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1 **From catastrophic collapse to multi-phase deposition: flow transformation, seafloor**
2 **interaction and triggered eruption following a volcanic-island landslide**

3

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18 **Abstract**

19 The current understanding of tsunamis generated by volcanic-island landslides is reliant on
20 numerical models benchmarked against reconstructions of past events. As the largest
21 historical event with timed tsunami observations, the 1888 sector collapse of Ritter Island,
22 Papua New Guinea provides an outstanding opportunity to better understand the linked
23 process of landslide emplacement and tsunami generation. Here, we use a combination of
24 geophysical imaging, bathymetric mapping, seafloor observations and sampling to
25 demonstrate that the Ritter landslide deposits are spatially and stratigraphically
26 heterogeneous, reflecting a complex evolution of mass-flow processes. The primary landslide
27 mass was dominated by well-bedded scoriaceous deposits, which rapidly disintegrated to
28 form an erosive volcanoclastic flow that incised the substrate over much of its pathway. The
29 major proportion of this initial flow is inferred to have been deposited up to 80 km from
30 Ritter. The initial flow was followed by secondary failure of seafloor sediment, over 40 km
31 from Ritter. The most distal part of the 1888 deposit has parallel internal boundaries,
32 suggesting that multiple discrete units were deposited by a series of mass-flow processes
33 initiated by the primary collapse. The last of these flows was derived from a submarine
34 eruption triggered by the collapse. This syn-collapse eruption deposit is compositionally
35 distinct from pre- and post-collapse eruptive products, suggesting that the collapse

36 immediately destabilised the underlying magma reservoir. Subsequent eruptions have been
37 fed by a modified plumbing system, constructing a submarine volcanic cone within the
38 collapse scar through at least six post-collapse eruptions. Our results show that the initial
39 tsunami-generating landslide at Ritter generated a stratigraphically complex set of deposits
40 with a total volume that is several times larger than the initial failure. Given the potential for
41 such complexity, there is no simple relationship between the volume of the tsunamigenic
42 phase of a volcanic-island landslide and the final deposit volume, and deposit area or run-out
43 cannot be used to infer primary landslide magnitude. The tsunamigenic potential of
44 prehistoric sector-collapse deposits cannot, therefore, be assessed simply from surface
45 mapping, but requires internal geophysical imaging and direct sampling to reconstruct the
46 event.

47

48 **Keywords:** Sector collapse, volcanic island, tsunami, landslide, Ritter Island, Papua New
49 Guinea

50

51 **Highlights:**

- 52 • Ritter Island's sector collapse provides an exemplar of volcanic tsunami hazards
- 53 • Deposit heterogeneity reflects erosion, secondary failure and a triggered eruption
- 54 • The volume of the distal deposit alone far exceeds the tsunamigenic failure
- 55 • A single catastrophic collapse led to stratigraphically complex distal deposits
- 56 • Accurate assessment of tsunami potential requires internal imaging and sampling

57

58 **1. Introduction**

59 The sector collapse of Ritter Island, Papua New Guinea (hereafter, Ritter), in 1888,
60 reduced a conical arc-volcanic island to a crescent shaped remnant (Johnson, 1987),
61 generating a tsunami that devastated local coastal communities and caused damage to
62 distances of ~600 km (Day et al., 2015). The collapse is the largest historically recorded
63 landslide at a volcanic island, and of similar scale to the Mount St. Helens sector collapse in
64 1980 (the headwall width of ~3 km and fall in vent elevation of ~1500 m compares with
65 values of 2 km and 1200 m, respectively, for Mount St Helen's; cf. Glicken, 1996). The
66 December 2018 collapse of Anak Krakatau was smaller, but emphasises the potential hazard

67 from such events. Sector collapses occur across all volcanic settings, and their deposits have
68 been identified around numerous arc (e.g., Deplus et al., 2001; Coombs et al., 2007; Silver et
69 al., 2009) and intraplate (e.g., Moore et al., 1989; Masson et al., 2002, 2008) volcanic islands.
70 Tsunami-generating volcanic-island landslides on the scale of Ritter (1-10 km³) have global
71 recurrence intervals of 100-200 years (Paris et al., 2014; Day et al., 2015), with all historical
72 examples occurring in subduction-zone settings. The next large volcanic-island landslide is
73 thus likely to be more comparable to the Ritter collapse than to the much larger events
74 evident from deposits offshore intraplate ocean islands (e.g., the Canary Islands), which have
75 nevertheless received more attention for their tsunamigenic potential (Ward and Day, 2001;
76 Løvholt et al., 2008). Although contemporaneous observations from Ritter are limited due to
77 its remote location, it is by far the largest volcanic-island landslide with timed eyewitness
78 accounts of the generated tsunami. Of particular significance is the observation of a single
79 tsunami wave-train, implying one phase of rapid failure and tsunami generation (Day et al.,
80 2015).

81 The current understanding of landslide-generated tsunami hazards from volcanic
82 islands is principally based on numerical models (cf. Løvholt et al., 2008; Abadie et al.,
83 2012). Such models require an accurate representation of landslide emplacement, but this is
84 challenging to represent due to changes in flow behaviour (for example, arising from mass
85 disaggregation and substrate interaction). Many interpretations of volcanic-island landslide
86 deposits are based on bathymetric observations (cf. Watt et al., 2014), providing poor
87 volumetric constraints and a limited understanding of mass distribution and emplacement
88 dynamics. Studies that draw on high-resolution geophysical imaging or direct sampling have
89 shown that landslide emplacement may involve significant seafloor-sediment incorporation
90 (Watt et al., 2012) or multi-stage deposition (Hunt et al., 2013), highlighting that accurate

91 modelling of tsunami hazards cannot draw on a simplistic extrapolation of landslide deposit
92 volumes.

93 In terms of advancing our understanding of landslide emplacement around volcanic
94 islands, the Ritter collapse is exceptional. The deposit's relatively small dimensions and
95 young age enable direct sampling and observations of both the scar and deposit, and spatially
96 dense imaging by geophysical data. Here, our objective is to use new seismic-reflection, sub-
97 bottom hydroacoustic and bathymetric data, remotely-operated vehicle (ROV) observations,
98 and petrographic and sedimentological analyses of seafloor samples to investigate evidence
99 of spatial and stratigraphic heterogeneity in the Ritter deposit. First, we draw on this array of
100 data types to describe the Ritter landslide from its source to its distal deposits. We then
101 demonstrate that compositional differences can be used to discriminate between pre- and
102 post-collapse volcanic units, and identify evidence that the collapse triggered a submarine
103 magmatic eruption. From these observations, we explain how a morphologically and
104 stratigraphically complex deposit resulted from a single catastrophic collapse. Finally, we
105 evaluate the implications of this for reconstructing volcanic-island sector collapses from their
106 submarine depositional record, which is essential for accurate tsunami hazard modelling.

107

108 **2. Previous work: 1888 landslide deposit facies**

109 Ritter is a relatively small and morphologically youthful edifice, rising from a ~7-km
110 wide base at ~1000 m beneath sea level. It lies between the larger islands of Umboi and Sakar
111 at the eastern end of the Western Bismarck arc (Fig. 1), which is associated with subduction
112 of the remnant Solomon Sea slab in an arc-continent collision environment on the north side
113 of New Guinea (Woodhead et al., 2010). Previous surface mapping (Johnson, 1987; Silver et
114 al., 2009; Day et al., 2015) shows that the 1888 landslide deposits can be divided into three
115 facies based on surface morphology. The proximal facies lies within a basin bounded by the

116 submarine flanks of Sakar and Umboi, and by two submarine volcanic ridges that formed a
117 partial barrier and constriction to the landslide as it flowed to the west. This facies has a
118 relatively flat surface in the north and an irregular, mounded morphology in its southern part
119 (Fig. 1). The mounds were interpreted by Day et al. (2015) as hummocks of the type
120 characteristic of many subaerial volcanic debris avalanches (Siebert, 1984).

121 Beyond the volcanic ridges, the Ritter landslide entered a deeper basin across a
122 relatively steep slope marked by sparse mounds, W of Sakar, that was interpreted by Day et
123 al. (2015) as the surface of a matrix-rich debris avalanche deposit (medial facies; Fig. 1). On
124 its downslope side, the medial facies is bounded by irregular scarps that cut into well-bedded
125 seafloor sediment. These were interpreted by Day et al. (2015) as marking sites of extensive
126 failure of the underlying seafloor sediment. Eroded grooves in this region continue more
127 distally (Fig. 1) across a smooth-surfaced facies, underlain by an acoustically transparent unit
128 of relatively even thickness (termed the debris-flow facies by Day et al., 2015). Tow-camera
129 observations and a dredge sample of cohesive intraclasts in a mud matrix (the only submarine
130 sample previously collected from any of the Ritter deposits; Day et al., 2015) suggested that
131 this unit comprised remobilised seafloor sediment, derived from the upslope sediment-failure
132 scarps.

133

134 **3. Methods**

135 A research expedition on the *RV Sonne* (SO-252; November-December 2016) collected
136 a range of data and observations around Ritter. This included a 3D seismic dataset (Karstens
137 et al., 2019) spanning the proximal facies (Fig. 1), 2D multichannel seismic profiles (two GI-
138 gun source with a 250-m long streamer), high-resolution sub-bottom echosounder profiles
139 (Parasound P70 system), multibeam echosounder bathymetric and backscatter data (EM122

140 and EM710), seafloor photography and direct sampling. Full operational, geophysical data
141 acquisition and processing details are provided in Berndt et al. (2017).

