). Both sides of the fault subside due to the influx of sediment, with the eastern side sinking
109 faster (1.0 vs. 0.4 mm/yr at 4.75 Myr). By 5 Myr, the eastern side of the fault has subsided almost
110 4 times more than the western side (3.6 vs. 1.0 km), rotating the strike-slip fault to a sub-vertical
111 dip. Following the halt of strike-slip motion and sedimentation, the subsidence rate lowers to 0.08
112 mm/yr as the sediment hill along the eastern boundary is distributed across the surface. By 10 Myr,
113 both sides have subsided another 0.4 km, and each basin is thickest next to the fault and thins
114 strike-perpendicularly.

115 The model indicates that flexural strike-slip basins emerge due to sedimentation above thin
116 lithosphere close to strike-slip faults. In contrast to classical half-grabens, these basins form
117 without a significant extensional component.

118 To test controls on flexural strike-slip basin formation, we run a series of models varying
119 in sedimentation rate, lithospheric thickness, and fault strength. We first focus on sedimentation
120 rate by altering the eastern side influx from 0 (i.e. only sediment rain) to 60 m²/yr (Fig. 3A-D).
121 With no side input, both sides subside evenly forming a synformal basin that is thickest at the fault
122 (Fig. 3A). This suggests that the asymmetry of the reference model basin is primarily affected by
123 the sedimentation, and not the initial perturbation. At higher side input, the eastern side sinks more,

124 from a maximum basement deflection of 0.9 km with no input to 5.7 km for 60 m²/yr of input (Fig.
 125 3D). The western side also shows greater deflection with higher eastern sediment input although
 126 it is not as pronounced (0.8 to 1.6 km), suggesting either that the sides are not fully decoupled or
 127 that more sediment reached the western side.

128 Next, we vary the lithospheric thickness from 60 to 30 km (Fig. 3E-H). The flexural
 129 basement deflection reduces for thicker, more rigid, lithosphere, from a maximum eastern

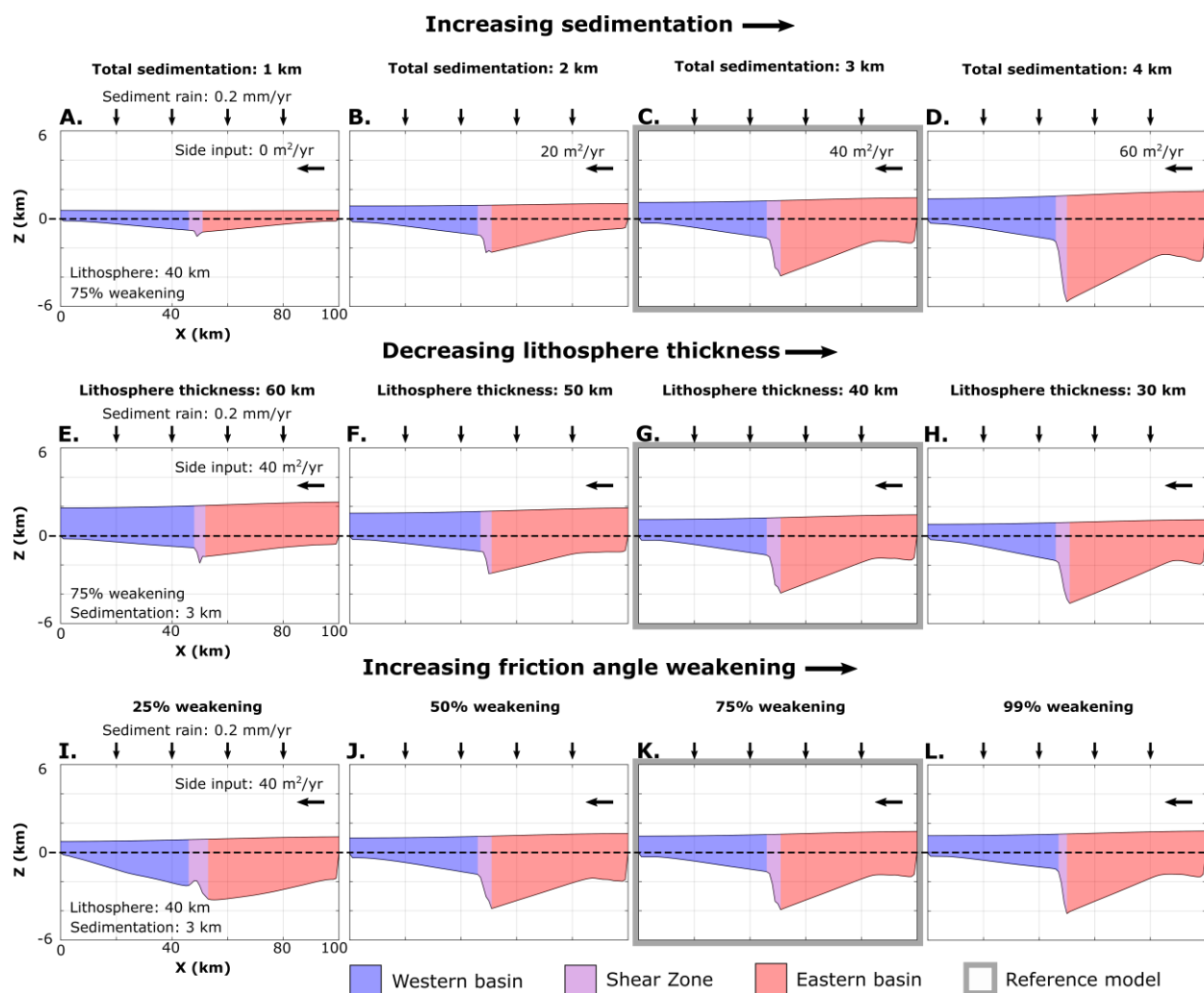


Figure 3. Basin formation when subject to variable sediment input (A-D), lithosphere thickness (E-H), and fault weakening (I-L). Dashed lines along 0 show the initial model elevation. Total sedimentation is the sediment thickness assuming even distribution across the model.

