

1 Subduction of trench-fill sediments beneath an
2 accretionary wedge: insights from sandbox
3 analogue experiments

4 Atsushi Noda^{*1}, Hiroaki Koge^{†2}, Yasuhiro Yamada³, Ayumu
5 Miyakawa¹, and Juichiro Ashi²

6 ¹Geological Survey of Japan, National Institute of Advanced Industrial Science and Technology
7 (AIST), Central 7, Higashi 1-1-1, Tsukuba, Ibaraki 305-8567, Japan

8 ²Atmosphere and Ocean Research Institute, The University of Tokyo, 5-1-5 Kashiwanoha,
9 Kashiwa-shi, Chiba 277-8564, Japan

10 ³Japan Agency for Marine-Earth Science and Technology (JAMSTEC), 3173-25 Showa-machi
11 Kanazawa, Yokohama 236-0001, Japan

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13 **ABSTRACT**

14 Ancient exhumed accretionary complexes are sometimes associated with high-pressure–
15 low-temperature (HP–LT) metamorphic rocks, such as psammitic schists, which are
16 derived from sandy trench-fill sediment. At accretionary margins, sandy trench-fill
17 sediments are rarely subducted to the depth of HP metamorphism because they are
18 commonly scraped off at the frontal wedge. This study uses sandbox analogue ex-
19 periments to investigate the role of seafloor topography in the transport of trench-fill

*E-mail: a.noda@aist.go.jp

†Current address: Geological Survey of Japan, National Institute of Advanced Industrial Science and Technology (AIST), Central 7, Higashi 1-1-1, Tsukuba, Ibaraki 305-8567, Japan

20 sediment to depth during subduction. The experiments were conducted with a detached,
21 rigid backstop to allow a topographic high (representing a seamount) to be subducted
22 through a subduction channel. In experiments without topographic relief, progressive
23 thickening of the accretionary wedge pushed the backstop down, leading to a step-
24 ping down of the décollement, narrowing the subduction channel, and underplating
25 the wedge with subducting sediment. In contrast, in experiments with a topographic
26 high, the subduction of the topographic high raised the backstop, leading to a stepping
27 up of the décollement and widening of the subduction channel. These results suggest
28 that the subduction of topographic relief is a possible mechanism for the transport of
29 trench-fill sediment from the trench to HP environments through a subduction channel.
30 A sufficient supply of sediment to the trench and topographic relief on the subducting
31 oceanic plate might enable trench-fill sediment to be accreted at various depths and
32 deeply subducted to become the protoliths of HP–LT metamorphic rocks.

33 INTRODUCTION

34 High-pressure–low-temperature (HP–LT) metamorphic rocks derived from terrigenous
35 sedimentary rocks are known to occur at subduction margins. Such metamorphic rocks
36 are exposed alongside low-grade accretionary rocks and fore-arc basin strata that include
37 coarse-grained sandy deposits with the same depositional ages as the metamorphic
38 rocks. For example, the Sanbagawa Metamorphic Complex in southwestern Japan
39 contains HP–LT psammitic and even conglomeratic schists (e.g., Wallis, 1998), and the
40 depositional ages and geochemical characteristics of the protolith are almost identical
41 to those of sandstone from the low-grade Shimanto Accretionary Complex (Kiminami
42 et al., 1999; Shibata et al., 2008; Aoki et al., 2012) and submarine fan turbidites
43 deposited in the associated fore-arc basin (Noda and Sato, 2018) (Figure 1). These
44 observations indicate that terrigenous trench-fill sediments were accreted in a shallow
45 subduction zone and were also subducted to >20 km depth. Other examples of such
46 subduction–accretion-related HP–LT metamorphic rocks can be seen in the Franciscan
47 Complex in California (e.g., Ernst, 2011; Jacobson et al., 2011; Dumitru et al., 2015;
48 Raymond, 2018), the Chugach terrane in Alaska (Plafker et al., 1994), the Central

49 Pontides in Turkey (Okay et al., 2006), and the Coastal Cordillera in Chile (Glodny
50 et al., 2005; Willner et al., 2004; Angiboust et al., 2018).

51 At typical sedimentary accretion zones, such as those in Cascadia (Gulick et al.,
52 1998; Booth-Rea et al., 2008; Calvert et al., 2011), Alaska (Moore et al., 1991; Ye et al.,
53 1997), Java (Kopp et al., 2009), southern Chile (Glodny et al., 2005; Melnick et al.,
54 2006), Sumatra (Singh et al., 2008; Huot and Singh, 2018), and Japan (Park et al., 2002;
55 Kimura et al., 2010), terrigenous trench-fill sediments are generally scraped off at the
56 frontal wedge, whereas hemipelagic-to-pelagic sediments underplate the base of the
57 accretionary wedge (e.g., Scholl, 2019). This may be because the increased structural
58 thickness of the wedge and progressive dewatering of subducting sediment causes
59 the décollement to step down and narrow the subduction channel (e.g., Sample and
60 Moore, 1987; Vannucchi et al., 2008). This suggests that the growth of the accretionary
61 wedge might inhibit the subduction of terrigenous sediment beyond the wedge through
62 the subduction channel. However, occurrences of HP–LT metasediment at some
63 accretionary margins demonstrate that terrigenous sediment can be subducted beneath
64 the wedge. One hypothesis is that a topographic high enables trench-fill sediment to be
65 subducted under the wedge (Figure 2). Subducting seamounts followed by subducting
66 material can be observed beneath the wedge along accretionary margins in southwestern
67 Japan (Moore et al., 2014), Alaska (Li et al., 2018), Barbados (Moore et al., 1995), and
68 Hikurangi (Barker et al., 2009; Bell et al., 2010).

69 The subduction of terrigenous material associated with the rough topography of a
70 subducting oceanic plate has been proposed to explain tectonic erosion of the wedge
71 (e.g., von Huene and Culotta, 1989; Lallemand et al., 1994; von Huene et al., 2004).
72 Sandbox analogue experiments have shown the potential for sediment transport be-
73 low the frontal wedge behind a subducting topographic high (Lallemand et al., 1992;
74 Dominguez et al., 2000). Numerical simulations show that in the wake of a subducting
75 seamount, there are unfaulted strata, large-offset thrust faults, increased fault spacing,
76 an oversteepened surface slope, and intense deformation along the base of the wedge
77 (Morgan and Bangs, 2017). In addition, recent seismic profiles across the accretionary
78 margins of the Nankai Trough (Bangs et al., 2006) and the Hikurangi Trench (Bell

79 et al., 2010) reveal that subducting seamounts or ridges and the surrounding sediment
80 are accommodated by a step-up in the décollement, and the surrounding sediment is
81 being transported to depth.

82 However, the influence of a subducting seamount beneath an accretionary wedge
83 on subduction and accretion fluxes is not well understood. In particular, the role of to-
84 pographic highs in modifying the décollement level and in maintaining or rejuvenating
85 the subduction channel as a conduit for sediment subduction needs to be explored. The
86 purpose of this study is (1) to investigate how the topographic roughness of the subduct-
87 ing plate interface influences material fluxes, including the accretion of sediment to the
88 wedge and the subduction of sediment along the subduction channel, and (2) to propose
89 a model that explains how terrigenous trench-fill sediment can be transported to depth.
90 We performed two types of sandbox analogue experiment, one with and one without a
91 topographic high. The novelty of these experiments is that they used a detached back-
92 stop to reproduce the subduction and underplating of sediment when a rigid topographic
93 high is subducted beneath an accretionary wedge. We also inserted two weak layers
94 within the sand, to reproduce the situation where the subducting sediment includes
95 several potential slip surfaces. Such multiple décollements are commonly found within
96 underthrust sediments or at the top of the oceanic crust, including at the Nankai (Moore
97 et al., 2001; Park et al., 2002), Hikurangi (Ghisetti et al., 2016; Plaza-Faverola et al.,
98 2016), and Barbados (Saffer, 2003) accretion zones.

99 **METHODS**

100 **Model Setup and Experimental Materials**

101 A scaled 2-D analogue modeling technique was used for this study so that the results
102 could be compared with naturally occurring geological structures (e.g., Buiter, 2012;
103 Graveleau et al., 2012). A glass-sided rectangular deformation rig with internal dimen-
104 sions of 100 cm × 30 cm × 20 cm was used (Figure 3). A steel plate was positioned at
105 one end as a fixed wall with a small open window at the bottom. A rigid wedge made
106 from wood was placed next to the steel plate but was not fixed to it. The wedge was

107 designed to behave like a static backstop that has a higher mechanical strength than the
108 accretionary wedge (e.g., Tsuji et al., 2015). A rigid backstop is used to ensure stability
109 during the experiments and for repeatability. The mobility of the backstop helped to
110 replicate the deformable nature of equivalent structures in natural geological systems,
111 and to allow topographic relief to be subducted. The backstop had a surface slope that
112 dips at 30° and is covered by sandpaper. A plastic (Mylar®) sheet was placed over
113 the rig's base plate and fixed to a roll that pulled the sheet using a stepper motor (on
114 the left side in Figure 3). The sheet was pulled beneath the rigid backstop at a rate of
115 0.5 cm/min, thereby compressing the experimental material above.

116 Two types of granular material were used for the experiments: Toyoura sand and
117 glass micro-beads. Dry granular materials like these are widely used as analogue
118 materials to simulate the brittle and frictional behavior of sedimentary rocks in accre-
119 tionary wedges because they display elastic–frictional plastic behavior and reproduce
120 the non-linear deformation of crustal rocks under brittle conditions (e.g., Dahlen, 1984;
121 Lohrmann et al., 2003; Graveleau et al., 2012). Toyoura sand, a standard testing mate-
122 rial commonly used by Japanese civil engineers, is a spherical quartz-rich sand with a
123 particle size of 0.14–0.26 mm ($D_{50} = 0.2$ mm), a density of approximately 1600 kg m^{-3} ,
124 an internal coefficient of friction, μ , of 0.59–0.68, and a cohesion, C , of 105–127 Pa
125 (Yamada et al., 2006; Dotare et al., 2016). The glass micro-beads are spherical and
126 0.045–0.063 mm in diameter, have a low internal coefficient of friction ($\mu = 0.47$) and
127 low cohesion (40 Pa), and are considered a suitable analogue for weaker layers (Yamada
128 et al., 2006, 2014).