142 Imagery was collected on ten dives around Ritter and the proximal landslide facies,
143 over a total distance of 14 km. Six of these dives used the OFOS high-resolution video
144 sledge, and a further four used OFOS mounted on the HyBIS ROV. Samples were collected
145 at 11 sites using a grab module mounted on HyBIS, capable of collecting up to 30 cm of
146 seafloor sediment. A heavier grab with a wider opening, capable of 40 cm penetration, was
147 used at one site (T1). Gravity coring was attempted at seven sites but failed to penetrate the
148 seafloor, only retrieving small amounts of sediment in the core catcher. Grab samples that
149 preserved intact stratigraphy were logged and subsampled as short cores. Analysed
150 volcanoclastic samples were wet-sieved at half-phi intervals, and for finer samples particle
151 size was determined by laser diffraction (Malvern Mastersizer 2000). Dried and sieved
152 samples were picked for componentry (1 mm to 500 μm fraction); separated grains were
153 mounted in resin and polished for textural imaging (scanning electron microscopy) and
154 compositional analysis (electron microprobe analysis; Jeol JXA8100 Superprobe and Oxford
155 Instruments AZtec system, Birkbeck College). The sites of all samples described in the text
156 are shown in Fig. 1.

157

158 **4. Landslide emplacement processes**

159 *4.1. The primary failure mass*

160 Prior to its collapse in 1888, Ritter was a steep-sided, conical volcanic island, and
161 numerous references in navigational reports (cf. Johnson, 1987) suggest a highly active
162 volcano, characterised by small-scale strombolian eruptions. This is consistent with
163 observations of the subaerial collapse headwall, which exposes interbedded scoria deposits
164 and thin, possibly spatter-fed, lava flows, intersected by cross-cutting dykes (Fig. 2A).

165 Similar lithologies make up the submerged headwall, which is dominated by scoriaceous,
166 bedded deposits in the upper part, with strongly brecciated hyaloclastite bodies becoming
167 more frequent in deeper exposures (Fig. 2B). A highly porphyritic, mafic lithology makes up
168 all the observed exposures. From these observations, we infer that the deeper parts of the
169 edifice were constructed by submarine explosive and effusive activity, and that the upper
170 flanks predominantly comprise scoriaceous material transported down the island flanks from
171 a subaerial vent. The structure of Ritter is thus relatively simple and dominated by poorly
172 consolidated coarse volcanoclastic units, bedded on a metre scale, alongside brecciated lavas
173 and numerous dykes. Our observations imply that the landslide mass was relatively weak,
174 likely to have rapidly disintegrated, and to have been dominated by sand- to cobble-sized
175 clasts that reflect the primary grain-size range of the edifice.

176

177 *4.2. Proximal landslide facies*

178 Seafloor observations across the proximal facies support the interpretation of a weak,
179 disintegrative landslide mass. At the mouth of the collapse amphitheatre, a prominent angular
180 mound with well-developed parallel internal reflections (Fig. 3) is interpreted as an intact
181 portion of the volcano flank (or toreva block). The toreva block's surface exposes a chaotic
182 arrangement of metre-scale dense volcanic blocks (Fig. 2C.i). However, beyond this region
183 the seafloor is ubiquitously smooth and draped in hemipelagic mud. Volcanic blocks up to 50
184 cm across protrude infrequently from the mud in the mounded part of the proximal landslide
185 facies (Figs. 1 & 2C.ii). This provides clear evidence of deposition from the primary
186 landslide mass, but seafloor observations do not suggest that individual mounds contain large
187 (metre to decimetre scale) fragments of the edifice. Furthermore, seismic reflection profiles
188 show that the proximal facies is underlain by folded and thrust-faulted packages of well-
189 bedded sediment (Fig. 3A; Karstens et al., 2019). This implies that the mounded morphology

190 in the proximal landslide facies does not reflect the transport of large, volcanic blocks of the
191 type evident in many subaerial debris avalanche deposits (Siebert, 1984), but reflects the
192 interplay of in-situ seafloor deformation with emplacement of an extensively disaggregated
193 primary landslide mass. A model for how this process occurred and a description of sediment
194 deformation in the proximal region is provided by Karstens et al. (2019), and is not discussed
195 further here.

196 The margins of the mounds in the proximal landslide facies are defined by a network of
197 channels that deepen towards the southwestern outflow of the basin (Fig. 1C), indicating
198 incision into the deformed substrate. The summits of the mounds are deeper than the surface
199 of the flatter area to the north, implying a maximum eroded volume of 1.6 km^3 , estimated by
200 projecting the flat northern surface across the mounded region. We infer that a mass flow
201 derived from the disintegrating primary collapse (2.4 km^3 entirely evacuated from the
202 collapse scar; Karstens et al., 2019) drove this erosion. Although some of this primary mass
203 was deposited proximally and may have infilled a more irregular surface in the northern part
204 of the basin, it does not form a seismically resolvable unit. We thus infer that much of the
205 primary failure mass travelled beyond the proximal facies, implying that up to 4 km^3 of
206 sediment – a volcanoclastic-hemipelagic mixture derived from Ritter and the eroded basin fill
207 – exited the southwestern outflow of the basin (a volume closer to 3 km^3 is more likely,
208 allowing for metre-scale deposition of the primary mass across $\sim 50 \text{ km}^2$ of the proximal
209 facies, and overestimation of channel erosion).

210

211 *4.3. The medial facies*

212 West of the submarine volcanic ridges (Fig. 1), the seafloor between Umboi and Sakar
213 is marked by prominent mounds up to several hundred metres across, which seismic
214 reflection profiles show are blocks rooted within partially buried, seismically-transparent

215 packages (Fig. 3). These are interpreted as volcanic debris avalanche deposits derived from
216 either Sakar or Umboi (based on the distribution of large blocks) and have very similar
217 seismic characteristics to examples offshore other arc islands (Watt et al., 2012). They are
218 interbedded with multiple thinner, tapering deposits. Given their localised extent and
219 proximity to the steep flanks of Sakar and Umboi, we infer that this package of sediment is
220 derived from small mass-wasting events on Sakar and Umboi.

221 The Ritter 1888 deposits cannot be traced into the medial facies in 2D seismic
222 reflection profiles (and there is a lack of sub-bottom echosounder penetration), indicating that
223 the deposit, if it exists, has a thickness of <6 m (the vertical resolution of the 2D seismic
224 data). Direct observations here showed a smooth seafloor draped in hemipelagic mud, with
225 no evidence of coarse clasts protruding at the surface. The lack of a seismically resolvable
226 package, along with backscatter characteristics (general high backscatter, with linear streaks
227 and scour around blocks rooted in deeper landslide deposits), suggests that the Ritter 1888
228 landslide was erosional through this area, potentially accelerating on the observed steeper
229 gradients and due to flow constriction between Umboi and Sakar. This implies that most of
230 the mass exiting the proximal region was ultimately deposited further downslope, in the
231 debris-flow facies identified by Day et al. (2015). We thus reinterpret the matrix-rich facies
232 of Day et al. as an area largely affected by erosion. The division into block- and matrix-rich
233 facies identified at other volcanic debris avalanche deposits (e.g., Glicken, 1996) may not be
234 applicable to the Ritter deposits, principally because of the weak, clastic nature of the primary
235 failure mass.

236

237 *4.4. Erosion and deposition in the debris-flow facies*

238 Beyond the marginal flanks of Sakar and Umboi, seismic profiles show that the
239 stratigraphy is characterised by parallel-bedded sediment that has accumulated on very low

240 gradients in a basin extending ~60 km to the northwest (Fig. 3). A near-seafloor acoustically
241 transparent deposit is observed in sub-bottom profiles across this region (Figs. 4 & 5). We
242 infer that this represents the 1888 landslide based on its seafloor position and its continuity
243 with seafloor erosional fabrics that extend downslope from the proximal Ritter facies. There
244 is no evidence of any internal reflections or structure throughout the deposit, a characteristic
245 typical of debris flow deposits (cf. Damuth, 1980). We thus retain the debris-flow facies
246 terminology of Day et al. (2015) to refer to this unit specifically.

247 The debris-flow deposit forms a lobe approximately 15 km across and up to 16 m in
248 thickness (Figs. 1A, 3B & 4), thickening towards the western margin of the basin (Fig. 5A)
249 and extending slightly up the western edge. At the base of the deposit, stepped incisions cut
250 into the seafloor by several metres (particularly in the proximal part of the deposit), across
251 lateral distances of several kilometres (Fig. 5). This unequivocal evidence of seafloor erosion
252 adds to the identification of seafloor failure scarps at the upslope margin of the facies by Day
253 et al. (2015). Further downslope, the debris-flow deposit thins across a gradient that levels
254 out towards a distinct break in slope (Fig. 1B). This break in slope lies above the buried distal
255 margin of a large landslide deposit (inferred from its distribution to be from Umboi or Sakar;
256 Fig. 3), and beyond this point the deposit forms a second lobe, of similar dimensions to that
257 further upslope.

258 The two lobes of the debris flow deposit have a complex surface morphology,
259 suggestive of spreading and subsequent erosion (Figs. 1A and 1B). The flat surface of the
260 proximal lobe is marked by irregular furrows, which result in an angular, slab-like
261 morphology. These are not slabs or blocks in a strict sense, since there is no evidence of
262 internal boundaries (Fig. 5) or compositional variation within the deposit (i.e. the slabs are
263 certainly not intact fragments of stratified sediment). The position of the furrows also shows
264 no correlation with erosional steps at the base of the deposit, suggesting that the surficial and

265 basal morphologies of the debris-flow facies are unrelated. The morphology may reflect
266 across-flow velocity differences or extension affecting the deposit as it came to rest (cf.
267 comparable morphologies in delta front debris-flow deposits; Prior et al., 1984).

268 The surface of the debris-flow deposit is overprinted by an erosional fabric, comprising
269 sub-parallel grooves that extend northeast (Fig. 1A) and curve to follow the topography along
270 the eastern margin of the basin. The direction of this fabric is slightly oblique to the
271 maximum slope and may reflect erosion by a turbidity current deflected by topography north
272 of Umboi (Fig. 1A). Erosional features are also evident on the surface of the distal debris-
273 flow lobe, in the form of meandering channels and marginal incision (Fig. 1B). This erosion
274 is inferred to represent the final phase of movement associated with the 1888 deposits.

