Various lithospheric deformation patterns derived from rheological contrasts between continental terranes: Insights from 2-D numerical simulations

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Abstract. Continents are formed by the amalgamation of numerous micro-terranes and island arcs, so they have spatially varying lithosphere strengths. The Crème brûlée (CB) model and the Jelly sandwich (JS) model have been commonly used to describe continental lithosphere strength-depth variations. Depending on the strength of continental lower crust, the CB and JS models can be further subdivided into two subclasses, in which the I subclass (CB-I and JS-I) and II subclass (CB-II and JS-II) respectively have a strong or weak lower crust. During continental collision, lithosphere deformation is the byproduct of the comprehensive interaction of multiple terranes. Here we used 2-D thermo-mechanical numerical models that contain three continental terranes to systematically explore the effects of terranes with various strengths on continental deformation, and studied the effects of different rheological assumptions on terrane deformation. We found four types of lithosphere deformation patterns: collision, subduction, thickening and delamination, and replacement. These simulation patterns are seen in observed deformation patterns and structures in East Asia, suggesting they are likely to be naturally occurring modes of intracontinental orogenesis.

1. Introduction

Continents have undergone multiple break-up and assembly events during the past ~2 billion years, with the assembly events often being associated with the accretion and deformation of numerous micro-terranes (Mitchell et al., 2021). Accreted terranes have different ages ranging from ~3500 – 3000 Ma to 50 – 0 Ma, and diverse compositions and structures linked to their diverse continental, arc, or

oceanic origin, which often leads to them having distinct initial lithospheric thicknesses and strengths (Artemieva, 2006; Audet and Bürgmann, 2011; Pasyanos et al., 2014; Morgan and Vannucchi, 2022). The lithosphere of ancient continental terranes like cratons are usually thick and strong, while younger lithosphere of continental margins and tectonically active regions is thin and weak (Audet and Bürgmann, 2011; Burov, 2011), and deeply buried former oceanic fragments can have temperature and strengths that vary over ~0.5 Gyr timescale (Morgan and Vannucchi, 2022).

Continental lithosphere strength conventionally been represented by two prevailing rheology models —the Crème brûlée (CB) and the Jelly sandwich (JS) idealizations (Chen and Molnar, 1983; Jackson, 2002; Burov and Watts, 2006; Bürgmann and Dresen, 2008; Burov, 2011). The Crème brûlée scenario suggests that lithosphere strength resides entirely in the crust, with the lithospheric mantle being much weaker (with this strength contrast being the explanation for why little seismicity is typically seen in the continental mantle, despite rock-mechanics arguments that it should usually be stronger than its overlying crust). In contrast, the Jelly sandwich model is based on conventional rock mechanics arguments which imply that in general the continental middle and lower crust should be weaker than overlying cooler upper crust and underlying further-from-solidus lithospheric mantle (Figure 1a). The rheology of the continental lower crust can also differ strongly in different continental terranes due to the varieties in composition, temperature, water content, stress, and tectonic environment (Bürgmann and Dresen, 2008; Hacker et al., 2015; Morgan and Vannucchi, 2022). Therefore, the CB and JS conceptualizations can be further subdivided into CB-I and CB-II, JS-I, and JS-II subclasses that reflect potentially variable strengths of the lower crust: CB-I and JS-I, CB-II and JS-II have strong and weak continental lower crust, respectively (Fig. 1a). Observations in Eastern Asia show a wide variability in terrane deformation styles that argue for the potential feasibility of all four of these rheological models (Figure 1b).

Several previous numerical modelling studies have discussed the effects of rheological contrasts between terranes in lithosphere deformation in a collisional system. Studies containing two terranes have explored contrasts in crustal rheology, and found that this can greatly change the morphology, size and deep lithosphere structure of collisional orogenic belt (Chen, 2021; Chen et al., 2017; Cook and Royden, 2008; Faccenda et al., 2008; Sun and Liu, 2018; Vogt et al., 2018; Xie et al., 2021). Strong crust also has the potential to protect its underlying lithospheric mantle from deformation and destruction (Heron and Pysklywec, 2016). Studies containing three or more terranes in their models

59　have usually focused on the middle terrane which can play a crucial role in lithosphere deformation in a

60　collisional system (Kelly et al., 2016, 2020; Li et al., 2016; Huangfu et al., 2018, 2022; Sun and Liu,

61　2018; Xie et al., 2023). A weak middle terrane is easy to be thickened, to the point where eventually its

62　lithospheric mantle can be delaminated from the crust; while a moderate-strength middle terrane can

63　induce far-field orogenesis; and a strong middle terrane may prevent propagation of deformation and

64　facilitate underthrusting of the advancing terrane. In addition, some studies have also stressed the

65　importance of local pre-existing weak zones which can change the order and style of lithosphere

66　deformation (Chen et al., 2020; Heron et al., 2016; Sokoutis and Willingshofer, 2011; Xie et al., 2021).

67　Large-scale continental collisional system often involves the multiple units of an indenting terrane, a

68　middle terrane, and far-end backwall terranes. These terranes have different lithosphere rheologies and

69　thicknesses, and they collectively contribute to several styles of continental deformation (Artemieva,

70　2006; Audet and Bürgmann, 2011; Pasyanos et al., 2014; Morgan and Vannucchi, 2022). Here, we use a

71　2-D thermo-mechanical numerical modeling method to systematically study the effects of terranes with

72　various rheological properties on continental deformation. Our numerical models simulate a

73　continent-continent collisional system that contains three continental terranes. They explores the effects

74　of four groups of lithosphere deformation patterns linked to the four rheological idealizations of CB-I,

75　CB-II, JS-I, and JS-II applied to each terrane. We will summarize the rheological features for each

76　deformation pattern, and then apply the simulations to better understand ongoing and past deformation

77　histories of various orogenic belts in eastern Asia, such as the eastern Tien Shan orogenic belt, the

78　Tibetan Plateau and the Early Paleozoic Orogen in Southeastern China.

79　**2. Numerical modelling method and model setup**

80　**2.1. Numerical modelling method**

81　Our thermo-mechanical models were performed with the I2VIS code of Gerya and Yuen (2003a),

82　previously used in Xie et al. (2021, 2023). This code combines finite differences with marker-in-cell

83　techniques to solve the mass, momentum, and energy conservation equations for incompressible flow.

84　It incorporates the non-Newtonian visco-plastic rheologies for the lithosphere, as well as the possibility

85　to include parameterizations of the effects of surface processes like sedimentation and erosion.

2.1.1. Governing equations

The mass conservation equation for incompressible flow is:

$$\frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} = 0 \,, \tag{1}$$

The momentum conservation equations (Stokes equations) are:

$$\begin{aligned}
\frac{\partial \sigma'_{xx}}{\partial x} + \frac{\partial \sigma'_{xy}}{\partial y} &= \frac{\partial P}{\partial x} \\
\frac{\partial \sigma'_{yy}}{\partial y} + \frac{\partial \sigma'_{xy}}{\partial x} &= \frac{\partial P}{\partial y} - g\rho
\end{aligned} \,, \tag{2}$$

The energy (heat) conservation equation is:

$$\begin{aligned}
\rho C_p \frac{DT}{Dt} &= -\frac{\partial q_x}{\partial x} - \frac{\partial q_y}{\partial y} + H_r + H_s + H_a + H_L \\
q_x &= -k\frac{\partial T}{\partial x} \\
q_y &= -k\frac{\partial T}{\partial y} \\
H_a &= T\alpha\frac{DP}{Dt} \\
H_s &= \sigma'_{xx}\dot{\varepsilon}_{xx} + \sigma'_{yy}\dot{\varepsilon}_{yy} + 2\sigma'_{xy}\dot{\varepsilon}_{xy}
\end{aligned} \,, \tag{3}$$

where x and y represent the horizontal and vertical coordinate directions, and v_x and v_y are the corresponding velocity components, respectively. σ'_{ij} and $\dot{\varepsilon}_{ij}$ ($i, j = x, y$) are deviatoric stress and strain-rate tensors, respectively; g is the gravitational acceleration; ρ is density. In the heat conservation equation, q_x and q_y are the horizontal and vertical components of the heat flux, respectively; C_p is heat capacity, and H_r, H_a, H_s, and H_L denote the radioactive, adiabatic, shear, and latent heat production, respectively; k is the thermal conductivity.

The rheological constitutive relationship connects the deviatoric stress and strain rate:

$$\begin{aligned}
\sigma'_{xx} &= 2\eta_{eff}\dot{\varepsilon}_{xx}, & \dot{\varepsilon}_{xx} &= \frac{\partial v_x}{\partial x} \\
\sigma'_{xy} &= 2\eta_{eff}\dot{\varepsilon}_{xy}, & \dot{\varepsilon}_{xy} &= \frac{1}{2}\left(\frac{\partial v_x}{\partial y} + \frac{\partial v_y}{\partial x}\right), \\
\sigma'_{yy} &= 2\eta_{eff}\dot{\varepsilon}_{yy}, & \dot{\varepsilon}_{yy} &= \frac{\partial v_y}{\partial y}
\end{aligned} \tag{4}$$

101   where η_{eff} is the effective viscosity.

