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1 **Porosity and permeability evolution in the Tuaheni Landslide Complex**
2 **at Hikurangi margin from IODP Sites U1517 and U1519**

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11 Several mechanisms involving excess pore pressure related to gas hydrate have
12 been proposed to explain active creeping at landslides such as the north
13 Hikurangi margin Tuaheni Landslide Complex (TLC). Cores and logging data
14 were retrieved by the International Ocean Discovery Program (IODP)
15 Expeditions 372 and 375 from the South TLC at Site U1517. Here, the evolutions
16 of porosity, pore structure and permeability are determined to assess the
17 compaction state in the landslide and compare it with that of the undeformed
18 sequence at Site U1519. Although no evidence of gas hydrate in the landslide at
19 Site U1517 was identified by the cruise, zones suspected to host gas hydrates
20 below the landslide or at Site U1519 are characterised by higher porosity, pore
21 diameter and permeability. We show that most of the sedimentary section is in
22 hydrostatic conditions, except the base of the TLC at Site U1517 and a zone
23 below the base of the gas hydrate stability zone at Site U1519. These zones might
24 be candidates for excess pore pressure build-up. There is no obvious evidence of
25 the involvement of gas hydrate in active creeping at the TLC, which is more
26 likely induced by hydrogeomechanical processes.

27 Keywords: porosity, permeability, pore pressure, gas hydrates, MICP, NMR,
28 CEC, landslide, Hikurangi, IODP

29 **Introduction**

30 The Tuaheni Landslide Complex (TLC) is situated on the upper slope of the
31 relatively sediment-starved north Hikurangi margin (Mountjoy et al., 2009), where the
32 Pacific Plate obliquely subducts, at a ~5.0 cm/y rate, beneath the Australian Plate
33 (Wallace et al., 2004) (Figure 1a). In the study area, the Pacific Plate is composed of the
34 Cretaceous-aged volcanic Hikurangi Plateau protruded by seamounts and blanketed by a
35 ~1.0–1.5 kilometres-thick Cenozoic to Mesozoic sedimentary cover. The TLC is
36 composed of Quaternary shelf-edge clinoform sequences (Pedley et al., 2010), with
37 wedge-shaped thinning-upwards siliciclastic sediments (Barnes et al., 1991; Alexander
38 et al., 2010), above Cenozoic and Late Mesozoic sediments (Field et al., 1997; Barnes et
39 al., 2002; Mountjoy and Barnes, 2011). Based on morphological differences, Mountjoy
40 et al. (2009) divided the TLC into a north (North TLC) and a south component (South
41 TLC) (Figure 1b), both feeding the Tuaheni Basin with Quaternary sediments
42 (Mountjoy et al., 2009; 2014). The South TLC also contributes to filling the Paritu
43 Basin with debris. Arcuate internal scarps characterise the North TLC. The South TLC,
44 on which this study is focused, is constituted by irregular rough landslide debris
45 downslope of large scarps. The South TLC is thought to be divided into two mass-
46 transport deposit (MTD) units (Mountjoy et al., 2014; Böttner et al., 2018; Gross et al.,
47 2018; Kuhlmann et al., 2019). These units are separated by an intra-debris negative-
48 polarity seismic reflector, interpreted as a décollement layer that may localise excess
49 pore pressure (Gross et al., 2018).

50 Based on bathymetry data, Mountjoy et al. (2009) suggested that the South TLC
51 initially formed as a catastrophic failure and is now actively creeping, rather than
52 growing through repeated failures leading to downslope transport of new source
53 material, as commonly admitted for submarine landslides (Mulder and Cochonat, 1996;

54 Canals et al., 2004). However, the mechanisms triggering this active slow deformation
55 remain unclear. First, Mountjoy et al. (2009) related creeping with high pore pressure
56 induced by gas expulsion, sediment loading or earthquakes. Such creeping would
57 conduct to small-scale compressional and extensional features across the landslide,
58 which have not been observed. On the contrary, the South TLC is marked by a
59 transition from a compressive upper sediment-supplied mass to an extensive part near
60 the toe where sediments are removed (Figure 2a) (Mountjoy et al., 2014; Barnes et al.,
61 2019a). Later, Mountjoy et al. (2013, 2014) observed that the upper limit of creeping
62 occurs downslope of where a hydrate-related bottom-simulating-reflector (BSR) (and
63 thus the base of gas hydrate stability (BGHS)) cutting through the landslide reaches the
64 seafloor. This zone also corresponds to the transition between compressional and
65 extensional deformation of the South TLC. They suggested that gas hydrates may be
66 involved in ongoing slow deformation and proposed three possible mechanisms: 1) gas
67 hydrate dissociation at the BGHS; 2) a hydrate pressure valve model implying the
68 repeated hydrofracturing and weakening of the gas hydrate zone with a transfer of the
69 overpressure accumulated underneath low-permeability gas hydrate-bearing sediments
70 to the landslide mass where the BGHS approaches the seafloor and 3) a hydrate glacier
71 model with gas hydrate-bearing sediments showing creeping behaviour (Mountjoy et
72 al., 2013, 2014; Barnes et al., 2019a).

73 The International Ocean Discovery Program (IODP) Expedition 372 drilled,
74 logged and cored Site U1517 in the extensional zone of the south component (Figures
75 1b and 2) to investigate the mechanisms behind creeping in the TLC and their possible
76 links with gas hydrate occurrence. Gas hydrates were identified beneath the landslide
77 based on drilling results, discarding the hydrate glacier model hypothesis (Barnes et al.,
78 2019c; Sreaton et al., 2019). Two other sites were initially planned to be drilled in the

79 compressive part of the landslide and a reference site outside the landslide mass (Pecher
80 et al., 2014). Even if this second drilling phase was not achieved (Barnes et al., 2019a),
81 the outcomes from Site U1517 (Barnes et al., 2019b) can be interpreted in light of the
82 properties of the sedimentary section at Site U1519 (Barnes et al., 2019a). Site U1519
83 was drilled, logged and partially cored ~15 kilometres at the northeast of Site U1517, in
84 the western part of the Tuaheni basin fed by the north component of the TLC (Figures
85 1b and 3) (Barnes et al., 2019c). Gas hydrate occurrence is inferred from direct
86 observations at Site U1519 (Barnes et al., 2019c).

87 Here, we document the evolution of compaction state from porosity and
88 permeability from measurements on core samples and logging-while-drilling (LWD)
89 data at Site U1517 through the extensional part of the South TLC and beneath, and at
90 Site U1519. Compaction state may be interpreted in terms of pore pressure
91 development. Elevated pore pressure is commonly associated with a local anomalously
92 high porosity compared to a reference porosity-effective stress curve (Conin et al.,
93 2011) determined assuming hydrostatic fluid pressure and uniaxial compaction
94 conditions (Screaton et al., 2002). However, the porosity routinely measured on core
95 samples during IODP expeditions cannot be directly used for compaction state analysis
96 because it is based on the total water content of samples. Total water content includes
97 interstitial water, which is contained in pores, and water bound in hydrous minerals such
98 as smectite (Brown and Ransom, 1996; Henry, 1997). Smectite-bound water is mainly
99 released in the pore space by dehydration to illite, which is generally assumed to occur
100 in a temperature range of 60–150°C (Freed and Peacor, 1989; Kastner et al., 2014). In
101 contrast, interstitial water is expelled from the pore structure by pore collapse induced
102 by compaction as sediments are buried (Moore and Vrolijk, 1992). Thus, interstitial
103 water is truly representative of the compaction state (Henry, 1997; Henry and

104 Bourlange, 2004; Conin et al., 2011). Hence, we quantify interstitial porosity by
105 correcting total porosity measured on samples from the smectite-bound water content to
106 determine the compaction state at both sites. We also compare permeability deduced
107 from pore size distribution with the permeability measured on two samples (cores 17H6
108 and 29F3 at 97.43 and 157.83 mbsf respectively) in a triaxial cell under effective
109 hydrostatic confining pressure approaching in-situ stress conditions. Finally, we discuss
110 the porosity and permeability dataset in terms of pore pressure development at both
111 Sites U1517 and U1519. We address the hypotheses (i.e. gas hydrates are involved in
112 creeping in the TLC either through gas hydrate dissociation at the BGHS or hydrate
113 pressure valve effect) and antitheses (i.e. creeping is induced by other mechanisms such
114 as repeated small-scale seafloor failure or liquefaction of coarse silt beds) suggested
115 during IODP Expedition 372 (see Barnes et al., 2019a for details) as possible
116 mechanisms triggering slow deformation in the TLC.

117 **Material and data**

118 *Site U1517*

119 Site U1517 provided core samples and in-situ geophysical data including LWD
120 data, and temperature and pressure measurements down to ~205 mbsf (Figure 4a). Five
121 lithological, seismic and logging units were identified onboard (Barnes et al., 2019b).
122 Lithological Unit I (~0–3 mbsf) corresponds to Holocene-aged greenish-grey silty clay.
123 Units II, III and IV are Late Pleistocene-aged. Unit II (~3–41 mbsf) is composed of a
124 succession of stacked turbidites with alternating layers of mud (i.e. clayey silt to silty
125 clay) and greenish-grey sand beds. Unit III (~41–67 mbsf) corresponds to matrix-
126 supported MTDs with thin to medium bedded convoluted mud layers and alternating
127 layers of greenish-grey silt and clay. Unit IV (~67–103 mbsf) is a greenish-grey to dark

128 greenish grey massive clayey silty hemipelagic sequence with MTDs at the top and
129 punctuated with silty turbidites at the base. Unit V (~103–187 mbsf) is a Middle
130 Pleistocene mixed slope sequence including clayey to clayey–silt hemipelagites, sandy
131 turbidites, MTDs and ash layers near the bottom of the hole.

132 Above the upper part of the preslide interval (~67–203 mbsf), the South TLC
133 mass was interpreted to be composed of Unit II (upper debris unit) and Unit III (lower
134 debris unit) (Mountjoy et al., 2014; Gross et al., 2018; Barnes et al., 2019b; Luo et al.,
135 2020), with the décollement for creeping ~37 mbsf corresponding to the intra-debris
136 negative polarity reflector of Gross et al. (2018). Mountjoy et al. (2014) suggested that
137 the upper and lower units were genetically linked. This model has been recently
138 challenged by studies that highlight different source and/or emplacement mechanisms.
139 Luo et al. (2020) suggest that the landslide was formed by two different depositional
140 events corresponding to Units II and III, while Couvin et al. (2020) propose that the
141 landslide is only composed of the upper Unit II.

142 The base of the gas hydrate stability zone (BGHS) has been interpreted to occur
143 ~162 mbsf prior to drilling (Gross et al., 2018; Barnes et al., 2019b; Sreaton et al.,
144 2019, 2020) or between ~85 and 128 mbsf (Sultan, 2020). Indication for gas hydrate
145 occurrence was only found in Unit V (Figure 4) (Barnes et al., 2019b; Sreaton et al.,
146 2019, 2020; Sultan, 2020).