275 Seafloor observations in this region indicate a smooth, featureless muddy surface. Site
276 H6, located on the proximal debris flow lobe, recovered a disturbed 12-cm thick sample. This
277 preserves a medium-grained volcanoclastic sand (see Section 5.3), above a silt that potentially
278 represents the top of the debris flow deposit (Fig. 6). Site H5 was sampled within an eroded
279 part of the distal debris-flow lobe, and comprised a homogeneous fluid mud, rich in
280 foraminifera and containing cohesive silt and fine-sand intraclasts up to several centimetres
281 across. This sample may entirely comprise remobilised hemipelagic mud (with fine-sand
282 interbeds), without any material derived from Ritter itself. Although it is not necessarily
283 representative of the entire debris flow facies, it suggests that pre-existing seafloor sediment
284 formed a substantial component of this part of the 1888 deposits.

285

286 *4.5. Distal turbidite deposition*

287 The most distal part of the 1888 deposits form a unit contiguous with the debris-flow
288 facies in sub-bottom profiles, but distinctive in having an extremely smooth surface and a
289 sheet-like morphology ponded within the distal basin topography (Figs. 1, 3, 4 & 5C). The

290 unit is acoustically distinct from underlying bedded sediment in having higher amplitude but
291 more laterally discontinuous internal reflections, and a base characterised by a continuous
292 high amplitude reflector. The unit contains at least three internal reflections (Fig. 5C), parallel
293 to the unit base and surface, and its base can be mapped across a single reflector around the
294 margins of the debris flow facies, pinching out at the basin margins. Based on these
295 morphological characteristics and its internal boundaries we interpret this unit as a stack of
296 turbidites derived from the 1888 collapse.

297 The turbidite facies is up to 10 m thick in the deepest part of the basin (Fig. 4). The top
298 few centimetres of the facies were sampled at H4 (Fig. 1), recovering a well-sorted, fine-
299 grained volcanoclastic sand (Fig. 6). This sample is not representative of the full unit,
300 particularly given its internal divisions. A sand sampled at H6, in the debris flow facies,
301 correlates with H4 based on sorting and fining relationships and compositional similarities
302 (see Section 5). Both sands have a high bioclast content (Fig. 7), indicating seafloor sediment
303 incorporation. The presence of mud intraclasts further supports this, and corroborates our
304 previous inference that turbidity currents formed the erosional fabric across the debris flow
305 facies. Although it is slightly coarser, the turbidite sand at H6 is only a few centimetres thick,
306 suggesting that deposition from this turbidity current largely bypassed the debris-flow facies.
307 This may also be true of the earlier turbidity currents that formed the lower units of the
308 turbidite facies, providing a potential origin for the seafloor erosion observed at the base of
309 the proximal debris flow lobe.

310

311 **5. Impacts of collapse on volcanic processes**

312 *5.1. Post-1888 volcanism at Ritter*

313 Submarine eruptions since the 1888 collapse (cf. Saunders and Kuduon, 2009) have
314 formed a cone in the centre of the collapse scar (Fig. 1), ~500 m high and with a summit

315 crater 250 m in diameter and 200 m below sea level. The cone's surface is composed of loose
316 scoriaceous gravel (Fig. 2). A sample of this gravel (H2) shows that it comprises pale and
317 dark vesicular components (Fig. 7), with similar phenocryst assemblages. Clinopyroxene is
318 the dominant phase (spanning the diopside-augite field), but plagioclase is also abundant
319 (maximum An₉₁), and both orthopyroxene and olivine are present (phenocryst cores have a
320 maximum forsterite composition of Fo₈₀).

321

322 5.1.1. Compositional differences with pre-collapse samples

323 The cone samples are compositionally distinct from pre-collapse Ritter rocks sampled
324 in the proximal landslide facies. Porphyritic lava blocks from the base of the headwall (H3;
325 Fig. 1) are dominated by coarse (up to 5mm across), equant clinopyroxene (augite, with
326 occasional diopside cores), with variable proportions of olivine and plagioclase. Olivine
327 phenocrysts have dominant core compositions of Fo₇₉₋₈₀ (Fig. 7), but highly forsteritic grains
328 (Fo₈₉) are also present, which we interpret as xenocrysts. Plagioclase phenocrysts show both
329 normal and reverse zoning, reaching a maximum anorthite content of An₉₀; orthopyroxene is
330 absent. Sand and cobble sized clasts from the mounded region of the proximal facies (H9;
331 Fig. 1) have a similar phenocryst assemblage: clinopyroxene compositions extend slightly
332 further into the diopside compositional field; orthopyroxene is absent; and plagioclase
333 phenocrysts span a near identical range (An₇₆₋₈₉, with one outlier at An₉₁). Olivine is rare, but
334 the only analysed olivines were of the high-forsterite type (maximum Fo₈₉). Groundmass
335 glass analyses from several different scoriaceous clasts have a dominant silica content of
336 54.5-56 wt% (Fig. 8). An absence of bioclasts in the sieved fraction of sample H9 (Fig. 7)
337 suggests that this material did not extensively mix with seafloor sediment.

338 The H3 sample does not represent the full spectrum of pre-collapse Ritter compositions
339 (for example, some subaerial pre-collapse rocks contain orthopyroxene; Johnson et al., 1972),

340 but given the sparseness of our sampling, the petrographic similarities between H3 and H9
341 are notable. Although H9 was transported 10 km from Ritter, it is lithologically
342 homogeneous. We interpret that both H3 and H9 represent submarine parts of the pre-1888
343 Ritter edifice.

344

345 5.1.2. Post-1888 deposits across the proximal facies

346 In the smooth-surfaced proximal facies, 7 km WNW of the modern cone, a grab sample
347 (T1; Fig. 1) recovered 45 cm of intact stratigraphy, comprising normally-graded mid- to dark-
348 grey volcanoclastic sand beds, with thicknesses of 1 to >15 cm (Fig. 6). At least six well-
349 sorted unimodal sands are present, separated by up to 20 mm of cohesive mud. These sands
350 are compositionally and texturally similar to sample H2 from the Ritter post-1888 submarine
351 cone: all samples contain a dark aphyric basaltic component (glass SiO₂ 50-52 wt%) with
352 rounded vesicles, as well as porphyritic vesiculated components with variable glass
353 compositions (extending up to 65 wt% SiO₂ in pale vesiculated clasts, which are present in all
354 the T1 beds but particularly common in T1-E (Figs. 6 & 8)). Porphyritic dark-coloured clasts
355 are also present, and commonly have a microcrystalline groundmass, resulting in highly
356 irregular, non-spherical vesicles (Fig. 9). The paler vesicular clasts have a glassy groundmass.
357 Despite the broad range in glass compositions and physical appearance, the phenocryst
358 assemblage is similar across all T1 units and clast types. Clinopyroxene and plagioclase
359 (maximum An₉₂, with rare outliers to An₉₄) dominate, with infrequent orthopyroxene and
360 olivine (maximum Fo₈₁).

361 The surficial position of the T1 volcanoclastic sands and their compositional similarities
362 to H2 suggest that they are derived from submarine explosive eruptions at the post-1888
363 Ritter vent. Inter-bed differences in glass and crystal compositions, and in the relative
364 abundance of pale and dark vesiculated clasts (Fig. 7), suggest that each T1 bed represents a

365 discrete eruption. The products of these eruptions differ from pre-collapse rocks (H3 and H9)
366 in both the glass and mineral compositions of the mafic components (Figs. 7 & 8), and in
367 their textural variety and the presence of an evolved component. Comparable compositions
368 have not been recognised in any pre-collapse subaerial or submarine samples. The T1 and H2
369 samples are also compositionally distinct from the scoria cones west of Ritter, which contain
370 very primitive olivines and mantle xenoliths (Tollan et al., 2017), an observation confirmed
371 by sample H10 (FO_{91} ; Figs. 1 & 7).

372 Sample H8, collected 8 km SW of Ritter (Fig. 1) is compositionally similar to the T1
373 sands, leading us to conclude that an exposure of well-stratified decimetre-scale sand and
374 mud beds (Fig. 2) at this site also represents deposits from post-1888 submarine eruptions at
375 Ritter. 30 km west of Ritter, sample H7 preserves the top 11 cm of seafloor sediment and
376 contains nine layers of grey silt, 1-10 mm in thickness, interbedded with cohesive mud (Fig.
377 6). The colour, thickness, grain-size and bedding characteristics of the silts are consistent
378 with them being the distal equivalent of the post-1888 volcanoclastic sands at T1 and H8.

379 The 250-m wide crater at the summit of the post-1888 cone and the loose scoriaceous
380 debris on its flanks implies that the cone has been constructed through submarine explosive
381 eruptions powerful enough to eject material far above the vent. We speculate that pyroclastic
382 material falling back from the submarine eruption column fed sediment density flows that
383 formed the deposits identified across the basin west of Ritter. These deposits form a metre-
384 scale veneer across the proximal landslide facies, which is not resolvable at the resolution of
385 the seismic reflection data. We thus infer that the surface morphology of the proximal facies
386 is a primary fabric developed during emplacement of the 1888 landslide, and has not been
387 significantly modified by subsequent volcanism.