102   **2.1.2. Rheology**

103   Here we make the conventional assumption that the crust and mantle have a visco-plastic rheology.

104   Viscous deformation is determined as a combination of diffusion and dislocation creep that depends on

105   temperature, pressure, and strain rate, expressed as (Gerya, 2019):

106
$$\eta_{disl} = \frac{1}{2} \frac{1}{(A_D)^{-1/n} (\dot{\varepsilon}_{II})^{(n-1)/n}} \exp\left(\frac{E_a + V_a P}{nRT}\right) * S$$
$$\eta_{diff} = \frac{1}{2} \frac{A_D}{\sigma_{cr}^{(n-1)}} \exp\left(\frac{E_a + V_a P}{RT}\right) * S$$
                            (5)

107   For mineral aggregates, both dislocation and diffusion creep occur simultaneously, with a combined

108   effective viscosity given by:

109
$$\frac{1}{\eta_{ductile}} = \frac{1}{\eta_{disl}} + \frac{1}{\eta_{diff}} ,$$
                            (6)

110   where η_{disl} and η_{diff} are viscosities for dislocation and diffusion creep, respectively. σ_{cr} is the

111   critical stress for the dislocation to diffusion stress transition, and the parameters A_D, E_a, V_a, and n are a

112   material constant, activation energy, activation volume, and stress exponent, respectively, and R is the

113   universal gas constant. The strength scaling factor, S, is introduced as a simple parameter to vary the

114   lithospheric viscosity.

115   Plasticity is implemented using a conventional pseudo-viscous yield criterion first used to study rifting

116   (e.g. Chen and Morgan, 1990) that is extended to include a strain-weakening-like parameterization of

117   fracture-related strain weakening (Gerya et al., 2010; Vogt et al., 2017):

$$\eta_{plastic} = \frac{\sigma_{yield}}{2\dot{\varepsilon}_{II}}$$

$$\sigma_{yield} = C + P\phi$$

118
$$C = \begin{cases} C_a + (C_b - C_a) \times \dfrac{\gamma}{\gamma_{cr}}, & \text{if } \gamma \leq \gamma_{cr} \\ C_b, & \text{if } \gamma \geq \gamma_{cr} \end{cases}, \tag{7}$$

$$\phi = \begin{cases} \phi_a + (\phi_b - \phi_a) \times \dfrac{\gamma}{\gamma_{cr}}, & \text{if } \gamma \leq \gamma_{cr} \\ \phi_b, & \text{if } \gamma \geq \gamma_{cr} \end{cases}$$

119 where σ_{yield} is yield stress, P is dynamic pressure, γ is the integrated plastic strain, and γ_{cr} is the

120 upper strain limit for fracture-related weakening. C and ϕ are cohesion and friction angle that depend

121 on the plastic value. C_a and ϕ_a are the initial and C_b and ϕ_b are final strength values, respectively.

122 This involves making the rheological assumption that deeply percolating fluids and high pore fluid

123 pressures can significantly lower the plastic strength of fractured rocks.

124 The final effective viscosity is determined by the minimum value between the ductile and plastic

125 viscosities (Ranalli, 1995):

126
$$\eta_{eff} = \min(\eta_{ductile}, \eta_{plastic}). \tag{8}$$

127 **2.1.3. Surface processes**

128 Topography in our models evolves according to a transport equation that is solved at each time step,

129 with a crude local parameterization of effects of accounts for sedimentation and erosion:

130
$$\frac{\partial y_{es}}{\partial t} = v_y - v_x \frac{\partial y_{es}}{\partial x} - v_s + v_e. \tag{9}$$

131 Where y_{es} is the vertical position of the surface as a function of horizontal distance x; and v_x and v_y are

132 the corresponding velocity components, respectively. v_s and v_e are the sedimentation and erosion rates,

133 respectively, conforming to the relation:

134 $v_s = 0$ mm/yr, $v_e = 0.3$ mm/yr when $y_{es} > 5$ km;

135 $v_s = 0.3$ mm/yr, $v_e = 0$ mm/yr when $y_{es} < 5$ km.

136 Owing to surface processes are not our focuses in this study, for the aim of simplification, we adopt a

137 small erosion and sedimentation rates of 0.3 mm/yr, which are similar to previous studies (Gerya and

138 Yuan, 2003b; Bian et al., 2020). As well, we simply choose a very large value of 5 km as the threshold

139 for initiating denudation and sedimentation to further weaken the influences of surface processes on the

140 evolutions of our model.

141 **2.2. Model Setup**

142 The 2-D numerical model covers a rectangular computational domain of 3000 km × 700 km and

143 consists of 1360 × 400 non-uniform grid cells with dozens of mobile markers in each grid cell to

144 transport physical properties (Figure 2a). Above 300 km, the cell-size of the grid in the middle of

145 model (X = 1300 – 2200 km) is 1 km × 1 km, and gradually widens towards the two sides to finally

146 become 5 km × 1 km. From the 300 km depth to the model bottom, each grid is stretched to 5 km in the

147 vertical direction. As a result, the grid in the middle of the model (X = 1300 – 2200 km) is 1 km × 5 km

148 and is 5 km × 5 km in the other regions. Changing resolutions in different model regions can ensure the

149 model can finely depict lithosphere deformation in the region of interest while improving the

150 calculation's efficiency.

151 In the initial configuration, the model comprises three continental terranes — the Pro-, Mid- and

152 Retro-terrane — which refer to the indenting 'Pro-' terrane driven by plate convergence, an

153 intermediate 'Mid-' terrane, and a far-end backwall 'Retro-' terrane, respectively (Figure 2a). For the

154 purpose of simplification, the three terranes are assumed to have the same initial crustal structure with

155 20 km thick upper and lower crust, respectively. In the meanwhile, to simulate lateral structure

156 differences within continental lithosphere (Pasyanos et al., 2014), thicknesses of the initial lithospheric

157 mantle of the Pro-, Mid- and Retro-terrane are 160 km, 90 km, and 120 km, respectively. The rest of

158 the region is filled by asthenosphere except along the model top, where a 20 km thick layer of "sticky

159 air" with low viscosity (1×10^{18} Pas) and low density (1 kg/m^3) is placed to simulate the effects of a

160 free surface (Schmeling et al., 2008). Flow laws and material properties for each lithospheric layer are

161 listed in Table 1.

162 Mechanical boundary conditions of the model are that the top and sides are free-slip boundaries which

163 mean that the vertical velocity at the top boundary and horizontal velocity at the side boundaries are all

164 zero. The bottom is assumed to be a somewhat non-physical 'permeable boundary' that was developed

165 to reduce the required depth of the computational region (Burg and Gerya, 2005). For top-driven flows

166 like those considered here, this approximation has been shown to not affect deformation in the upper

167   parts of the region (Burg and Gerya, 2005). Finally, a constant convergence rate of 20 mm/yr is

168   assigned to the Pro-terrane (X = 1000 km) to drive the model.

169   Initial temperature conditions are set as follows: the model top is set to $0\,°C$, the two side boundaries

170   are adiabatic boundaries with zero horizontal heat fluxes, and the model bottom has an initial

171   temperature of 1593 $°C$, and can dynamically adjust as the model evolves. The initial thermal gradient

172   in the crust is 15 $°C$/km in the three terranes, so their Moho temperature is 600 $°C$. A temperature of

173   1330 $°C$ is applied at the bottom of the lithospheric mantle of the three terranes, which leads to the Pro-

174   and Mid-terrane having minimum and maximal thermal gradients in the lithospheric mantle,

175   respectively (see the right plane in Figure 2a). An adiabatic thermal gradient of 0.5 $°C$/km is assumed

176   within the asthenosphere. The temperature field would evolve over time, thus, although the three

177   terranes are not in thermal equilibrium at the start of the experiments, it has few effects on model

178   evolution. The initial setup of lithosphere structure and temperature field make the Mid-terrane weakest

179   when same rheology model is used for the three terranes.

180   **3. Simulation Results**

181   The rheological models of CB-I, CB-II, JS-I, and JS-II result from different strength scaling factors for

182   the upper crust, the lower crust, and the lithospheric mantle in our numerical models (Figure 2b). We

183   systematically test the effects of these rheological assumptions on the deformation of the Pro-, Mid-

184   and Retro-terranes. According to different behaviors of lithosphere deformation, these simulation

185   results can be categorized into four basic modes of collision, subduction, thickening and delamination,

186   and replacement (Figures S1 and S2). In the deformation mode of collision, the lithospheric mantle of

187   the Mid-terrane is extruded out and the lithospheric mantles of the two bounding terranes meet and

188   collide together. In the deformation mode of subduction, the lithospheric mantle of one of the bounding

189   terranes subducts into the deep mantle below the Mid-terrane while the other one keeps almost

190   undeformed. In the deformation mode of thickening and delamination, one of the bounding terranes is

191   shortened by compression, and delamination may come on the heels of thickening of lithosphere due to

192   gravitational instability in some cases. In the deformation mode of replacement, the bottom of weak

193   and thick lithospheric mantle of the bounding terrane is scraped off by the strong lithospheric mantle of

194   the Mid-terrane, and replaced by the latter. Here, we select a typical case for each mode of lithosphere

195   deformation to describe more details of these modes of model evolution.

3.1. Case 1: Lithosphere Collision

197   Case 1 represents the scenario of lithosphere collision (Figure 3). In this model, the assumed

198   rheological models for the Pro-, Mid- and Retro-terrane are JS-I, JS-II, and JS-I, respectively, which

199   means that the Mid-terrane has a significantly weaker lower crust relative to the Pro- and

200   Retro-terranes. The lithospheric mantle of the Mid-terrane is also slightly weaker due to its thinner

201   lithosphere and correspondingly higher initial temperature field. The lithosphere strength profiles of the

202   three terranes are shown in Figure 3g.

203   The Mid-terrane is the first to deform when the Pro-terrane begins to collide, absorbing plate

204   convergence in the form of lithosphere thickening (Figures 3a and 3d). The upper crust of the

205   Mid-terrane breaks due to strain weakening, and several reverse faults are formed to absorb crustal

206   shortening. The lower crust folds, and strain diffusely distributes within it. Since the Retro-terrane is

207   relatively strong, it prevents crustal deformation from propagating into this terrane, and restricts the

208   bulk of deformation to the Mid-terrane. With continuous advance of the Pro-terrane and resistance of

209   the Retro-terrane, the crust of the Mid-terrane is intensively shortened, leading to more thrusting

210   structures in the upper crust and a "flower-like" structure in the lower crust (Figures 3b and 3e). Thrust

211   structures and crustal deformation also expand toward the Pro- and Retro-terranes at this stage.

