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26 the top surface of, the slide. Along the longitudinal shear zone, the shear bands and ridges of Cell A
27 were dragged downslope, due to Cell A impeding the movement of Cell B. This interaction suggests
28 that Cell B travelled faster, and/or further, than Cell A, due to the absence of any flow obstacles. The
29 abrupt cessation of Cell A is recorded by positive seabed relief, whose amplitude decreases updip. The
30 transport processes of the Gorgon Slide show how entrainment and abrasion of megaclasts induced
31 velocity perturbations during emplacement causing: (i) changes to the flow rheology, and (ii) initiation
32 and cessation of flow cells. A better understanding of how flow cells evolve during MTCs transport
33 may help to refine modelling of the potential impact of MTCs on submarine structures (e.g. pipelines,
34 cables, etc.).

35

INTRODUCTION

36 The degradation of submarine slopes drives emplacement of large mass-transport complexes (MTCs),
37 which are deposits of gravity-driven depositional processes that include slides, slumps and debris
38 flows (e.g. Dott 1963; Nardin et al. 1979; Nemeč 1991; Moscardelli and Wood 2008; Posamentier and
39 Martinsen 2011). Besides their role in continental margin evolution (e.g. Gamboa et al. 2010) and
40 petroleum system development (e.g. Weimer and Shipp 2004), MTCs also pose a significant geohazard
41 for coastal and offshore engineering structures (e.g. Parker et al. 2008; Randolph and White 2012;
42 Vanneste et al. 2013). Understanding the magnitude of this geohazard is essential and partly depends
43 on the emplacement processes of the MTCs (Masson et al. 2006). For example, the rheology of MTCs
44 can evolve during transport (Iverson 1997; Dykstra et al. 2011; Joanne et al. 2013; Ortiz-Karpp et al.
45 2017; Hodgson et al. 2018), which controls the amount of drag forces experienced by submarine
46 pipelines (e.g. Zakeri 2009).

47 Transport processes of an MTC are dynamic, where a large, single (first-order) flow cell can evolve
48 during its translation by (i) dilution through water ingestion (Fisher 1983; Talling et al. 2012; Sun et al.
49 2018); and/or (ii) formation of smaller, second-order (intra-MTC) flow cells due to internal velocity

50 variations (Alsop and Marco 2014). In contrast to flow transformation, the formation processes of flow
51 cells remain poorly documented. Although the nature of flow cells have been inferred from outcrop
52 studies (Farrell 1984; Alsop and Marco 2014), limited outcrop extent invariably hampers full 3D
53 analysis. Studies involving 3D seismic reflection data have also documented the presence of intra-MTC
54 flow cells (Gee et al. 2005; Bull et al. 2009; Steventon et al. 2019). However, the mechanisms
55 responsible for the initiation, translation, and cessation of these flow cells remain poorly understood.
56 This work demonstrates that individual cells move at different speeds and/or at different times, as
57 indicated by flow fabrics and longitudinal shears within, and on the top surface of, MTCs (*sensu* Bull
58 et al. 2009).

59 Here, we study a recent MTC, the Gorgon Slide (hereafter the 'slide') that contains flow cells (Fig. 1).
60 We use five, high-quality 3D seismic reflection datasets from the Exmouth Plateau, offshore NW
61 Australia (Figs. 1B-C). The 3D seismic reflection datasets cover most of the Gorgon Slide area, including
62 evacuation and deposition zones, which enable us to characterise the slide from its head to toe (Fig.
63 1C). As the slide is at, or just below, the seabed, detailed seismic attribute analysis allows us to: (i)
64 document kinematic indicators on basal-shear and top surfaces, and within internal body, of the slide;
65 (ii) reconstruct emplacement processes of the slide whose kinematic indicators serve as evidence of
66 flow cells; (iii) infer the impact of flow cells on flow behaviour; (iv) discuss potential factors controlling
67 the formation of the flow cells.