129 Layers of sand and glass micro-beads with a total thickness of 3.4 cm were used
130 in the experiments. The sand and glass were sprinkled into the rig from a height of
131 approximately 30 cm above the rig floor (Figure 3). Alternating layers of blue, red,
132 and black sand were laid down to help visualize the cross-sectional geometry of the
133 models, without influencing the mechanical homogeneity. Mechanically weak layers
134 were created by adding two thin layers of glass micro-beads, each 3 mm thick.

135 Experiment A (Exp. A) investigated the subduction of a smooth oceanic plate
136 beneath a static backstop (Figure 3a). Experiment B (Exp. B) investigated the subduction

137 of topographic relief (e.g., a seamount), using a block that was attached to the plastic
138 sheet (Figure 3b). The height of the relief was 1.6 cm, approximately half of the
139 total thickness of the sediment. The height of the relief was chosen to avoid drastic
140 deformation of the accretionary wedge. The surface of the topographic relief was
141 covered by a Teflon[®] sheet. The total amount of horizontal shortening was 30 cm for
142 Exp. A and 35 cm for Exp. B.

143 After each 2 cm increment of shortening, we sprinkled dry sand from at least
144 10 cm above the surface of the accretionary wedge to fill the topographic lows that had
145 developed (Figure 4). This sand was used to replicate sedimentation in fore-arc/slope
146 basins that form on the surfaces of accretionary wedges. A total of 1129 g of sand was
147 added over the course of Exp. A and 910 g during Exp. B. The volumes of sand added
148 during Exp. A and B were 706 and 569 cm³, respectively.

149 In addition to investigating wedge morphology, we studied temporal variations in
150 sediment influx/outflux. The sediment influx and outflux (cm²) were calculated using
151 the thicknesses (cm) of the trench-fill sediments (influx) and the subduction channel
152 underneath the backstop (outflux), which were multiplied by the rig width (30 cm) and
153 divided by the length of shortening (cm). Input and output (cm³) are here defined to
154 be the integrals of influx and outflux, respectively, with respect to shortening length
155 (cm). Time-lapse digital images were taken through the transparent side glass at 5 s
156 intervals using a PC-based controller. The images were later analyzed to calculate
157 sediment influx/outflux and to study the cross-sectional geometry of the wedges. The
158 experiments did not account for the effects of isostatic compensation and erosion, which
159 would have contributed to the differences between our models and natural examples
160 (e.g., Schellart and Strak, 2016).

161 **Scaling**

162 Models used in laboratory experiments should be properly scaled so that the results
163 can be considered true analogues of geological processes (e.g., Hubbert, 1937). It is
164 assumed that brittle deformation will obey frictional Mohr–Coulomb-type laws. The
165 basic scaling relationship between the physical properties of a model and those in

166 nature, which relates the stress, σ , density, ρ , gravity, g , and length, l (Hubbert, 1937;
167 Schellart, 2000) is

$$\frac{\sigma_g}{\sigma_m} = \frac{l_g}{l_m} \times \frac{g_g}{g_m} \times \frac{\rho_g}{\rho_m}. \quad (1)$$

168 where the subscripts m and g indicate model and geological values, respectively. The
169 cohesion C can substitute for stress, σ (Schellart, 2000; Graveleau et al., 2012), and the
170 experiments are performed under normal gravity ($g_m/g_g = 1$); consequently, Eq. 1 can
171 be modified to give

$$\frac{l_g}{l_m} = \frac{C_g}{C_m} \times \frac{\rho_m}{\rho_g}. \quad (2)$$

172 For mean bulk density values of 2000–2500 kg m⁻³ and cohesion values of 5–
173 20 MPa, which are typical of sedimentary rocks in accretionary wedges (Schumann
174 et al., 2014), the length scale ratio ranges from approximately 3×10^4 to 1×10^5 . A
175 1 cm model layer in an experiment therefore corresponds to 300 m to 1 km in nature.
176 The 3.4-cm-thick sediment layers used in this experiment can be scaled to 1–3 km of
177 strata, which is a moderate thickness of trench-fill sediment for a modern accretionary
178 margin (Noda, 2016). The 5 cm width and 1.6 cm height of the topographic relief
179 used in Exp. B can be scaled to 1.5–5 km and 0.5–1.6 km, respectively. The scaled
180 dimensions of the topographic relief are comparable to many seamounts on the Pacific
181 plate. However, the height-to-radius ratio of 0.64 in the model is higher than that
182 of 0.21 for natural seamounts (Jordan et al., 1983; Smith, 1988). This high ratio is
183 used to enhance the effects of topography. The total amount of shortening during the
184 experiments was 30–35 cm, which is equivalent to 9–35 km of displacement. Assuming
185 a plate convergence rate of 5 cm/year, this in turn corresponds to $1.8\text{--}7 \times 10^5$ years. A
186 sediment supply to the topographic lows of 910–1129 g for 6×10^5 years is equivalent
187 to a sediment budget on the order of 10^6 t/year. The calculated sediment budget is the
188 same order of magnitude as the sediment load in many mountainous rivers in Japan
189 and New Zealand (Milliman and Syvitski, 1992), and the sedimentary influx into the
190 Kumano Basin during the last 4 Myr ($50 \text{ km} \times 70 \text{ km} \times 2 \text{ km}$).

191 **RESULTS**

192 **Experiment A: Subduction without a seamount**

193 During the first ~9 cm of shortening, high-frequency, low-amplitude forethrusts de-
194 veloped in front of the backstop (Stage 1, Figure 5a; 8 cm of shortening in Figure 6).
195 The wedge was uplifted quickly (uplift rate is 0.34 in Figure 5d), and thus the slope
196 increased rapidly, exceeding 12° by the end of Stage 1 (Figure 5c). After the emergence
197 of T_6 (Stage 2), the frequency of forethrust initiation and the uplift rate of the wedge
198 (0.10) were lower than during Stage 1, but the rate of wedge widening (0.22) remained
199 nearly constant (Figure 5). The slope of the wedge surface ranged from 8.5° to 13° , and
200 was 9.5° at the end of the experiment (Figure 5c).

201 Deformation was concentrated in the upper layer of glass beads, which acted as a
202 décollement, until 16 cm of shortening (Figure 6). At around 18 cm of shortening the
203 décollement stepped down to the lower layer of glass beads as the toe of the backstop
204 subsided below the upper layer of glass beads. During this stage, the footwall of
205 forethrust T_7 underthrust the wedge and the sand layer between the two layers of glass
206 beads underplated the wedge, creating a duplex structure (18–24 cm of shortening in
207 Figure 6). This underthrusting raised the hanging wall of T_7 and created a piggy-
208 back basin (trench-slope basin) on top of the wedge (22 cm of shortening). After the
209 activation of T_8 , with the lower layer of glass beads acting as a décollement, subducting
210 sediment was accreted to both the frontal and basal parts of the wedge with increasing
211 amount of underplating and thickness of the forethrust sheet of T_8 . The final forethrust,
212 T_9 , was initiated with the upper layer of glass beads acting as the detachment (30 cm
213 of shortening). The final wedge was nearly 30 cm in length and had a constant slope
214 of 9.5° (Figure 5). The toe of the backstop further subsided, to the lower layer of glass
215 beads (30 cm of shortening in Figure 6).

216 The outflux from the subduction channel (sediment subduction) gradually de-
217 creased, but its rate of change increased (Figure 5e). In particular, after the décollement
218 stepped down, the outflux dropped rapidly. Influx to the accretionary wedge (solid
219 dashed line in Figure 5e) increased to balance the total sediment influx. The output-to-

220 input ratio of the experiment was 0.36 (Table 1).

221 **Experiment B: Subduction With a Seamount**

222 Stage 1 of Exp. B was almost identical to that of Exp. A in terms of wedge progradation,
223 and the widening and uplift rates of the wedge (Figure 5a–d). Stage 2 started after the
224 initiation of forethrust T_5 , earlier than in Exp. A. T_5 was active for over 12.8 cm of
225 shortening, exceeding that of any other forethrust in either experiment (Figure 5a). This
226 long activity acted to reduce the width of the wedge and steepened its slope to 17.7°
227 (Figure 5b, c). The wedge progradation rate during Stage 2 was 0.10, nearly half that
228 of Exp. A (Figure 5a). The uplift rate varied from 0.06 to 0.29, but the mean rate was
229 the same as in Exp. A (Figure 5b).

230 The wedge deformation process during Stage 1 of Exp. B was similar that in Exp. A
231 (0–6 cm of shortening in Figure 7). However, at 7 cm of shortening, the seamount
232 triggered the first forethrust of Stage 2 at 10 cm from the toe of the wedge (T_5 in Figure 7).
233 The subduction of the seamount led to an undeformed layer underthrusting the wedge,
234 and then uplifted the hanging wall as a trench-slope basin to create accommodation
235 space (10–16 cm of shortening in Figure 7).

236 A décollement was formed in the upper layer of glass beads on the landward side of
237 the seamount and in T_5 on the trenchward side during the period between the initiation
238 of T_5 and collision of the seamount with the backstop (8–12 cm of shortening in
239 Figure 7). Just prior to the collision (12–18 cm of shortening), both the upper and lower
240 layers of glass beads were sliding and the sand layer between two layers of glass beads
241 underplated and was injected into T_5 . The décollement stepped up from the lower layer
242 of glass beads to T_5 when the seamount passed. In addition, following the collision
243 the seamount raised the backstop and opened a subduction channel beneath it (>20 cm
244 of shortening in Figure 7). The subsequent forethrusts, T_6 and T_7 , were rooted in a
245 décollement in the upper layer of glass beads. Finally, the toe of the backstop subsided
246 slightly, causing the lower layer of glass beads to act as a décollement.

247 Sediment outflux gradually decreased (blue line in Figure 5f), as it did during Exp. A,
248 until the seamount reached the backstop. After the seamount raised the backstop, at

249 around 17–20 cm of shortening (Figure 7), sediment outflux fully recovered and even
250 exceeded its initial rate (Figure 5d). Outflux soon decreased again as the seamount
251 subducted farther landward and the backstop subsided (Figure 5f). The output-to-input
252 ratio of the experiment was 0.46 (Table 1).