212   Topography also grows towards the two bounding terranes (Figure 7a). Ultimately, the weak

213   lithospheric mantle of the Mid-terrane is squeezed out, and the Pro- and Retro-terrane's lithospheric

214   mantles meet and so start to collide beneath the overlying crust of the Mid-terrane (Figures 3c and 3f).

3.2. Case 2: Lithosphere Subduction

216   Case 2 shows lithosphere subduction of the Pro-terrane (Figure 4). In this model, the assumed

217   rheological models for the Pro-, Mid- and Retro-terrane are JS-II, JS-I, and JS-I, respectively. The

218   Mid-terrane has a stronger lower crust compared with the Pro-terrane, but its lithospheric mantle is a

219   little weaker than the Pro-terrane due to higher temperature field resulting from its thinner lithosphere

220   structure (Figure 4g). When convergence begins, the weak lower crust of the Pro-terrane is blocked by

221   the stronger lower crust of the Mid-terrane. This induces it to stack in a collisional front to form a

222   remarkable folding structure (Figures 4a and 4d). The strong lithospheric mantle of the Pro-terrane

223 continues to move forward and underthrusts beneath the Mid-terrane. As the Pro-terrane advances, its

224 crust gradually enters the Mid-terrane, inducing shortening and thickening of the upper crust of the

225 Mid-terrane, while the strong lower crust of the Mid-terrane almost keeps undeformed (Figures 4b and

226 4e).

227 Meanwhile, the lithospheric mantle of the Pro-terrane continues to underthrust scraping off part of the

228 lithospheric mantle of the Mid-terrane. Eventually, the crust of the Pro-terrane wedges a long distance

229 into the Mid-terrane, and the lithospheric mantle of the Pro-terrane subducts into the deeper mantle

230 (Figures 4d and 4f). In this example, crustal deformation and topography gradually propagate from the

231 Pro-terrane to the Mid-terrane, whereas the Retro-terrane remains nominally 'undeformed' at all times

232 (Figure 7b). In some experiments, the lithospheric mantle of the Retro-terrane can subduct beneath the

233 Mid-terrane (Figure S1).

234 **3.3. Case 3: Lithosphere Thickening and Delamination**

235 Case 3 illustrates the thickening and delamination of the lithospheric mantle of the Pro-terrane (Figure

236 5). In this case, the rheological models for the Pro-, Mid- and Retro-terranes are CB-II, JS-I, and JS-I,

237 respectively (Figure 5j). The Pro-terrane has a rheologically weaker lower crust and lithospheric mantle,

238 making it relatively easy to deform once the collision has started. The lithosphere of the Pro-terrane is

239 first thickened, and the crust starts to fold in two discrete zones (Figures 5a and 5f). The lower part of

240 the thickened lithospheric mantle is denser than its ambient mantle owing to lower temperature, which

241 causes it to drip downwards (Figures 5b–5h). After delamination of the thickened lithosphere,

242 subduction of the Pro-terrane's lithospheric mantle along one of the deformation localization zones

243 absorbs the plate convergence (Figures 5e and 5i). Crustal deformation is restricted in the Pro-terrane

244 until lithosphere delamination, after which crustal strain and topography rapidly spread from the

245 Pro-terrane to the Mid-terrane (Figure 7c). Like case 2, the Retro-terrane stays essentially undeformed

246 at all times.

247 **3.4. Case 4: Lithosphere Replacement**

248 Case 4 illustrates how the lithospheric mantle of the Pro-terrane is replaced by that of a neighboring

249 stronger Mid-terrane (Figure 6). In this case, the rheological models for the Pro-, Mid- and

250 Retro-terranes are CB-I, JS-II, and JS-I, respectively. The Pro-terrane has a strong lower crust and a

251 thick and weak lithospheric mantle, while the Mid-terrane has a weaker lower crust and a strong

252 lithospheric mantle (Figure 6g). This lithosphere configuration between the Pro- and the Mid-terrane

253 causes deformation to be primarily distributed in the Pro-terrane's lithospheric mantle and the

254 Mid-terrane's crust. As a result, the Mid-terrane's crust becomes intensely shortened by fold and thrust

255 structures, but its strong lithospheric mantle wedges into the Pro-terrane's thick and weak lithospheric

256 mantle (Figures 6a, 6b, 6d and 6e). The strong lithospheric mantle of the Mid-terrane scrapes off the

257 lower part of the weak lithospheric mantle of the Pro-terrane and so replaces it (Figures 6c and 6f).

258 Similar to case 1, crustal deformation and topography expand from the Mid-terrane towards its side

259 terranes (Figure 7d). The lithospheric mantle of the Retro-terrane can also be replaced in some cases

260 (e.g. Figure S1).

261 **4. Discussion**

262 **4.1. Rheological Characteristics for Distinct Lithosphere Deformation Patterns**

263 Distinct lithosphere deformation patterns in our simulations arise from rheological contrasts between

264 neighboring continental terranes. Figure 8 summarizes the rheological characteristics of these distinct

265 deformation patterns. When the Mid-terrane's lithospheric mantle is weakest (typified by models in

266 which the rheological model of the Mid-terrane is CB-II), it is easy for its mantle to be extruded out,

267 leading to collision between the lithospheric mantles of its surrounding Pro- and Retro-terranes. When

268 one of the two bounding terranes has extremely weak lithospheric mantle, its lithosphere is first to be

269 thickened by compression, and delamination may follow due to density-driven instability. When the

270 lower crust of the Mid-terrane is relatively strong (CB-I or JS-I), while the lower crust is weaker in the

271 Pro- or Retro-terrane (CB-II or JS-II), the lithospheric mantle of the Pro- or Retro-terrane will tend to

272 subduct into the deep mantle, e.g. leading to intracontinental subduction. Finally, when the Mid-terrane

273 has a weak lower crust and strong lithospheric mantle (JS-II), while the Pro- or Retro-terrane has a

274 strong lower crust and weak lithospheric mantle (CB-I), the lithospheric mantle of the former may

275 replace the lithospheric mantle of the latter.

276 When the deformation patterns involve the collision and replacement of lithosphere, continental

277 deformation involves all three terranes (Figures 3 and 6). In contrast, the other deformation patterns

278 only involve two terranes, the Pro- or Retro-terrane and the Mid-terrane (Figures 4 and 5). The

279   rheological properties of the Mid-terrane are responsible for these differences. Like previous numerical

280   studies (Kelly et al., 2016, 2020; Li et al., 2016; Huangfu et al., 2018, 2022; Sun and Liu, 2018), our

281   simulations show that a weak Mid-terrane is easier to deform, and that in this case lithosphere

282   deformation will expand from center to its neighboring sides; while a relatively strong Mid-terrane

283   prevents deformation from propagating far, so that lithosphere deformation is constrained to occur

284   within two terranes..

285   Although our multi-terrane numerical models mainly focus on the impact of the lateral strength

286   differences between different terranes in a continental collisional system, rheological models of CB-I,

287   CB-II, JS-I, JS-II also involve vertical rheological variation (Figure 1a). It seems difficult to summarize

288   how vertical strength variation affects lithosphere deformation of the continental collisional system.

289   For example, in some cases, only changing the rheological models of the Pro- or Retro-terranes may

290   produce distinct deformation modes such as collision, subduction, thickening and delamination, and

291   displacement (e.g., the first and third rows, the third column in the upper left panel in Figure 8 and the

292   third column in the lower right panel in Figure 8). However, changing the rheological models of the

293   Pro- or Retro-terranes seems to have less impact on the deformation mode of the continental collisional

294   system, according to the simulation results of models which are connected by several cross-shaped

295   solid lines with different colors in Figure 8. Thus, it is difficult to determine whether the horizontal

296   strength contrasts between terranes or the vertical strength variation of a single terrane plays the

297   dominant role in a multi-terrane collisional system. This is also the significance and necessity of our

298   study.

299   **4.2. Influences of Lithosphere Structure**

300   Lithospheric thickness is one of the critical factors that control its strength (Burov, 2011), and it can

301   strongly vary between tectonic regions (Pasyanos et al., 2014). In our models, we assume different

302   lithospheric thicknesses for the Pro-, Mid- and Retro-terrane to explore these effects. Complex effects

303   are seen. When changing the lithospheric thicknesses of the Mid-terrane, or of all three terranes,

304   remarkable variations in lithosphere deformation appear in cases 1 and 2, but smaller variations are

305   seen for cases 3 and 4 (Figure 9). Cases 1 and 2 assume a Jelly sandwich rheology for the Pro-, Mid-

306   and Retro-terrane, so the strength of the lithospheric mantle of three terranes is comparable. Strength

307   variations produced by differences in lithospheric thickness may alter the relative strength of the three

308 terranes, resulting in distinct lithospheric deformations. For example, if the Pro- and Mid-terranes have

309 same lithosphere thickness, deformation mode in Case 2 would change from subduction to thickening

310 (subplot 3 vs. subplot 8 in Figure 9); if the Mid-terrane is thickest or the three terranes have same

311 thickness of lithosphere, deformation mode in Case 1 would change from collision to replacement

312 (subplot 2 vs. subplot 17 and 22 in Figure 9), and the polarity of the subduction of Pro-terrane's

313 lithospheric mantle would be reversed in Case 2 (subplot 3 vs. subplot 18 and 23 in Figure 9). Instead,

314 in cases 3 and 4, the Pro-, Mid- and Retro-terrane have two regions with stronger Jelly-sandwich-like

315 rheological structures and one with a weaker Crème brûlée structure, and deformation preferentially

316 concentrates in the weaker terrane. In comparison to the large strength difference implied for the

317 lithospheric mantle between the Crème brûlée and Jelly Sandwich rheological models, the strength

318 variations associated with the differences in lithosphere thickness are relatively small. Therefore,

319 changing the thicknesses of the lithosphere has much smaller effects on the lithosphere deformation, as

320 seen in cases 3 and 4 (also see the subplots in 3rd and 4rd rows of Figure 9).