68 **DATA AND METHODOLOGY**

69 We use five, high-quality post-stack time-migrated (PSTM) 3D seismic reflection datasets that image
70 the evacuation and most of the deposition zones of the Gorgon Slide (Figs. 1B-C). Only a small part of
71 the slide (7%, i.e. 166 km² of 1760 km² total area) is not imaged by the 3D seismic reflection data.
72 Thus, three 2D seismic reflection lines were used to infer the downdip limit of the deposition zone
73 (green lines in Fig. 1C). The vertical resolution of the 3D seismic reflection data at the base of the slide

74 (c. 500 mbsf) ranges from 8-11 m, based on near seabed sediment velocity and dominant frequency
75 of 1824 m/s and 40-60 Hz, respectively. Bin spacing of the 3D seismic volumes ranges from 12.5 x
76 18.75 m to 20 x 25 m (see Appendix 1 for details). Depth conversion of seabed and basal-shear surface
77 time structure maps was conducted by using average velocities of water (1519 m/s) and near seabed
78 sediment (1824 m/s), respectively (Appendix 1). The average water velocity is constrained by ten
79 industry wells (Fig. 1B), and the near seabed sediment velocity data is available from well ODP 762
80 (see Figs. 1A and 2).

81 The seabed and basal-shear surface of the Gorgon Slide were mapped to gain insights on the
82 kinematics of the slide during transport. We also employed an iso-proportional slicing method (Zeng
83 et al. 1998), midway between the seabed and basal-shear surface of the slide, to visualise
84 heterogeneity of internal seismic facies. Several seismic attributes were used in this analysis,
85 particularly: (i) *variance* to better image discontinuities (Chopra and Marfurt 2007), including grooves
86 on the basal-shear surface of an MTC (e.g. Bull et al. 2009); (ii) *RMS Amplitude* to better delineate
87 features that have distinct positive or negative amplitudes resulting from an acoustic (velocity and
88 density) contrast (Brown 2011), such as megaclasts encased within transparent debrite (e.g. Ortiz-
89 Karpf et al. 2017); (iii) *dip* to better image rugosity of a surface (Brown 2011), including seabed relief
90 (e.g. Bull et al. 2009); and (iv) *spectral decomposition* of seismic reflection volume was used to
91 highlight internal heterogeneities of a geological body (Partyka et al. 1999; Eckersley et al. 2018).

92

GEOLOGICAL SETTING

93 The Exmouth Plateau is a part of North Carnarvon Basin (NCB), a basin that experienced multiple rifting
94 events (Fig. 2) (Tindale et al. 1998; Longley et al. 2002). Deposition during the subsequent thermal
95 subsidence phase since the Cretaceous was initially dominated by fine-grained siliciclastic sediments,
96 and become carbonate-dominated as the Australian plate drifted northward towards the equator (e.g.
97 Apthorpe 1988; Hull and Griffiths 2002). Progradation of carbonate-dominated clinoforms has

98 persisted from the Oligocene to the present-day (Fig. 2B) (Cathro et al. 2003; Moss et al. 2004). A
99 collision between the Australian and Eurasian plates (Miocene to present-day) has reactivated
100 optimally oriented, pre-existing rift-related faults, forming inversion structures such as the NE-SW
101 trending Exmouth Plateau Arch (Fig. 2) (Keep et al. 1998). Emplacement of MTCs is widespread across
102 the plateau, especially during this inversion period (Boyd et al. 1993; Hengesh et al. 2013; Scarselli et
103 al. 2013), although pelagic carbonate sediments presently dominate the stratigraphy with
104 sedimentation rates as low as 20 m/Ma (Golovchenko et al. 1992). We focus on the Gorgon Slide,
105 which extends from the seabed (blue) down to its basal-shear surface (yellow, see Fig. 2B). Beneath
106 the headwall of the Gorgon Slide, there is a rift-related horst block, which was drilled by Bluebell-1
107 (Fig. 2B) (McCormack and McClay 2013). This horst contains the giant Gorgon gas field, containing 11
108 tcf of gas in place (Clegg et al. 1992).

109 **THE GORGON SLIDE**

110 *General Characteristics*

111 The Gorgon Slide was evacuated from a failed slope that defines part of the present-day shelf-edge of
112 NW Australia (Fig. 1). It was deposited in the adjacent Kangaroo Syncline, forming a lenticular, NW-SE
113 trending feature that wedges-out to the SE (against the continental slope) and to the NW (against the
114 eastern margin of the Exmouth Plateau Arch) (see Figs. 1B-C and 2). The slide has a maximum runout
115 distance of c. 70 km. It is up to 500 m-thick and has a total volume of c. 500 km³ (Fig. 3A) (Nugraha et
116 al. 2019a). The slide terminates against two lateral margins (to the NE and SW), it is c. 30 km-wide in
117 the central part and abruptly narrows to c. 18 km at its frontal end (Fig. 3A). These width changes form
118 two frontal margins (eastern and western), that are separated by a NW-SE trending, c. 10 km-long
119 lateral margin (Fig. 3A). The central and frontal parts of the slide display a notable along-strike change
120 in seabed rugosity, which defines two distinctive regions: Areas A and B (Fig 3B). Area A comprises a
121 highly rugose seabed, indicated by frequent changes of dip, and is bound by the NE lateral and eastern
122 frontal margins (Fig. 3B). Area B comprises a relatively less rugose seabed and is bound by the SW