147 ***Site U1519***

148 Site U1519 was drilled and logged down to 650 mbsf across three seismic units:
149 1) horizontally layered basin–fill including stacked MTDs and turbiditic sequences (~0–
150 281 mbsf), the upper MTD being associated with the North TLC (Mountjoy et al., 2009,
151 2014; Barnes et al., 2019c), 2) southeast–dipping layered slope sequences of MTDs
152 truncated at the unconformity between seismic Units 1 and 2 (~281–546 mbsf) and the

153 top (~546–650 mbsf) of a northwest-dipping and eroded hanging-wall of an inactive
154 northwest-dipping splay fault that lies ~500 metres beneath the bottom of Site U1519
155 (Figure 3). Site U1519 crosscuts a weak reflector at ~560 mbsf interpreted as a possible
156 BSR. Several intervals were cored: ~0–23 mbsf at hole D, ~0–86 mbsf at hole E and
157 ~10–164 mbsf, ~250–288 mbsf and ~518–640 mbsf at hole C. Based on core
158 examination, two lithological units were identified during the cruise (Barnes et al.,
159 2019c) (Figure 5). Unit I corresponds to Holocene to late middle Pleistocene dark
160 greenish grey mud to mudstone (i.e. silty clay(stone) to clayey silt(stone)) associated
161 with hemipelagic settling with an ash layer occurring few meters below the seafloor.
162 Unit I is correlated with seismic Unit 1 and logging Unit 1. Seismic and logging Units 2
163 and 3 correspond to lithologic Unit II, including middle to early Pleistocene light
164 greenish-grey mudstone with scattered interbeds of silt and very fine sand, medium to
165 coarse sand, and contorted mudstone interpreted as MTDs. Gas hydrate-bearing
166 sediments were sampled in the lower part of the hole in core 25R4 (~627 mbsf, Figure
167 5) (Barnes et al., 2019c).

168 **Methods**

169 This study is based on 1) onboard data including LWD data and measurements
170 on core samples and 2) post-cruise analysis of 24 and 40 core samples retrieved at Site
171 U1517 and U1519, respectively. Samples were selected near the samples that were used
172 for pore water composition analysis onboard. Samples were conditioned for shipping
173 and stored at chilled temperature (2–8°C) in vacuum-sealed plastic bags with a sponge
174 saturated with seawater to retain moisture.

175 **Porosity**

176 *Total connected porosity*

177 We calculated the total connected porosity (ϕ_t) with a Micromeritics® AccuPyc
178 II 1340 helium-displacement pycnometer using the procedure after Blum (1997) that
179 was also followed onboard (Wallace et al., 2019; Pecher et al., 2019).

180 *Bound water content*

181 Clay bound water includes 1) water molecules absorbed in the interlayer space
182 that are usually organised into layers and 2) water molecules adsorbed on smectite
183 external surfaces and are electrostatically bounded to hydrated cations compensating the
184 negative charge of smectite layers. Clay-bound water content (ϕ_b) was calculated using
185 Equation (1):

186
$$\phi_b = n \frac{M_w}{\rho_w} CEC \rho_g (1 - \phi_t) \quad (1)$$

187 based on the cation exchange capacity (CEC, in mol/kg) that is a proxy for hydrous to
188 non-hydrous minerals content ratio, the water molar mass ($M_w = 0.018\text{kg/mol}$) and
189 density ($\rho_w = 1024 \text{ kg/m}^3$), the grain density (ρ_g in kg/m^3) and n the average number of
190 water molecules per cation charge related to the number of water layers in the clay
191 interlayer space. The average number of water molecules per cation charge is typically
192 determined from the relationship between the CEC and the volume of chloride-free
193 water per volume of grains (Figure 6) that is derived from the soluble chloride content
194 of the sample (Henry and Bourlange, 2004; Conin et al., 2011; Daigle et al., 2015a;
195 Dutilleul et al., 2020a; 2020b). The average number of water molecules per cation
196 charge commonly ranges 6–8, 12–16, and 18–24 for one, two or three water layers
197 respectively (Henry, 1997). All the samples at Sites U1517 and U1519 show low CEC
198 values $<0.2 \text{ mol/kg}$ and thus do not show a specific bound water ratio versus CEC trend

199 from which n can be inferred (see Henry and Bourlange, 2004; Dutilleul et al., 2020a
200 for further details). We used the average value of $n=15$ which is typical of smectite with
201 two water layers occurring in marine sediments (Kastner et al., 2014), following the
202 results of Dutilleul et al. (2020b).

203 *Interstitial porosity and compaction trends*

204 Interstitial porosity (ϕ_i) was obtained by correcting the total connected porosity (ϕ_t) for the clay-bound water content (ϕ_b) following Equation (2):

$$206 \quad \phi_i = \phi_t - \phi_b \quad (2)$$

207 The ‘onboard interstitial porosity’ was determined by extrapolating bound water
208 content measured on samples to correct the total connected porosity values that were
209 measured onboard on 153 samples at Site U1517 and 278 samples at Site U1519. At
210 Site U1517, where extensive coring was implemented, we used onboard interstitial
211 porosity data to determine compaction curves (Figure 7). Instead, at Site U1519 where
212 coring was restricted to limited intervals, we used resistivity-derived porosity calibrated
213 to onboard interstitial porosity.

214 *Resistivity-derived porosity*

215 For clay-rich materials, which characterised by high electrical surface
216 conductivity, we estimated resistivity-derived porosity from LWD resistivity using the
217 model by Revil et al. (1998). In this model, the resistivity-derived porosity (ϕ) is related
218 to the formation factor F based on Archie’s law (Archie, 1942) as expressed by
219 Equation (3) where a and m are constants.

$$220 \quad F = a \phi^{-m} \quad (3)$$

221 Conin et al. (2011) and Dutilleul et al. (2020b) have shown that in siliciclastic and
 222 volcanoclastic clay-rich sediments, resistivity-derived porosity fits interstitial porosity
 223 using $a=1$ and adjusting the cementation factor (m) between 1 and 3.5.
 224 F was determined with Equation (4) based on the bulk electrical conductivity (σ), the
 225 electrical surface conductivity (σ_s) and the electrical conductivity of the interstitial fluid
 226 (σ_{if}) as:

$$227 \quad F = \frac{\sigma_{if}}{\sigma} \left[1 + 2 \frac{\sigma_s}{\sigma_{if}} \left(\frac{\sigma_{if}}{\sigma} - 1 \right) \right] \quad (4)$$

228 since the hypotheses of Bussian (1983) and Bourlange et al. (2003) are valid at Site
 229 U1517 where $\frac{\sigma_s}{\sigma_{if}} \ll 1$.

230 We used ring resistivity which has the highest vertical resolution for the depth
 231 penetration (Barnes et al., 2019c) to determine σ .

232 σ_{if} was determined using Equation (5) from the concentrations of Cl^- , Na^+ , K^+ , Ca^{2+} ,
 233 Mg^{2+} and SO_4^{2-} in interstitial water (C_{iws}^i) and seawater (C_{sw}^i), the ionic mobility in the
 234 fluid (β_f^i), the number of charges of ions (Z_i) and the seawater conductivity (σ_{sw}).

$$235 \quad \sigma_{if} = \sigma_{sw} \frac{\sum_i (\beta_f^i \times Z_i \times C_{iws}^i)}{\sum_j (\beta_f^j \times Z_j \times C_{sw}^j)} \quad (5)$$

236 The seawater conductivity was calculated using Equation (6) based on temperature T (in
 237 $^{\circ}\text{C}$).

$$238 \quad \sigma_{sw} = 5.32(1 + 0.02(T - 25)) \quad (6)$$

239 At Site U1517, $T(^{\circ}\text{C}) = 5.32 + 39.8 \times 10^{-3} z$ (7) where z is the depth (in mbsf) (Barnes et
 240 al., 2019b) while $T(^{\circ}\text{C}) = 5.70 + 24.3 \times 10^{-3} z$ (8) at Site U1519 (Barnes et al.,
 241 2019c).

242 The model developed by Revil et al. (1998) assumes a major contribution of the Stern
243 layer to surface electrical conduction, spherical grains and a linear temperature
244 dependency of the exchangeable cation mobility (β_s) to quantify σ_s with Equation (9):

$$245 \quad \sigma_s = \frac{2}{3} \rho_g CEC \beta_s \quad (9)$$

246 ***Pore size distribution***

247 *Mercury injection capillary pressure (MICP)*

248 The distribution of the size of pore throats from 360 μm to 5.7 nm was
249 determined by MICP experiments performed with a Micromeritics® AutoPore IV 9500
250 at room temperature (20°C). Samples were formerly oven-dried at 105°C \pm 5°C for 24h.
251 During the low-pressure analysis, the samples were first degassed under vacuum. The
252 cell was then gradually intruded by a measured volume of mercury up to a mercury
253 pressure of \sim 0.2 MPa. Next, we used a high-pressure analysis sequence whereby we
254 measured the volume of intruded mercury filling the cell and the sample at a known
255 pressure that was increased stepwise to 220 MPa. This stage provides the distribution of
256 the size of pore throats using the Young-Laplace Equation (10) (Purcell, 1949):

$$257 \quad r = \frac{2 \sigma_{Hg} \cos \theta_{Hg}}{P_{Hg}} \quad (10)$$

258 from the pore throat radius (r in m), the air-mercury interfacial tension (σ_{Hg} =0.485
259 N/m), the mercury-sediment contact angle (θ_{Hg} =140°) and the mercury injection
260 pressure (P_{Hg} in Pa).

261 *Nuclear magnetic resonance (NMR)*

262 A Bruker® Minispec Mq20 was used to perform proton NMR measurements at
263 room temperature (20°C) and atmospheric pressure on core samples with a diameter of

264 8 mm. Proton NMR records the magnetisation decay of the hydrogen nuclei of water
 265 molecules in a saturated porous sample when static and pulsed magnetic fields are
 266 applied. We measured the transverse relaxation time (T_2) that provides insight into void
 267 size and porosity (Daigle et al., 2014; Bossennec et al., 2018) applying the Carr-Purcell-
 268 Meiboom-Gill (CPMG) sequence (Carr and Purcell, 1954; Meiboom and Gill, 1958).
 269 We used a recycle delay of 0.1 s, a half-echo time τ of 0.04 ms (the minimum
 270 obtainable for this equipment), gain ranging around 70–80%, 200 echoes per scan and
 271 128 scans were stacked (Dutilleul et al., 2020b). The raw T_2 exponential decay
 272 measured was inversed in a smoothed T_2 distribution using UpenWin© software. We
 273 used the Equation (11) following Marschall et al. (1995) to correlate the T_2 (in s)
 274 measured by LWD NMR to MICP pore throat radius (here, in μm) based on the
 275 effective relaxivity (ρ_e , in $\mu\text{m/s}$):

$$r = 2\rho_e T_2 \quad (11)$$

277 We extrapolated the effective relaxivity across coring gaps where no MICP data were
 278 available to determine an average pore radius profile across Sites U1517 and U1519
 279 based on LWD NMR data.