321 In addition, the weak zones that suture two terranes are generally preserved during continental

322 amalgamation (Burker et al., 1977; Vink et al., 1984; Yin and Harrison, 2000). These local pre-existing

323 lithosphere weaknesses would be preferentially activated if the continental lithosphere were subjected

324 to compression, and could play a key role in concentrating deformation, adjusting deformation

325 sequences, and inducing lithosphere subduction (Sokoutis and Willingshofer, 2011; Heron et al., 2016;

326 Chen et al., 2020; Xie et al., 2021, 2023). Comparing the simulation results of models with and without

327 weak zone, we find that a weak zone will facilitate lithosphere subduction in earlier stages of model

328 evolution, resulting in more diverse lithosphere deformation patterns during the later stage (Figure 10).

329 **4.3. Implications for the Tectonics of East Asia**

330 **4.3.1. Lithosphere Collision beneath the Eastern Tien Shan**

331 The eastern Tien Shan is an ideal region to study the deformation patterns linked to long-term

332 lithosphere collision (Figure 11a). The eastern Tien Shan is bounded by the Tarim Basin to the south,

333 and the Junggar Basin to the north. It is composed of a series of former island arcs and small

334 continental blocks that amalgamated during the late Paleozoic (Han and Zhao, 2017). The lithosphere

335 of the eastern Tien Shan is weaker and thinner in comparison to its neighboring Tarim Basin and

336 Junggar Basin (Kumar et al., 2005; Lei and Zhao et al., 2007; Zhang et al., 2013; Deng and Tesauro,

2016). At ~20 – 25 Ma, the eastern Tien Shan became a reactivated orogeny in response to ongoing India-Asia collision (Yin et al., 1998). Compression linked to the India-Asia collision induced the Tarim lithosphere to underthrust northward (Xu et al., 2002; Guo et al., 2006; Lei and Zhao et al., 2007; Lü et al., 2019; Hapaer et al., 2022; Sun et al., 2022). In the northern part of the eastern Tien Shan, significant high-velocity anomalies and Moho overlap are also imaged, which are conventionally explained as being due to the southward underthrusting of the Junggar lithosphere (Xu et al., 2002; Guo et al., 2006; Li et al., 2016; Lü et al., 2019). High-velocity anomalies in the Tarim and Junggar lithosphere appear to connect beneath the eastern Tien Shan, suggesting the lithosphere of the Tarim and Junggar Basins has converged and collided together in this region (Figure 11b; Lü et al., 2019). Bidirectional underthrusting of the Tarim and Junggar lithosphere leads to intense crustal shortening and thrust faults on both flanks over the adjacent basins, as well as attendant fold and reverse fault zones along the range fronts (Yin et al., 1998; Wang et al., 2004).

4.3.2. Lithosphere Thickening and Delamination in the Tibetan Plateau

The deformation pattern arising from lithosphere thickening and delamination has been applied to the Tibetan Plateau (Figure 11c). Tibetan lithosphere may have been significantly weakened by hydration, metasomatism, and partial melting of the lithospheric mantle during a series of oceanic closure and terrane accretion events before the India-Asia collision (Yin and Harrison, 2000; Zhang et al., 2014; Ma et al., 2021). It was then pushed northward by the Indian craton and was blocked by the Tarim/Qaidam craton during India-Asia collision, leading to double crustal thickness (Zhao and Morgan, 1985; Zhang et al., 2011). The lithosphere beneath the Tibetan Plateau does not thicken significantly like its crust, especially beneath northern Tibet (Owens and Zandt, 1997; Tunini et al., 2016). Numerous observations instead suggest that the Tibetan lithosphere has been detached from the crust and has sunk into deeper mantle, consistent with the presence of high-velocity regions in the deep mantle in western, southern and southeastern Tibet (Li et al., 2008; Chen et al., 2017; Feng et al., 2021). A significant depression of the 660-km discontinuity beneath the Himalaya terrane and the uplift of 410-km discontinuity in western Tibet have also attributed to the presence of delaminated Tibetan lithosphere (Wu et al., 2022). In northern Tibet, anomalously high temperatures are assumed to be linked to a region of inefficient S_n propagation indicating a thin or absent lithospheric mantle lid in this region, while a remarkable low-velocity zone in the mantle and ultra-potassic volcanics also suggest

lithosphere thinning (Barazangi and Ni, 1982; Turner et al., 1996; Owens and Zandt, 1997; Guo et al., 2006; Liang et al., 2012; Tunini et al., 2016). After lithosphere thinning commenced in the Miocene, the Tibetan Plateau rapidly grew outwards (Lu et al., 2018 and references therein; Molnar et al., 1993; Xie et al., 2023).

4.3.3. Lithosphere Subduction in Southeastern China

An example of intracontinental subduction is the Early Paleozoic Orogen in Southeastern China which appears to have not been preceded by oceanic subduction (Figure 11d; Faure et al., 2009). The northeasterly trending Early Paleozoic Orogen of Southeastern China is located on the Wuyi-Yunkai Fold Belt which welds the Cathaysian Block to the south and the Yangtze Block to the north. Two groups of models of collisional belt (Guo et al., 1989; Hsü et al., 1990) and intercontinental orogen (Faure et al., 2009; Charvet et al., 2010; Li et al., 2010) have been proposed to explain the Early Paleozoic Orogen in Southeastern China. Arguments against it being a collisional orogenic belt are its lack of preserved ophiolites, a magmatic arc, subduction complexes, and high-pressure metamorphism; instead, structural, metamorphic, and sedimentary elements indicate that this orogen was an intracontinental orogen controlled by the northward subduction of Cathaysian Block (Faure et al., 2009; Charvet et al., 2010; Li et al., 2010). At ~465 Ma, the Cathaysian Block underthrust beneath the Yangtze Block along the Jiangshan–Shaoxing Fault, in which the lithospheric weaknesses inherited from previous tectonic event of Nanhua rift at 800 – 850 Ma played an important role (Faure et al., 2009; Charvet et al., 2010). During continental subduction, north-south horizontal shortening is accommodated by ductile decollement zones in the Cathaysian Block, causeing remarkable south-directed crustal folding and thrusting structures related to both thin-skinned and thick-skinned tectonics in the Wuyishan proper and its southern border, and north-directed structures to the west of Ganjiang Fault and north of Jiangshan–Shaoxing Fault, where only thin-skinned tectonics is visible (Li et al., 1998; Shu et al., 1999; Charvet et al., 2010). The tectonic development of the Early Paleozoic Orogen in Southeastern China appears similar to the deformation mode of lithosphere subduction (Figure 4, 10a and 10c).

So far, we have yet to find a suitable region to apply the model deformation pattern of lithosphere replacement. In this deformation pattern, crustal deformation and topographic evolution are similar to those in the deformation pattern of lithosphere collision (Figures 7a and 7c). Thus, it is not easy to

395 identify this pattern by geological and geophysical techniques when the replaced and original

396 continental lithosphere has similar properties. Improved imaging observations with better resolution

397 may allow this deformation pattern to be identified in the future.

5. Model Limitations

399 Although we can obtain four deformation modes of continental lithosphere by changing the rheologies

400 of different terranes in a collisional model, we must keep in mind that our results are based on some

401 simplifications and assumptions, which may affect the model results. For example, in our model three

402 terranes are directly collaged together, but in nature different terranes are often connected through weak

403 sutures which may preferentially deform when they are subjected to compression (Burker et al., 1977;

404 Yin and Harrison, 2000). These local pre-existing weak zones have non-negligible influences on

405 lithospheric deformation, and their role were widely discussed in previous studies (Sokoutis and

406 Willingshofer, 2011; Heron et al., 2016; Chen et al., 2020 ; Xie et al., 2021, 2023). We also discussed

407 the effects of local pre-existing weak zones in Section 4.2. In addition, lithosphere thicknesses of the

408 Pro-, Mid- and Retro-terranes are chosen arbitrary in our models, but they also have important

409 influences on lithosphere deformation (Figure 9). Some studies suggest that differences in crustal

410 strength will also cause different lithospheric deformation (Faccenda et al., 2008; Vogt et al., 2017,

411 2018), but the three terranes are set same crustal structure in our model for the aim of simplification. As

412 well, some studies believe that the convergence rate will greatly affect the deformation of orogenic

413 belts (Chen et al., 2016; Vogt et al., 2017), but in this study, the impact of the convergence rate is not

414 been discussed.

6. Conclusions

416 The continental lithosphere is likely to have strong lateral variations in its strength. We explored 2-D

417 numerical models that contain three diverse types of continental terranes to study the responses of

418 continental terranes with different strengths to compression. Four rheological models were respectively

419 applied to each of the Pro-, Mid- and Retro-terranes, and simulation results can be grouped into four

420 distinct deformation styles: lithosphere collision, subduction, thickening and delamination, and

421 replacement. These deformation styles arise from the rheological contrasts between the terranes: (1)

422 when the middle terrane is the weakest, its lithosphere is easily extruded, which leads to lithosphere

423 collision between its two bounding terranes; (2) when the middle terrane has a strong lower crust, while

424 the lower crust of a bounding terrane is weak, then subduction of the lithosphere of the bounding

425 terrane will occur; (3) when a bounding terrane is the weakest, its lithosphere would tend to be

426 thickened by lateral compression, followed by lithosphere delamination due to the resulting

427 density/gravitational instability; (4) when a bounding terrane has a strong lower crust and weak

428 lithospheric mantle, while the middle terrane has a weak lower crust and strong lithospheric mantle,

429 then lithosphere replacement will occur. These simulation patterns are seen in observed deformation

430 patterns and structures in the eastern Tien Shan, and the Tibetan Plateau, the Early Paleozoic Orogen of

431 Southeastern China, suggesting they are likely to be naturally occurring modes of intracontinental

432 orogenesis.