130 deflection of 4.6 km at 30 km thick lithosphere to 1.4 km at 60 km thick lithosphere. This suggests
131 that it is unlikely for deep flexural basins to form in regions with thick lithosphere.

132 Finally, we vary the amount of friction angle weakening (Fig. 3I-L). This shows that fault
133 strength affects flexural subsidence (4.2 vs. 3.3 km deflection at 99% and 25% weakening,
134 respectively). At high strength (less weakening), the two lithosphere sides are more coupled and
135 subside together. At low strength, the sides are decoupled and more sensitive to asymmetric
136 sedimentation.

137 **FLEXURAL STRIKE-SLIP BASINS IN THE ANDAMAN SEA**

138 Seismic data suggest the EAB as an asymmetric basin that spans both sides of the ABCF
139 (Mahattanachai et al., 2021). On the western side, basin thickness is relatively uniform (1-2 km;
140 Fig. 1D). Along the fault on the eastern side the basin is substantially thicker (~5 km) and thins to
141 the east towards the sediment source areas of the MR and Peninsular Thailand.

142 The Gulf of Moattama Basin (GMB) formed along the active SF, and similarly does not fit
143 the pull-apart basin criteria. The GMB basin depocenter is located along the fault, however, unlike
144 the EAB, sediment thicknesses are relatively symmetric across the fault (e.g., Morley & Alvey,
145 2015). While the EAB received sediment perpendicular to the fault trace, the GMB received
146 sediment axially from the northern Ayerwaddy and Salween rivers (Morley & Alvey, 2015).

147 We suggest that the primary requirement for flexural strike-slip basin formation is weak or
148 thin lithosphere in an area of high sedimentation. There are two basin types controlled by the
149 sedimentation pattern: 1) when sedimentation is symmetric and both sides receive a similar
150 sediment load (Fig. 3A), a *symmetric basin* forms; 2) for asymmetric sedimentation, two distinct
151 basin sides may form that subside depending on the sediment load they receive (*asymmetric basin*;

152 Fig. 3C). In both types, the maximum flexure and basin depocenter occur along the fault, and the
153 basin thins strike-perpendicularly.

154 The Andaman sea provides likely examples for both flexural strike-slip basin types. 1) The
155 GMB, where northern axial sedimentation provided relatively even sedimentation to each side of
156 the fault forming a *symmetric flexural basin*. While this basin would not have experienced purely
157 uniform sedimentation, it shows a similar synformal geometry with the depocenter along the fault
158 as in Fig. 3A. 2) The EAB (Fig. 1), where perpendicular sedimentation from the MR forced greater
159 flexure on the eastern side of the fault. This sedimentation pattern led to an *asymmetric flexural*
160 *basin*. The EAB and reference model basin both have a change in sediment thickness across the
161 fault, and basin thinning toward the sediment source. Furthermore, basin thicknesses (excluding
162 post-kinematic sediment) along the fault's eastern side (4.5 vs. 4.8 km in the model and EAB,
163 respectively) and western side (1.8 vs. 1.6 km) are comparable.

164 Despite the similarities, there are some discrepancies between the modeled basin and the
165 EAB. Eastward thinning is less pronounced in the model. As the basement slope is primarily
166 affected by the lithosphere thickness and sediment load, three possible explanations are: 1) The
167 ABCF is capped by a regional unconformity with the post-kinematic sediments (Morley, 2017;
168 Srisuriyon & Morley, 2014), and the fault may have received more sediment while active than
169 expected from the seismic data. 2) The ABCF formed in a necking zone and, as the lithosphere
170 thickness is not well-constrained, it varied spatially and may have been thinner than the 40 km
171 value used here. 3) A more significant syn-strike-slip extensional component would have further
172 deepened the basin along the fault (Sobolev et al., 2005).

173 Here we focused on the Andaman Sea, but the key requirements for flexural strike-slip
174 basins – thin lithosphere, focused sedimentation, and a weak fault – are possibly met in the New

175 Guinea Basin in the Bismarck Sea (Fig. S3; Martinez and Taylor, 1996) or the Yinggehai Basin in
176 the South China Sea (Fig. S4; Clift and Sun, 2006), but new seismic data are needed to test this
177 hypothesis. Another candidate is the Gyeongsang Basin in SE Korea, which formed in a
178 transtensional back-arc that shifted into strike-slip motion (Cheon et al., 2020). Smaller-scale
179 flexural strike-slip basins may form within larger pull-apart regions. Examples of these could be
180 the Flat Basin along the East Anatolian Fault (Garcia Moreno et al., 2011) or the Navassa Basin in
181 the Jamaica Passage (Fig. S5; Corbeau et al., 2016), which are faultward dipping strike-slip basins
182 that are not located between offset segments.