280 NMR porosity (ϕ) (Daigle et al., 2014) was determined using Equation (12) with the
 281 volume of dry grains (V_g) from the helium-displacement pycnometer and the volume of
 282 water (V_w) in the sample.

$$\phi_{NMR} = \frac{V_w}{V_w + V_g} \quad (12)$$

284 V_w was quantified using a calibration (Equation (13)) in which the maximum signal
 285 amplitude A_0 (corrected for the gain) was recorded during the T_2 measurement for
 286 known volume of water.

$$V_w = 19.762A_0 - 0.092 \quad (R^2=0.94) \quad (13)$$

288 NMR porosity was also measured onboard using the Schlumberger ProVisionPlus
289 logging tool.

290 ***Permeability***

291 *Permeability measurements under confining pressure*

292 Permeability measurements were conducted at room temperature, in a 200 MPa
293 hydrostatic pressure cell equipped with a pore fluid pressure circuit (Reuschlé, 2011).
294 Experiments were run on two cylindrical samples (20 mm in diameter and 40 mm in
295 length) from the preslide interval at Site U1517 at 97.43 mbsf (core 17H6, lithological
296 Unit IV) and 157.83 mbsf (core 29F3, Unit V) respectively. Permeability was measured
297 on the initial saturated samples using a pulse decay method (Brace, 1984): an initial
298 imposed differential pore fluid pressure (ΔP_p) is applied to equilibrate through the
299 sample. Its decay is approximately exponential and the decay time is inversely
300 proportional to the permeability following Equations (14) and (15) (Hsieh et al., 1981;
301 Reuschlé, 2011):

$$302 \quad \Delta P_p(t) \propto \exp(-\alpha t) \quad (14)$$

303 and

$$304 \quad \alpha = [Ak(C_u + C_d)]/(\mu LC_u C_d) \quad (15)$$

305 with the time (t), the cross-sectional area (A) and the length (L) of the sample, the
306 viscosity of the pore fluid ($\mu=10^{-3}$ Pa.s at 20°C), the permeability (k in m^2), and the
307 compressive storages of the upstream and downstream pore pressure circuits ($C_u =$
308 3.957×10^{-9} m^3/MPa and $C_d = 4.828 \times 10^{-9}$ m^3/MPa , respectively).

309 *Permeability from mercury injection capillary pressure (MICP)*

310 Pore throats size distribution measured by MICP was used to determine Katz-
311 Thompson permeability K_{KT} (Katz and Thompson, 1986, 1987; Nishiyama and
312 Yokoyama, 2014) with Equation (16):

$$313 \quad K_{KT} = \frac{1}{89} \phi_i \frac{(l_i^h)^3}{l_c} f(l_{max}^h) \quad (16)$$

314 with the pore throat diameter (l_c) at the inflection point of the cumulative MICP curve
315 $f(l)$ of fractional volume of connected pore throats with diameters of l and larger, the
316 pore throat diameter (l_{max}^h) corresponding to the optimum path for permeability when
317 $l^3 f(l)$ is maximum and $l = l_{max}^h$ (Katz and Thompson, 1986, 1987).

318 **Results**

319 ***Site U1517***

320 At Site U1517, CEC is very low to low (~0.05–0.20 mol/kg), independent of
321 lithological units, and averages 0.15 mol/kg (Figure 4c). The range of CEC values
322 suggests that the relatively homogeneous (~32%) total clay mineral content across the
323 section (Figure 4b) is composed of a mix of smectite, illite, kaolinite and or/chlorite. In
324 particular, the comparison of CEC values measured in siliciclastic clay-rich samples
325 from the North Sumatra margin (Duttilleul et al., 2020a) and other sites from the North
326 Hikurangi margin (e.g. Duttilleul et al., 2020b and in press) with the clay mineral
327 assemblage determined by XRD in near-by samples (e.g. Rosenberger et al., 2020;
328 Underwood, 2021) suggests that at Site U1517, smectite content in bulk sediment
329 ranges ~5%-30% (Figure 8). These values agree with the post-cruise XRD
330 characterisation of the clay mineral assemblage of Underwood and Dugan (2021). This
331 study shows that smectite and illite contents in bulk sediment range ~5%-29% and

332 ~5%-21% respectively while chlorite and kaolinite contents are <5% and <1%
333 respectively (S. Cardona, pers. comm.)

334 Consequently, bound water content is low (~2%–8%, ~6% on average) and
335 relatively homogeneous across the sedimentary section at Site U1517. Hence, total
336 connected porosity and interstitial porosity show similar variation across the section
337 drilled, in this case, an overall decrease with increasing depth (Figure 4d). However, no
338 general compaction trend can be defined at Site U1517. In detail, interstitial porosity
339 exponentially decreases from ~64% at the seafloor to ~34% at the base of the South
340 TLC. The interstitial porosity versus vertical effective stress (σ'_v) curve determined by
341 least-square fitting is $\phi_i = 45.6 e^{-0.57\sigma'_v}$ ($R^2=0.40$, as shown in Figure 7) for the South TLC
342 interval. A strong contrast appears with the top of lithological Unit IV where interstitial
343 porosity is ~45% and decreases with increasing scatter following $\phi_i = 47.6 e^{-0.14\sigma'_v}$ (least-
344 square fitting with $R^2=0.23$) down to ~40% at ~185 mbsf. The interstitial porosity tracks
345 the LWD NMR porosity very well but it does not track the LWD neutron porosity
346 which shows higher values (Figure 4d). Larger neutron porosity values are typically
347 detected in clay-rich sediments (e.g. Expedition 314 Scientists, 2009; Expedition 317
348 Scientists, 2011; Expedition 334 Scientists, 2012a, b; Dutilleul et al., 2020b) due to the
349 high concentration of hydrogen characterising the hydroxyls of the mineral structure of
350 the clay, with a possible additional effect of trace elements acting as thermal absorbers
351 (Ellis, 1986; Ellis and Singer, 2007). NMR porosity measured on samples generally
352 matches LWD NMR porosity, although the latter T_2 distribution is truncated at values
353 between 1 and 1000 ms (Barnes et al., 2019b) in contrast to laboratory measurements.
354 An exception is Unit IV, where NMR porosity tracks LWD neutron porosity.

355 Overall, resistivity-derived porosity fits interstitial porosity with $a = 1$ and $m = 1.7$
356 in the TLC and $m = 1.9$ below (Figure 4i) except in some intervals (~20–30, ~96–106,
357 ~135–165 mbsf) as discussed below.

358 Figure 9 shows the permeability data obtained on cores 17H6 and 29F3 under a
359 constant pore fluid pressure of 8 and 8.5 MPa respectively corresponding to their in-situ
360 pore pressure conditions and an increasing effective hydrostatic confining pressure from
361 1 to 20 MPa. The lowest effective pressure is slightly higher than the in-situ pressure for
362 sample 17H6 (0.87 MPa) and slightly lower for sample 29F3 (1.39 MPa). The
363 progressive closure of cracks can explain the decrease of permeability k with effective
364 pressure P_{eff} . Walsh (1981) proposed a fracture closure model that suggests that $k^{1/3}$
365 should be linearly related to $\log P_{\text{eff}}$. Permeability plotted as $k^{1/3}$ is shown as a function
366 of P_{eff} in a linear-log graph in Figure 10 for samples 17H6 and 29F3. The data are found
367 to follow a linear trend, in agreement with Walsh's model. They correspond to the over-
368 consolidation of the samples with an initially measured permeability ($1.3 \cdot 10^{-18} \text{ m}^2$ for
369 core 17H6 and $1.4 \cdot 10^{-17} \text{ m}^2$ for core 29F3) close to the expected in-situ permeability.

370 MICP shows that the average size of pore throats decreases and is homogenised
371 with increasing depth from $0.3 \text{ }\mu\text{m}$ to $0.1 \text{ }\mu\text{m}$ in lithological Units I to IV (Figure 4e).
372 Lithological Unit V shows contrasted ranges of pore throat diameters with a general
373 trend toward higher values, increasing up to $\sim 1.4 \text{ }\mu\text{m}$ at ~ 165 mbsf, and dropping down
374 to $\sim 0.4 \text{ }\mu\text{m}$ in the lower part of the unit (Figures 4e and 11). The corresponding
375 permeabilities determined using the Katz-Thompson model are in the order of 10^{-17} m^2 .
376 These estimated permeability values are somewhat higher than those measured under
377 in-situ conditions (see, for example, core 17H6). The effective pressure effect may
378 explain the difference since the Katz-Thompson estimate holds for zero pressure
379 conditions and permeability has been shown to decrease at pressures lower than our

380 minimum of 1 MPa (Screaton et al., 2021). Another possible explanation may be the
381 lower water permeability in clay- and silt-rich rocks compared to gas permeability, due
382 to fluid-rock interactions that the Katz-Thompson model does not consider.

383 NMR T_2 measured on samples and LWD NMR T_2 are both well correlated to the
384 average size of pore throats but show contrasting values (Figure 4f). NMR T_2 measured
385 on samples are in the range of ~1–2 ms. In contrast, LWD NMR T_2 decreases from ~20
386 to ~6 ms from the seafloor to the bottom of Unit IV. It increases up to ~30 ms before
387 dropping in the lower part of Unit V. The average LWD relaxivity is ~9 $\mu\text{m/s}$ while the
388 average relaxivity measured on samples is ~50 $\mu\text{m/s}$.

389 ***Site U1519***

390 At Site U1519, CEC is low but in the high range measured at Site U1517
391 (~0.14–0.20 mol/kg). CEC is also independent of lithological units and averages 0.16
392 mol/kg (Figure 5c). This range of values suggests that the clay mineral assemblage,
393 which constitutes ~45% of the bulk mineralogy of sediments, is homogeneous (Figure
394 5b). Similarly to Site U1517, the CEC values suggest that the clay fraction is composed
395 of a mix of smectite (~5%-30% based on Figure 8), illite, kaolinite, and or/chlorite
396 taking into account that CEC and XRD measurements were not carried out on the same
397 samples, resulting in significant scattering ($\pm 15\%$). This result cannot be confirmed
398 since no characterisation of the clay mineral assemblage was undertaken at Site U1519.

399 The resulting bound water content is low (~5%–8%) and relatively
400 homogeneous across the section, averaging 7%. Hence, as observed at Site U1517, total
401 connected porosity and interstitial porosity vary similarly (Figure 5d). Across the
402 limited intervals where coring was attempted, the interstitial porosity decreases with
403 increasing depth from 68% below the seafloor to ~33% at the bottom of Unit I, it shows
404 slightly lower values at the bottom of logging and seismic Units 2 (~31%) before

405 increasing across logging and seismic Units 3 up to ~ 35% with a maximum ~42%
406 around 605 mbsf. In the first unit, strong porosity contrast occurs based on grain size,
407 with a general trend toward higher porosity values (+5% in average) in mudstones than
408 in shallower clayey siltstones/silty claystones. In contrast to Site U1517, the interstitial
409 porosity at Site U1519 is in better agreement with the LWD neutron porosity than with
410 the NMR porosity, which shows lower values. The NMR porosity measured on samples
411 exceeds the LWD NMR porosity and matches both interstitial and LWD neutron
412 porosities.