123 lateral and western frontal margins (Fig. 3B). Areas A and B are separated by a linear, NW-SE trending
124 feature (see zoomed-in image in Fig. 3B), which is subparallel to both lateral margins of the slide. This
125 linear feature is narrow (c. 170-300 m-wide) and extends for c. 26 km, from its updip limit to the point
126 where it merges with Area B lateral margin (see Figs. 3B-C). This feature marks the change of seabed
127 relief between Areas A and B, where the vertical difference of relief between the two areas are
128 variable downdip, ranging from c. 10 m to c. 20 m (Fig. 3C).

129 The Gorgon Slide originated from a steeply-dipping (c. 30°), c. 350 m-high headwall scarp, which forms
130 a c. 18 km-long evacuation zone (Fig. 4A). The frontal margin of Area A is clearly marked by positive
131 seabed relief (c. 30 m) relative to the smooth seabed bounding pre-existing strata immediately
132 downdip. The Gorgon Slide and the pre-existing strata share similar chaotic and transparent seismic
133 facies. Thus, the pre-existing strata are interpreted as an older MTC (Fig. 4A). A strike seismic section
134 of the Gorgon Slide exhibits the following along-strike lateral boundaries (Fig. 4B): (i) the SW lateral
135 margin defines the pinch-out of the slide onto substrate, and (ii) the NE lateral margin that marks the
136 boundary with the older MTC. Sub-parallel, continuous reflections on top of the older MTC are cross-
137 cut and overlain by overspill of the slide. Thus, these reflections are interpreted as 'paleo-seabed' (Fig.
138 4B). The morphology of the basal-shear surface appears to mimic that of the underlying substrate in
139 most of Area B in the SW (Fig. 4B). In Area A, the basal-shear surface truncates underlying seismic
140 reflections and its gradient becomes flatter adjacent to the NE lateral margin.

141 The linear zone on the seabed (see Figs. 3B-C) is interpreted as a longitudinal shear zone (*sensu* Bull
142 et al. 2009), which records internal variations of transport velocity within an MTC. Occasionally, the
143 longitudinal shear zone not only defines the change of seabed relief between Areas A and B, it could
144 also coincide with positive seabed relief (see Fig. 4B).

145 The internal character of the Gorgon Slide is variable, and a seismic facies classification is used to
146 capture this internal heterogeneity (Fig. 5). This classification captures variations in both strain and

147 degree of internal disaggregation (i.e. flow rheology) within the MTCs (e.g. Alves et al. 2014). It builds
148 on previous studies that have calibrated seismic reflection data with lithology from well data (e.g.
149 Sawyer et al. 2009), as well as seismic reflections forward modelling using outcrops (Dykstra et al.
150 2011). Five seismic facies (SF) are defined in this study (Fig. 5A) based on variability in internal
151 reflection configurations, both in cross-section and plan-view (see Figs. 5B-G): (i) SF-1 - mostly
152 transparent with low-to-variable amplitude reflections, which are interpreted as debrites (cf.
153 Posamentier and Kolla 2003; Posamentier and Martinsen 2011; Ortiz-Karpf et al. 2017); (ii) SF-2 -
154 contains low-to-medium amplitude, discontinuous folded reflections in cross-section that occasionally
155 forms sinuous lineations in plan-view, and also interpreted as debrites with partially disaggregated
156 material; (iii) SF-3 - contains high-amplitude folded reflections that are offset by thrusts, interpreted
157 as fold and thrust systems; (iv) SF-4 is characterised by isolated packages of coherent, sub-parallel
158 reflections within a matrix composed of SF-1 and 2, interpreted as megaclasts transported within
159 debritic matrix (cf. McGilvery 2004; Bull et al. 2009; Jackson 2011; Ortiz-Karpf et al. 2017; Hodgson et
160 al. 2018); and (v) SF-5 is composed of sub-parallel and continuous reflections that characterise non-
161 MTC seismic facies, i.e. progradation beneath the shelf and layered slope strata within the evacuation
162 zone (see Fig. 4A) (e.g. Prélat et al. 2015). Internal seismic facies of the longitudinal shear zone is
163 generally characterised by SF-1 (Fig. 3C), indicating highly disaggregated materials within the zone due
164 to intense shearing (cf. Ogata et al. 2014; Bull and Cartwright 2019; Omeru and Cartwright 2019).
165 However, its relatively subtle width means that the internal seismic facies are often undifferentiated
166 from adjacent MTC bodies (Figs. 4B, and 5B-E).