183 **CONCLUSION**

184 This study suggests a new class of flexural basins that form along strike-slip faults. These
185 basins are characterized by a fault-parallel depocenter and sediment that thins strike-
186 perpendicularly. The basins can be classified in two types, which are both represented in the
187 Andaman Sea: 1) *symmetric flexural basins*, where axial sedimentation causes a synformal shape,
188 as seen in the GMB, and 2) *asymmetric flexural basins*, where asymmetric sedimentation forces
189 one basin side to subside more than the other, as seen in the EAB.

190 Flexural strike-slip basins form due to a strike-slip fault that acts as a weak zone facilitating
191 differential subsidence due to sediment loading. The fault decouples the lithosphere sides, allowing
192 them to respond independently to the sediment load they receive. This decoupling makes the
193 sedimentation pattern important for the resulting basin symmetry. For a flexural strike-slip basin
194 to form two criteria must be met: the strike-slip fault must 1) cut through thin lithosphere and 2)
195 be subjected to a sufficient tectonic load.

196

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1 Supplementary Information

2 Flexural strike-slip basins

3
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26 **Additional supporting information (uploaded separately)**

27 Video S1

28 Video S2

29

30 **Text S1: ASPECT Methods**

31 **1.1 Governing equations**

32 We perform numerical simulations of a 3D strike-slip system using the open source finite-element
 33 code ASPECT (Advanced Solver for Problems in Earth's ConvecTion, version 2.3.0-pre; Heister
 34 et al., 2017; Kronbichler et al., 2012; Rose et al., 2017; Bangerth et al., 2019). ASPECT solves the
 35 following incompressible conservation equations assuming an infinite Prandtl number (i.e. without
 36 the inertial term),

37
$$-\nabla \cdot (2\eta\dot{\epsilon}) + \nabla P = \rho \mathbf{g}, \quad (1)$$

38
$$\nabla \cdot (\mathbf{u}) = 0, \quad (2)$$

39
$$\bar{\rho} C_p \left(\frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T \right) - \nabla \cdot \mathbf{k} \nabla T = \bar{\rho} H \quad (3)$$

40
$$+ \alpha T (\mathbf{u} \cdot \nabla P),$$

41
$$\frac{\partial c_i}{\partial t} + \mathbf{u} \cdot \nabla c_i = q_i, \quad (4)$$

42 where equation (1) represents the conservation of momentum, with η the effective viscosity, $\dot{\epsilon}$ the
 43 deviator of the strain rate tensor (defined as $\frac{1}{2}(\nabla \mathbf{u} + (\nabla \mathbf{u})^T)$), \mathbf{u} the velocity, P the pressure, ρ the
 44 density, and \mathbf{g} gravity. Equation (2) describes the conservation of volume. Equation (3) represents
 45 the conservation of energy where $\bar{\rho}$ is the reference adiabatic density, C_p the specific heat capacity,
 46 T the temperature, \mathbf{k} the thermal conductivity, H the radiogenic heating, and α the thermal
 47 expansivity. As right-hand-side heating terms, we include radioactive heating and adiabatic
 48 heating from top to bottom, respectively. Finally, we solve the advection equation (4) for each
 49 compositional field c_i (e.g., upper crust, lower crust, and accumulated plastic strain) with reaction
 50 rate q_i nonzero only for the plastic strain field.

51 1.2 Rheology

52 We use a visco-plastic rheology (Glerum et al., 2018), which additionally includes plastic
 53 weakening based on accumulated plastic strain. In the viscous regime, we use a composite of
 54 diffusion and dislocation creep (Karato and Wu, 1993), formulated as:

$$55 \quad \eta_{\text{eff}}^{\text{diff|dis}} = \frac{1}{2} A_{\text{diff|dis}}^{-1} d^m \dot{\epsilon}_e^{\frac{1-n}{n}} \exp\left(\frac{(E_{\text{diff|dis}} + PV_{\text{diff|dis}})}{nRT}\right), \quad (5)$$

56 where A is a scalar prefactor, d the grain size, $\dot{\epsilon}_e$ the square root of second invariant of the
 57 deviatoric strain rate, E the activation energy, P the pressure, V the activation volume, R the gas
 58 constant, T the temperature, and n the stress exponent. For diffusion, $n = 1$ and the equation
 59 becomes independent of strain rate. For dislocation creep, the grain size exponent m vanishes,
 60 rendering dislocation creep independent of grain size. Values for A , E , V , and n used in our models
 61 are composition-dependent and found in supplementary Table S1.

62 In the plastic regime, when viscous stresses exceed the yield stress, we use the Drucker-Prager
 63 yield criterion (Davis and Selvadurai, 2002). The effective plastic viscosity is given by

$$64 \quad \eta_{\text{eff}}^{\text{pl}} = \frac{\frac{6C \cos\phi}{\sqrt{3}(3-\sin\phi)} + \frac{6P \sin\phi}{\sqrt{3}(3-\sin\phi)}}{2\dot{\epsilon}_e}, \quad (6)$$

65 where C is the cohesion and ϕ the internal angle of friction. The accumulation of plastic strain is
 66 tracked as a compositional field. This field is used to linearly weaken ϕ from an initial value of
 67 30° to a final value of 7.5° over the accumulated plastic strain interval of 0 to 1. The time-integrated
 68 value of the strain reaction rate q_i is approximated as $\dot{\epsilon}_e \cdot dt$ when plastic yielding occurs (with dt
 69 the current timestep size).