388

389 *5.2. The distal volcanoclastic turbidite*

390 The volcanoclastic turbidite sampled at H4 and H6, in the distal Ritter deposits, has
391 some unexpected compositional characteristics. Rather than being mixed, as might be
392 expected from a landslide-derived turbidite, both samples contain two discrete volcanoclastic
393 components. Slightly over half the clasts are vesicular, mafic grains, generally aphyric and
394 with a single dominant glass composition, identical to the post-collapse basaltic grains in T1
395 (and distinct from pre-collapse mafic clasts; Fig. 8). Although phenocrysts are relatively
396 infrequent in these grains, clinopyroxene and plagioclase dominate, orthopyroxene is rare,
397 and olivine was not analysed. Phenocryst compositions overlap with analyses of the more
398 proximal Ritter samples (Fig. 7). A second component, making up ~25% of the samples, is a
399 white, highly vesicular pumice (Figs. 7 & 9). The pumice glass composition is rhyolitic but
400 relatively scattered, and more evolved than the T1 pale-coloured vesicular clasts. The pumice
401 contains abundant phenocryst amphibole (equant, unrimmed magnesiohornblende) and
402 plagioclase (extending to An₆₀; Fig. 7), with minor apatite and Ti-magnetite. Very rare clasts
403 of this low-density pumice, up to 2 cm across, were found in disturbed parts of the T1 grab
404 sample (7 km WNW of Ritter), indicating that coarser grains of the hornblende-pumice are
405 also present near Ritter.

406

407 5.2.1. Origin of the volcanic components

408 The narrow compositional range of the mafic component is consistent with an origin
409 from a single eruption. A juvenile origin is also supported by its compositional similarity
410 with post-1888 basaltic grains, and its dissimilarity with pre-collapse samples (Fig. 8). The
411 hornblende-pumice defines a similarly discrete population. Although the pumice could have
412 been incorporated from a seafloor deposit (and bioclasts and mud-intraclasts certainly
413 indicate hemipelagic sediment incorporation), this would imply a significant volumetric
414 addition from a single, near-surface deposit. This seems unlikely to have occurred without

415 any mafic volcanoclastic incorporation, which would have produced a wider compositional
416 range in the mafic component. Furthermore, no comparable pumice is known from local
417 volcanoes. Hornblende is an infrequent phenocryst phase in regional magmas, and evolved
418 magma compositions are extremely rare (Johnson et al., 1972) throughout the Western
419 Bismarck arc, which is dominated by mafic magmas. The coarseness of the pumice clasts at
420 site T1, as well compositional dissimilarities, means that regional caldera-forming eruptions
421 (Long Island, Witori and Dakataua; Fig. 1) can also be discounted as sources (Machida et al.,
422 1996; Blong et al., 2017). The only similar local rock is a hornblende-andesite that crops out
423 near the summit of Sakar (Johnson et al., 1972). Rocks from Sakar and Umboi are otherwise
424 very similar to the augite-rich Ritter samples, but the Sakar andesite demonstrates that more
425 evolved and diverse magma compositions are present in nearby magmatic systems. No
426 pumice deposits are known from Sakar itself or in any local subaerial exposures.

427 The samples containing the hornblende pumice only represent the uppermost turbidite
428 sub-unit. It is very likely that the Ritter landslide also produced a far-reaching turbidity
429 current, and we therefore suggest that the deeper turbidite sub-units contain mixed
430 volcanoclastic sediment from the pre-collapse edifice. The absence of pre-collapse material in
431 the uppermost turbidite, and the erosion that overprints the debris-flow facies, suggests a time
432 gap between the collapse and the event generating the uppermost turbidite, sufficient for the
433 resulting sediment-flows to remain separate.

434

435 5.2.2. A collapse-triggered eruption?

436 Given the above observations, we suggest that the hornblende pumice is a juvenile
437 magmatic component from a compositionally bimodal submarine eruption that followed the
438 1888 collapse. This eruption must have been substantially larger than subsequent (post-1888)
439 eruptions at Ritter, given that its products were transported to much greater distances. The

440 absence of a second documented tsunami either implies that the eruption followed the
441 collapse within minutes (with both events contributing to a single wave train), or that any
442 tsunami generated by the eruption was too small to be identifiable on a regional scale (no
443 observers were present on nearby islands). The eruption vent would have been ~700 m below
444 sea level (inferred from the base of the post-1888 cone). The occurrence of a relatively deep
445 submarine eruption is consistent with the lack of an observed explosive eruption column
446 following the 1888 collapse. Water-rich melts can still form pumice at depths exceeding 1000
447 metres (e.g., Rotella et al., 2015), and it is possible that pumice reported to have washed up
448 on nearby beaches was derived from the eruption (Anonymous, 1888; Steinhauser, 1892). A
449 submarine eruption, following rather than accompanying the collapse, is also consistent with
450 the distribution of eruptive products only to the west of Ritter (sample 2A-CC, on the east
451 flank of Ritter (Fig. 1), contains no pale vesiculated or pumiceous clasts).

452 Although it is possible that magma ascent preceded Ritter's sector collapse and
453 destabilised the edifice, we suggest that the collapse itself may have triggered magma ascent
454 due to decompression of the underlying magma reservoir. There are other examples of
455 compositionally anomalous eruptions occurring after major sector collapses (Watt, revised),
456 and the multiple compositional modes of younger post-collapse eruptions at Ritter (Figs. 8 &
457 9) point to ongoing tapping of a complex plumbing system, erupting more evolved melts
458 than are apparent in pre-collapse rocks.

459

460 **6. Summary of emplacement processes**

461 *6.1. Flow transformation and multi-phase deposition – a conceptual model*

462 The Ritter 1888 landslide began with a single stage of tsunamigenic collapse, but the
463 erosional patterns, morphological and compositional heterogeneities of the resultant deposits
464 can only be explained through multiple styles and phases of mass movement. This

465 complexity resulted from flow transformation, seafloor interaction and eruptive activity.
466 Aspects of this process remain ambiguous, but its principal features can be constrained (Fig.
467 10). The initial landslide mass was highly disintegrative, and while a proportion of this
468 material remains within the proximal region, the remaining fraction travelled through the
469 constriction between Umboi and Sakar. In the medial facies, further downslope, there is no
470 strong evidence of a Ritter deposit (Figs. 1 & 3). We infer that the initial volcanoclastic
471 density current was erosive through this region. Further downslope, the leading part of this
472 flow must have remained erosional despite a much reduced gradient, as indicated by
473 widespread seafloor erosion (Fig. 5). We thus infer that much of the primary volcanoclastic
474 mass that exited the proximal region was deposited in the most distal part of the Ritter
475 deposits.

476 The substrate eroded by the initial flow is buried by a homogeneous debris-flow facies.
477 An absence of internal structure within this unit suggests an extensively disaggregated mass,
478 whose emplacement must have followed the initial erosive event. The deposit thickness and
479 lobate forms suggest a relatively strong debris flow (cf. Talling, 2013), but the deposit does
480 not extend upslope towards the proximal landslide facies. Our sampling of the distal debris
481 flow lobe suggests that the major constituent of the debris-flow deposit is relatively local
482 hemipelagic sediment. Failure of this material may have followed erosion by the initial
483 volcanoclastic flow, retrogressively forming the scarps that mark the upslope margin of the
484 debris flow facies (Day et al., 2015). This process may have been enhanced through
485 deposition and loading by later volcanoclastic material exiting the proximal region. Our
486 sampling doesn't allow us to test this, but the lack of internal boundaries within any parts of
487 the debris-flow facies suggest its emplacement essentially involved a single phase of failure
488 and deposition. Finally, the surface erosional fabric across the distal deposits is most easily

489 explained by a late stage turbidity current derived from a submarine explosive eruption at
490 Ritter, forming the compositionally-distinctive uppermost turbidite sub-unit.

491 The above explanation implies heterogeneity in both the type and distribution of
492 material across the Ritter deposits. A substantial fraction of the primary failure mass was
493 likely transported over 80 km, but deposition of this material bypassed much of the flow
494 pathway. Conversely, much of the sediment in the debris-flow facies may have only travelled
495 a few kilometres and be locally derived.

496 The different phases of movement that formed the Ritter deposits were not necessarily
497 separated by time gaps (with the possible exception of the final eruption-generated turbidity
498 current), but can be explained by a combination of flow transformation and secondary
499 sediment failure. As well as forming a major component of the proximal mass affected by the
500 landslide (Fig. 3), pre-existing sediment makes up much of the distal deposit, and the general
501 prevalence of seafloor sediment interaction across the Ritter deposits replicates observations
502 around other volcanic islands (Watt et al., 2012). The relative proportions of volcanoclastic
503 and hemipelagic sediment in the distal deposit cannot be constrained, but it has a total volume
504 of 5.0 km³ (Fig. 4). This exceeds the volume of material exiting the proximal region (3–4
505 km³, which itself includes up to 1.6 km³ of eroded seafloor sediment), and is also a minimum
506 estimate, since it excludes any deposition closer to Umboi and Sakar, as well as turbidite
507 deposits outside the mapped region. The additional volume must primarily comprise seafloor
508 sediment eroded further downstream, although a further, unconstrained component is that of
509 the eruption-generated turbidity current.

510

511 *6.2. Implications for interpreting volcanic-island landslide deposits*

512 In a broad sense, the 1888 Ritter deposits comprise all material deformed or transported
513 as a consequence of the volcano's sector collapse, with a total volume of ~15 km³. However,

514 this includes material in the proximal facies that was affected by in-situ deformation, and
515 distal deposits from secondary sediment failure and a post-collapse eruption. Referring to this
516 entire volume as a landslide deposit is thus somewhat misleading without an understanding of
517 its structural and compositional heterogeneities. For landslides involving such complex
518 emplacement processes, there is no simple relationship between deposit area and primary
519 landslide magnitude. The processes identified here may be additional factors influencing the
520 apparently high mobility (i.e. total downslope extent of associated deposits) of many
521 landslides around volcanic islands (cf. Hürlimann et al., 2000). Similar emplacement
522 complexities may be important in other submarine settings, but evidence of secondary
523 downslope failures may be much harder to identify when the primary landslide mass is not
524 compositionally distinctive from surrounding seafloor sediment (Gee et al., 1999).

525 An understanding of how the Ritter landslide was emplaced has only been achieved
526 through internal geophysical imaging in combination with sampling, and it is reasonable to
527 extend this conclusion to landslide deposits around other volcanic islands, where
528 morphological complexities are often apparent (Deplus et al., 2001; Coombs et al., 2007;
529 Watt et al., 2014). If landslide scars are obscured by subsequent volcanism, both geophysical
530 data and sampling are likely to be required for accurate estimates of primary failure volumes,
531 which is the critical parameter for modelling associated tsunamis.