413 The resistivity-derived porosity fits the interstitial and onboard interstitial
414 porosities with $a = 1$ and $m = 1.9$ in lithological Unit I, $m = 2.3$ in the lower part of
415 seismic and logging Units 2 (top of lithological Unit II), and $m = 2.6$ in the top of
416 logging and seismic Units 3 (bottom of Unit II) except in a zone ~600–620 mbsf (Figure
417 5i). A general compaction trend $\phi_i = 39.3 e^{-0.03\sigma'_v}$ ($R^2 = 0.16$) can be defined at Site U1519
418 (excluding the washout interval ~148-227 mbsf) based on fitting the resistivity-derived
419 porosity to the onboard interstitial porosity. In detail, three different compaction trends
420 can be defined to best fit the data (Figure 7): the first one for clayey siltstones/silty
421 claystones in lithological Unit I, the second one for mudstones of the same unit, and the
422 third one for Unit II. We notice that the compaction trend for shallow clayey
423 siltstones/silty claystones in Unit I at Site U1519 determined from the onboard
424 interstitial porosity is equivalent to that of the TLC at Site U1517 where the same type
425 of lithology or even coarser lithologies have been identified.

426 MICP shows that the average size of pore throat diameters decreases and
427 becomes more homogeneous with increasing depth from ~0.45 μm to ~0.15 μm near the
428 bottom of the hole (Figures 5e and 11). As observed at Site U1517, the permeabilities
429 determined using the Katz-Thompson model are in the order of 10^{-17} m^2 . NMR T_2

430 measured on samples and LWD NMR T_2 show diverging values (~ 10 – 2.30 ms and
431 ~ 19 – 3 ms, respectively), although their evolution correlates with that of the average
432 pore throat size measured by MICP (Figure 5e; f). The average LWD relaxivity is ~ 14
433 $\mu\text{m/s}$ while the average relaxivity measured on samples is ~ 90 $\mu\text{m/s}$.

434 **Discussion**

435 *Physical properties of gas hydrates bearing intervals*

436 When not directly observed onboard, gas hydrate occurrence (green rectangle
437 and shaded zones in Figure 4 and green rectangle in Figure 5) was inferred from 1)
438 negative anomaly of pore water content (Figures 4j et 5j) due to gas hydrate dissociation
439 during core retrieval (Paull and Ussler, 2001), or 2) positive peaks of LWD P-wave
440 velocity (Figures 4g and 5g) and resistivity (Figures 4h and 5h) (Barnes et al., 2019b).
441 However, there are uncertainties regarding the precise location of gas hydrate-bearing
442 intervals estimated from the salinity detection method. Indeed, other phenomena,
443 including changes in the salinity of the water mass, diagenetic reactions or the pore
444 water extraction method used onboard may also cause negative chloride anomalies
445 (Fitts and Brown, 1999; Barnes et al., 2019b, c; Sultan, 2020). At Site U1517, chloride
446 data mark a zone in the preslide interval among ~ 135 and ~ 165 mbsf with gas hydrate
447 saturation ranging 2%–68% (Barnes et al., 2019b), with the BGHS at ~ 162 mbsf
448 (Screaton et al., 2019) (Figure 4). Sultan (2020) reinterpreted chloride anomalies and
449 proposed a BGHS between 85 and 128 mbsf, although this is not supported by Screaton
450 et al. (2019, 2020). LWD data evidenced three intervals potentially hosting gas hydrates
451 (green shaded zones on Figure 4) between ~ 112 – 114 , ~ 117 – 121 and ~ 128 – 145 mbsf
452 (Barnes et al., 2019b). At Site U1519, chloride data mark a damage zone (Figure 5k)
453 between 520 and 630 mbsf with elevated gas hydrate concentration with a discrete

454 negative anomaly ~628 mbsf (Figure 5j) where the occurrence of gas hydrates has been
455 directly evidenced by core IR scanning in silt layers with higher CEC at ~627 mbsf.

456 Gas hydrate distribution depends on the initial characteristics of the porous
457 medium and the effective stress that control the growing space, thus porosity and
458 permeability (Dai et al., 2012, He et al., 2013) which vary with hydrate formation and
459 dissociation (Misyura, 2016; Wang, 2020). When hydrates form in the pore space, the
460 permeability of gas hydrate saturated sediments decreases (Nimblett and Ruppel, 2003;
461 Dai and Seol, 2014; Daigle et al., 2015b; Wei et al., 2021) while it increases when
462 hydrates dissociate (Wang, 2020). Sediments characterised by a high initial porosity are
463 more sensitive to permeability changes associated with increasing gas hydrates
464 saturation (Shen et al., 2020).

465 Although there are uncertainties concerning the precise location of gas hydrate-
466 bearing intervals (e.g. Barnes et al., 2019b, c; Sreaton et al., 2019 and 2020; Sultan,
467 2020), our core-based dataset shows evidence that the sediment matrix in intervals
468 likely to host gas hydrates, as inferred from chloride anomalies, is characterised by a
469 general trend of higher interstitial porosities, average pore diameters, and Katz-
470 Thompson permeabilities (Figures 4 and 5). These characteristics are favorable to the
471 growth of gas hydrates. This behaviour is more visible at Site U1517 (~135-160 mbsf)
472 than at Site U1519 (~520-630 mbsf). On LWD data, intervals hosting gas hydrates are
473 typically characterised by high electrical resistivity (Goldberg et al., 2010) and P-wave
474 velocity (Waite et al., 2009) values and apparent lower NMR porosity (Kleinberg,
475 2003). Similar values, for zones thought to host gas hydrates, can be observed at Site
476 U1517 in the intervals between ~112–114, ~117–121 and ~128–145 mbsf or in some
477 intervals at the meter scale between ~135 and 160 mbsf at Site U1517 (Figure 4h) and
478 ~520-630 mbsf at Site U1519 (Figure 5h).

479 ***Overview of the compaction state and creep mechanisms at the South TLC***

480 Several zones at both sites show anomalously high values of resistivity-derived
481 porosity relative to the compaction curve or the onboard interstitial porosity. Two
482 intervals at Site U1517 (~20–30 mbsf and ~96–106 mbsf) and one at Site U1519 (~148–
483 227 mbsf) represented by the grey shaded zones in Figures 4 and 5 with anomalously
484 high resistivity-derived porosity correspond to washout zones (Barnes et al., 2019b, c).
485 Peaks of high resistivity-derived porosity related to the occurrence of gas hydrates in the
486 pore space also occur in the zone ~135-165 mbsf at Site U1517. In this section, the
487 rapid dissociation of gas hydrates when cores are brought to the surface results in the
488 freshening of the pore water extracted from cores, impacting resistivity-derived porosity
489 (Barnes et al., 2019b) (Figures 4j and 5j). These zones excluded, interstitial porosity or
490 resistivity-derived porosity can be examined to determine compaction state and discuss
491 mechanisms behind creeping at the South TLC.

492 At Site U1517, interstitial porosity versus effective vertical stress curve (referred
493 to as compaction trend) and cementation factor m vary across the TLC (~0-67 mbsf)
494 and in the preslide interval beneath (Figure 7). In particular, although Units II and III
495 (considered here as corresponding to the upper and lower units of the landslide) may not
496 be genetically linked and show variations in lithology and emplacement mechanisms
497 (Gross et al., 2018; Couvin et al., 2020; Luo et al., 2020), their porosity values plot on
498 the same compaction trend (Figure 7). This compaction trend is similar to that of the
499 upper sedimentary section at Site U1519, at Site U1518 nearby the deformation front
500 (Dutilleul et al., in press), and at Site U1520 in the Hikurangi Trough (Dutilleul et al.,
501 2020b). This suggests that mechanical compaction of shallow poorly consolidated
502 sediments is relatively uniform across the margin, although siliciclastic lithology may
503 slightly vary in terms of grain size between these different locations. The ‘negative

504 polarity intra-debris reflector' at the interface between Units II and III identified by
505 Gross et al. (2018) (~37 mbsf, marker D in Figure 4) is not associated with an
506 anomalously high interstitial porosity nor resistivity-derived porosity. Such anomaly
507 was expected since this reflector has been inferred to represent an excess pore pressure
508 zone (Gross et al., 2018) coinciding with the décollement for active creeping (Barnes et
509 al., 2019a, b).

510 However, onboard interstitial porosity increases by 11% and resistivity-derived
511 porosity by 7% at the base of the landslide at Site U1517 (~67 mbsf, marker B in Figure
512 4). Three hypotheses can explain the distinct porosity patterns observed in the landslide
513 and the preslide intervals with strong porosity shift at the interface:

- 514 • Hypothesis 1: a local excess pore pressure.
- 515 • Hypothesis 2: differences in lithology, emplacement mechanisms, and
516 deformation history result in distinct compaction trends in the landslide and
517 preslide intervals.
- 518 • Hypothesis 3: a combination of hypotheses 1 and 2.

519 Hypothesis 1 is supported by the rapid burial of sediment due to the emplacement of the
520 landslide (Screaton et al., 2019), which is likely to release fluids through rapid
521 compaction of the preslide interval. Although our dataset does not clearly show a
522 contrast of permeability at the base of the landslide, the clay-rich upper part of the
523 preslide interval is prone to the accumulation of the released fluids, thus to excess pore
524 pressure development which may also be triggered by focused fluid flow (Carey et al.,
525 2019). On the other hand, the lithology contrast between the landslide, characterised by
526 numerous sand and silt-rich turbidites, and the clay-rich upper part of the preslide
527 interval supports hypothesis 2 (Gamage et al., 2011). Hypothesis 2 is also supported by
528 the strong differences in emplacement mechanisms and thus the deformation history of

529 the preslide interval characterised by a mixed slope sedimentary succession (Barnes et
530 al., 2019b) and the landslide interval. As a consequence of the quick deposition of the
531 landslide, the preslide interval should be more consolidated than the lower landslide
532 unit. In fact, it is less consolidated, which supports hypothesis 1. Hence, we suggest that
533 hypothesis 3 is the most plausible.