253 **DISCUSSION**

254 **Décollement Step-down and Underplating**

255 The gradual decrease of the outflux in Exp. A (Figure 5e) increased the influx to the
256 accretionary wedge, which increased its growth rate. During the time the upper layer of
257 glass beads acted as a décollement, the sediment above it was accreted to the wedge front.
258 As the slip switched to the lower layer of glass beads, the sediment between the two
259 layers of glass beads underplated the wedge, and frontal accretion continued. Similar
260 results have been reported in previous analogue experiments; i.e., underplating becomes
261 significant when the outflux from the subduction channel (sediment subduction) is
262 smaller than the influx (Kukowski et al., 1994; Albert et al., 2018). The results of
263 our experiment support the conclusion that a narrowing of the subduction channel and
264 a decrease in outflux can lead to sediment underplating the wedge and faster wedge
265 growth.

266 If we assume that sand above the upper layer of glass beads is terrigenous sediment,
267 and that sand below this layer is hemipelagic–pelagic sediment, the former can be
268 scraped off at the wedge front and the latter may be underplated below the wedge
269 (see Figure 8). This occurs because terrigenous and hemipelagic sediments tend to be
270 detached as a result of variations in diagenetic alteration (Moore, 1975) or smectite
271 content (Vrolijk, 1990; Deng and Underwood, 2001), or existence of weak smectitic
272 pelagic clay (Moore et al., 2015). This can be observed in the Nankai Trough, where
273 there is a step-down in the décollement at 1–3 km depth, in the transitional region
274 between the aseismic and seismic zones (cf. Park et al., 2002; Kimura et al., 2007),
275 which could be due to the different physical properties of these rock types.

276 The stepping down of the décollement in this study was associated with subsidence

277 of the backstop, which was probably linked to increased overburden stress caused by
278 thickening of the wedge. Increased overburden stress may inhibit the subduction of
279 terrigenous sediment to great depth. Underplating related to subsidence of the backstop
280 (inner wedge) also occurs along erosive margins. For example, thick (> 2 km) sediment
281 cover suggests subsidence of the inner wedge of the Ecuador–Colombia margin (Collot
282 et al., 2008). Seismic profiles indicate underplating between listric splay faults and the
283 basal décollement beneath the apex of the inner wedge, but the total mass flux at the
284 plate interface is negative (Collot et al., 2008), and material at the base of the inner
285 wedge is eroded.

286 **Décollement Step-up and Sediment Subduction**

287 In Exp. B, the subduction of a seamount shifted the décollement from the glass bead
288 layers into forethrust T₅ along the leading flank of the seamount. While T₅ was active
289 as a “top décollement” (cf. Lallemand et al., 1994), incoming undeformed layered sand
290 in the wake of the seamount was underthrust below the accretionary wedge. This is
291 similar to what is seen in seismic profiles from the Nankai (Bangs et al., 2006) and
292 Hikurangi margins (Bell et al., 2010), which show a décollement with a step-up caused
293 by seamount subduction.

294 Another effect of seamount subduction in Exp. B is that raising the backstop widened
295 the subduction channel, allowing thick layers of sand to subduct below the backstop
296 through the subduction channel. In nature, if an oceanic plate with sufficiently large
297 topographic highs subducts under a static backstop (cf. Tsuji et al., 2015), trench-fill
298 terrigenous sediment accompanying the highs could be transported through the sub-
299 duction channel to a higher-pressure environment than sediment on a smooth oceanic
300 plate. Exp. B could be analogous to the transport mechanism of the protolith of the an-
301 cient Sanbagawa Metamorphic and Shimanto Accretionary complexes of southwestern
302 Japan.

303 We propose a schematic model for the subduction of terrigenous sediment under
304 an accretionary wedge (Figure 8). A progressive thickening of the wedge increases the
305 overburden on the décollement that develops along weak layers in the cover sediment

306 deposited on the subducting oceanic plate. This overburden results in dewatering
307 and diagenetic alteration of the subducting sediment, which increases its mechanical
308 strength, leading to a step-down in the décollement (Figure 8a, b). The reduction of
309 sediment outflux due to narrowing of the subduction channel increases the mass of
310 sediment underplated beneath the wedge and the rate of frontal accretion. When a
311 topographic high (e.g., a seamount or an aseismic ridge) subducts under the wedge,
312 the décollement steps up to the forethrust along the leading flank of the seamount
313 (Figure 8b). This likely enables the subduction of terrigenous sediment beneath the
314 wedge. Further subduction of the topographic high would raise the backstop and open
315 the subduction channel for terrigenous sediment to be subducted into a high-pressure
316 environment (Figure 8c). After the topographic high passes the inner wedge or backstop,
317 the décollement under the accretionary wedge returns to the plate boundary or a weak
318 layer within the trench-fill sediments.

319 **Further Implications**

320 Excess pore pressure is important in maintaining subduction channels along the plate
321 interface (e.g., Saffer and Bekins, 2006). If the excess pore pressure drops below the
322 overburden pressure, the physical conditions in the subduction channel may resemble
323 those in the accretionary wedge (cf., Nankai and Barbados; Saffer, 2003). This probably
324 accelerates both the stepping down of the décollement and underplating (Strasser et al.,
325 2009; Kimura et al., 2011). In contrast, numerical simulations predict that the raising
326 of the wedge due to the subduction of a seamount could delay the release of fluid from
327 subducting sediment (Baba et al., 2001; Ruh et al., 2016). Low-velocity layers observed
328 in the wake of subducting seamounts could provide evidence of under-compacted sedi-
329 ment with potentially high excess pore pressures (e.g., Sage et al., 2006). Furthermore,
330 the seismic reflection characteristics of the Hikurangi subduction margin also suggest
331 localized reductions in effective stress associated with seamount subduction (Bell et al.,
332 2010). In addition to topographic relief, excess pore pressure could allow subduction
333 channels to persist for longer than would otherwise be possible. Our experiments can-
334 not currently incorporate the effects of excess pore pressure; consequently, we need to

335 consider ways to include these effects.

336 Where the trench-fill sediments are insufficient to fully cover the topographic relief
337 of the subducting oceanic crust, tectonic erosion may dominate and the accretionary
338 wedge cannot grow, as seen in northeastern Japan, Costa Rica, and Ecuador (von Huene
339 et al., 2004; Collot et al., 2011). Therefore, a sediment-rich subduction zone is required
340 for terrigenous sediments to be transported from shallow depths (e.g., the Shimanto
341 accretionary complex) to the depth of HP metamorphism (e.g., the Sanbagawa meta-
342 morphic complex).

343 **CONCLUSIONS**

344 We conducted a series of analogue experiments to investigate how terrigenous sed-
345 iment is subducted under an accretionary wedge. The results yielded the following
346 conclusions.

- 347 1. An increase in overburden stress due to progressive thickening of the accre-
348 tionary wedge leads the décollement to step down and narrows the subduction
349 channel. This accelerates the growth of the wedge through underplating and
350 frontal accretion.
- 351 2. When a topographic high subducts under the wedge, the décollement steps up
352 from a weak detachment layer within the incoming sediment to the forethrust
353 along the landward flank of the seamount. This enables terrigenous sediment in
354 the wake of the seamount to be underthrust beneath the wedge.
- 355 3. If a topographic high is rigid enough to uplift the backstop, it can widen the
356 subduction channel to transport the terrigenous sediment that follows toward
357 deeper environments.
- 358 4. A sufficient sediment supply to the trench and a rough oceanic crust surface
359 are necessary for simultaneous shallow accretion, underplating of the wedge,
360 and transportation of sediment to deeper settings as the protolith of HP-LT
361 metamorphic rocks.

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369 available on Figshare (<http://doi.org/10.6084/m9.figshare.10263674>).

370 **REFERENCES CITED**

- 371 Albert, F., Kukowski, N., Tassara, A., and Oncken, O., 2018, Material transfer and
372 subduction channel segmentation at erosive continental margins: Insights from
373 scaled analogue experiments: *Tectonophysics*, v. 749, p. 46–61, [https://doi.org/](https://doi.org/10.1016/j.tecto.2018.10.019)
374 [10.1016/j.tecto.2018.10.019](https://doi.org/10.1016/j.tecto.2018.10.019).
- 375 Angiboust, S., Cambeses, A., Hyppolito, T., Glodny, J., Monié, P., Calderón, M., and Ju-
376 liani, C., 2018, A 100-m.y.-long window onto mass-flow processes in the Patagonian
377 Mesozoic subduction zone (Diego de Almagro Island, Chile): *Geological Society of*
378 *America Bulletin*, v. 130, no. 9–10, p. 1439–1456, <https://doi.org/10.1130/B31891.1>.
- 379 Aoki, K., Iizuka, T., Hirata, T., Maruyama, S., and Terabayashi, M., 2007, Tectonic
380 boundary between the Sanbagawa belt and the Shimanto belt in central Shikoku,
381 Japan: *Journal of the Geological Society of Japan*, v. 113, no. 5, p. 171–183,
382 <https://doi.org/10.5575/geosoc.113.171>.
- 383 Aoki, K., Isozaki, Y., Yamamoto, S., Maki, K., Yokoyama, T., and Hirata, T., 2012, Tec-
384 tonic erosion in a Pacific-type orogen: Detrital zircon response to Cretaceous tecton-
385 ics in Japan: *Geology*, v. 40, no. 12, p. 1087–1090, <https://doi.org/10.1130/G33414.1>.
- 386 Baba, T., Hori, T., Hirano, S., Cummins, P.R., Park, J.O., Kameyama, M., and Kaneda,
387 Y., 2001, Deformation of a seamount subducting beneath an accretionary prism:

- 388 Constraints from numerical simulation: *Geophysical Research Letters*, v. 28, no. 9,
389 p. 1827–1830, <https://doi.org/10.1029/2000GL012266>.
- 390 Bangs, N.L.B., Gulick, S.P.S., and Shipley, T.H., 2006, Seamount subduction erosion
391 in the Nankai Trough and its potential impact on the seismogenic zone: *Geology*,
392 v. 34, no. 8, p. 701–704, <https://doi.org/10.1130/G22451.1>.
- 393 Barker, D.H.N., Sutherland, R., Henrys, S., and Bannister, S., 2009, Geome-
394 try of the Hikurangi subduction thrust and upper plate, North Island, New
395 Zealand: *Geochemistry, Geophysics, Geosystems*, v. 10, no. Q02007, [https://doi.org/](https://doi.org/10.1029/2008GC002153)
396 [10.1029/2008GC002153](https://doi.org/10.1029/2008GC002153).
- 397 Bell, R., Sutherland, R., Barker, D.H.N., Henrys, S., Bannister, S., Wallace, L., and
398 Beavan, J., 2010, Seismic reflection character of the Hikurangi subduction inter-
399 face, New Zealand, in the region of repeated Gisborne slow slip events: *Geophys-
400 ical Journal International*, v. 180, no. 1, p. 34–48, [https://doi.org/10.1111/j.1365-](https://doi.org/10.1111/j.1365-246X.2009.04401.x)
401 [246X.2009.04401.x](https://doi.org/10.1111/j.1365-246X.2009.04401.x).
- 402 Booth-Rea, G., Klaeschen, D., Grevemeyer, I., and Reston, T., 2008, Heteroge-
403 neous deformation in the Cascadia convergent margin and its relation to thermal
404 gradient (Washington, NW USA): *Tectonics*, v. 27, no. TC4005, [https://doi.org/](https://doi.org/10.1029/2007TC002209)
405 [10.1029/2007TC002209](https://doi.org/10.1029/2007TC002209).
- 406 Buiter, S.J.H., 2012, A review of brittle compressional wedge models: *Tectonophysics*,
407 v. 530–531, p. 1–17, <https://doi.org/10.1016/j.tecto.2011.12.018>.
- 408 Calvert, A.J., Preston, L.A., and Farahbod, A.M., 2011, Sedimentary underplating at
409 the Cascadia mantle-wedge corner revealed by seismic imaging: *Nature Geoscience*,
410 v. 4, no. 8, p. 545–548, <https://doi.org/10.1038/ngeo1195>.
- 411 Collot, J.Y., Agudelo, W., Ribodetti, A., and Marcaillou, B., 2008, Origin of a crustal
412 splay fault and its relation to the seismogenic zone and underplating at the erosional
413 north Ecuador–south Colombia oceanic margin: *Journal of Geophysical Research*,
414 v. 113, no. B12102, <https://doi.org/10.1029/2008JB005691>.

- 415 Collot, J.Y., Ribodetti, A., Agudelo, W., and Sage, F., 2011, The South Ecuador subduc-
416 tion channel: Evidence for a dynamic mega-shear zone from 2D fine-scale seismic
417 reflection imaging and implications for material transfer: *Journal of Geophysical*
418 *Research*, v. 116, no. B11, <https://doi.org/10.1029/2011JB008429>.
- 419 Dahlen, F.A., 1984, Noncohesive critical coulomb wedges: An exact solution: *Jour-*
420 *nal of Geophysical Research*, v. 89, no. B12, p. 10,125–10,133, [https://doi.org/](https://doi.org/10.1029/JB089iB12p10125)
421 [10.1029/JB089iB12p10125](https://doi.org/10.1029/JB089iB12p10125).
- 422 Deng, X., and Underwood, M.B., 2001, Abundance of smectite and the loca-
423 tion of a plate-boundary fault, Barbados accretionary prism: *Geological Soci-*
424 *ety of America Bulletin*, v. 113, no. 4, p. 495–507, [https://doi.org/10.1130/0016-](https://doi.org/10.1130/0016-7606(2001)113<0495:AOSATL>2.0.CO;2)
425 [7606\(2001\)113<0495:AOSATL>2.0.CO;2](https://doi.org/10.1130/0016-7606(2001)113<0495:AOSATL>2.0.CO;2).
- 426 Dominguez, S., Malavieille, J., and Lallemand, S.E., 2000, Deformation of accretionary
427 wedges in response to seamount subduction: Insights from sandbox experiments:
428 *Tectonics*, v. 19, no. 1, p. 182–196, <https://doi.org/10.1029/1999TC900055>.
- 429 Dotare, T., Yamada, Y., Adam, J., Hori, T., and Sakaguchi, H., 2016, Initiation of a
430 thrust fault revealed by analog experiments: *Tectonophysics*, v. 684, p. 148–156,
431 <https://doi.org/10.1016/j.tecto.2015.12.023>.
- 432 Dumitru, T.A., Ernst, W.G., Hourigan, J.K., and McLaughlin, R.J., 2015, Detrital
433 zircon U–Pb reconnaissance of the Franciscan subduction complex in northwestern
434 California: *International Geology Review*, v. 57, no. 5–8, p. 767–800, [https://doi.org/](https://doi.org/10.1080/00206814.2015.1008060)
435 [10.1080/00206814.2015.1008060](https://doi.org/10.1080/00206814.2015.1008060).
- 436 Ernst, W.G., 2011, Accretion of the Franciscan Complex attending Jurassic–Cretaceous
437 geotectonic development of northern and central California: *Geological Society of*
438 *America Bulletin*, v. 123, no. 9–10, p. 1667–1678, <https://doi.org/10.1130/B30398.1>.
- 439 Geological Survey of Japan, AIST, ed., 2015, Seamless digital geological map of Japan
440 1:200,000, May 29, 2015 version: Geological Survey of Japan, National Institute of
441 Advanced Industrial Science and Technology, [Cited 06 Dec 2019]. Available from:
442 https://gbank.gsj.jp/seamless/index_en.html.

- 443 Ghisetti, F.C., Barnes, P.M., Ellis, S., Plaza-Faverola, A.A., and Barker, D.H.N., 2016,
444 The last 2 Myr of accretionary wedge construction in the central Hikurangi margin
445 (North Island, New Zealand): Insights from structural modeling: *Geochemistry, Geo-*
446 *physics, Geosystems*, v. 17, p. 2661–2686, <https://doi.org/10.1002/2016GC006341>.
- 447 Glodny, J., Lohrmann, J., Echtler, H., Gräfe, K., Seifert, W., Collao, S., and Figueroa,
448 O., 2005, Internal dynamics of a paleoaccretionary wedge: insights from combined
449 isotope tectonochronology and sandbox modelling of the South-Central Chilean
450 forearc: *Earth and Planetary Science Letters*, v. 231, no. 1, p. 23–39, [https://doi.org/](https://doi.org/10.1016/j.epsl.2004.12.014)
451 [10.1016/j.epsl.2004.12.014](https://doi.org/10.1016/j.epsl.2004.12.014).
- 452 Graveleau, F., Malavieille, J., and Dominguez, S., 2012, Experimental modelling of
453 orogenic wedges: A review: *Tectonophysics*, v. 538–540, p. 1–66, [https://doi.org/](https://doi.org/10.1016/j.tecto.2012.01.027)
454 [10.1016/j.tecto.2012.01.027](https://doi.org/10.1016/j.tecto.2012.01.027).
- 455 Gulick, S.P.S., Meltzer, A.M., and Clarke, S.H., 1998, Seismic structure of the southern
456 Cascadia subduction zone and accretionary prism north of the Mendocino triple
457 junction: *Journal of Geophysical Research*, v. 103, no. B11, p. 27,207–27,222,
458 <https://doi.org/10.1029/98JB02526>.
- 459 Hara, H., and Hara, K., 2019, Radiolarian and U–Pb zircon dating of Late Cretaceous
460 and Paleogene Shimanto accretionary complexes, Southwest Japan: Temporal vari-
461 ations in provenance and offset across an out-of-sequence thrust: *Journal of Asian*
462 *Earth Sciences*, v. 170, p. 29–44, <https://doi.org/10.1016/j.jseaes.2018.10.016>.
- 463 Hara, H., Nakamura, Y., Hara, K., Kurihara, T., Mori, H., Iwano, H., Danhara, T.,
464 Sakata, S., and Hirata, T., 2017, Detrital zircon multi-chronology, provenance, and
465 low-grade metamorphism of the Cretaceous Shimanto accretionary complex, east-
466 ern Shikoku, Southwest Japan: Tectonic evolution in response to igneous activity
467 within a subduction zone: *Island Arc*, v. 26, no. 6, p. e12,218, [https://doi.org/](https://doi.org/10.1111/iar.12218)
468 [10.1111/iar.12218](https://doi.org/10.1111/iar.12218).
- 469 Hubbert, M.K., 1937, Theory of scale models as applied to the study of geologic

- 470 structures: Geological Society of America Bulletin, v. 48, no. 10, p. 1459–1519,
471 <https://doi.org/10.1130/GSAB-48-1459>.
- 472 Huot, G., and Singh, S.C., 2018, Seismic evidence for fluid/gas beneath the Mentawai
473 Fore-Arc Basin, central Sumatra: Journal of Geophysical Research: Solid Earth, v.
474 123, no. 2, p. 957–976, <https://doi.org/10.1002/2017JB014849>.
- 475 Jacobson, C.E., Grove, M., Pedrick, J.N., Barth, A.P., Marsaglia, K.M., Gehrels, G.E.,
476 and Nourse, J.A., 2011, Late Cretaceous–early Cenozoic tectonic evolution of the
477 southern California margin inferred from provenance of trench and forearc sediments:
478 Geological Society of America Bulletin, v. 123, no. 3–4, p. 485–506, [https://doi.org/](https://doi.org/10.1130/B30238.1)
479 [10.1130/B30238.1](https://doi.org/10.1130/B30238.1).
- 480 Jordan, T.H., Menard, H.W., and Smith, D.K., 1983, Density and size distribution
481 of seamounts in the eastern Pacific inferred from wide-beam sounding data: Jour-
482 nal of Geophysical Research, v. 88, no. B12, p. 10,508–10,518, [https://doi.org/](https://doi.org/10.1029/JB088iB12p10508)
483 [10.1029/JB088iB12p10508](https://doi.org/10.1029/JB088iB12p10508).
- 484 Kiminami, K., Hamasaki, A., and Matsuura, T., 1999, Geochemical contrast between the
485 Sanbagawa psammitic schists (Oboke unit) and the Cretaceous Shimanto sandstones
486 in Shikoku, Southwest Japan and its geologic significance: Island Arc, v. 8, no. 3, p.
487 373–382, <https://doi.org/10.1046/j.1440-1738.1999.00248.x>.
- 488 Kimura, G., Kitamura, Y., Hashimoto, Y., Yamaguchi, A., Shibata, T., Ujiie, K., and
489 Okamoto, S., 2007, Transition of accretionary wedge structures around the up-dip
490 limit of the seismogenic subduction zone: Earth and Planetary Science Letters, v.
491 255, no. 3–4, p. 471–484, <https://doi.org/10.1016/j.epsl.2007.01.005>.
- 492 Kimura, G., Moore, G.F., Strasser, M., Sreaton, E., Curewitz, D., Streiff, C., and
493 Tobin, H., 2011, Spatial and temporal evolution of the megasplay fault in the Nankai
494 Trough: Geochemistry, Geophysics, Geosystems, v. 12, no. Q0A008, [https://doi.org/](https://doi.org/10.1029/2010GC003335)
495 [10.1029/2010GC003335](https://doi.org/10.1029/2010GC003335).
- 496 Kimura, H., Takeda, T., Obara, K., and Kasahara, K., 2010, Seismic evidence for active