532 Internal reflections show that the Ritter turbidite comprises multiple sub-units. The
533 uppermost unit is composed of juvenile material from a post-collapse eruption, and we infer
534 that deeper sub-units are derived from the primary failure mass. Divisions within these deeper
535 units may reflect both compositionally discrete phases (e.g., secondary seafloor failure) and
536 complexities introduced by transport pathways and seafloor topography (e.g., current
537 reflection). Multi-part turbidites in several landslide-derived deposits around the Canary
538 Islands have been inferred to represent multi-stage landslide failure (Hunt et al., 2013),

539 implying a reduced tsunami-generating capacity. In the case of Ritter, a multi-part turbidite
540 appears to have been generated from a single initial collapse. However, we caution against a
541 simplistic interpretation that all landslides on volcanic islands proceed in a similar fashion.
542 Historical sector collapses at subaerial arc volcanoes have involved a single rapid phase of
543 movement (e.g., Mount St. Helens; Glicken, 1996), and it is unsurprising that collapses on
544 their island equivalents are similar. This may be typical of landslides at arc volcanic islands,
545 which have moderate dimensions (1–10 km³) and represent the majority of tsunamigenic
546 volcanic landslides. Major landslides in ocean-island settings are much larger (potentially far
547 exceeding 100 km³), but these islands are also structurally and morphologically very different
548 to island-arc volcanoes, and their landslides may also be more shallowly seated (cf. Watt et
549 al., 2014). It should not, therefore, be assumed that the typical style of collapse is identical
550 across arc and ocean-island settings.

551

552 **7. Conclusions**

553 Our sedimentological and petrological analysis of seafloor samples, combined with
554 interpretation of geophysical data and seafloor imagery, has allowed us to reconstruct the
555 emplacement of deposits generated by the sector collapse of Ritter Island in 1888. Our results
556 show that the primary landslide mass disintegrated rapidly and is principally distributed
557 within a proximal basin and in the distal turbidite. Parts of the intervening region contain
558 debris-flow deposits comprising a major proportion of disaggregated hemipelagic sediment,
559 the failure of which appears to have been triggered following erosion by the initial
560 volcanoclastic flow. The initial landslide was also shortly followed by a magmatic eruption,
561 erupting both basaltic magma and a distinctive, evolved pumiceous component. The eruption
562 of this pumice implies that more varied magma compositions exist beneath Ritter than is
563 apparent from the stable, basaltic-andesitic pre-collapse magmatism. This is supported by the

564 bimodal nature of repeated eruptions that have constructed a submarine cone in the collapse
565 scar. At least six discrete eruptions have occurred since 1888, suggesting output levels that
566 are at least comparable to Ritter's pre-collapse activity, and indicating that Ritter remains one
567 of the region's most active volcanoes.

568 The total volume involved in the Ritter 1888 sector collapse (up to 15 km³) is several
569 times larger than the initial failure volume. The distal deposits alone (40-85 km from Ritter)
570 have a volume far exceeding that of the initial collapse. Lateral and stratigraphic
571 heterogeneity in the Ritter deposits results from a combination of different flow behaviours,
572 proximal and distal seafloor erosion, secondary failure of seafloor sediment, and a magmatic
573 eruption triggered by the initial collapse. Such complexities have only been revealed through
574 a combination of sampling and geophysical imaging; accurate assessments of primary
575 collapse volumes, and thus the tsunamigenic potential of ancient volcanic landslides, must
576 therefore be based on detailed deposit characterisation and cannot be achieved by surface
577 mapping alone.

578 The observation of a single tsunami wave-train at Ritter implies that the initial landslide
579 was the principal tsunami generating mechanism, and that the triggered submarine eruption
580 was less significant in this context. Our results also demonstrate that a single-stage collapse
581 can result in a highly complex set of deposits. This complexity is recorded in the Ritter
582 turbidite facies, which has internal reflections and which we conclude is stratigraphically
583 varied (in part due to the triggered eruption). This has implications for how primary sector-
584 collapse complexities are inferred from distal turbidite stratigraphies, although we note that
585 sector-collapse processes should not be assumed to follow similar mechanisms across both
586 arc and intraplate settings, given significant differences in magnitude and island structure.

587

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599

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699

700 **Figure captions**

701 **Figure 1**

702 Bathymetric map of the seafloor around Ritter Island, Papua New Guinea, showing major
703 morphological features of the 1888 landslide deposits (SO-252 bathymetry merged with
704 GEBCO bathymetry outside the survey area and SRTM elevation data on land). **A** and **B**
705 show bathymetry of two areas of the distal deposit, highlighting erosional and depositional
706 features. The shaded relief map in **C** is coloured according to our interpreted morphological
707 characteristics of the landslide scar and proximal facies, distinguishing pre-existing volcanic
708 topography and the post-1888 submarine cone. Sample sites are shown in bold and image
709 localities (Fig. 2) in italics. **D** shows a regional location map of the Bismarck arc (cf. Johnson
710 et al., 1972; Woodhead et al., 2010).
711

712 **Figure 2**

713 Photographs of the Ritter landslide headwall, deposits and post-collapse units (locations in
714 Fig. 1). **A**: The central portion of the subaerial headwall, showing interbedded scoriaceous
715 deposits and parallel-bedded thin lavas dissected by cross-cutting dykes. Insets show finer-
716 scale detail (photograph: T. Bierstedt). **B**: The submerged headwall, comprising small
717 submarine lava bodies with poorly-developed pillow structures (B.i) or forming dense
718 hyaloclastite breccias with alteration rinds (B.ii; arrows), interbedded with reworked
719 volcanoclastic deposits (B.iii). **C**: The proximal facies seafloor: Dense lava blocks with planar
720 fractured surfaces on the surface of the Toreva block at the mouth of the collapse scar (C.i);
721 infrequent exposures of angular gravel- to cobble-sized lava clasts (rarely up to 50 cm across)
722 protruding on the steep margins of a mound, draped by yellow hemipelagic mud (C.ii). **D**:
723 Post-1888 deposits: loose scoriaceous gravel on the surface of the post-1888 submarine cone
724 (D.i); volcanoclastic sands interbedded with cohesive hemipelagic mud, exposed near site H8
725 in the walls of shallow, rounded depressions (D.ii).
726

727 **Figure 3**

728 2D seismic reflection profile along the 1888 deposits (line position in Fig. 1), with annotated
729 interpretations. The profile highlights the relationship between topographic changes and the
730 deposit facies. **A** shows typical structures in the proximal landslide facies. The bold dotted
731 line shows the base of the proximal deposits defined by Karstens et al. (2019); **B** shows
732 details of the distal deposits and underlying stratigraphy. Selected reflectors have been picked
733 as black lines to highlight deposit structures.
734

735 **Figure 4**

736 Thickness map of the distal 1888 Ritter deposits based on interpolated sub-bottom
737 echosounder profiles (bathymetry contours at 50 m intervals). The mapped region has a
738 volume of 4.97 km³ (based on a sediment velocity of 1600 m/s) and extends as far as the unit
739 can be imaged with high-frequency hydroacoustic data. The deposit shows two distinct

740 acoustically-transparent lobes (interpreted as debris-flow deposits), beyond which lies a flat
741 surfaced deposit of very regular thickness.

742

743 **Figure 5**

744 Sub-bottom echosounder profiles through the distal 1888 Ritter deposits (line positions in
745 Fig. 4). **A** and **B** show erosional features and changes in internal and morphological
746 characteristics across the debris flow facies (note the vertical exaggeration). **C** highlights
747 internal reflections within the distal, flat-surfaced deposit, interpreted as a multi-unit turbidite
748 ponded in the distal basin.

749

750 **Figure 6**

751 Logs of grab samples preserving shallow intact sections of seafloor stratigraphy (locations in
752 Fig. 1). The volcanoclastic sands at H6 and H4 are the uppermost part of the syn-collapse
753 turbidite. Both contain mud intraclasts and are very well sorted. Volcanoclastic sands and silts
754 at T1 and H7 result from post-1888 eruptions at the submarine cone, and are relatively finer
755 and thinner. The beds are unimodal and normally-graded, and interbedded with cohesive
756 mud. The photograph shows an example of the stratigraphy. Grain-size measurements are
757 from wet-sieving at half-phi intervals or laser-diffraction measurements (H7).

758

759 **Figure 7**

760 Sediment components and phenocryst compositions of submarine volcanoclastic samples
761 (locations in Fig. 1). H-10C, from an outlying scoria cone, contrasts with all other samples,
762 which are interpreted as being from Ritter. **A**: Componentry based on grain-counting of
763 sieved 500 μm to 1 mm grain-size fractions. H4-B and H6-G (from the uppermost syn-
764 collapse turbidite) are notable for their bioclast content. All post-collapse samples contain
765 variable proportions of pale vesicular clasts. **B**: Plagioclase phenocryst core compositions
766 from picked crystals or vesicular clasts. Individual analyses are shown as crosses; the grey
767 bars mark the 5th to 95th percentile range (black line: median) for each sample group; post-
768 collapse phenocrysts extend to higher anorthite values. A pumiceous component in the syn-
769 collapse turbidite contains much less anorthitic phenocrysts (similar grains occur infrequently
770 in post-collapse samples). **C**: Olivine phenocryst core compositions. Olivine is frequent in
771 pre-collapse samples and based on two clustered populations we have interpreted a distinct,
772 highly forsteritic xenocrystal population. This population was not found in post-collapse
773 samples, where olivine is much less abundant but extends to slightly more forsteritic
774 compositions than the pre-collapse phenocrysts.