167 To reconstruct emplacement processes, the Gorgon Slide is synthesized in terms of its kinematic
168 indicators and internal seismic facies from four domains: (i) headwall; (ii) upper translation (UTD); (iii)
169 lower translation (LTD); and (iv) toe (see Fig. 3B).

170

Headwall domain

171 Besides the main headwall scarp, features observed in the headwall domain (see Fig. 6) include a small
172 scarp, circular depressions, and linear depressions updip from the main headwall scarp (see zoomed-
173 in image in Fig. 6A). The small scarp (c. 10 m-high) is likely to be older than the main headwall scarp
174 due to their cross-cutting relationship (Fig. 6B). The circular depressions have a diameter of c. 100 to
175 300 m, which are mainly distributed within the small scarp area. They are interpreted as pockmarks
176 (e.g. Hengesh et al. 2013; Scarselli et al. 2013), which would indicate active fluid venting. The linear
177 depressions are c. 3 to 5 km-long and c. 15 m-deep, which are interpreted as crown cracks possibly
178 marking the location of future slope failure (Varnes 1978; Frey-Martinez et al. 2005).

179 In between the main headwall scarp and the evacuation-deposition zone boundary, within the
180 evacuation zone, there are c. 5-16 km-long elongate features (Figs. 6A-B) with v-shaped geometries
181 (c. 150-300 m-wide and c. 10-25 m-deep, see zoomed-in image in Fig. 6B). We interpret them as
182 grooves (*sensu* Bull et al. 2009) that were formed due to tooling of failed materials (e.g. megaclasts)
183 into the substrate during transport (Posamentier and Martinsen 2011; Ortiz-Karpf et al. 2017;
184 Hodgson et al. 2018; Sobiesiak et al. 2018). Based on their orthogonal relationship with the headwall
185 scarp, these grooves are a reliable indicator of the translation pathway of the slide through the
186 evacuation zone.

187 The initial failed volume of the Gorgon Slide that was removed from the headwall domain ranges from
188 31 to 43 km³, which is 12-16 times smaller than the deposited volume (c. 500 km³) (Nugraha et al.
189 2019a). This volume discrepancy is interpreted as a result of significant erosion and substrate
190 entrainment of the carbonate ooze substrate during transport (see Nugraha et al. 2019a).

191

Upper translation domain

192 **Basal-shear surface.**– Grooves observed in the updip part of the upper translation domain (Fig. 7A),
193 are a continuation of those within the evacuation zone (see Fig. 6), displaying similar dimensions and
194 geometries (see Headwall domain section). However, the grooves in this domain converge downslope
195 towards the NE lateral margin (Fig. 7A), which contrasts to the more commonly described downslope-
196 diverging grooves (e.g. Posamentier and Kolla 2003; McGilvery 2004; Ortiz-Karpf et al. 2017). This
197 geometry implies that the pathway of the slide was focused towards the steep, NE lateral margin. As
198 a result, the slide is thickest adjacent to this lateral margin (see Fig. 3A). The pathway was likely
199 controlled by the morphology of the basal-shear surface that broadly follows the morphology of the
200 underlying substrate (see Fig. 4B). This supports the observations of Ortiz-Karpf et al. (2017), who
201 stress the impact of precursor morphology on the emplacement of MTCs.

202 In the central part of the basal-shear surface, there is a pair of NW-SE trending, curved lineations
203 across the upper translation domain (Fig. 7A). They bound an area that is shallower than its
204 surrounding area (shaded grey in Fig. 7A). These lineations appear to mark subtle changes (c. 10 m) in
205 the depth of the basal-shear surface. We interpret these lineations as 'ramps' bounding an area called
206 a 'flat' (Trincardi and Argnani 1990; Lucente and Pini 2003; Frey-Martinez et al. 2005; Bull et al. 2009).
207 The ramps record basal erosion by the overlying slide that are commonly expressed by truncated
208 reflections of underlying substrate by a basal-shear surface (e.g. Bull et al. 2009). However, as the
209 ramps in this domain represent relatively small steps (i.e. 10 m), the basal-shear surface does not
210 truncate more than one reflector.