433 *Code availability*

434 Requests for the numerical code I2VIS should be sent to the main developer

435 (taras.gerya@erdw.ethz.ch).

436 *Data availability*

437 Numerical modeling data and the model evolution animations of Cases 1 – 4 are all provided in

438 Zenodo (https://doi.org/10.5281/zenodo.8354366 and https://doi.org/10.5281/zenodo.10731981).

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Figures and captions
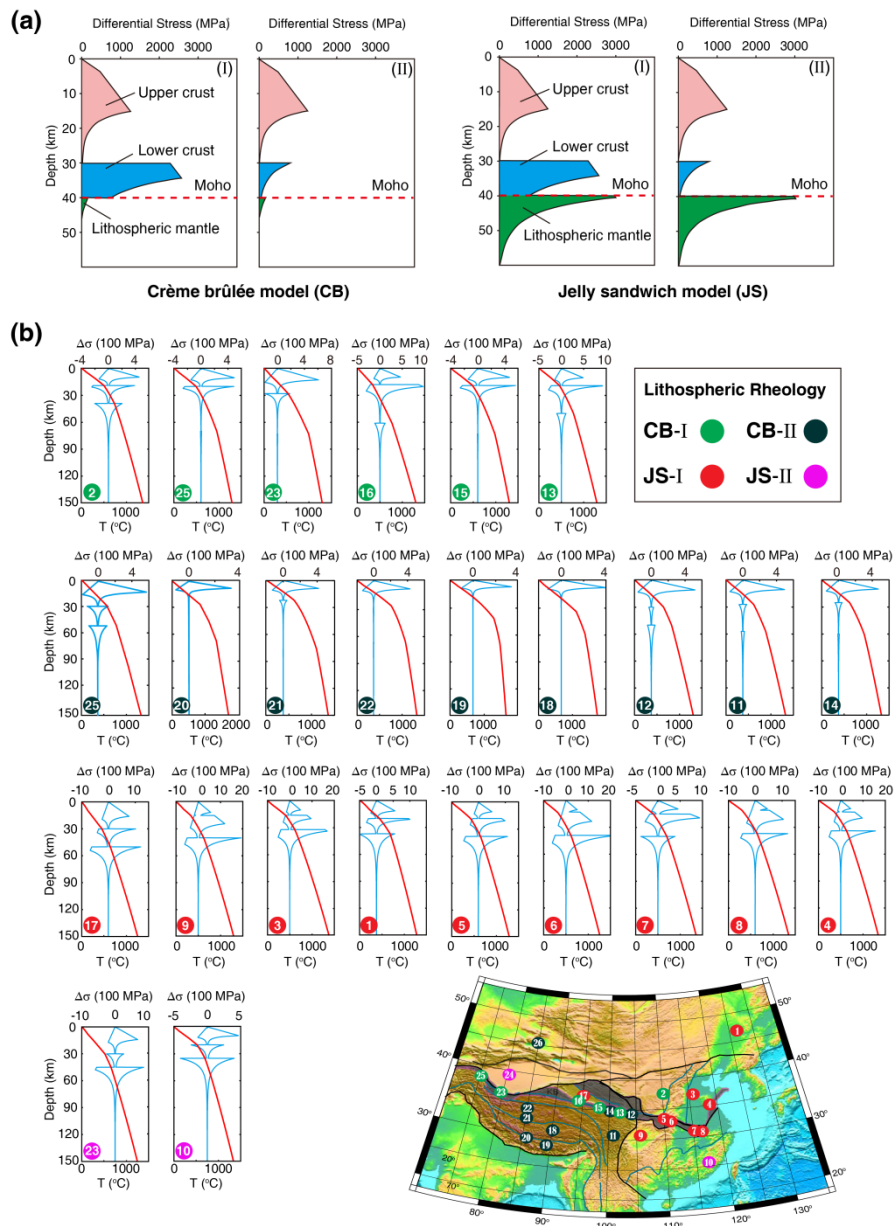
Figure 1. Four rheological models of continental lithosphere. (a) Crème brûlée model (CB) and Jelly sandwich model (JS). The two rheological models can be further subdivided into CB-I, CB-II, JS-I, and JS-II according to the strength of the lower crust (modified from Jackson, 2002). (b) Observations of four distinct lithosphere rheological structures implied for East Asia (modified from Zhang et al., 2013). Locations of strength profiles are pointed out by dots with numbers in the topography map. Dots filled with different colors indicate different models of lithospheric rheology. These strength profiles are calculated based on observed geothermal structure and lithospheric structure, and assumed that compositions of the upper and lower crust and lithospheric mantle are wet quartzite, undried granulite and dry olivine, respectively. Variations of temperature and lithospheric compositions

lead to a diverse suite of strength profiles vs. depth.
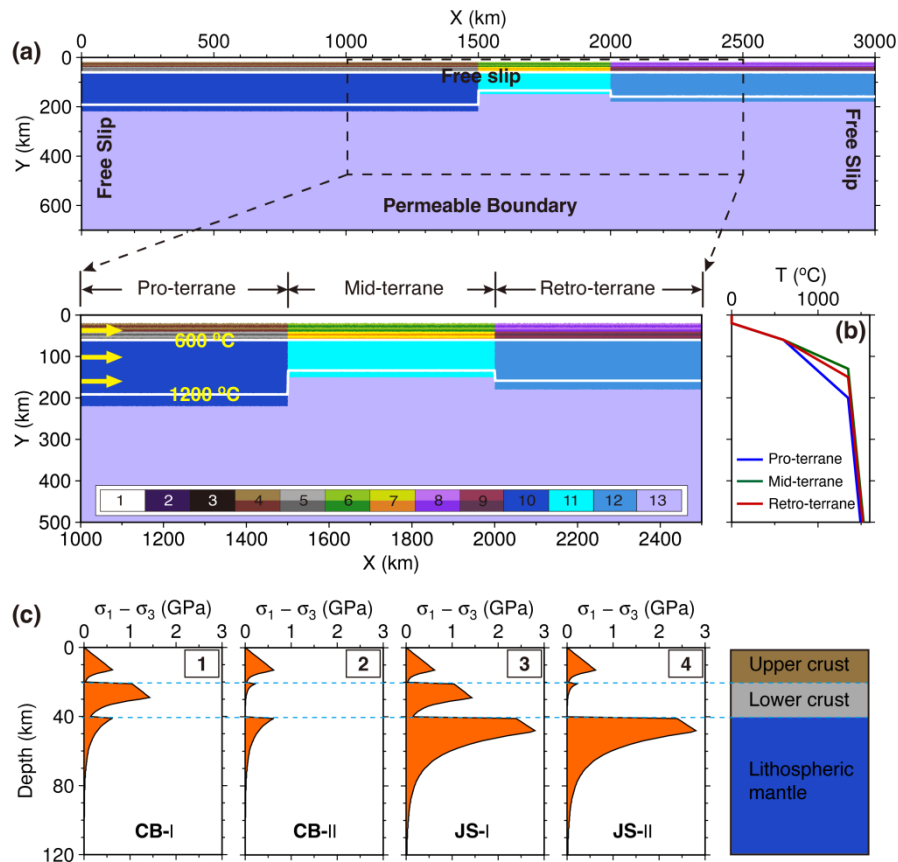
Figure 2. Initial model setup. (a) Initial model configuration. The model size is 3000 km $\times$ 700 km, and size of study region is 1500 km $\times$ 500 km. Three continental terranes of the Pro-, Mid- and Retro-terranes are contained in the numerical model, and they are 200 km, 130 km, and 160 km thick, respectively. White lines are isotherms with an interval of 600 °C. Yellow arrows indicate the convergence rate of 20 mm/yr. Colored grids: 1 – sticky air; 2 – sediments; 3 – weak zone; 4, 6, 8 – the upper crust of the Pro-, Mid- and Retro-terranes, respectively; 5, 7, 9 – the lower crust of the Pro-, Mid- and Retro-terranes, respectively; 10, 11, 12 –lithospheric mantle of the Pro-, Mid- and Retro-terranes, respectively; 13 – asthenosphere. (b) Initial temperature structure for the three terranes. The Pro- and Mid-terranes respectively have a coldest and warmest lithospheric mantle due to their differences of lithosphere thicknesses. (c) Four rheological models with contrasting lithospheric strength profiles. These are derived from different strength scaling factor (*S*) combinations for the upper crust, lower crust, and lithospheric mantle (Table S1). Strength profiles are calculated based on the Pro-terrane's initial lithospheric structure, composition, and temperature field. The prescribed strain rate is 1×10^{-14} s^{-1}. CB-I and CB-II, the crème brûlée model with strong and weak lower crust, respectively; JS-I and JS-II, the jelly sandwich model with strong and weak lower crust, respectively.