534 IODP Expedition 372 drilling results and Sreaton et al. (2019) have shown that gas
535 hydrates are not involved in active creeping at the South TLC interval where they are
536 absent. In the preslide interval where gas hydrates are expected to occur, the lack of
537 excess pore pressure at the BGHS, evidenced by our dataset, shows that neither the
538 hydrate dissociation model nor the hydrate pressure valve model, linking gas hydrates to
539 slow deformation, is possible. The results of this study allow the examination of
540 mechanisms, which are not related to gas hydrates, but may explain active creeping in
541 the South TLC interval. Our findings support the repeated small-scale failure
542 mechanism associated with cycles of excess pore pressure as initially proposed by
543 Mountjoy et al. (2009), since no (hypothesis 2) or moderate (hypotheses 1 and 3) excess
544 pore pressure is found at Site U1517 in the extensional zone of the South TLC. This
545 mechanism could be validated in the case that a larger excess pore pressure is evidenced
546 in the compressional zone of the landslide, which has not yet been drilled (Barnes et al.,
547 2019a).

548 The sedimentary section at Site U1519 is in hydrostatic conditions. However, ~600
549 mbsf (~5.9 MPa in Figure 7), there is a zone where onboard interstitial porosity is 4–
550 11% higher than the value predicted by the compaction curve of Unit II. This
551 discrepancy of porosity values could be related to an excess pore pressure in the order
552 of ~1.5–5 MPa. The origin of this potential excess pore pressure zone is unclear. It
553 could be related to concentrated fluid flow in a high permeability sand-rich interval

554 sandwiched between lower permeability MTDs. Another possibility is a relation
555 between the presence of free gas and the nearby location of the BGHS. However, the
556 latter is not well constrained with, on the one hand, a possible BSR ~560 mbsf or below,
557 and on the other hand, gas hydrates sampled in a core at ~627 mbsf (Pecher et al.,
558 2019c).

559 **Conclusions**

560 We quantify interstitial porosity at the south (Site U1517) and the north (Site
561 U1519) TLC by correcting the total connected porosity measured on core samples for
562 bound water content determined from cation exchange capacity. At Site U1517,
563 interstitial porosity decreases by ~30% across the TLC (~64–34%), following a
564 compaction trend different from that of the preslide interval beneath. This porosity
565 decrease is associated with a reduction of pore throats from ~0.3 to ~0.1 μm . At Site
566 U1519 where coring was limited, we use the interstitial porosity to calibrate the high-
567 resolution resistivity-derived porosity. It decreases from ~68% to ~42% following three
568 different compaction trends, corresponding to a reduction in the average size of pore
569 throats from ~0.45 to ~0.15 μm across the sedimentary section. At both sites, the in-situ
570 permeability measured with a triaxial cell and deduced from pore size measurements
571 ranges from 10^{-18} to 10^{-17} m^2 . The dataset shows that the sedimentary intervals suspected
572 to host gas hydrates ~112–114, ~117–121 and ~128–145 and ~135 and 160 mbsf at Site
573 U1517 and between ~520 and 630 mbsf at Site U1519 exhibit higher interstitial
574 porosity, pore diameter and permeability favorable to the growth of gas hydrates. These
575 results suggest hydrostatic conditions across most of the two sedimentary sections.
576 However, at Site U1519, excess pore pressure may occur in a zone below the base of the
577 gas hydrate stability zone. At Site U1517, a porosity shift of ~+10% occurs at the base
578 of the landslide. We suggest that this substantial difference in the porosity profile

579 between the landslide and preslide intervals results from a combination of 1) contrasted
580 physical properties due to differences in lithology, emplacement mechanisms and
581 compaction history and 2) moderate excess pore pressure associated with the trapping of
582 fluids at the base of the landslide. The results of IODP Expedition 372 discredit the role
583 of gas hydrates in active creeping at the TLC. Our results open the way to the study of
584 other mechanisms at the TLC, such as potentially repeated small-scale failure associated
585 with charge and discharge of excess pore pressure, or at analogous basins where
586 submarine landslides and gas hydrates coexist. Future work, such as in situ pore
587 pressure analysis recorded at Site U1519, is required to understand the mechanisms of
588 creep at the TLC.

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599 **Data**

600 This research used data provided by the International Ocean Discovery Program
601 (IODP) and freely available on the LIMS Report Interface Page at
602 web.iodp.tamu.edu/LORE or on the log database at

603 mlp.ldeo.columbia.edu/logdb/scientific_ocean_drilling. Post-cruise data including
604 corrected porosity, CEC, exchangeable cation composition, MICP and NMR are
605 available in the OTELo Research Data Repository ([https://doi.org/10.24396/ORDAR-](https://doi.org/10.24396/ORDAR-40)
606 [40](https://doi.org/10.24396/ORDAR-40) for Site U1519 and <https://doi.org/10.24396/ORDAR-41> for Site U1517).

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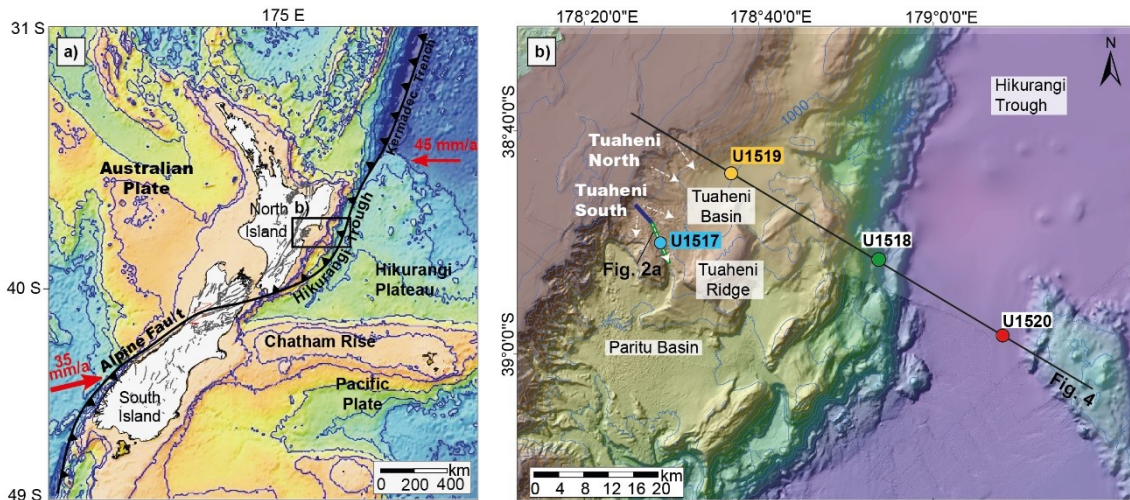
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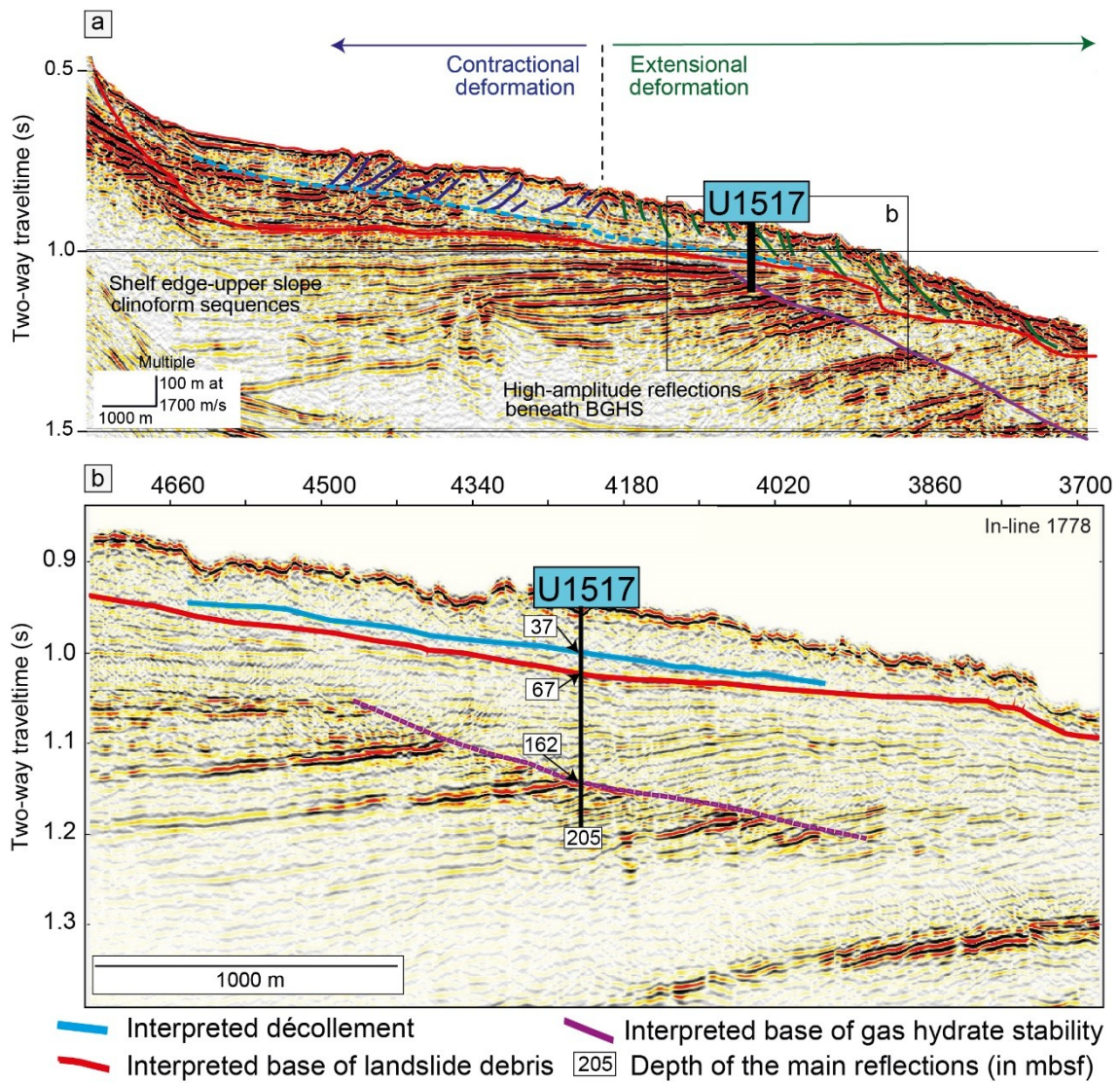
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1006

1007 **Figures**



1008
1009 Figure 1. a) Tectonic setting of the Hikurangi margin with plate motion indicated by red
1010 arrows (modified after Barnes et al., 2019a). b) Bathymetric map of the IODP
1011 Expeditions 372/375 study area offshore Gisborne located on a) (modified after
1012 Mountjoy et al., 2014 and Saffer et al., 2019). The thick line represents the seismic
1013 profile TAN1114-10b shown in Figure 2a with the location of Site U1517. The blue
1014 (respectively, green) section of this line indicates the contractional (respectively,
1015 extensional) zone of the slide represented in Figure 2a. The black line represents the
1016 seismic profile 05CM-04 across the margin with the location of Site U1519 and of other
1017 Sites from IODP Expeditions 372-375. White dashed arrows display the direction of
1018 movement within the landslides of the North and South Tuaheni (after Mountjoy et al.,
1019 2014).