70 **Text S2: FastScape Methods**

71 FastScape is a landscape evolution code that changes the topographic surface through uplift,
72 advection, the stream-power law (SPL), and hillslope diffusion (Braun and Willett, 2013). It can
73 additionally deposit fluvial sediment (Yuan et al., 2019a) and include a marine component that
74 handles marine sediment (sand/silt) transport and deposition, and layer compaction based on
75 sand/silt porosity (Yuan et al., 2019b). For simplicity, we here assume that the entire model surface
76 is submarine, with uniform properties (i.e., sand and silt transport coefficients are the same), and
77 that there is no compaction (porosity is zero). Hence, FastScape deforms the surface through the
78 uplift rate and marine diffusion equation only as

$$79 \quad \frac{dh}{dt} = \mathbf{U} + K_m \nabla^2 h, \quad (7)$$

80 where h is the topographic elevation, \mathbf{U} the uplift rate and K_m the marine diffusion coefficient.

81 **Text S3: ASPECT/FastScape coupling**

82 In this paper we use a two-way coupling of the tectonic ASPECT code and the landscape evolution
83 FastScape code. For this coupling a FastScape shared library is called by an ASPECT plugin to
84 deform its surface. The plugin has 3 main components: 1) Copy the height and velocity values
85 from ASPECT. 2) Initialize and run FastScape. If it is the first timestep of the tectonic model,
86 FastScape will be initialized using height and velocity values from ASPECT. In subsequent
87 timesteps, as FastScape runs separately, it can be at a higher resolution than ASPECT, only the
88 velocity values from ASPECT are transferred to FastScape. Before running FastScape, the initial

89 topography values are saved. After running FastScape, the new and initial topography are
 90 compared to determine a nodal Z velocity,

$$91 \quad \mathbf{V}_z = \frac{h_f - h_i}{dt_a}, \quad (8)$$

92 where h_i is the initial surface, h_f the final surface, and dt_a the ASPECT timestep. 3) Use the Z
 93 velocity field to determine the displacement of the ASPECT mesh surface and interior. This
 94 displacement is then interpolated onto the ASPECT surface using the overarching mesh
 95 deformation plugin.

96 The FastScape mesh includes an additional element-size layer of ghost nodes compared to the
 97 ASPECT surface mesh. The values in these nodes are not considered when interpolating the
 98 surface back to ASPECT and are used primarily to avoid FastScape boundary artifacts being sent
 99 to the ASPECT model (e.g., the boundaries do not uplift from advected topography). To avoid
 100 possible erroneous sediment flux out or into the model from artificial slopes, each timestep the
 101 ghost nodes are updated with the topography and velocity values of the nearest inward node
 102 (ASPECT boundary node).

103 Besides passing ASPECT's uplift velocities, we use the plugin's FastScape interface to supply
 104 additional input to the surface process model in two ways: 1) to add marine background
 105 sedimentation via the sediment rain effect, and 2) to add a boundary sediment flux using the ghost
 106 nodes. For the sediment rain, at each nodal point we update FastScape with a flat height increase
 107 every ASPECT timestep equivalent to a rate of 0.2 mm/yr. Second, through the diffusion
 108 component in equation (7) we prescribe a constant sediment flux at the boundary assuming that,

$$109 \quad \mathbf{Q} = K_m S, \quad (9)$$

110 where Q is the sediment flux and S the slope. Since K_m and Q are user-set parameters, to achieve
111 this we alter S by uplifting the boundary ghost nodes every ASPECT timestep so that Q remains
112 constant.

113 **Text S4: Model setup**

114 In this study we examine how a strike-slip fault responds to sedimentation. We therefore set up a
115 3D box model with dimensions $100 \times 8 \times 120$ km (X , Y , and Z , where Z is depth) and 5 compositions
116 representing a wet quartzite upper crust (Rutter and Brodie, 2004), wet anorthite lower crust
117 (Rybacki et al., 2006), dry olivine lithospheric mantle, wet olivine asthenosphere (Hirth &
118 Kohlstedt, 2003), and a sediment layer that has rheologic parameters identical to wet quartzite, but
119 with density and temperature parameters consistent with sediment (Sippel et al., 2017). The total
120 crustal thickness is set to 8 km (4 km upper crust, 4 km lower crust) based on crustal estimates of
121 the area (7-10 km; Mahattanachai et al., 2021). The lithospheric mantle extends between the Moho
122 and the lithosphere-asthenosphere boundary (LAB) at 40 km depth. The LAB depth, like the crust,
123 has been perturbed by a previous extensional period. The remaining material beneath the LAB is
124 considered asthenosphere (Fig. S1). While there is no initial sediment layer, the top boundary is
125 fixed to a sediment composition so that any top-inflow of material due to topography changes other
126 than uplift is sediment.

127 The ASPECT model mesh consists of two element sizes: 1 km and 2 km. The upper 8 km of the
128 model is refined at 1 km to best resolve the crust and forming sediment layer. This high-resolution
129 area is additionally extended to a depth of 35 km from $X = 42$ km to $X = 52$ km to better resolve
130 the strike-slip fault. All other areas are refined at 2 km.

131 The initial temperature above the LAB is determined by a steady-state geotherm (Turcotte and
132 Schubert, 2002), and below by a mantle adiabat. For simplicity, an initial weak zone is seeded

133 through a small perturbation: we raise the LAB locally by 10% of the lithospheric mantle thickness.
134 We fix the top boundary temperature at 0 °C, and the bottom boundary at the temperature initially
135 determined from the mantle adiabat. All other boundaries are set to zero heat-flux.