775

776 **Figure 8**

777 Matrix glass compositions in vesicular grains separated from volcanoclastic samples
778 (locations in Fig. 1). H9-B, representing pre-collapse material, spans a narrow range of
779 basaltic andesite glass compositions. Material in the uppermost syn-collapse turbidite is
780 distinctive from this: mafic clasts are dominated by basaltic glass, while pumiceous
781 porphyritic clasts have a rhyolitic matrix. Post-1888 (T1) deposits are multimodal, with an
782 aphyric basaltic component and porphyritic basaltic-andesitic to dacitic components.

783

784 **Figure 9**

785 Examples of volcanic clast textures from volcanoclastic samples. All samples contain mafic
786 vesicular components; aphyric, glassy clasts have rounded vesicles and are basaltic in the
787 uppermost syn-collapse turbidite and post-collapse deposits, but basaltic-andesite in the
788 proximal landslide sample (H9-B). Porphyritic mafic clasts are common in T1 samples and
789 typically have microcrystalline groundmasses, resulting in very irregular vesicles (T1-C

790 image), but in some cases are glassy, with a basaltic-andesite composition (T1-I image). Pale
791 vesicular clasts are also frequent in T1 samples and have more evolved glass compositions.
792 All clast types contain highly anorthitic and overlapping plagioclase populations ($An_x\%$
793 values). In contrast, the pale vesicular clasts in the syn-collapse turbidite (H6-G image)
794 contain less anorthitic phenocrysts and hornblende, and have a very open, pumiceous texture.
795

796 **Figure 10**

797 A summary of processes involved in the emplacement of the 1888 Ritter landslide deposits.
798 The topographic profile shows a transect along the flow pathway (approximately equivalent
799 to that shown in Fig. 3), with variable vertical exaggeration.
800

801 **Supplementary Data**

- 802 **1. Site coordinates and descriptions of all sampling sites and of images used in the text,**
803 **with sample descriptions and images (.pdf).**
804 **2. Mineral compositions for data plotted in Figs. 7 and 8 and referred to in text.**

Figure 1

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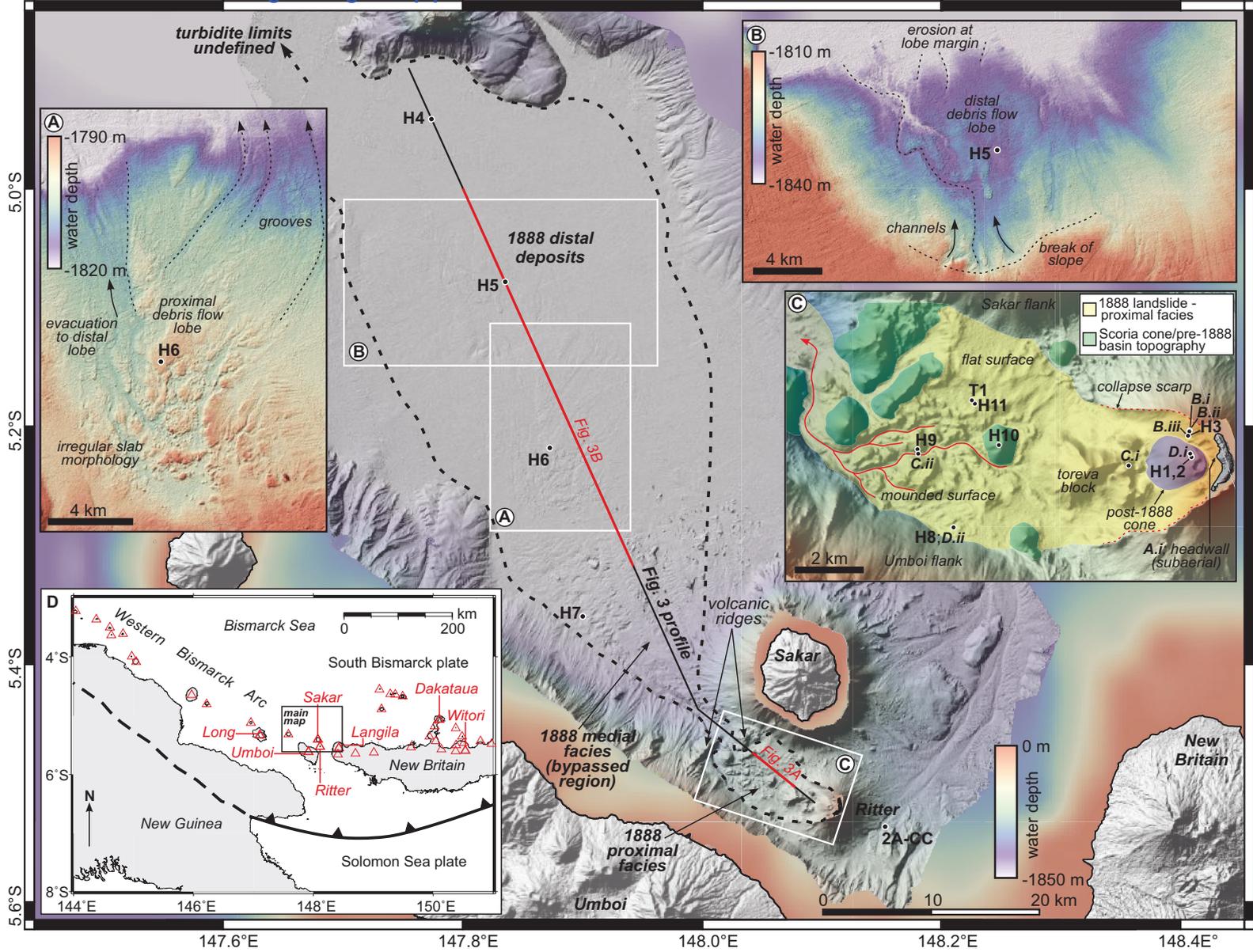
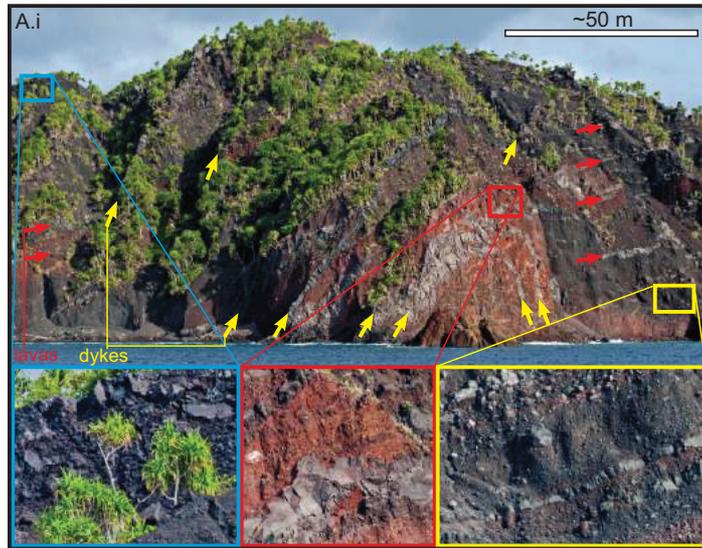