- 497 underplating below the megathrust earthquake zone in Japan: *Science*, v. 329, no.
498 5988, p. 210–212, <https://doi.org/10.1126/science.1187115>.
- 499 Kopp, H., Hindle, D., Klaeschen, D., Oncken, O., Reichert, C., and Scholl, D., 2009,
500 Anatomy of the western Java plate interface from depth-migrated seismic images:
501 *Earth and Planetary Science Letters*, v. 288, no. 3–4, p. 399–407, [https://doi.org/](https://doi.org/10.1016/j.epsl.2009.09.043)
502 [10.1016/j.epsl.2009.09.043](https://doi.org/10.1016/j.epsl.2009.09.043).
- 503 Kukowski, N., von Huene, R., Malavieille, J., and Lallemand, S.E., 1994, Sediment
504 accretion against a buttress beneath the Peruvian continental margin at 12° S as
505 simulated with sandbox modeling: *Geologische Rundschau*, v. 83, no. 4, p. 822–831,
506 <https://doi.org/10.1007/BF00251079>.
- 507 Lallemand, S.E., Malavieille, J., and Calassou, S., 1992, Effects of oceanic ridge
508 subduction on accretionary wedges; experimental modeling and marine observations:
509 *Tectonics*, v. 11, no. 6, p. 1301–1313, <https://doi.org/10.1029/92TC00637>.
- 510 Lallemand, S.E., Schnürle, P., and Malavieille, J., 1994, Coulomb theory applied to
511 accretionary and nonaccretionary wedges: Possible causes for tectonic erosion and/or
512 frontal accretion: *Journal of Geophysical Research*, v. 99, no. B6, p. 12,033–12,055,
513 <https://doi.org/10.1029/94JB00124>.
- 514 Li, J., Shillington, D.J., Saffer, D.M., Bécel, A., Nedimović, M.R., Kuehn, H., Webb,
515 S.C., Keranen, K.M., and Abers, G.A., 2018, Connections between subducted sed-
516 iment, pore-fluid pressure, and earthquake behavior along the Alaska megathrust:
517 *Geology*, v. 46, no. 4, p. 299–302, <https://doi.org/10.1130/G39557.1>.
- 518 Lohrmann, J., Kukowski, N., Adam, J., and Oncken, O., 2003, The impact of analogue
519 material properties on the geometry, kinematics, and dynamics of convergent sand
520 wedges: *Journal of Structural Geology*, v. 25, no. 10, p. 1691–1711, [https://doi.org/](https://doi.org/10.1016/S0191-8141(03)00005-1)
521 [10.1016/S0191-8141\(03\)00005-1](https://doi.org/10.1016/S0191-8141(03)00005-1).
- 522 Melnick, D., Bookhagen, B., Echtler, H.P., and Strecker, M.R., 2006, Coastal defor-
523 mation and great subduction earthquakes, Isla Santa María, Chile (37°S): *Geolog-*

- 524 ical Society of America Bulletin, v. 118, no. 11–12, p. 1463–1480, [https://doi.org/](https://doi.org/10.1130/B25865.1)
525 10.1130/B25865.1.
- 526 Milliman, J.D., and Syvitski, J.P.M., 1992, Geomorphic/tectonic control of sediment
527 discharge to the ocean: The importance of small mountainous rivers: *Journal of*
528 *Geology*, v. 100, no. 5, p. 525–544, <https://doi.org/10.1086/629606>.
- 529 Moore, G.F., Kanagawa, K., Strasser, M., Dugan, B., Maeda, L., Toczko, S., and the
530 IODP Expedition 338 Scientific Party, 2014, IODP Expedition 338: NanTroSEIZE
531 Stage 3: NanTroSEIZE plate boundary deep riser 2: *Scientific Drilling*, v. 17, p.
532 1–12, <https://doi.org/10.5194/sd-17-1-2014>.
- 533 Moore, G.F., Zhao, Z., Shipley, T.H., Bangs, N., and Moore, J.C., 1995, Structural
534 setting of the Leg 156 area, northern Barbados Ridge accretionary prism, *in* Shipley,
535 T.H., Ogawa, Y., Blum, P., and et al., eds., *Proceedings of the Ocean Drilling*
536 *Program, Initial Reports*, v. 156: College Station, TX, Ocean Drilling Program, p.
537 13–27, <https://doi.org/10.2973/odp.proc.ir.156.102.1995>.
- 538 Moore, G.F., others, Klaus, A., and Becker, L., 2001, New insights into deformation
539 and fluid flow processes in the Nankai Trough accretionary prism: *Results of Ocean*
540 *Drilling Program Leg 190: Geochemistry, Geophysics, Geosystems*, v. 2, no. 1058,
541 <https://doi.org/10.1029/2001GC000166>.
- 542 Moore, J.C., 1975, Selective subduction: *Geology*, v. 3, no. 9, p. 530–532,
543 [https://doi.org/10.1130/0091-7613\(1975\)3<530:SS>2.0.CO;2](https://doi.org/10.1130/0091-7613(1975)3<530:SS>2.0.CO;2).
- 544 Moore, J.C., others, Fisher, M.A., and Sample, J., 1991, EDGE deep seismic reflection
545 transect of the eastern Aleutian arc-trench layered lower crust reveals underplating and
546 continental growth: *Geology*, v. 19, no. 5, p. 420–424, [https://doi.org/10.1130/0091-](https://doi.org/10.1130/0091-7613(1991)019<0420:EDSRTO>2.3.CO;2)
547 [7613\(1991\)019<0420:EDSRTO>2.3.CO;2](https://doi.org/10.1130/0091-7613(1991)019<0420:EDSRTO>2.3.CO;2).
- 548 Moore, J.C., Plank, T.A., Chester, F.M., Polissar, P.J., and Savage, H.M., 2015, Sediment
549 provenance and controls on slip propagation: Lessons learned from the 2011 Tohoku
550 and other great earthquakes of the subducting northwest Pacific plate: *Geosphere*,
551 v. 11, no. 3, p. 533–541, <https://doi.org/10.1130/GES01099.1>.

- 552 Morgan, J.K., and Bangs, N.L., 2017, Recognizing seamount-forearc collisions at
553 accretionary margins: Insights from discrete numerical simulations: *Geology*, v. 45,
554 no. 7, p. 635–638, <https://doi.org/10.1130/G38923.1>.
- 555 Nagata, M., Miyazaki, K., Iwano, H., Danhara, T., Obayashi, H., Hirata, T., Yagi, K.,
556 Kouchi, Y., Yamamoto, K., and Otoh, S., 2019, Timescale of material circulation in
557 subduction zone: U–Pb zircon and K–Ar phengite double-dating of the Sanbagawa
558 metamorphic complex in the Ikeda district, central Shikoku, southwest Japan: *Island
559 Arc*, v. 28, no. 4, p. e12,306, <https://doi.org/10.1111/iar.12306>.
- 560 Noda, A., 2016, Forearc basins: Types, geometries, and relationships to subduction
561 zone dynamics: *Geological Society of America Bulletin*, v. 128, no. 5–6, p. 879–
562 895, <https://doi.org/10.1130/B31345.1>.
- 563 Noda, A., Danhara, T., Iwano, H., and Hirata, T., 2017, LA-ICP-MS U–Pb and fission-
564 track ages of felsic tuff beds of the Takikubo Formation, Izumi Group in the Kan-onji
565 district, eastern Shikoku, southwestern Japan: *Bulletin of the Geological Survey of
566 Japan*, v. 68, no. 3, p. 119–130, <https://doi.org/10.9795/bullgsj.68.119>.
- 567 Noda, A., and Sato, D., 2018, Submarine slope–fan sedimentation in an ancient forearc
568 related to contemporaneous magmatism: The Upper Cretaceous Izumi Group, south-
569 western Japan: *Island Arc*, v. 27, no. 2, p. e12,240, <https://doi.org/10.1111/iar.12240>.
- 570 Okay, A.I., Tüysüz, O., Satır, M., Özkan-Altiner, S., Altiner, D., Sherlock, S., and
571 Eren, R.H., 2006, Cretaceous and Triassic subduction-accretion, high-pressure–
572 low-temperature metamorphism, and continental growth in the Central Pontides,
573 Turkey: *Geological Society of America Bulletin*, v. 118, no. 9–10, p. 1247–1269,
574 <https://doi.org/10.1130/B25938.1>.
- 575 Otoh, S., Shimojo, M., Aoki, K., Nakama, T., Maruyama, S., and Yanai, S., 2010,
576 Age distribution of detrital zircons in the psammitic schist of the Sanbagawa Belt,
577 Southwest Japan: *Journal of Geography (Chigaku Zasshi)*, v. 119, no. 2, p. 333–346,
578 <https://doi.org/10.5026/jgeography.119.333>.