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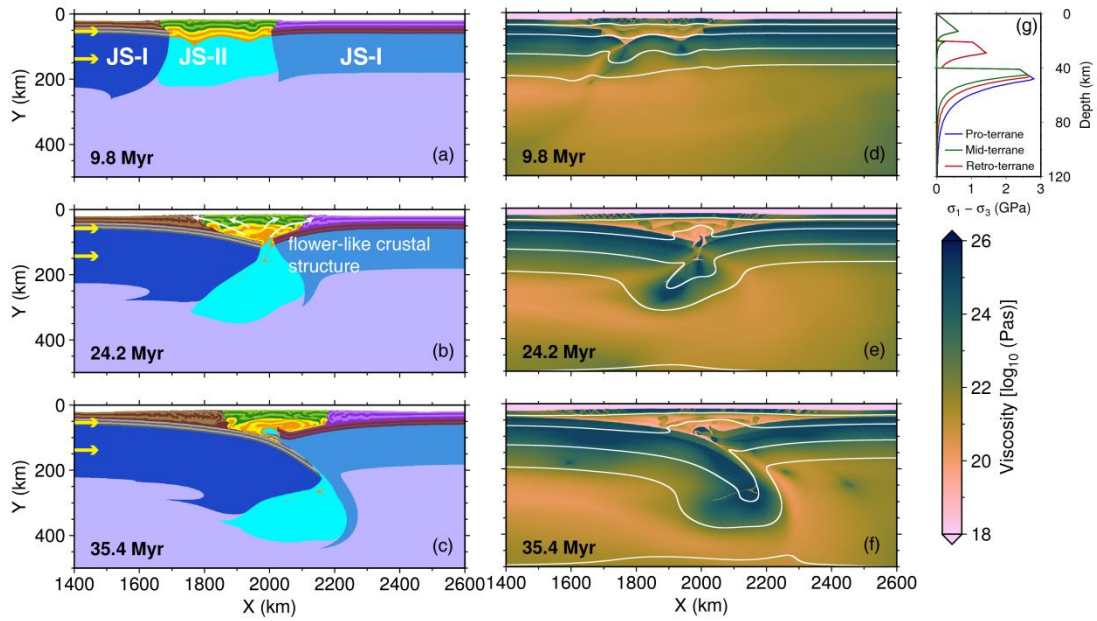
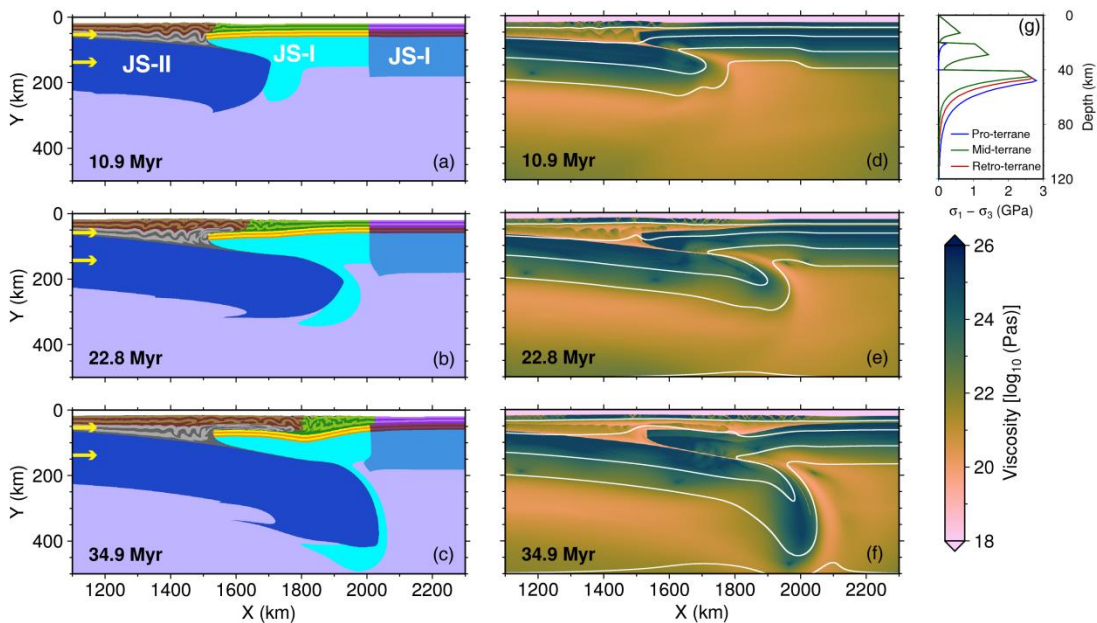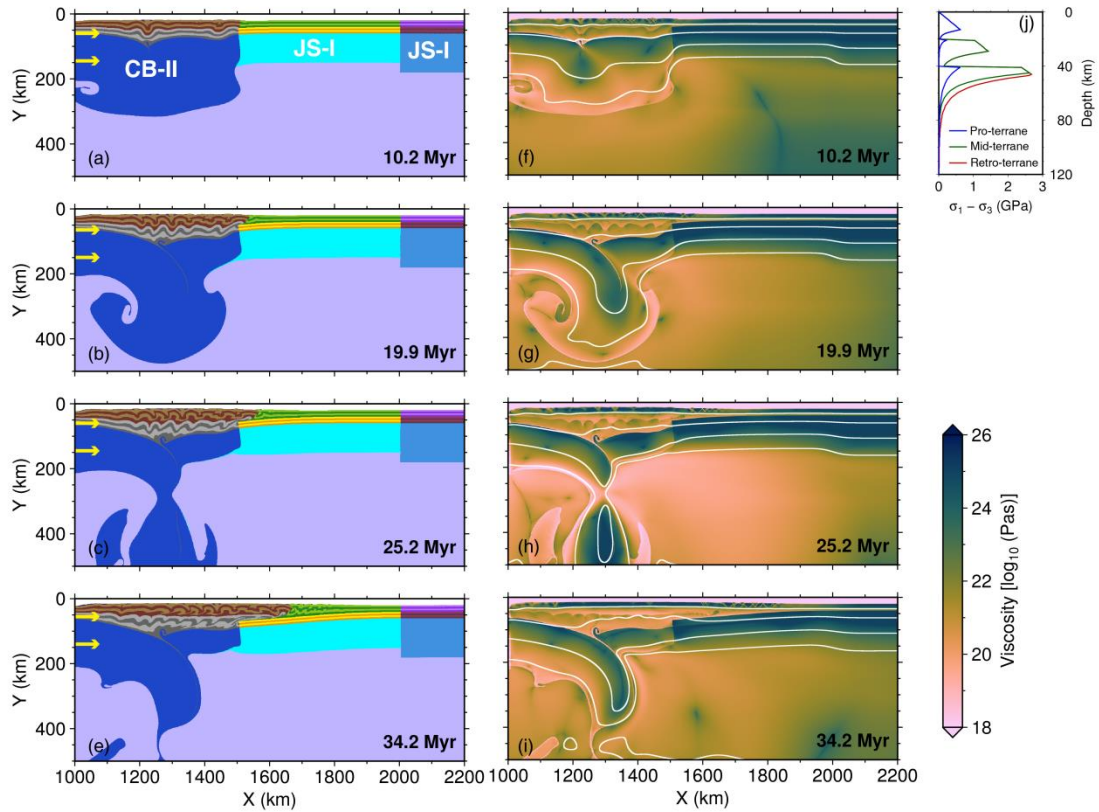

Figure 3. Collision of the lithospheres of the Pro- and Retro-terranes. Rheological models for the Pro-, Mid- and Retro-terranes are JS-I, JS-II, and JS-I, respectively, as shown in (g). The left panel shows compositional fields at 9.8 Myr, 24.2 Myr, and 35.4 Myr, respectively. Yellow arrows indicate the convergence rate. The right panel shows the corresponding viscosities. White lines are isotherms with an interval of 300 °C.

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Figure 4. Subduction of the lithosphere of the Pro-terrane. Rheological models for the Pro-, Mid- and Retro-terranes are JS-II, JS-I, and JS-I, respectively. See Figure 3 for plotting conventions.

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Figure 5. Thickening and delamination of the lithosphere of the Pro-terrane. Rheological models for the Pro-,

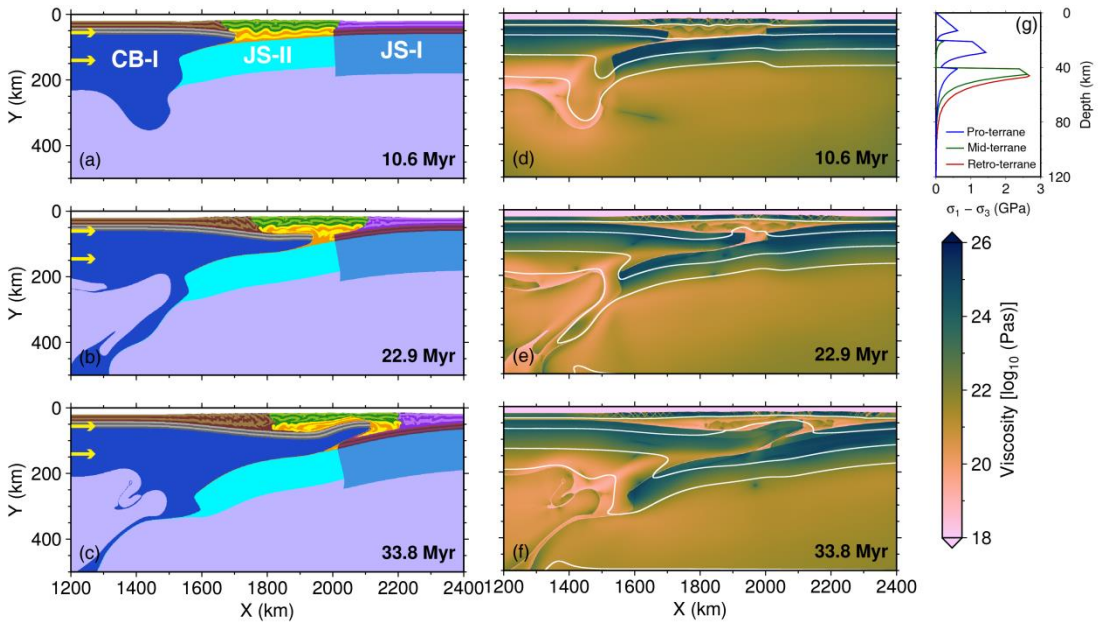Mid- and Retro-terranes are CB-II, JS-I, and JS-I, respectively. See Figure 3 for plotting conventions.