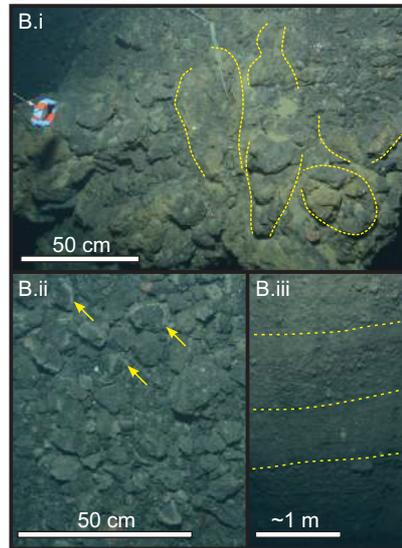
Figure2

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A: Subaerial scar



B: Submarine scar



C: Proximal landslide deposit



D: Post-collapse deposits

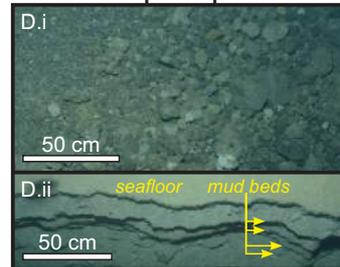


Figure3

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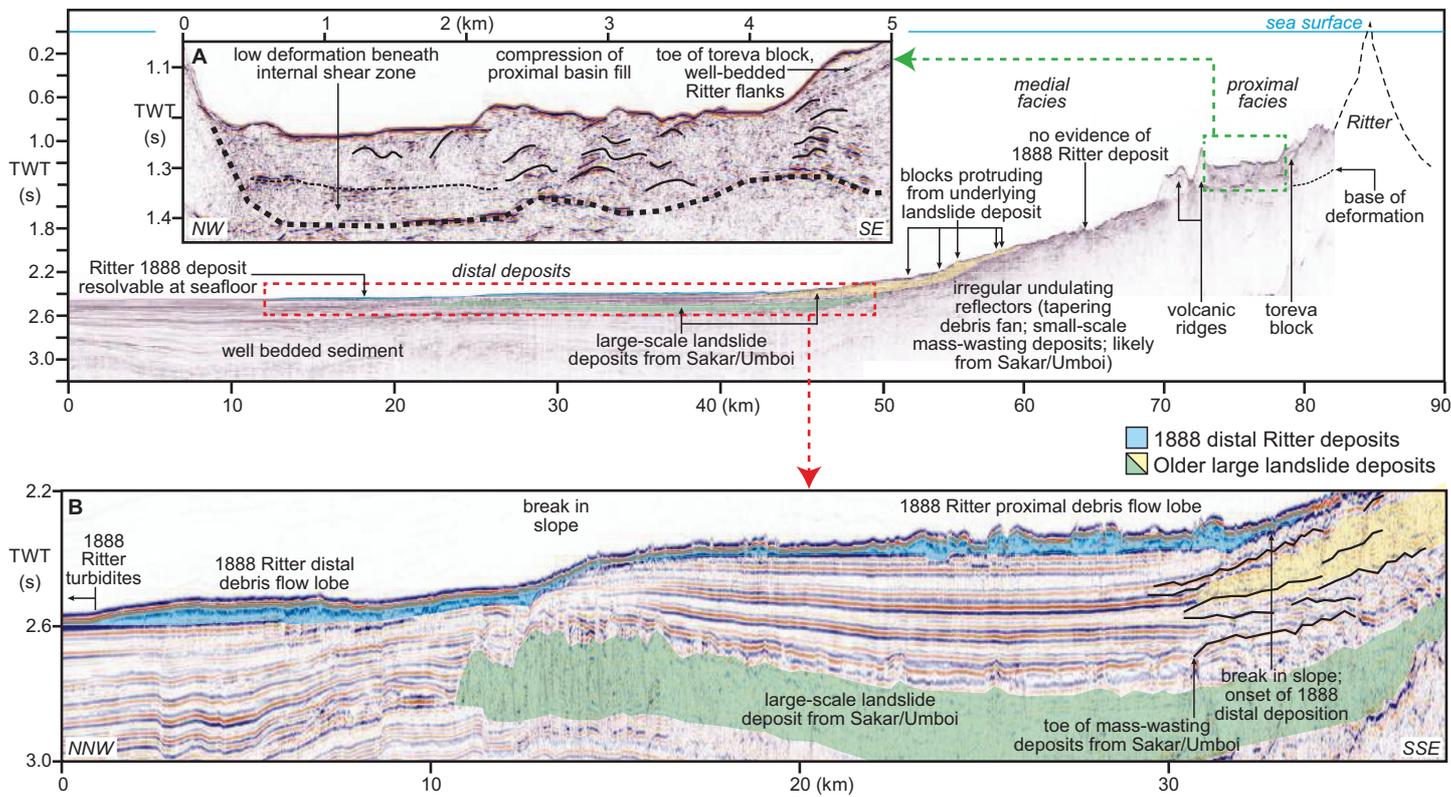


Figure4

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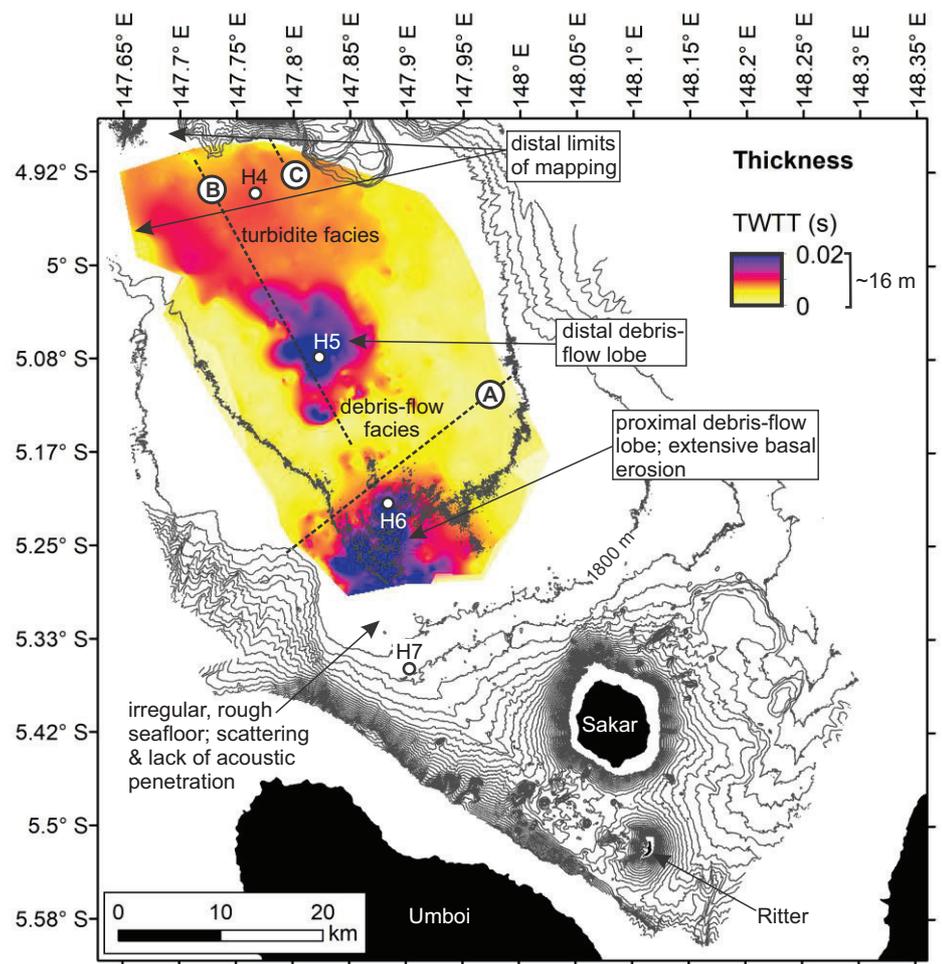


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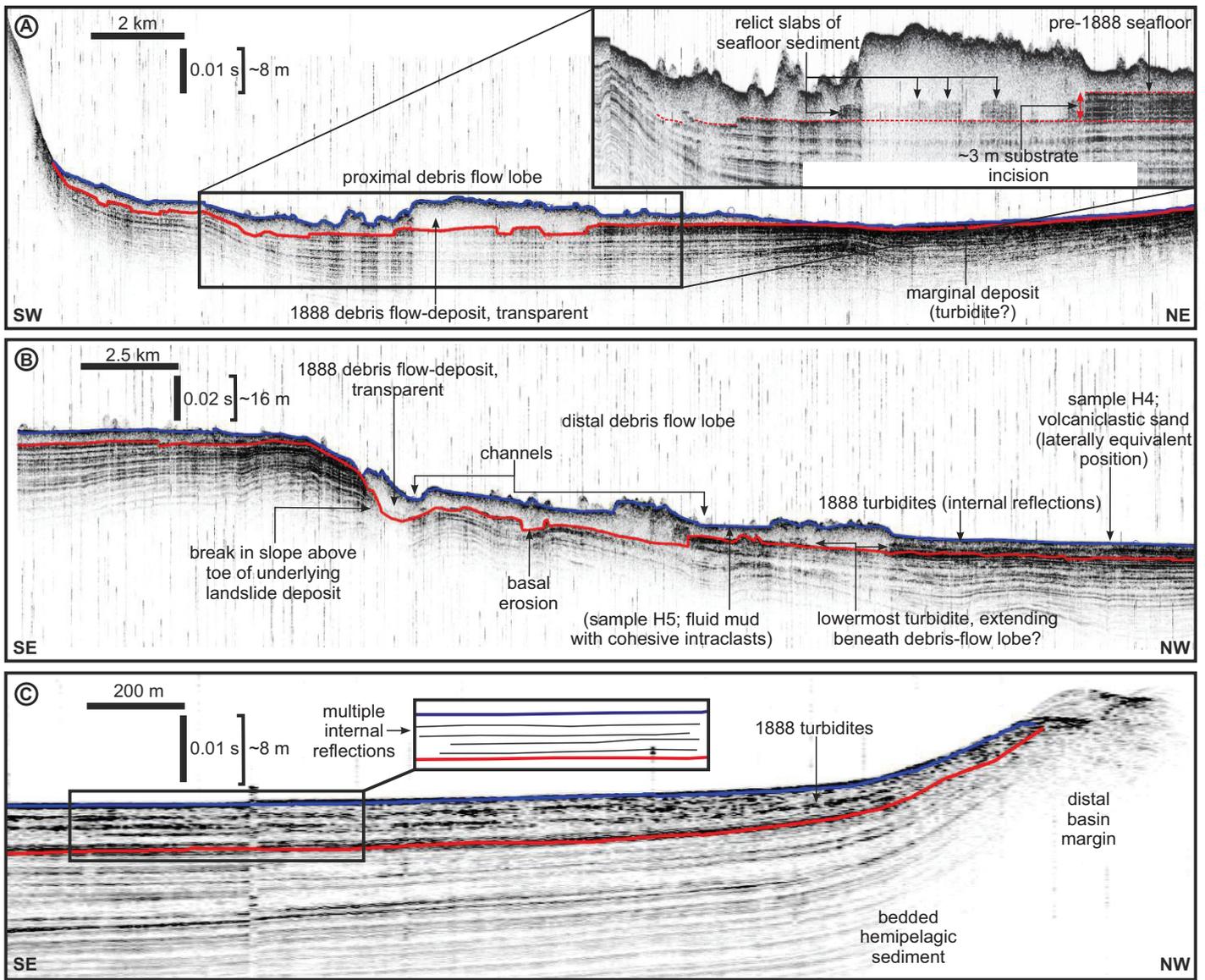