- 579 Park, J.O., Tsuru, T., Takahashi, N., Hori, T., Kodaira, S., Nakanishi, A., Miura, S., and
580 Kaneda, Y., 2002, A deep strong reflector in the Nankai accretionary wedge from
581 multichannel seismic data: Implications for underplating and interseismic shear
582 stress release: *Journal of Geophysical Research*, v. 107, no. B4, [https://doi.org/
583 10.1029/2001JB000262](https://doi.org/10.1029/2001JB000262).
- 584 Plafker, G., Moore, J.C., and Winkler, G.R., 1994, Geology of the Southern Alaska
585 margin, *in* Plafker, G., and Berg, H.C., eds., *The geology of Alaska, The Geology
586 of North America*, v. G-1: chap. 12, Boulder, Colorado, United States, Geological
587 Society of America, p. 389–449, <https://doi.org/10.1130/DNAG-GNA-G1.389>.
- 588 Plaza-Faverola, A., Henrys, S., Pecher, I., Wallace, L., and Klaeschen, D., 2016, Splay
589 fault branching from the Hikurangi subduction shear zone: Implications for slow
590 slip and fluid flow: *Geochemistry, Geophysics, Geosystems*, v. 17, p. 5009–5023,
591 <https://doi.org/10.1002/2016GC006563>.
- 592 Raymond, L.A., 2018, What is Franciscan?: revisited: *International Geology Review*,
593 v. 60, no. 16, p. 1968–2030, <https://doi.org/10.1080/00206814.2017.1396933>.
- 594 Ruh, J.B., Sallarès, V., Ranero, C.R., and Gerya, T., 2016, Crustal deformation dynamics
595 and stress evolution during seamount subduction: High-resolution 3-D numerical
596 modeling: *Journal of Geophysical Research: Solid Earth*, v. 121, p. 6880–6902,
597 <https://doi.org/10.1002/2016JB013250>.
- 598 Saffer, D.M., 2003, Pore pressure development and progressive dewatering in under-
599 thrust sediments at the Costa Rican subduction margin: Comparison with northern
600 Barbados and Nankai: *Journal of Geophysical Research*, v. 108, no. B5, 2261,
601 <https://doi.org/10.1029/2002JB001787>.
- 602 Saffer, D.M., and Bekins, B.A., 2006, An evaluation of factors influencing pore
603 pressure in accretionary complexes; implications for taper angle and wedge me-
604 chanics: *Journal of Geophysical Research*, v. 111, no. B04101, [https://doi.org/
605 10.1029/2005JB003990](https://doi.org/10.1029/2005JB003990).

- 606 Sage, F., Collot, J.Y., and Ranero, C.R., 2006, Interplate patchiness and subduction-
607 erosion mechanisms: Evidence from depth-migrated seismic images at the central
608 Ecuador convergent margin: *Geology*, v. 34, no. 12, p. 997–1000, [https://doi.org/
609 10.1130/G22790A.1](https://doi.org/10.1130/G22790A.1).
- 610 Sample, J.C., and Moore, J.C., 1987, Structural style and kinematics of an un-
611 derplated slate belt, Kodiak and adjacent islands, Alaska: *Geological Soci-
612 ety of America Bulletin*, v. 99, no. 1, p. 7–20, [https://doi.org/10.1130/0016-
613 7606\(1987\)99<7:SSAKOA>2.0.CO;2](https://doi.org/10.1130/0016-7606(1987)99<7:SSAKOA>2.0.CO;2).
- 614 Schellart, W.P., 2000, Shear test results for cohesion and friction coefficients for dif-
615 ferent granular materials: scaling implications for their usage in analogue mod-
616 elling: *Tectonophysics*, v. 324, no. 1–2, p. 1–16, [https://doi.org/10.1016/S0040-
617 1951\(00\)00111-6](https://doi.org/10.1016/S0040-1951(00)00111-6).
- 618 Schellart, W.P., and Strak, V., 2016, A review of analogue modelling of geodynamic
619 processes: Approaches, scaling, materials and quantification, with an application to
620 subduction experiments: *Journal of Geodynamics*, v. 100, p. 7–32, [https://doi.org/
621 10.1016/j.jog.2016.03.009](https://doi.org/10.1016/j.jog.2016.03.009).
- 622 Scholl, D.W., 2019, Seismic imaging evidence that forearc underplating built the ac-
623 cretionary rock record of coastal North and South America: *Geological Magazine*,
624 v. ??, p. 1–14, <https://doi.org/10.1017/S0016756819000955>.
- 625 Schumann, K., Behrmann, J.H., Stipp, M., Yamamoto, Y., Kitamura, Y., and Lempp, C.,
626 2014, Geotechnical behavior of mudstones from the Shimanto and Boso accretionary
627 complexes, and implications for the Nankai accretionary prism: *Earth, Planets and
628 Space*, v. 66, no. 1, p. 129, <https://doi.org/10.1186/1880-5981-66-129>.
- 629 Shibata, T., Orihashi, Y., Kimura, G., and Hashimoto, Y., 2008, Underplating of
630 mélangé evidenced by the depositional ages: U–Pb dating of zircons from the Shi-
631 manto accretionary complex, southwest Japan: *Island Arc*, v. 17, no. 3, p. 376–393,
632 <https://doi.org/10.1111/j.1440-1738.2008.00626.x>.

- 633 Singh, S.C., others, Tapponnier, P., and Hananto, N.D., 2008, Seismic evidence for
634 broken oceanic crust in the 2004 Sumatra earthquake epicentral region: *Nature*
635 *Geoscience*, v. 1, no. 11, p. 777–781, <https://doi.org/10.1038/ngeo336>.
- 636 Smith, D.K., 1988, Shape analysis of Pacific seamounts: *Earth and Planetary Science*
637 *Letters*, v. 90, no. 4, p. 457–466, [https://doi.org/10.1016/0012-821X\(88\)90143-4](https://doi.org/10.1016/0012-821X(88)90143-4).
- 638 Strasser, M., others, Kimura, G., and Kitamura, Y., 2009, Origin and evolution of a
639 splay fault in the Nankai accretionary wedge: *Nature Geoscience*, v. 2, no. 9, p.
640 648–652, <https://doi.org/10.1038/NGEO609>.
- 641 Tsuji, T., Ashi, J., Strasser, M., and Kimura, G., 2015, Identification of the static
642 backstop and its influence on the evolution of the accretionary prism in the Nankai
643 Trough: *Earth and Planetary Science Letters*, v. 431, p. 15–25, [https://doi.org/](https://doi.org/10.1016/j.epsl.2015.09.011)
644 [10.1016/j.epsl.2015.09.011](https://doi.org/10.1016/j.epsl.2015.09.011).
- 645 Vannucchi, P., Remitti, F., and Bettelli, G., 2008, Geological record of fluid flow and
646 seismogenesis along an erosive subducting plate boundary: *Nature*, v. 451, no. 7179,
647 p. 699–703, <https://doi.org/10.1038/nature06486>.
- 648 von Huene, R., and Culotta, R., 1989, Tectonic erosion at the front of the Japan Trench
649 convergent margin: *Tectonophysics*, v. 160, p. 75–90, [https://doi.org/10.1016/0040-](https://doi.org/10.1016/0040-1951(89)90385-5)
650 [1951\(89\)90385-5](https://doi.org/10.1016/0040-1951(89)90385-5).
- 651 von Huene, R., Ranero, C.R., and Vannucchi, P., 2004, Generic model of subduction
652 erosion: *Geology*, v. 32, no. 10, p. 913–916, <https://doi.org/10.1130/G20563.1>.
- 653 Vrolijk, P., 1990, On the mechanical role of smectite in subduction
654 zones: *Geology*, v. 18, no. 8, p. 703–707, [https://doi.org/10.1130/0091-](https://doi.org/10.1130/0091-7613(1990)018<0703:OTMROS>2.3.CO;2)
655 [7613\(1990\)018<0703:OTMROS>2.3.CO;2](https://doi.org/10.1130/0091-7613(1990)018<0703:OTMROS>2.3.CO;2).
- 656 Wallis, S., 1998, Exhuming the Sanbagawa metamorphic belt: the importance of
657 tectonic discontinuities: *Journal of Metamorphic Geology*, v. 16, no. 1, p. 83–95,
658 <https://doi.org/10.1111/j.1525-1314.1998.00072.x>.

- 659 Willner, A.P., Glodny, J., Gerya, T.V., Godoy, E., and Massonne, H.J., 2004, A
660 counterclockwise PTt path of high-pressure/low-temperature rocks from the Coastal
661 Cordillera accretionary complex of south-central Chile: constraints for the earliest
662 stage of subduction mass flow: *Lithos*, v. 75, no. 3, p. 283–310, [https://doi.org/
663 10.1016/j.lithos.2004.03.002](https://doi.org/10.1016/j.lithos.2004.03.002).
- 664 Yamada, Y., Baba, K., Miyakawa, A., and Matsuoka, T., 2014, Granular experiments
665 of thrust wedges: Insights relevant to methane hydrate exploration at the Nankai
666 accretionary prism: *Marine and Petroleum Geology*, v. 51, p. 34–48, [https://doi.org/
667 10.1016/j.marpetgeo.2013.11.008](https://doi.org/10.1016/j.marpetgeo.2013.11.008).
- 668 Yamada, Y., Kaneda, K., and Matsuoka, T., 2006, Influences of material properties on
669 analogue model experiments of geologic structures: *Journal of the Society of Ma-
670 terials Science, Japan*, v. 55, no. 5, p. 452–457, <https://doi.org/10.2472/jsms.55.452>,
671 in Japanese with English abstract.
- 672 Ye, S., Flueh, E.R., Klaeschen, D., and von Huene, R., 1997, Crustal structure along
673 the EDGE transect beneath the Kodiak shelf off Alaska derived from OBH seis-
674 mic refraction data: *Geophysical Journal International*, v. 130, no. 2, p. 283–302,
675 <https://doi.org/10.1111/j.1365-246X.1997.tb05648.x>.

Table 1. Total sediment input and output, and their ratio. Asterisk (*) indicates output includes the volume of the seamount.