211 Adjacent to the NE lateral margin, there is an area comprising highly discontinuous reflections on the
212 variance map (Fig. 7A). Some of these discontinuous reflections form lineations oriented oblique to
213 the NE lateral margin. This area is characterised by low-to-medium amplitude, discontinuous
214 reflections at, and immediately beneath, the basal-shear surface (Fig. 7D). We interpret the substrate

215 in this area to have been compressionaly deformed due to stress exerted by the slide, forming a
216 'basal-shear zone' (Butler and McCaffrey 2010; Hodgson et al. 2018; Cardona et al. 2020).

217 On top of the older MTC, there is a series of lineations (0.5 to 1.5 km-long) that originate from the NE
218 lateral margin (Figs. 7A and D). These lineations are oriented at c. 45° relative to the NE lateral margin.
219 We interpret these lineations as shear fractures (i.e. Riedel shears) that developed due to strike-slip
220 movement along the NE lateral margin. This implies that the Gorgon Slide exerted stress onto the
221 older MTC during transport (e.g. Fleming and Johnson 1989; Martinsen 1994; Fossen 2016).
222 Northwestwards transport of the Gorgon Slide implies that the NE lateral margin represents a dextral
223 strike-slip fault of the Gorgon Slide relative to the older MTC (Fig. 7A). Fleming and Johnson (1989)
224 suggest that this type of fractures is developed during an early stage of strike-slip faulting along lateral
225 margin of the MTCs, prior to the formation of through-going lateral margins. They recorded fractures
226 oriented at 45° clockwise from the trend of a dextral lateral margin, similar to the shear fractures
227 found in our study.

228 **Internal body.**– The proximal part of the upper translation domain is dominated by SF-1 that
229 surrounds scattered megaclasts (SF-4) (Fig. 7B). These megaclasts have elliptical to rectangular
230 geometry in map-view, with long-axis lengths ranging from c. 0.18 to 1 km and thickness of c. 70 to
231 140 m (Fig. 7B). Seismic sections show that these megaclasts are sometimes internally folded and
232 faulted (Fig. 7D). In the central part of this domain, the megaclasts are concentrated, forming a c. 15
233 km-long and 3 km-wide, convex-upslope cluster of megaclasts. This cluster is bound by a gradational
234 boundary with SF-1 in the E, and an abrupt boundary in the W, which is defined by the longitudinal
235 shear zone (Fig. 7B). Most of this cluster occurs within Area A, with a subsidiary megaclasts cluster
236 observed within Area B, c. 5 km downdip (Fig. 7B). These clusters share similar frequency expressions
237 on spectral decomposition map (Fig. 8A), and internal reflections and thicknesses (Figs. 8B-C), and are
238 separated by the longitudinal shear zone.

239 Immediately downdip from the central part of the Area A megaclast cluster, partially-disaggregated
240 materials contained within SF-2 are aligned to form a series convex-upslope bands (Fig. 7B). These
241 bands are sub-parallel to the geometry of the cluster (Figs. 7B and 8A). In contrast, downdip from the
242 eastern margin of the cluster, the bands show a convex-downslope geometry terminating at the NE
243 lateral margin (Figs. 7B and 8A). A NW-SE trending, narrow area (c. 500 m-wide and c. 10 km-long) of
244 SF-1 defines the boundary between these two sets of bands (Fig. 7B). There is a similar occurrence of
245 convex-downslope bands downdip from the cluster in Area B (Fig. 7B). In seismic section, these bands
246 are expressed as low-frequency, medium-amplitude folded reflections (Fig. 7D).

247 The scattered megaclasts in the proximal part of this domain are clustered, possibly due to downslope-
248 convergence of the pathway of the Gorgon Slide based on the orientation of the grooves on the basal-
249 shear surface (see Fig. 7A). The clusters of megaclasts in Areas A and B are interpreted to have been
250 initially emplaced as a single cluster. Subsequently, they were cross-cut by the longitudinal shear zone,
251 which was also initiated at this area (Fig. 7B).