Figure6

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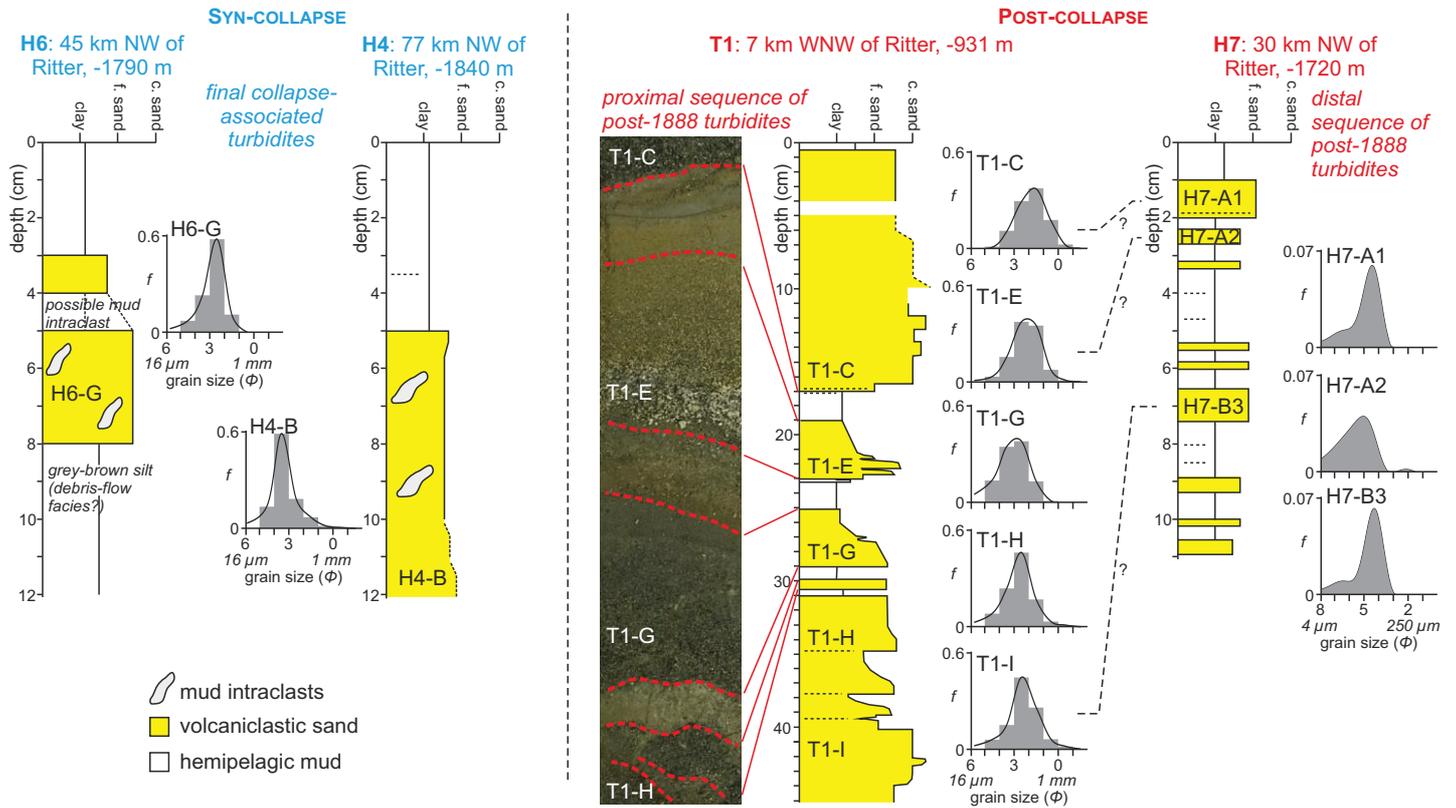


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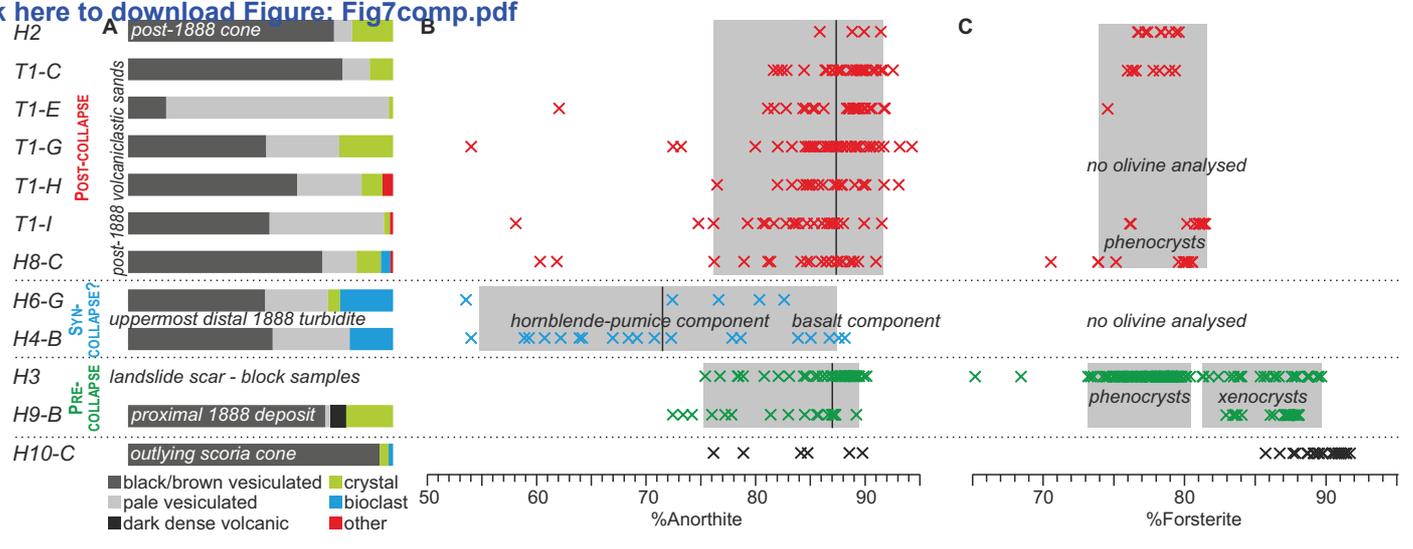


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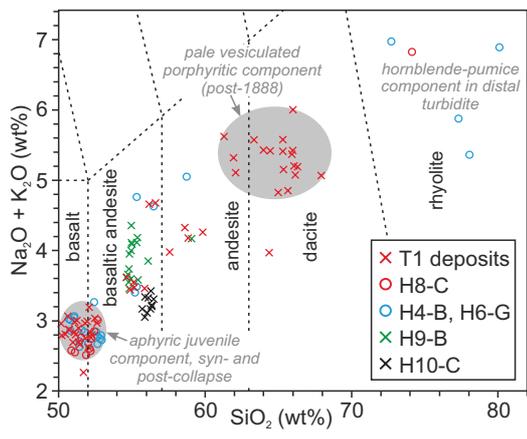


Figure9

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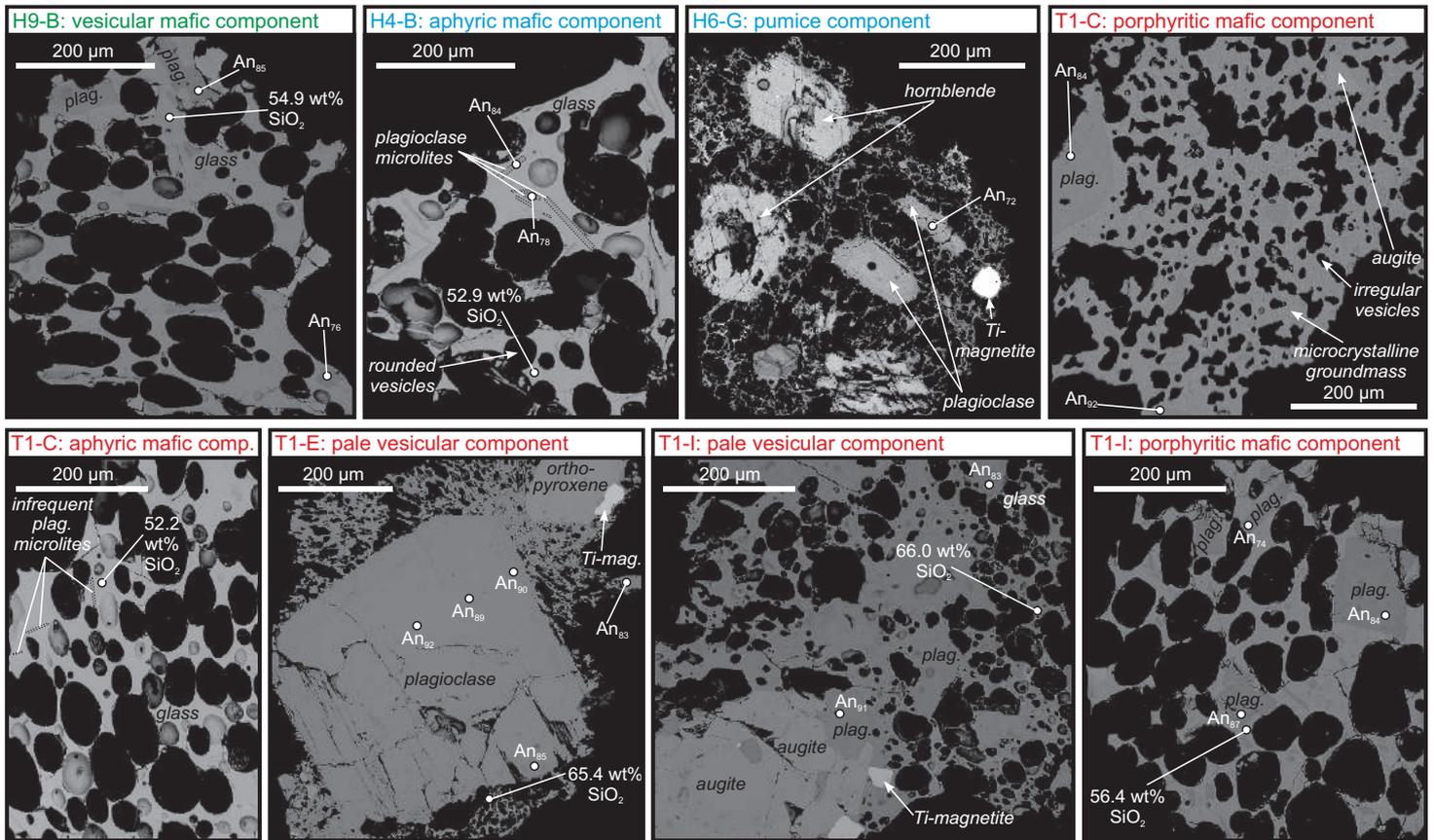


Figure 10
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