Exp.	Displacement (cm)	Input (cm ³)	Output* (cm ³)	Accretion (cm ³)	Output/Input Ratio
A	30.0	2,912	1,052	1,860	0.36
B	35.0	3,315	1,527	1,788	0.46

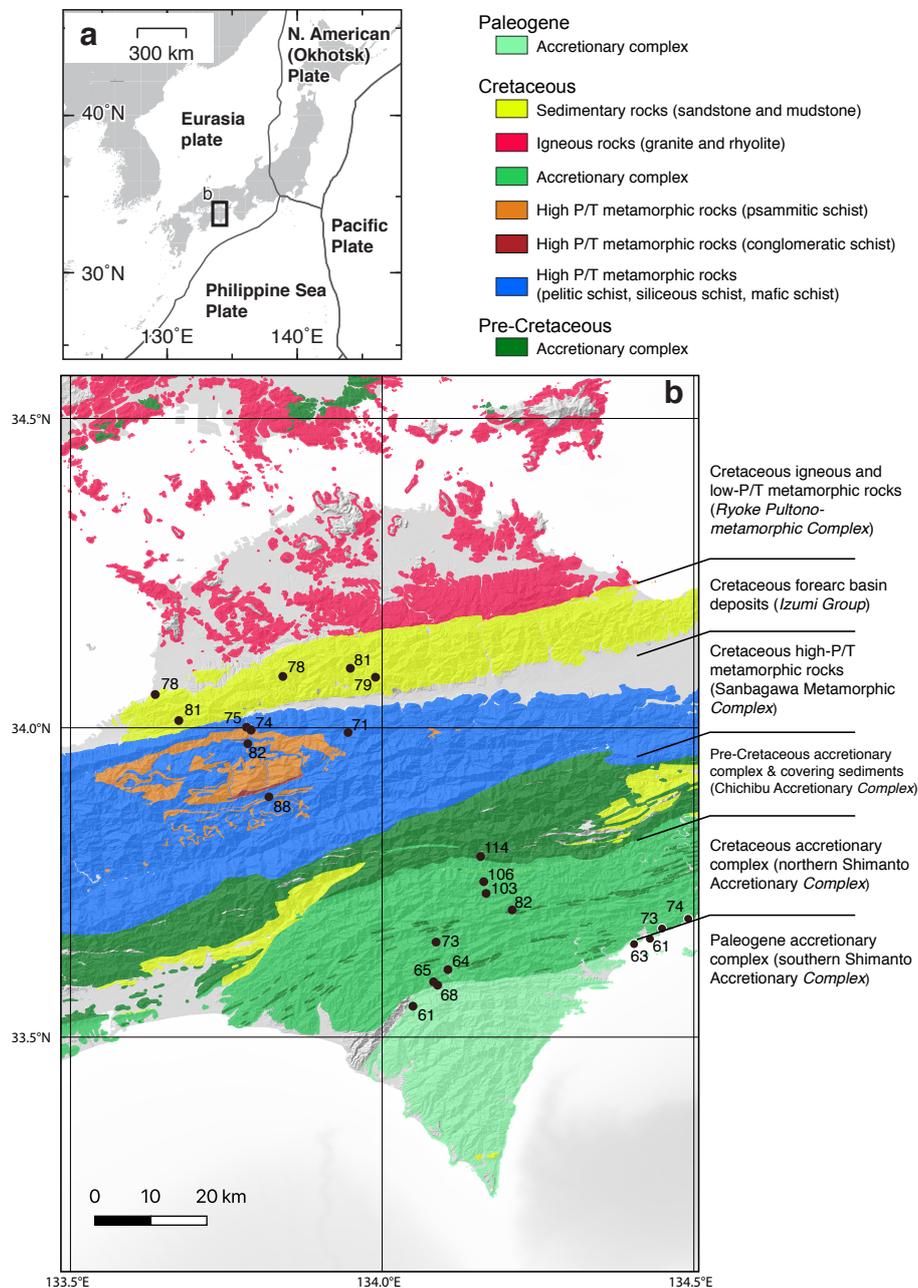


Figure 1. Generalized geological map of eastern Shikoku, southwestern Japan, reproduced from the Seamless Digital Geological Map of Japan (Geological Survey of Japan, AIST, 2015). Black dots are labeled with detrital zircon U–Pb ages (Ma) of felsic tuff beds in the Izumi Group, composed mainly of sandy turbidites and mudstone (Noda et al., 2017; ?), the psammitic schist of the Sanbagawa Metamorphic Complex (Aoki et al., 2007; Nagata et al., 2019; Otoh et al., 2010), and sandy turbidites in the northern Shimanto Accretionary Complex (Hara et al., 2017; Hara and Hara, 2019; Shibata et al., 2008).

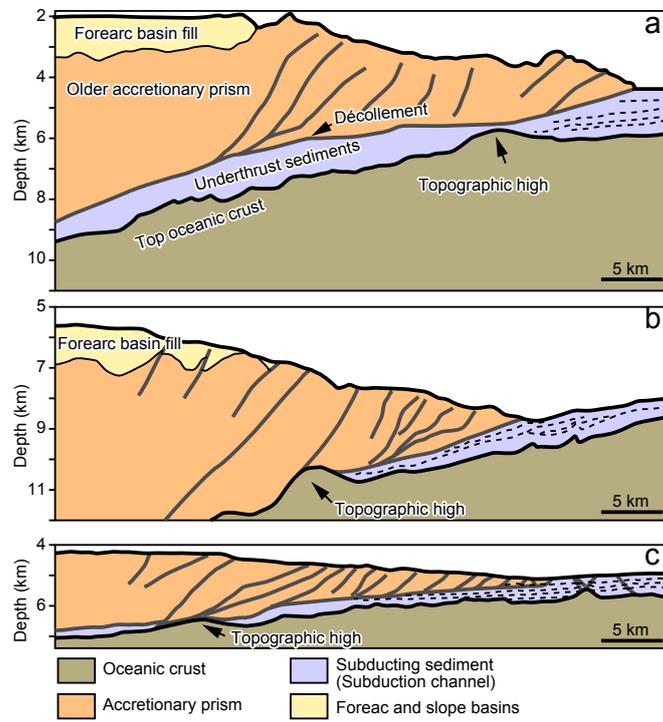


Figure 2. Representative cross-sections of accretionary margins with topographic highs. (a) Nankai Trough (Moore et al., 2014). (b) Southwestern Alaskan margin (Li et al., 2018). (c) Northern Barbados margin (Moore et al., 1995).

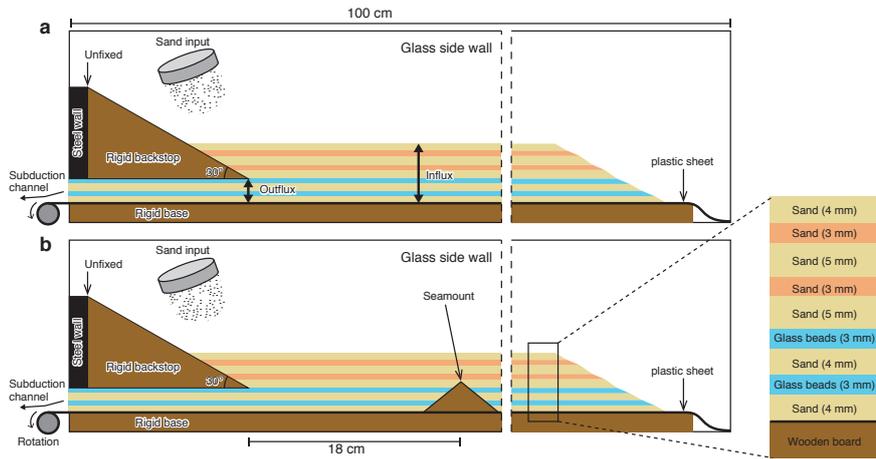


Figure 3. Experimental apparatus.

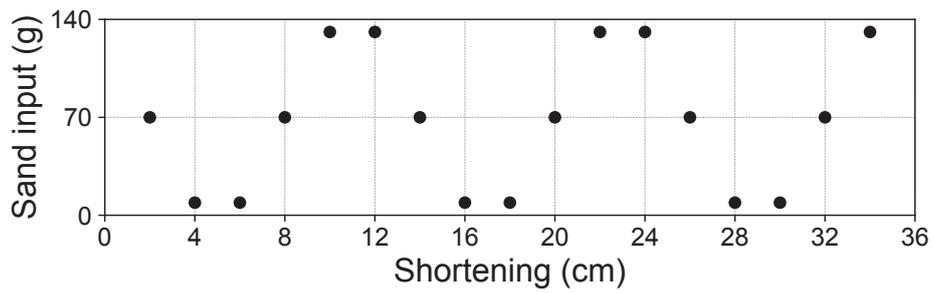


Figure 4. Amount of sand added to the topographic lows during the experiments.

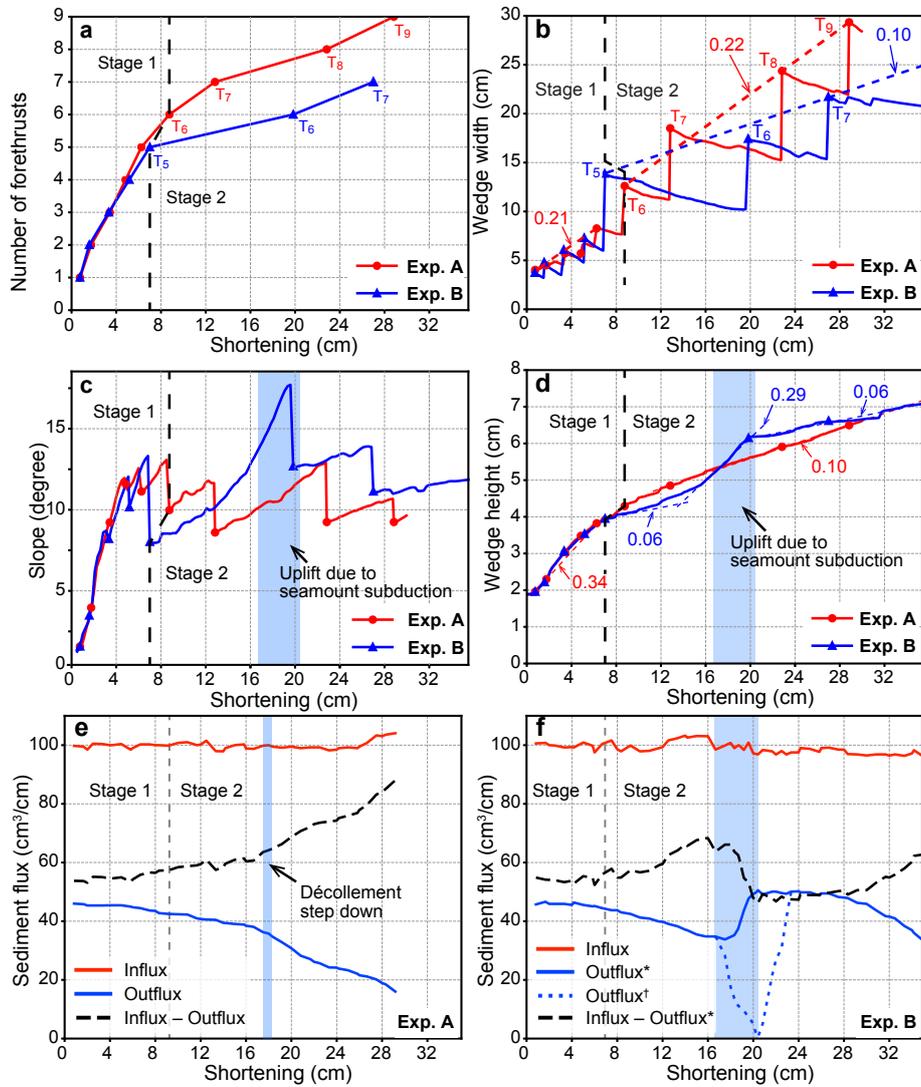


Figure 5. Geomorphic parameters of the wedges (a–d) and sediment fluxes (e–f). (a) Number of forethrusts. (b) Wedge width. Dashed lines are labelled with wedge progradation rates calculated from the amount of progradation (cm) divided by the amount of shortening (cm). (c) Wedge slope angle. (d) Wedge height. Dashed lines are labelled with uplift rates calculated from the amount of uplift (cm) divided by the amount of shortening (cm). (e) Sediment influx and outflux for Exp. A (without seamount). (f) Sediment influx and outflux for Exp. B (with seamount). Asterisk (*) and dagger (†) indicate outfluxes including and excluding the volume of the seamount, respectively.