136 The coupled model is run for 10 Myr, where the model in the first 5 Myr includes non-zero velocity
137 boundary conditions. During this time, the western boundary is given a strike-slip component of
138 20 mm/yr (in Y), and an extensional component of 0.2 mm/yr (in X), while Z is set to no-slip. This
139 gives a total of 100 km of dextral strike-slip motion and 1 km of extension. The small extensional
140 component is introduced to avoid compressional pop-ups that form at the shear zone as the
141 lithosphere subsides due to the sediment load (Fig. S2). The exact extensional value is chosen to
142 accommodate horizontal stress forces related to isostatic compensation. From 5-10 Myr, extension
143 and strike-slip motion stop as the western boundary is set to no-slip in all directions. All other
144 boundary conditions are constant for the entire model run, with the eastern boundary being no-slip
145 in all directions, the north and south boundaries set to periodic to simulate an infinitely long strike-
146 slip fault the initial lithostatic pressure computed at a reference location prescribed on the bottom
147 boundary to allow for outflow in response to sedimentation, and the top deformed through the use
148 of FastScape.

149 FastScape is set up with an arbitrarily high sea level so that the entire model is considered
150 submarine. This setup leads to a model with no acting stream power law, and sediment being
151 moved solely through marine sediment diffusion. For simplicity, we additionally assume that there
152 is no compaction and no difference between sand and silt. As such, we use a diffusion coefficient
153 of 500 m²/y for both, a value consistent with open marine environments in previous modelling
154 studies (e.g., Rouby et al., 2013). During the syn-strike-slip phase of the tectonic model (0-5 Myr)
155 we supply sediment to the model in two ways: 1) To account for pelagic/hemipelagic

156 sedimentation (sediment rain) we deposit at a constant sedimentation rate of 0.2 mm/yr. 2) We
157 assume there is an asymmetric off-model source of sediment, similar to the eastern Mergui Ridge
158 for the East Andaman Basin, that inputs sediment into the system from the eastern boundary at a
159 rate of 40 m²/yr. This is done utilizing equation (9), wherein we uplift the ghost nodes at each
160 timestep so that a constant flux is prescribed through marine diffusion. After this syn-tectonic
161 stage, sediment supply to the system is halted, although marine diffusion continues to work on the
162 topography.

163

Parameter	Symbol	Units	Sediment	Upper crust	Lower crust	Lithospheric mantle	Asthenosphere
Reference density (surface)	ρ_0	kg m ⁻³	2520	2700	2850	3280	3300
Thermal expansivity	α	K ⁻¹	$3.7 \cdot 10^{-5}$	$2.7 \cdot 10^{-5}$	$2.7 \cdot 10^{-5}$	$3.0 \cdot 10^{-5}$	$3.0 \cdot 10^{-5}$
Thermal diffusivity	κ	m ² s ⁻¹	$7.28 \cdot 10^{-7}$	$9.26 \cdot 10^{-7}$	$5.85 \cdot 10^{-7}$	$8.38 \cdot 10^{-7}$	$8.33 \cdot 10^{-7}$
Heat capacity	C_p	J kg ⁻¹ K ⁻¹	1200	1200	1200	1200	1200
Heat production	H	W m ⁻³	$1.2 \cdot 10^{-6}$	$1.5 \cdot 10^{-6}$	$0.2 \cdot 10^{-6}$	0	0
Cohesion	C	Pa	$20 \cdot 10^6$	$20 \cdot 10^6$	$20 \cdot 10^6$	$20 \cdot 10^6$	$20 \cdot 10^6$
Internal friction angle (unweakened)	ϕ	°	30	30	30	30	30
Strain weakening interval	-	-	[0,1]	[0,1]	[0,1]	[0,1]	[0,1]
Strain weakening factor	ϕ_{wf}	-	0.25	0.25	0.25	0.25	0.25
Creep properties			Sediment	Wet quartzite	Wet anorthite	Dry olivine	Wet olivine
Stress exponent (dis)	n	-	4.0	4.0	3.0	3.5	3.5
Constant prefactor (dis)	A_{dis}	Pa ⁻ⁿ s ⁻¹	$8.57 \cdot 10^{-28}$	$8.57 \cdot 10^{-28}$	$7.13 \cdot 10^{-18}$	$6.52 \cdot 10^{-16}$	$2.12 \cdot 10^{-15}$
Activation energy (dis)	E_{dis}	J mol ⁻¹	$223 \cdot 10^3$	$223 \cdot 10^3$	$345 \cdot 10^3$	$530 \cdot 10^3$	$480 \cdot 10^3$
Activation volume (dis)	V_{dis}	m ³ mol ⁻¹	0	0	$38 \cdot 10^{-6}$	$18 \cdot 10^{-6}$	$11 \cdot 10^{-6}$
Constant prefactor (diff)	A_{diff}	Pa ⁻¹ s ⁻¹	$5.79 \cdot 10^{-19}$	$5.79 \cdot 10^{-19}$	$2.99 \cdot 10^{-25}$	$2.25 \cdot 10^{-9}$	$1.5 \cdot 10^{-9}$
Activation energy (diff)	E_{diff}	J mol ⁻¹	$223 \cdot 10^3$	$223 \cdot 10^3$	$159 \cdot 10^3$	$375 \cdot 10^3$	$335 \cdot 10^3$
Activation volume (diff)	V_{diff}	m ³ mol ⁻¹	0	0	$38 \cdot 10^{-6}$	$6 \cdot 10^{-6}$	$4 \cdot 10^{-6}$
Grain size (diff)	d	m	0.001	0.001	0.001	0.001	0.001
Grain size exponent (diff)	m	-	2.0	2.0	3.0	0	0