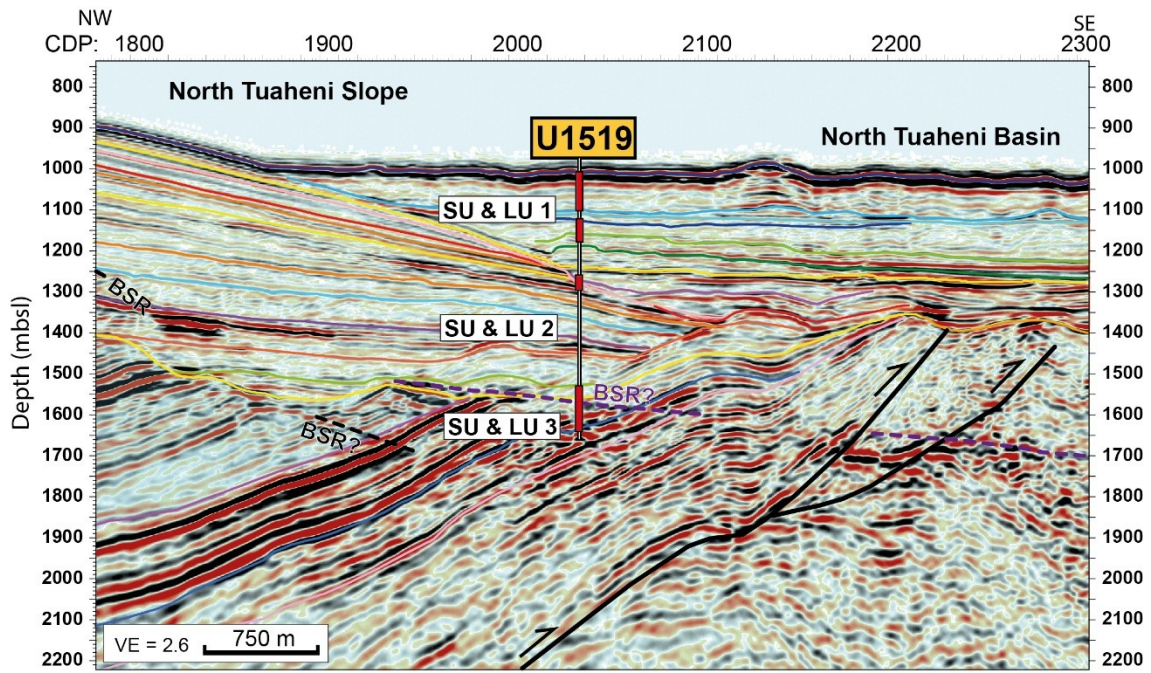
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1021 Figure 2. a) Seismic profile TAN1114-10b across the South Tuaheni landslide
 1022 (modified after Mountjoy et al., 2014 and Barnes et al., 2019a) with b) a zoom in on
 1023 Site U1517 in the extensional part and the main seismic reflectors (in-line 1778), with
 1024 the depth in metres below sea floor (modified after Barnes et al., 2019a). BGHS: base of
 1025 the gas hydrate stability zone.

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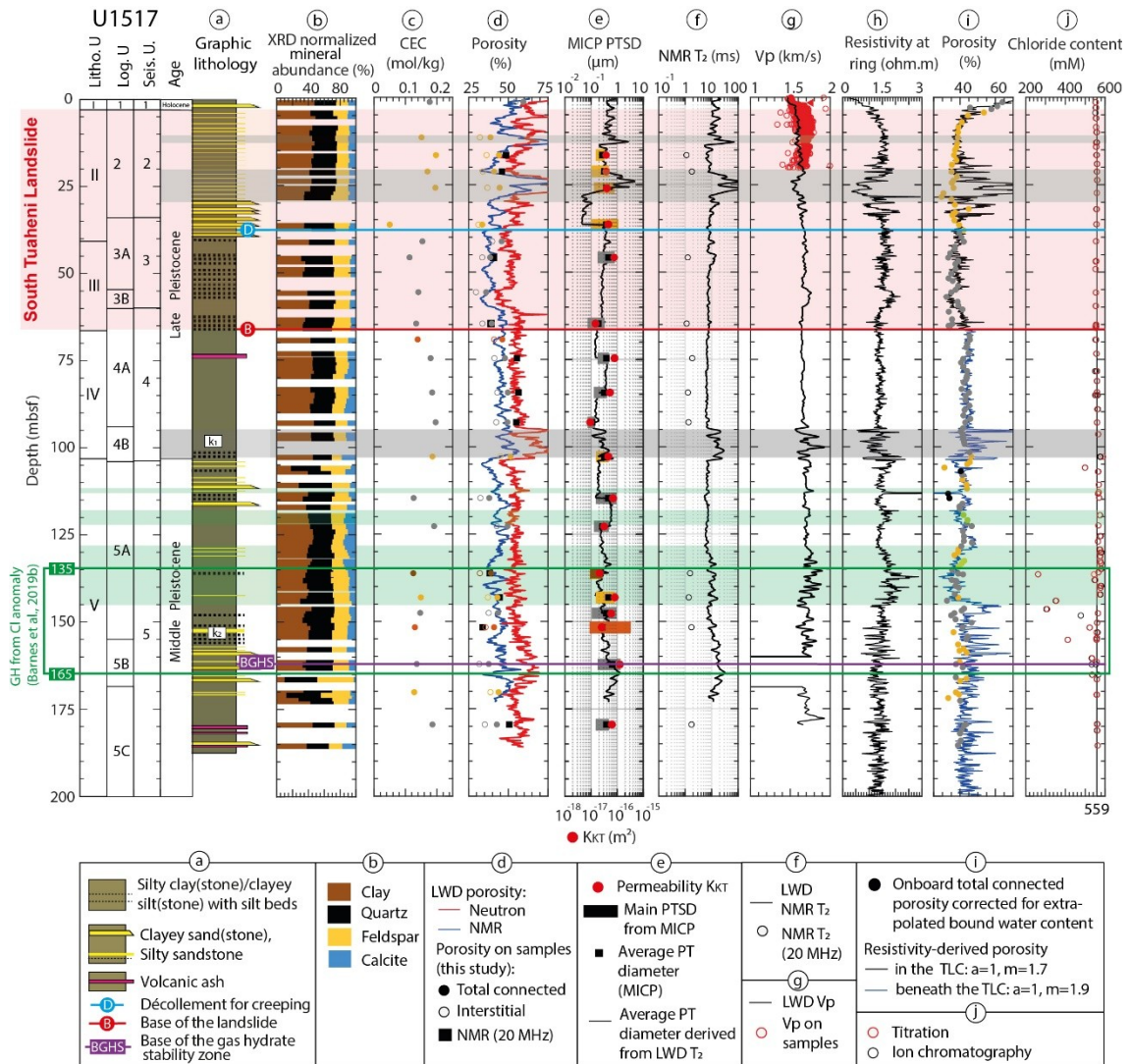
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1030 Figure 3. Zoom in on seismic profile 05CM-04 (Figure 1b) on Site U1519 showing the

1031 cored sections (red bars), logging units (LU) and seismic units (SU) (modified after

1032 Barnes et al., 2019c). VE: vertical exaggeration; BSR: bottom-simulating-reflector.

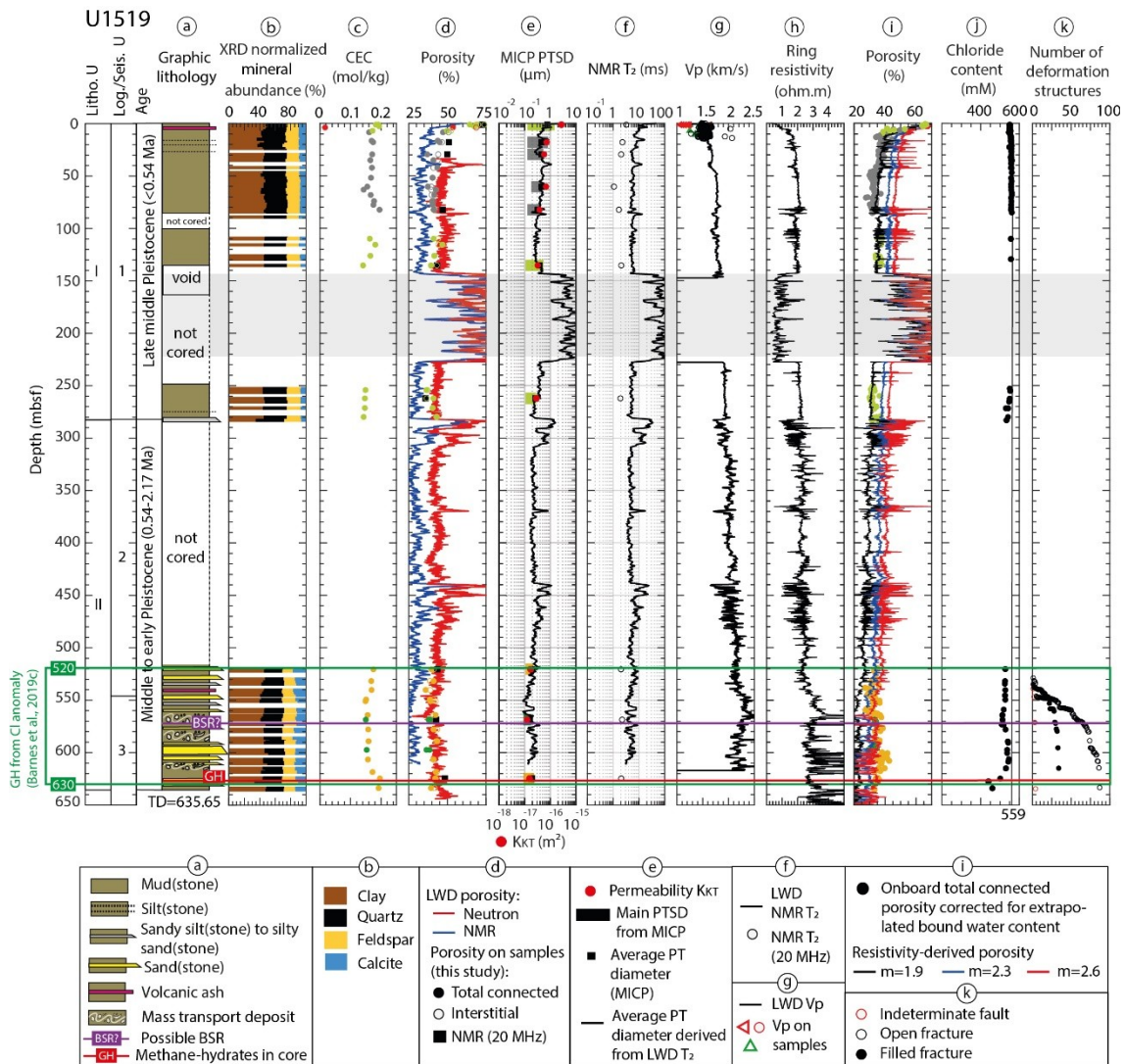
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1035 Figure 4. Summary of lithological, logging and seismic units and description (a),
 1036 mineralogical (b), chemical (c and j) and physical (d, e, f, g, h, i) properties at Site
 1037 U1517. For (c), (d), (e) and (i), the colour of data points or bars indicates the lithology
 1038 of the samples analysed: light green for clay(stone); black for silt(stone); gold for
 1039 alternating sand(stone) and mud(stone) layers; grey for silty clay(stone), clayey
 1040 silt(stone) or alternating silt and clay layers; orange for clayey silt(stone) with sand;
 1041 brown for silt(stone) with sand. k_1 and k_2 indicate the location of cores 17H6 and 29F3
 1042 respectively for permeability measurements. XRD: X-Ray diffraction; CEC: Cation
 1043 exchange capacity; Kkt: Katz–Thompson permeability; MICP PT(SD): mercury
 1044 injection capillary pressure pore throat (size distribution) determined based on Bertrand