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Figure 6. Replacement of the lithosphere of the Pro-terrane. Rheological models for the Pro-, Mid- and

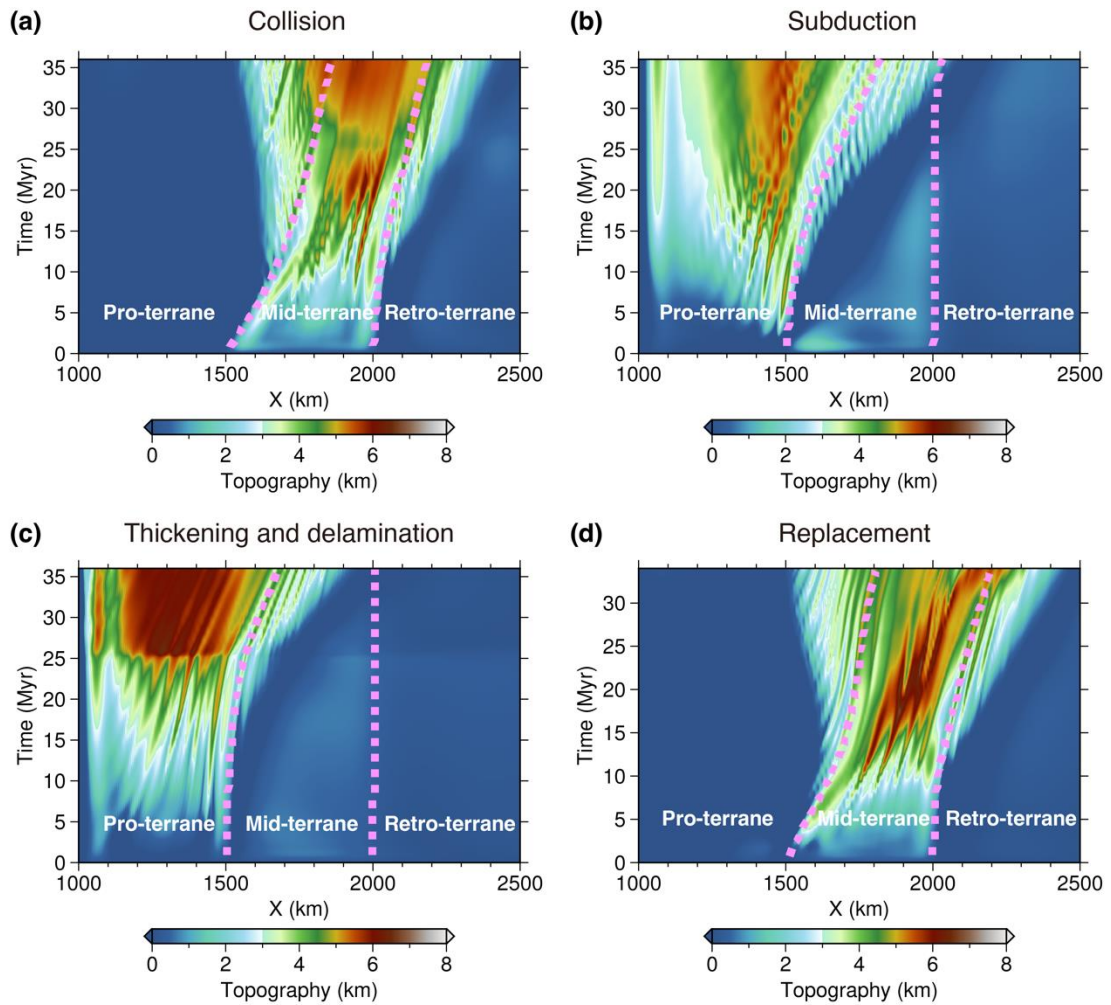Retro-terranes are CB-I, JS-II, and JS-I, respectively. See Figure 3 for plotting conventions.

Figure 7. Evolution of surface relief for the different deformation styles. The purple dashed lines indicate the boundaries between terranes. (a) – (d) Surface relief associated with the deformation patterns of lithosphere collision, subduction, thickening and delamination, and replacement, respectively.
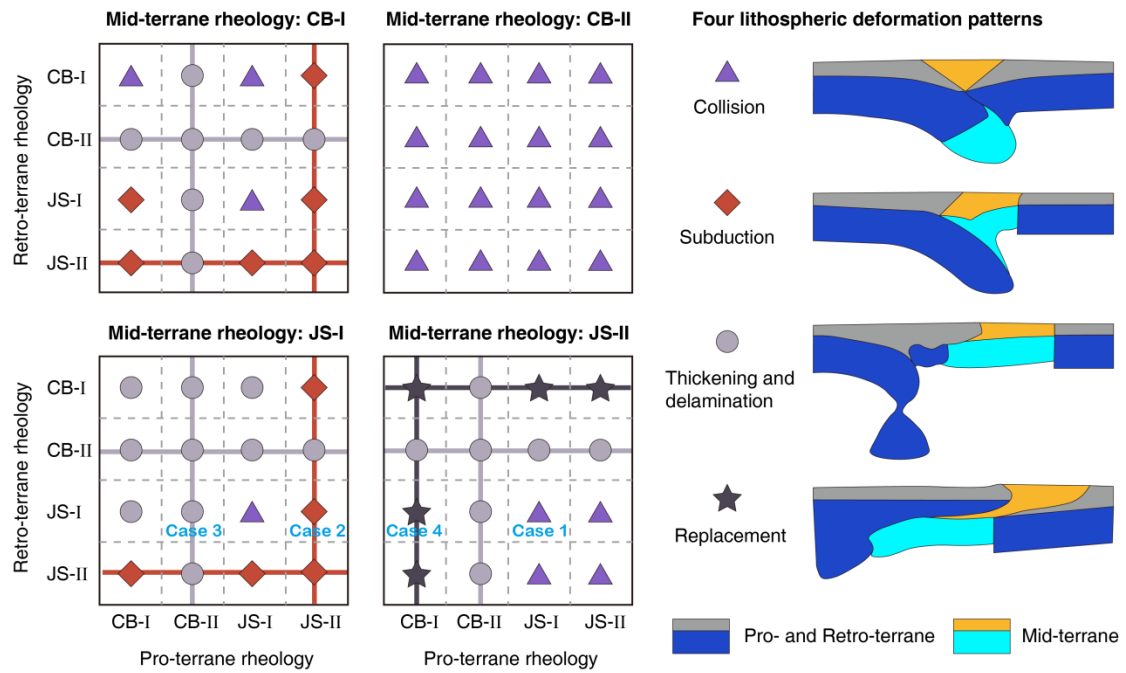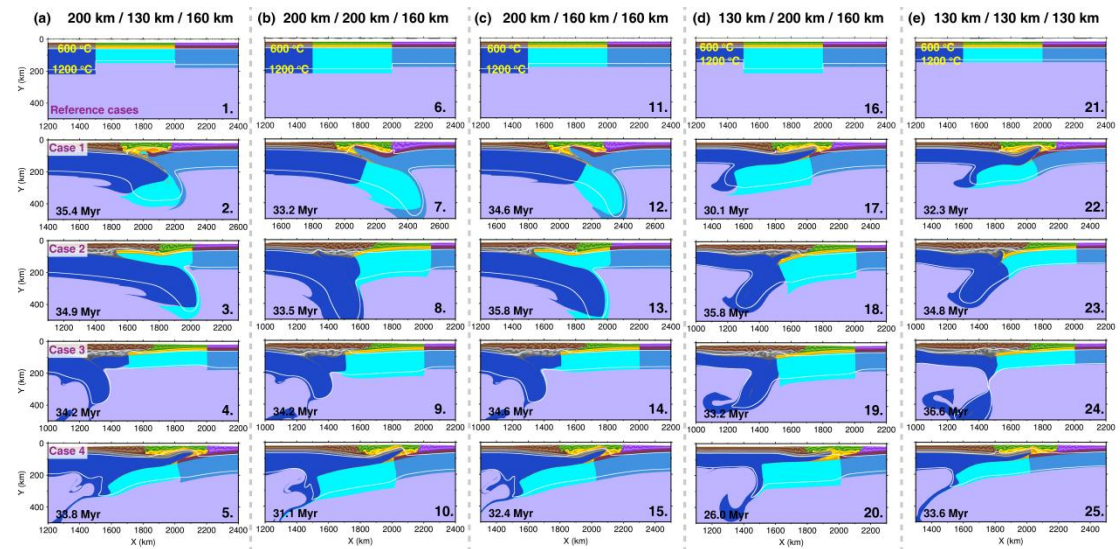
Figure 8. Four styles of lithosphere deformation patterns. Symbols with colors indicate different deformation patterns of the lithosphere. Cases 1 – 4 are the selected models chosen to illustrate details of these modes of compressional evolution.