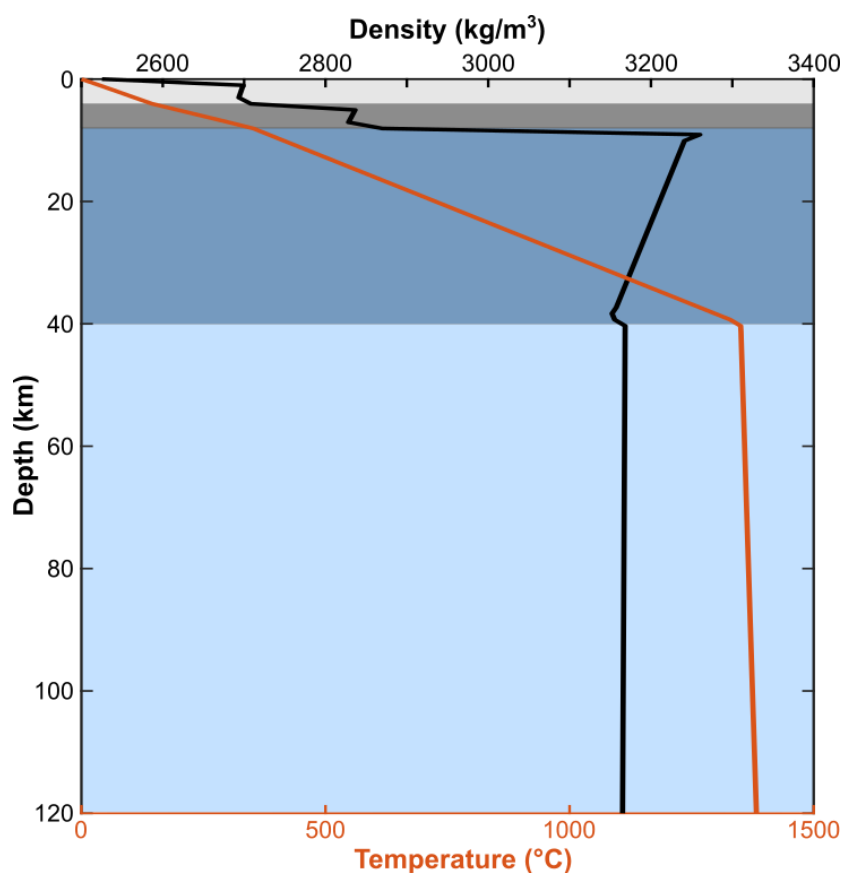
164 **Table S1:** ASPECT model parameters.

166

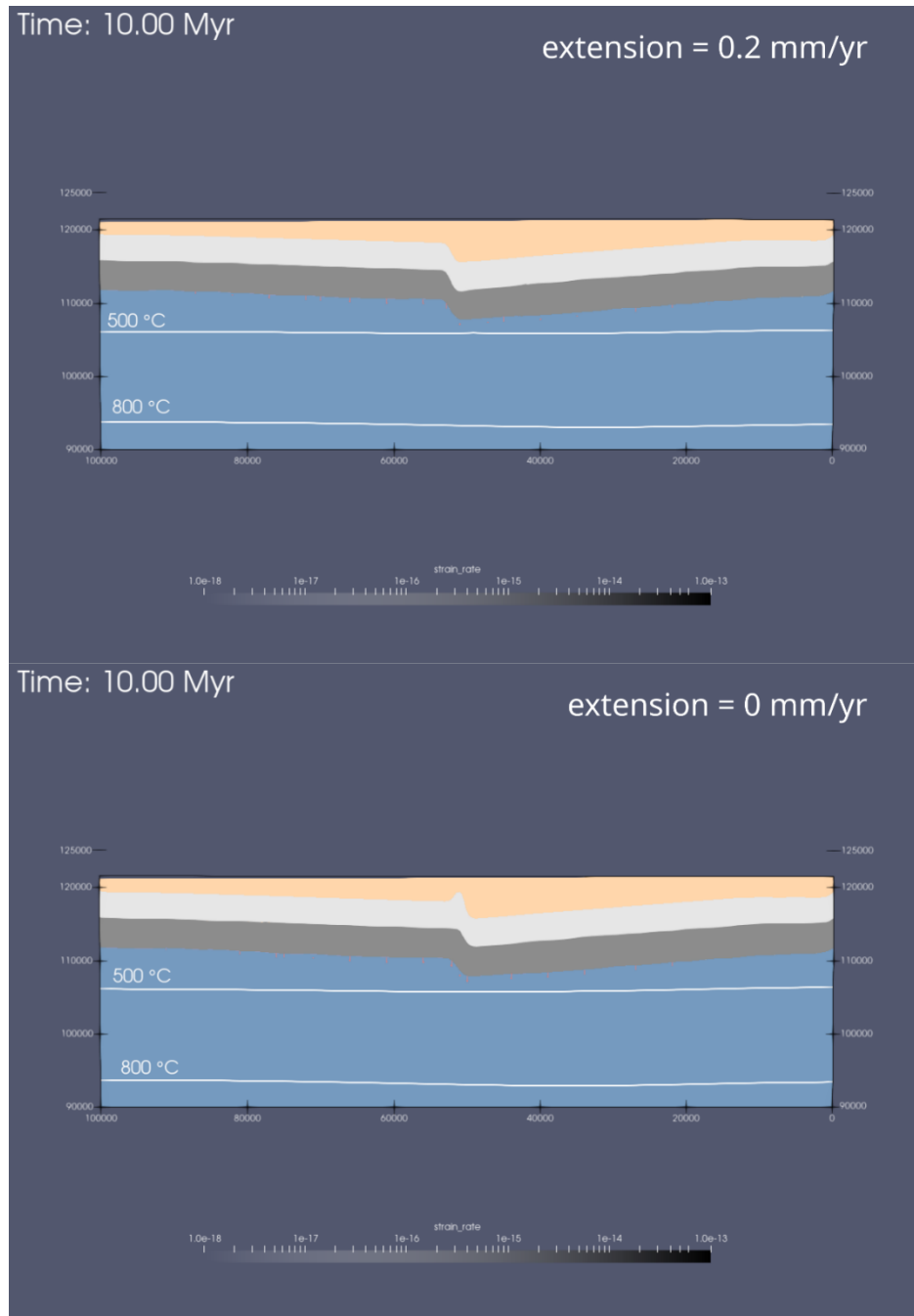
Parameter	Symbol	Unit	Value
Marine sand transport coefficient	K_{sand}	m^2/yr	500
Surface sand porosity	φ_{sand}	-	0
Sand e-folding depth	z_{sand}	m	0
Marine silt transport coefficient	K_{silt}	m^2/yr	500
Surface silt porosity	φ_{silt}	-	0
Silt e-folding depth	z_{silt}	m	0
Sand-shale ratio	F	-	1
Thickness of transport layer	L	m	100
Sea level	h_{sea}	m	5000