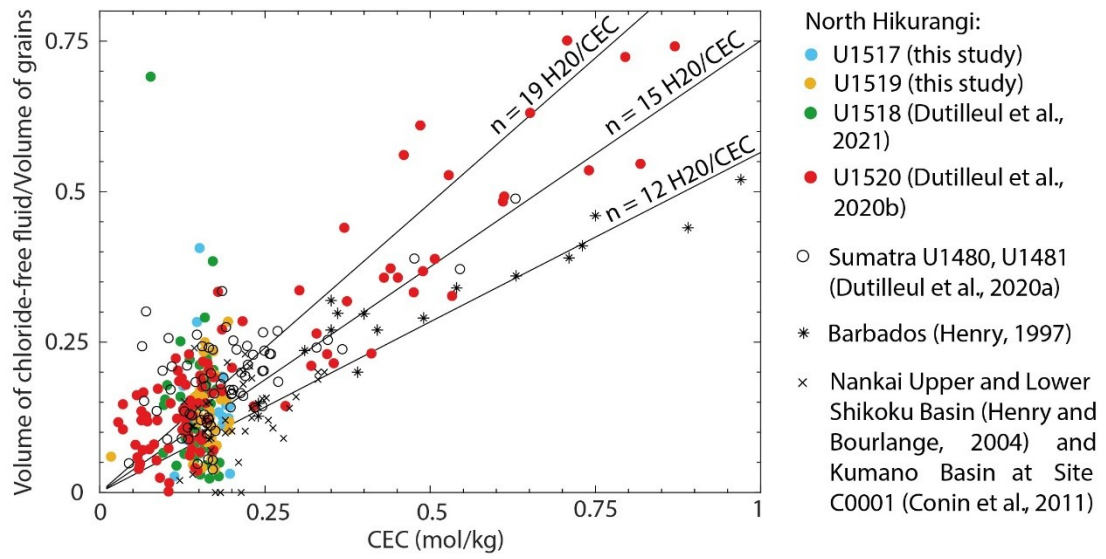
1045 et al. (2021); NMR: nuclear magnetic resonance; T_2 : transverse relaxation time; LWD:
1046 logging-while-drilling; V_p : P-wave velocity; TLC: Tuaheni Landslide Complex. The
1047 red shaded zone represents the landslide mass with marker horizons D and B. The three
1048 green shaded zones highlight zones expected to host gas hydrates based on LWD data.
1049 The green rectangle indicates the zone hosting gas hydrates from chloride anomalies
1050 (Barnes et al., 2019b). In j), 559 mM is the average chloride concentration in modern
1051 seawater. The base of the gas hydrate stability zone (BGHS) is located at 162 mbsf
1052 (Screaton et al., 2019). The grey shaded zones are interpreted onboard as washout zones
1053 between ~10-13 mbsf, ~21-28 mbsf and ~94-104 mbsf (Barnes et al., 2019a).
1054



1055

1056 Figure 5. Summary of lithological, logging and seismic units and description (a),
 1057 mineralogical (b), chemical (c and j), physical (d, e, f, g, h, i) and structural (k)
 1058 properties at Site U1519. For (c), (d), (e) and (i), the colour of data points or bars
 1059 indicates the lithology of the samples analysed: light green for clay(stone); green for
 1060 contorted mud(stone); grey for silty clay(stone); red for ash. XRD: X-Ray diffraction;
 1061 CEC: cation exchange capacity; Kkt: Katz–Thompson permeability; MICP PT(SD):
 1062 mercury injection capillary pressure pore throat (size distribution) determined based on
 1063 Bertrand et al. (2021); NMR: nuclear magnetic resonance; T_2 : transverse relaxation
 1064 time; LWD: logging-while-drilling; Vp: P-wave velocity. The green rectangle
 1065 highlights a zone of gas hydrate occurrence inferred from chloride data. The grey

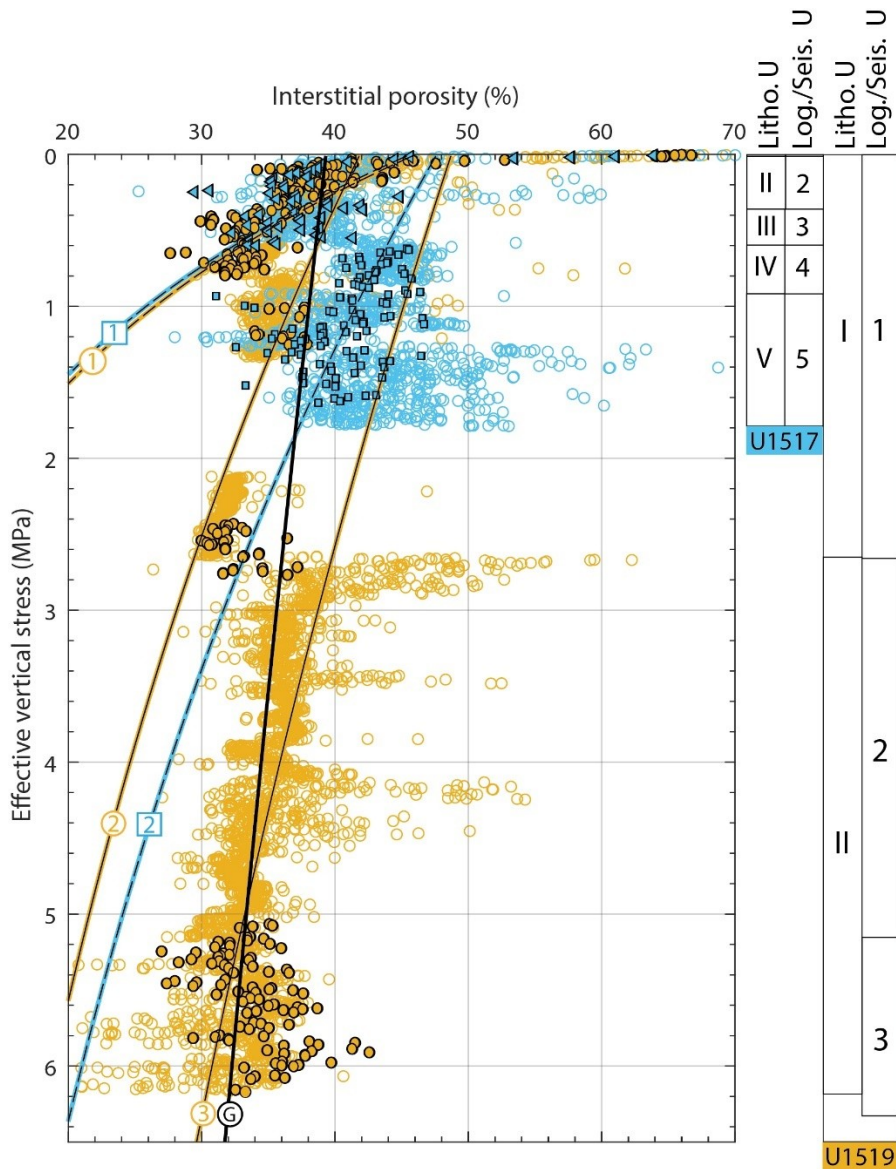
1066 shaded zone represents a washout interval ~143-227 mbsf identified onboard (Barnes et
1067 al., 2019c). In i), the average chloride concentration in modern seawater is 559 mM. A
1068 possible bottom-simulating-reflector (BSR) is located at 560 mbsf.
1069



1070

1071 Figure 6. Bound water ratio versus Cation Exchange Capacity (CEC) at north Hikurangi
 1072 margin including Site U1517 and U1519, Site U1520 (reference site across the
 1073 undeformed entering sedimentary section) and U1518 (active thrust at the deformation
 1074 front) and at other subduction zones from previous works.

1075



Core data: Onboard total connected porosity corrected from extrapolated bound water content at:

Site U1517:

▲ Lithological Units I-III, across the TLC

■ Lithological Units IV & V, below the TLC

● Site U1519

Resistivity-derived porosity (washout intervals excluded):

○ Site U1517

○ Site U1519

Compaction trends:

— from resistivity-derived porosity

- - from core data

U1517:

1 $\Phi_i = 45.6 \exp(-0.57 \sigma_v')$; $R^2 = 0.40$

2 $\Phi_i = 47.6 \exp(-0.14 \sigma_v')$; $R^2 = 0.23$

U1519:

1 $\Phi_i = 46.1 \exp(-0.55 \sigma_v')$; $R^2 = 0.45$

2 $\Phi_i = 41.88 \exp(-0.13 \sigma_v')$; $R^2 = 0.37$

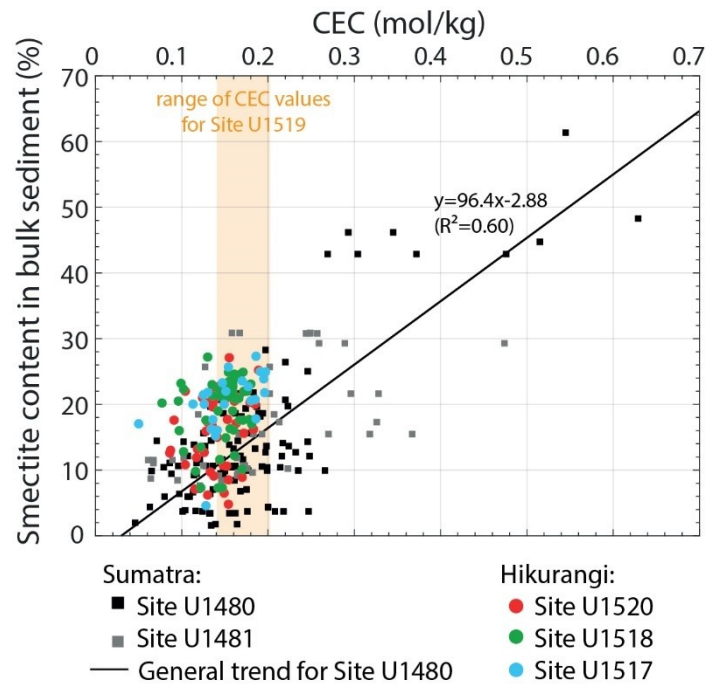
3 $\Phi_i = 48.7 \exp(-0.08 \sigma_v')$; $R^2 = 0.34$