Figure 9. Effects of lithosphere thicknesses of various terranes. (a) – (e) Final simulation results of models with the different lithosphere thicknesses of the Pro-, Mid-, and Retro-terranes, respectively. (a) Final simulation results of reference cases. Rheological models of the Pro-, Mid- and Retro-terranes in 2 – 4 rows are same with those in Cases 1 – 4, respectively.
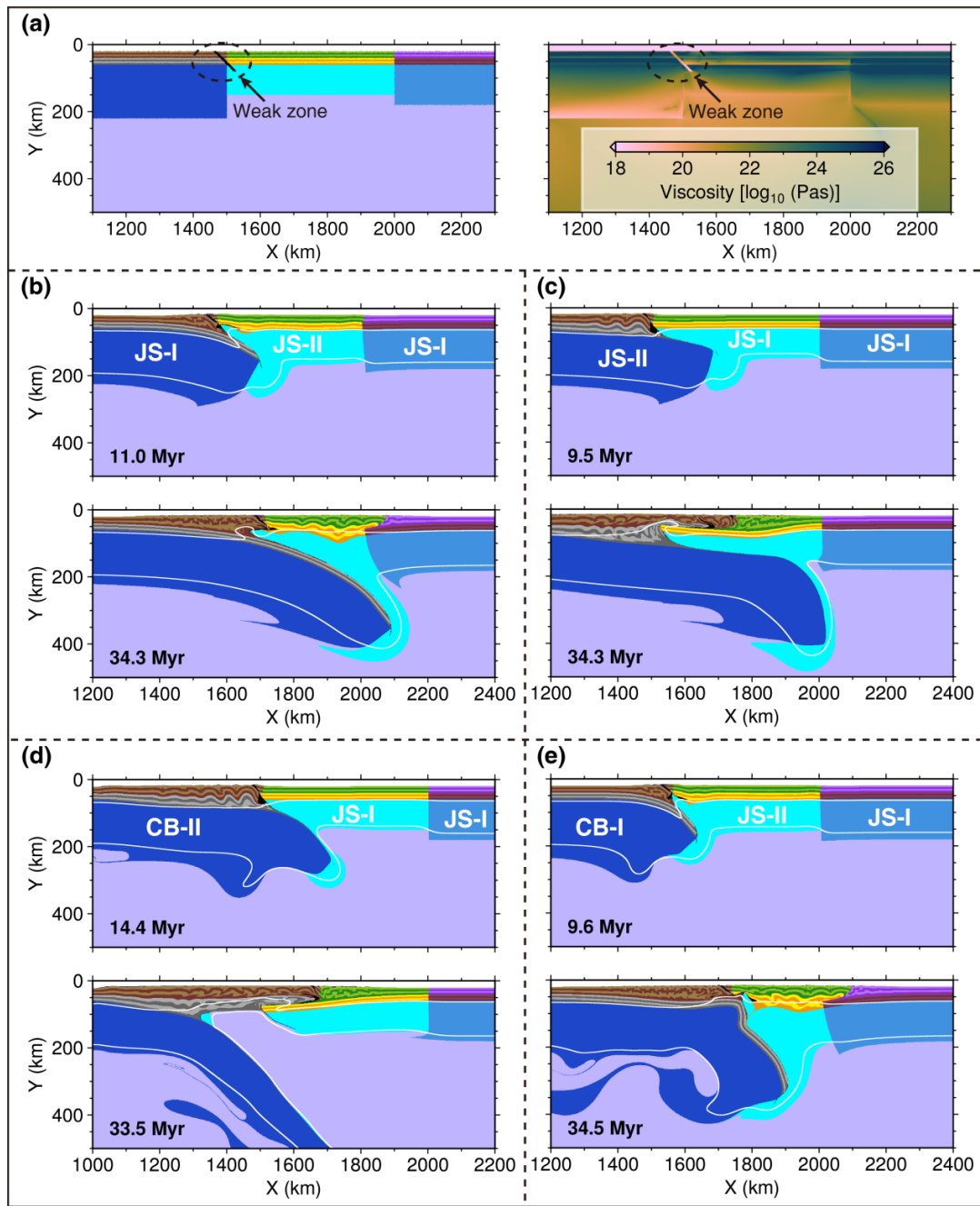
Figure 10. Effects of local weak zone on lithosphere deformation. (a) Details about the weak zone. (b) – (e) Final simulation results of models corresponding to Cases 1 – 4, respectively.

(a) Eastern Asia

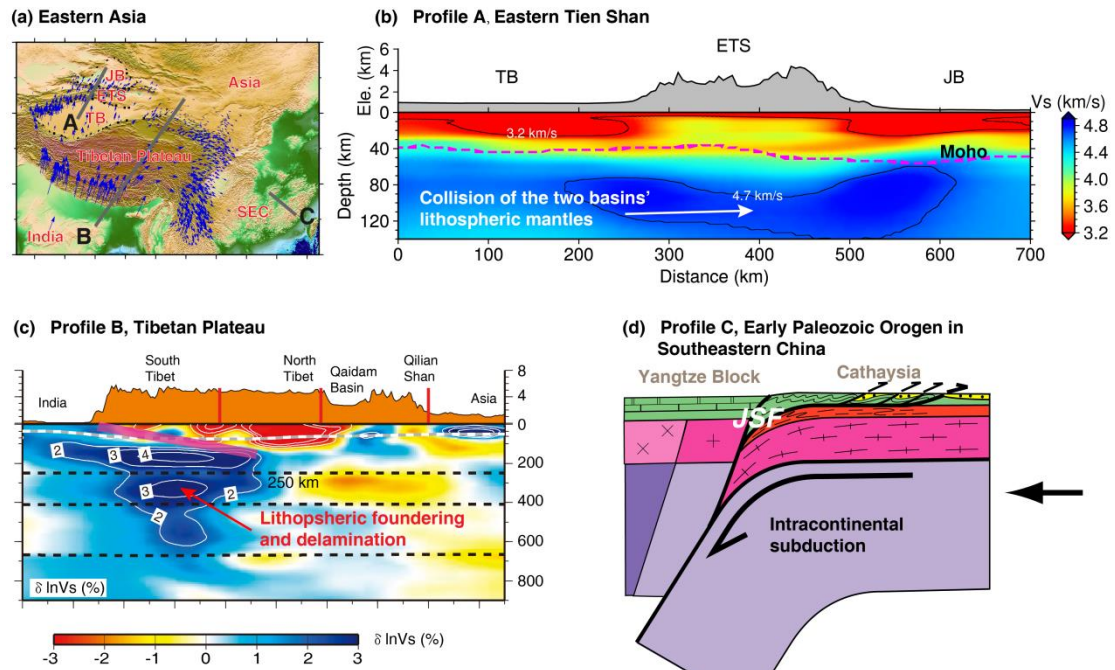**(b) Profile A, Eastern Tien Shan**

ETS

TB

JB

3.2 km/s

Moho

Collision of the two basins' lithospheric mantles

4.7 km/s

Vs (km/s)

(c) Profile B, Tibetan Plateau

South Tibet

North Tibet

Qaidam Basin

Qilian Shan

India

Asia

250 km

Lithopsheric foundering and delamination

δ lnVs (%)

δ lnVs (%)

(d) Profile C, Early Paleozoic Orogen in Southeastern China

Yangtze Block

Cathaysia

JSF

Intracontinental subduction

Figure 11. Implications of simulation results to East Asia. (a) Topography of East Asia. The three gray lines point out the locations of lithosphere profiles in (b), (c) and (d). TB, Tarim Basin; ETS, eastern Tien Shan; JB, Junggar Basin; SEC, Southeastern China. (b) Collision of the lithospheric mantle of Tarim Basin and Junggar Basin beneath the eastern Tien Shan (modified from Lü et al., 2019). (c) Lithospheric founding and delamination in the Tibetan Plateau (modified from Chen et al., 2017). (d) Intracontinental subduction in the Early Paleozoic Orogen in southeastern China (modified from Faure et al., 2009). JSF, Jiangshan–Shaoxing Fault.

763   **Table 1. Flow laws and material properties for different lithospheric layers.** ρ_0 is the initial density; it evolves

764   with time as $\rho = \rho_0\left(1-\alpha\left(T-T_0\right)\right)\left(1+\beta\left(P-P_0\right)\right)$, where $T_0 = 20\,°C$, $P_0 = 10^5$ MPa. Flow law: qtz. = quartzite, Plag.

765   = plagioclase, ol. = olivine.

Material properties	Sediment	Upper crust	Lower crust	Lithospheric mantle	Asthenosphere
ρ_0 (kg/m^3)	2600	2700	2800	3300	3300
Flow laws	Wet qtz.	Wet qtz.	Plag.	Dry ol.	Dry ol.
$1/A_D$ (Pa^n s)	1.97×10^{17}	1.97×10^{17}	4.80×10^{22}	3.98×10^{16}	3.98×10^{16}
n	2.3	2.3	3.2	3.5	3.5
E_a (KJ/mol)	154	154	238	532	532
V_a (J/bar)	0.8	0.8	1.2	1.2	1.2
$\phi = sin(\varphi)$	0.2 – 0.1	0.3 – 0.1	0.3 – 0.1	0.6 – 0.4	0.6 – 0.3
C (Pa)	$1\times10^{7\text{-}6}$	$1\times10^{7\text{-}6}$	$1\times10^{7\text{-}6}$	$1\times10^{7\text{-}6}$	$1\times10^{7\text{-}6}$
H_r (uW/m^3)	2.0	1.5	0.5	0.022	0.022
C_p (J/kg K)	1000	1000	1000	1000	1000
α (1/K)	3×10^{-5}	3×10^{-5}	3×10^{-5}	3×10^{-5}	3×10^{-5}
β (1/MPa)	1×10^{-5}	1×10^{-5}	1×10^{-5}	1×10^{-5}	1×10^{-5}
k (W/m/K)	$0.64+807/(T+77)$	$0.64+807/(T+77)$	$1.18+474/(T+77)$	$[0.73+1293/(T+77)]\times(1+0.00004P)$	$[0.73+1293/(T+77)]\times(1+0.00004)$

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