167 **Table S2:** FastScape model parameters.

168



169 **Figure S1:** Density (black) and temperature (red) profiles by depth. Colored backgrounds
 170 represent the compositions, with light gray (upper crust), dark gray (lower crust), dark blue
 171 (mantle lithosphere), and light blue (asthenosphere).

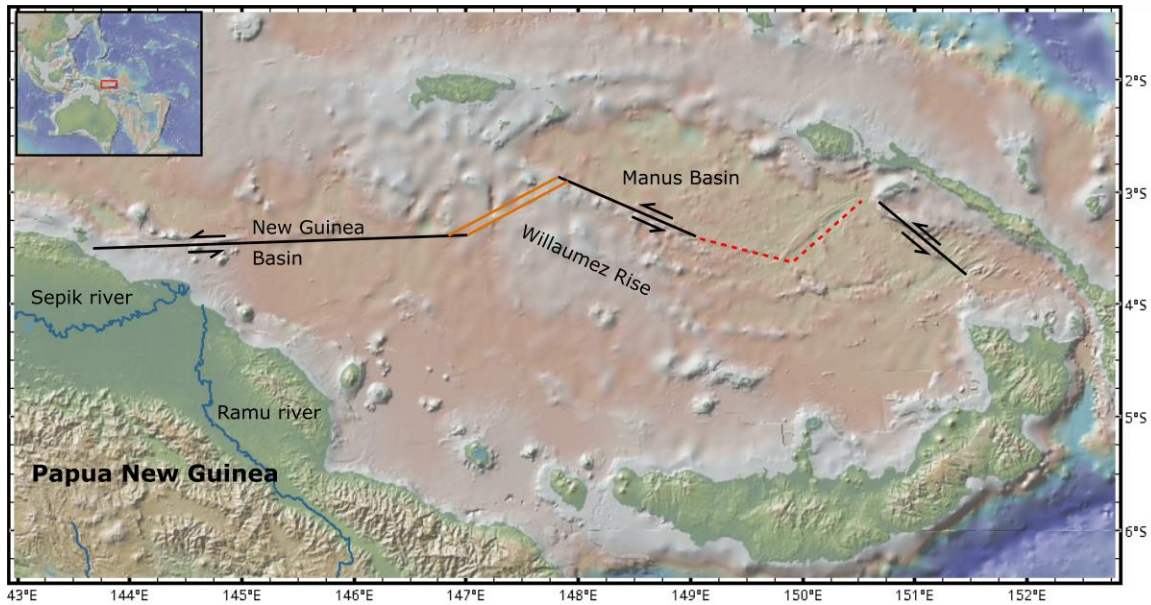


172

173 **Figure S2:** Comparison showing the reference model with A) a 0.2 mm/yr extensional component.

174 B) no extensional component, leading to the formation of a small compressional pop-up in the

175 center.



176

177 **Figure S3:** Regional map of the Manus back-arc region, fault locations based on Fig. 1 in Martinez