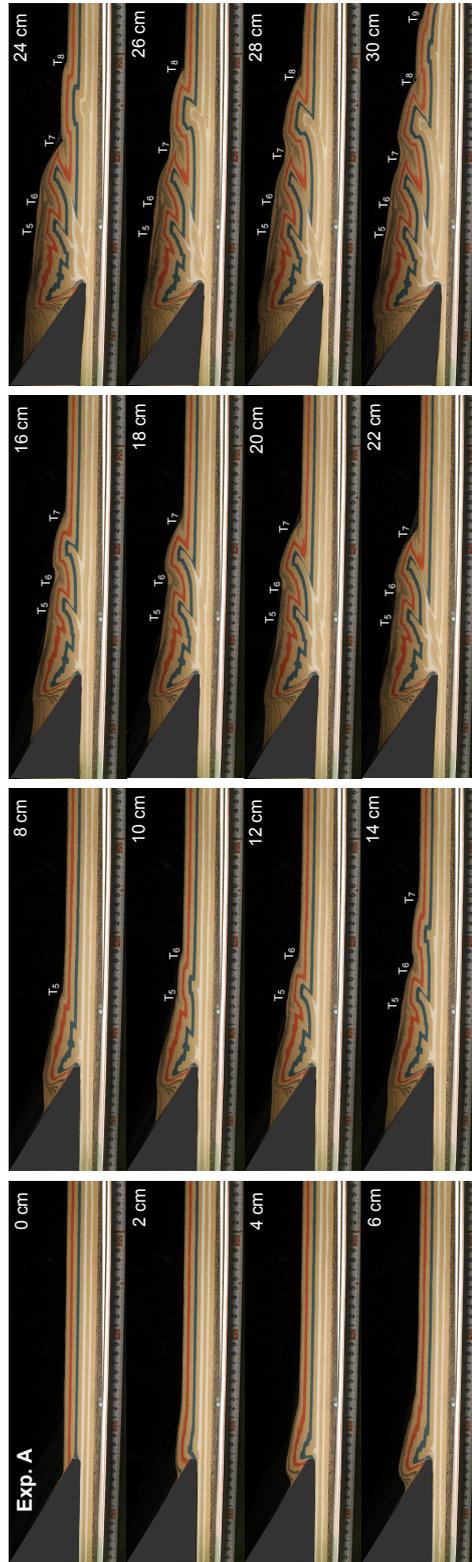


Figure 6. Representative images of Exp. A.

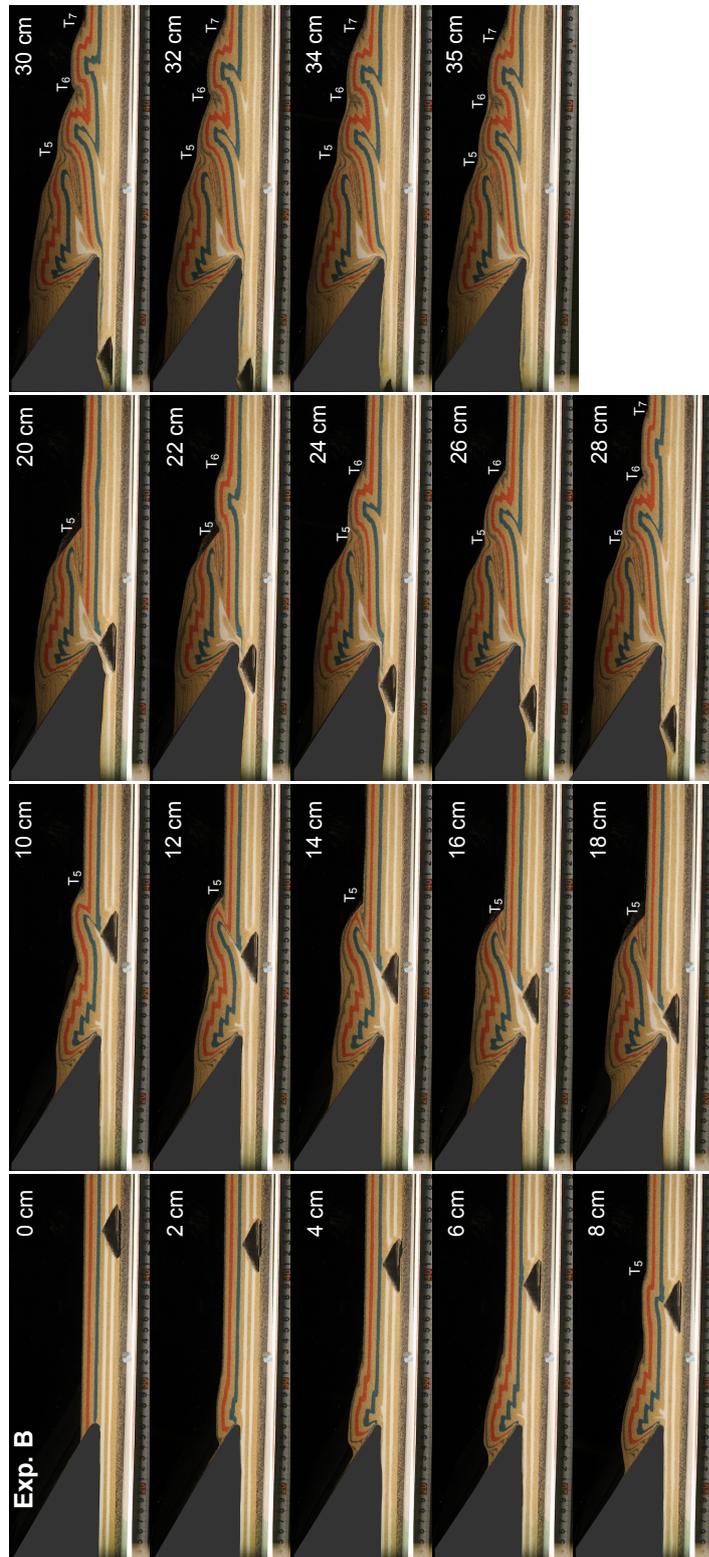


Figure 7. Representative images of Exp. B.

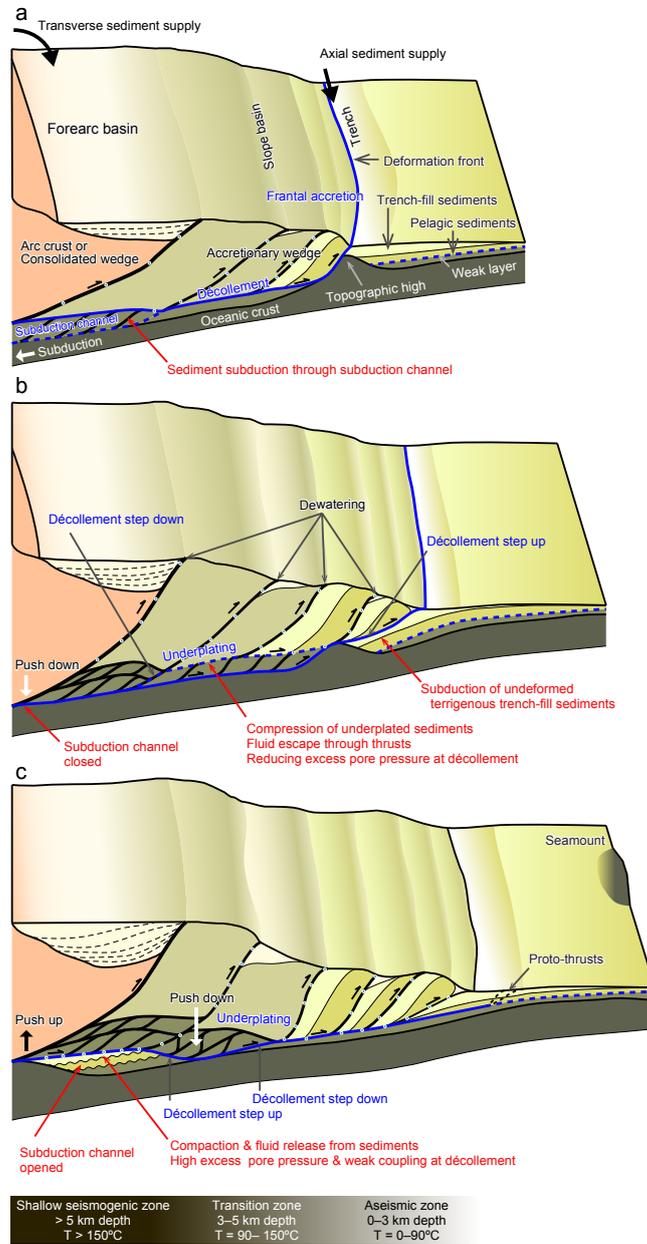


Figure 8. Schematic model of sediment subduction through a subduction channel beneath an accretionary wedge. (a) Subduction of a topographic high raises the décollement to accommodate the topographic high and the following trench-fill sediment. (b) An increase in overburden gravitational force under the inner wedge shifts the décollement downward and facilitates underplating. In the wake of the subducting seamount, terrigenous sediment is underthrust beneath the accretionary wedge. (c) The seamount raises the backstop, enabling the subduction of terrigenous sediment. After the passage of the seamount, the décollement returns to the original, lower position, and the subduction channel closes, resulting in underplating beneath the wedge.