ⓐ $\Phi_i = 39.3 \exp(-0.03 \sigma_v')$; $R^2 = 0.16$

1076

1077 Figure 7. Summary of ring resistivity-derived porosity (open circles) calibrated to
 1078 onboard interstitial porosity (i.e. total connected porosity corrected for extrapolated
 1079 bound water content) (triangles, squares or circles) and compaction curves at Site
 1080 U1517 (light blue) and U1519 (gold). Compaction curves determined from resistivity-

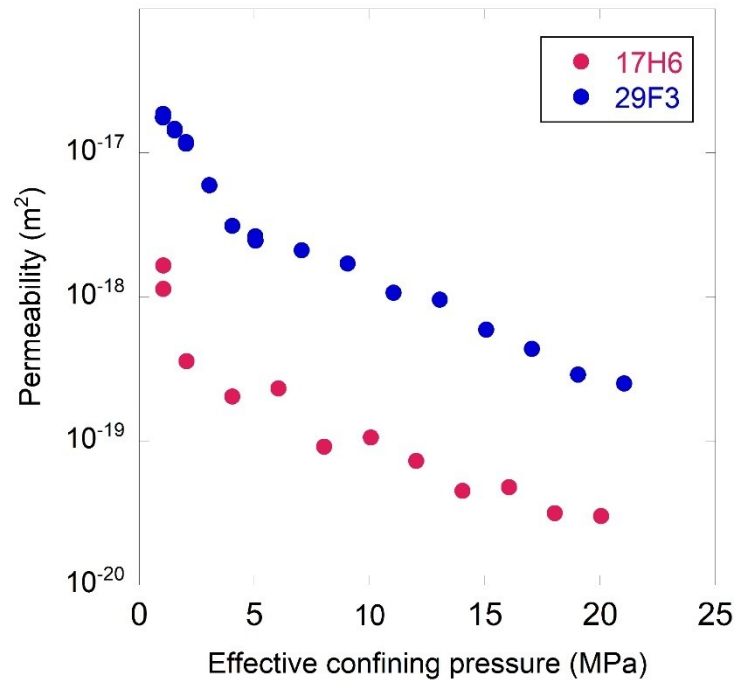
1081 derived porosity and ‘onboard interstitial porosity’ are shown with solid lines and
1082 dashed lines respectively. The general compaction curve (trend G) at Site U1519
1083 (excluding the washout interval ~148-227 mbsf) determined from resistivity-derived
1084 porosity is $\phi_i = 39.3 e^{-0.03\sigma_v}$ ($R^2=0.16$). At Site U1519, distinct compaction trends are
1085 given (gold line numbered in open circles): trend 1) for clayey siltstones/silty claystones
1086 in lithological Unit I, trend 2) for mudstones of the same unit and trend 3) for Unit II. At
1087 Site U1517 (blue line numbered in open squares), two compaction trends were
1088 determined from onboard interstitial porosity: trend 1) is the trend for the South TLC
1089 interval and trend 2) is valid for the preslide interval.
1090



1091

1092 Figure 8. Relation between smectite content in bulk sediment determined by X-ray
 1093 diffraction and cation exchange capacity (CEC) values at different IODP Sites in north
 1094 Sumatra margin and north Hikurangi margin. Smectite content in bulk sediments and
 1095 CEC are measured on near-by but different samples. Smectite content in bulk sediments
 1096 for Sumatra Sites U1480 and U1481 are calculated from the XRD analysis of the clay
 1097 fraction of Rosenberger et al. (2020) and the total clay mineral content of the nearest
 1098 sample given in McNeill et al. (2017a and b). Smectite content in bulk sediments for
 1099 Hikurangi Sites U1520 and U1518 are from Underwood (2021) and are from
 1100 Underwood and Dugan (2021) and S. Cardona (pers. comm.) for Site U1517. CEC data
 1101 for Sumatra Sites are from Dutilleul et al. (2020a). CEC data for Hikurangi Sites U1520
 1102 and U1518 are from Dutilleul et al. (2020b) and (in press) respectively.

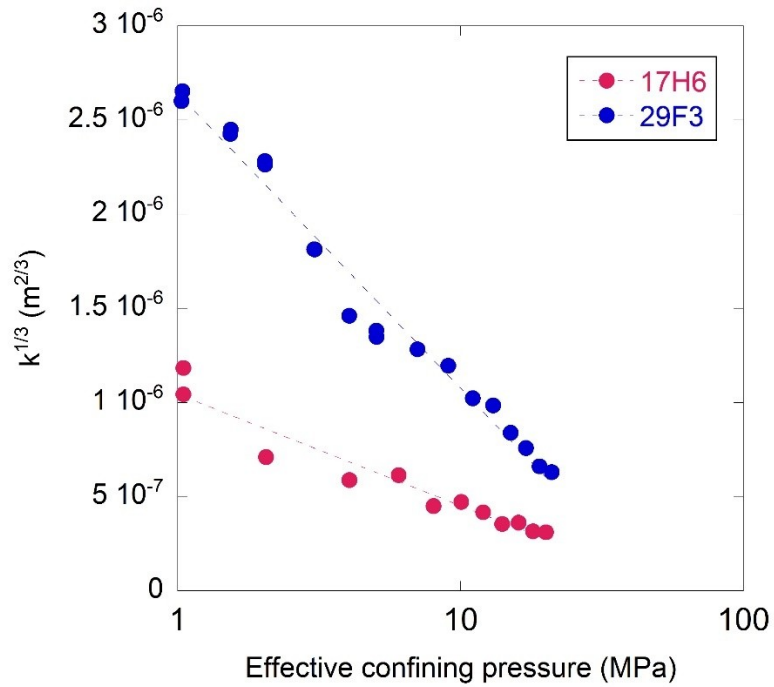
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1104

1105 Figure 9. Permeability k as a function of effective confining pressure P_{eff} at in situ pore
 1106 pressure conditions for samples 17H6 (97.43 mbsf; clayey silt) and 29F3 (157.83 mbsf;
 1107 alternating sand and mud layers) from Site U1517.

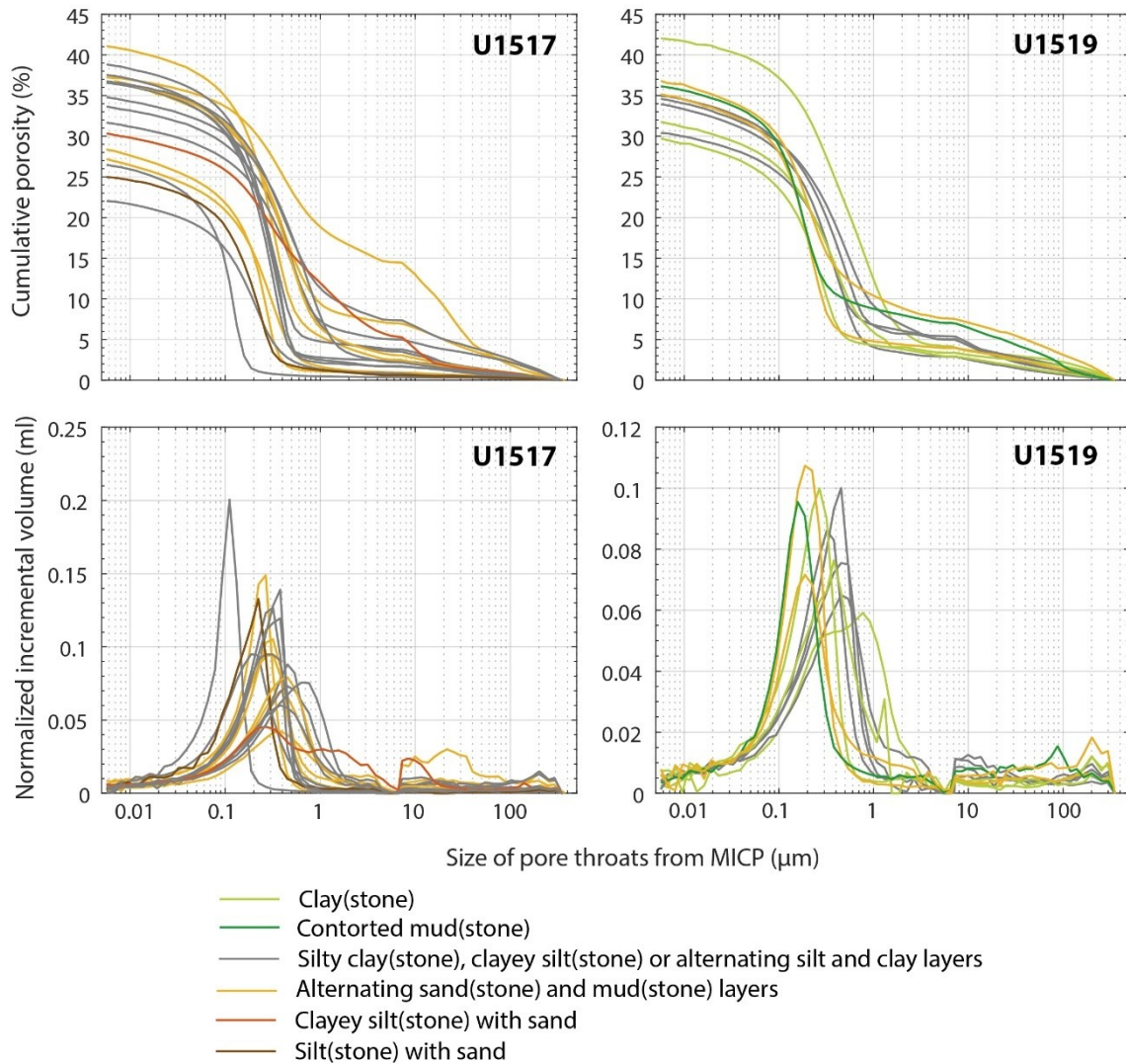
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1110 Figure 10. Permeability data of Figure 9 plotted as $k^{1/3}$ as a function of P_{eff} in a linear-log
 1111 graph showing a linear trend consistent with the fracture closure model proposed by
 1112 Walsh (1981) for samples 17H6 (97.43 mbsf; clayey silt) and 29F3 (157.83 mbsf;
 1113 alternating sand and mud layers) from Site U1517.

1114



1115

1116 Figure 11. MICP pore-throat size distribution spanning common lithologies at Site

1117 U1517 and U1519 versus cumulative porosity and normalised incremental injection

1118 volume.