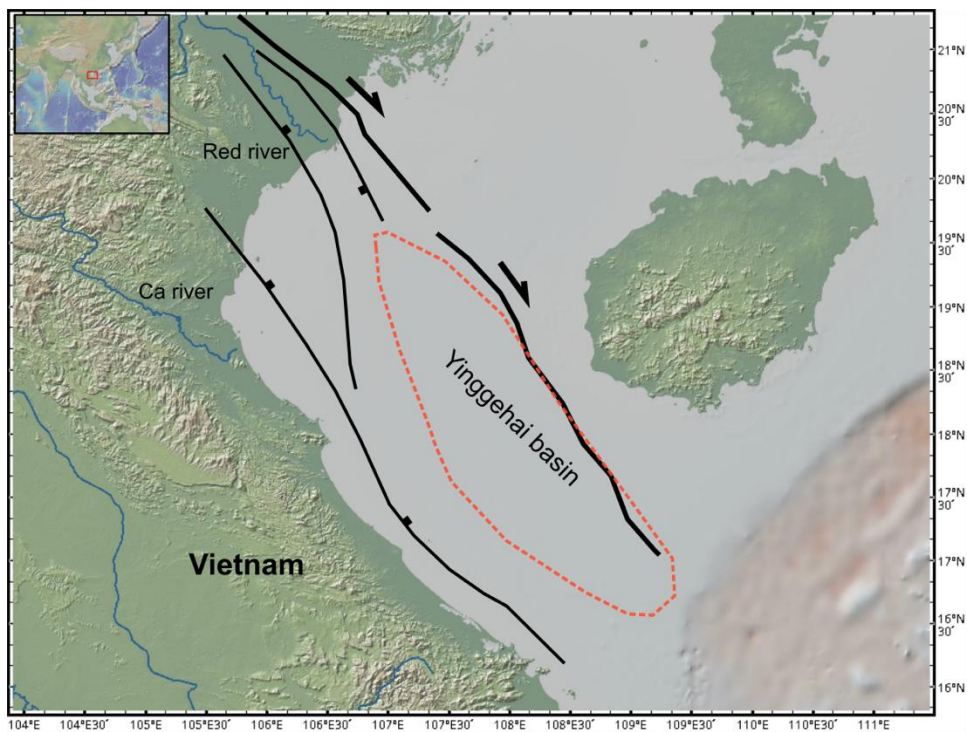
178 and Taylor, 1996. Black lines indicate strike-slip faults, parallel orange lines spreading centers,

179 dashed red lines lava fields, and blue lines major rivers. This figure was made using GeoMappApp

180 (www.geomappapp.org; Ryan et al., 2009).

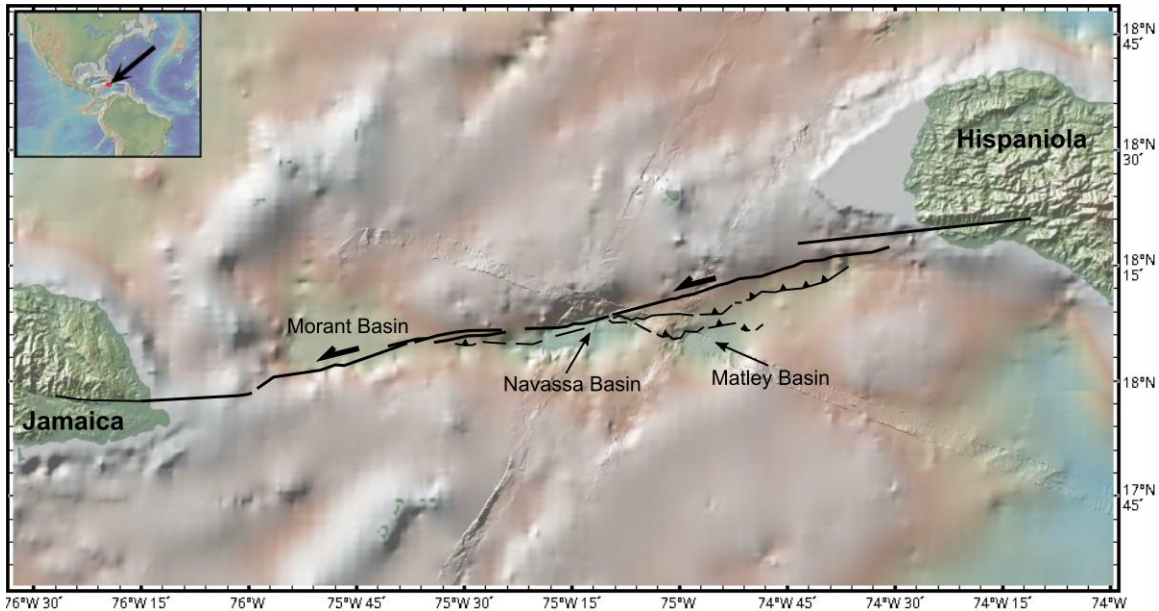
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183
184 **Figure S4:** Regional map showing the Red River Fault Zone, and location of the Yinggehai basin.
185 Fault locations based on Fig. 10 in Noda, 2013. Black lines show faults, blue lines major rivers,
186 and the Yinggehai basin is outline in the dashed orange circle. This figure was made using
187 GeoMappApp (www.geomappapp.org; Ryan et al., 2009).

188
189
190



191
 192 **Figure S5:** Regional map of the Jamaica Passage showing the Navassa strike-slip basin along the
 193 Enriquillo-Plantain-Garden Fault Zone. Fault locations based on Fig. 6 in Corbeau et al., 2016.
 194 This figure was made using GeoMappApp (www.geomapp.org; Ryan et al., 2009).

195

196 **Video S1:** Full evolution of the tectonic reference model (Fig. 2C,K,G). Colors represent
 197 composition where tan is sediment, light gray is upper crust, dark gray is lower crust, dark blue is
 198 mantle lithosphere, and light blue is the asthenosphere. The white lines are temperature contours,
 199 gray-scale the strain rate, and arrows indicate the total velocity magnitude.

200

201 **Video S2:** Evolution of the middle slice of the top 30 km of the reference tectonic model. Colors
 202 represent composition where tan is sediment, light gray is upper crust, dark gray is lower crust,
 203 dark blue is mantle lithosphere, and light blue is the asthenosphere. The white lines are temperature

204 contours, gray-scale the strain rate, and red arrows indicate the subsidence rate (Z velocity) along
 205 the 8 km depth contour.

206

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