252 The clusters of megaclasts were likely induced internal velocity perturbations within the slide during
253 transport. This internal velocity variation is evidenced by the convex-downslope shear bands, in both
254 Areas A and B, which are located downflow from the convex-upslope shear bands adjacent to the
255 cluster of megaclasts in Area A (Fig. 7B). Another indicator of internal velocity variations is the narrow
256 area within Area A that separates the convex-downslope and -upslope shear bands (Fig. 7B). This area
257 is interpreted as an 'internal shear zone' (cf. Ogata et al. 2014; Bull and Cartwright 2019; Omeru and
258 Cartwright 2019), containing disaggregated material due to intense shearing. Other studies have also
259 discussed how the entrainment and abrasion of megaclasts during transport of MTCs could affect flow
260 rheology (e.g. Joanne et al. 2013; Ortiz-Karpf et al. 2017; Hodgson et al. 2018; Sobiesiak et al. 2019),
261 and therefore variations in intra-MTC flow velocity.

262 **Top surface.**– The top surface of the Gorgon Slide sometimes enhances the appearance of some
263 features recorded within the internal body (Fig. 7C). For example, the internal shear bands are
264 expressed as ridges with positive seabed relief. These ridges are terminated at the longitudinal shear
265 zone between Areas A and B, and change their orientations (from convex-upslope to -downslope) at
266 the internal shear zone (see Figs. 7C and 8). These ridges are interpreted as a secondary flow fabric
267 (*sensu* Bull et al. 2009), suggesting flow velocity variation within the slide.

268 *Lower translation domain*

269 **Basal-shear surface.**– The majority of kinematic indicators observed in the upper translation domain
270 extend to this lower translation domain (i.e. ramp, deformed substrate and shear fractures; Fig. 9A),
271 apart from the absence of grooves. Here, the ramps are deeper (c. 20 m-deep, Fig. 9D), and lineations
272 within the deformed substrate area are more apparent on the variance map (Fig. 9A). Downflow from
273 the deformed substrate, there are SE-facing ramps (i.e. perpendicular to transport direction) that
274 merge with the ramp that extends from the upper translation domain (Fig. 9A). Adjacent to the
275 deformed substrate, there is a concentration of shear fractures that diminish downdip.

276 The ramps, deformed substrate, and shear fractures indicate that erosion and deformation also
277 occurred in this lower translation domain (Fig. 9A). There is a close spatial relationship between the
278 deformed substrate and the concentration of shear fractures (Fig. 9A), which also coincides with the
279 thickest slide occurrence in Area A (Fig. 3A). This could imply that the basal and lateral substrate
280 deformation was more severe due to increased stress exerted by the thickest part of the Gorgon Slide.
281 However, Cardona et al. (2020), concluded from an outcrop study that there is no statistical
282 correlation between the intensity of deformation of the basal-shear zone and the thickness of the
283 overlying MTC.

284 **Internal body.**— The convex-upslope shear bands, between the longitudinal and internal shear zones
285 within Area A, diminish downflow (Fig. 9B). In contrast, the convex-downslope shear bands within
286 Area B continue and are more prominent in this lower translation domain (Fig. 9B).

287 Adjacent to the NE lateral margin, there is another cluster of megaclasts (Fig. 9B). In seismic section
288 (Fig. 9D), this cluster contains megaclasts that have similar facies to those in the upper translation
289 domain (Figs. 7D and 8B-C). However, these megaclasts have shorter long-axis (*c.* 0.05 to 0.54 km) and
290 are thicker (*c.* 73 to 220 m) than those in the upper translation domain, i.e. *c.* 0.17 to 0.98 km-long
291 and *c.* 70 to 137 m-thick (see Fig. 10A). Also, the megaclasts in the upper and lower translation
292 domains show different trends of their long-axis (Fig. 10B). The megaclasts in the upper translation
293 domain are generally trending NE-SW, and the ones in the lower translation domain are trending
294 NNW-SSE. We also found megaclasts that are concentrated in the basal part of the slide, which are
295 (see Fig. 9D): (i) containing chaotic and transparent internal reflections, (ii) bound by folded top, and
296 (iii) underlain by a ramp. We name them as 'basal megaclasts'.

297 The presence of longitudinal and internal shear zones suggest that internal variation of flow velocity
298 continued to occur in this domain (Fig. 9B). Between these shear zones, the gradual downflow
299 disappearance of the convex-upslope shear bands suggests a decrease in internal velocity
300 perturbations induced by the cluster of megaclasts in upper translation domain (see Fig. 7B).

301 The cluster of megaclasts adjacent to the NE lateral margin (Figs. 9B and D) are located immediately
302 downflow from, and have similar width (2.5 km) to, the deformed substrate area (Fig. 9A). Thus, the
303 basal and lateral substrate deformations documented on the basal-shear surface (Fig. 9A) could be
304 related to this cluster of megaclasts, instead of reflecting the maximum thickness of the Gorgon Slide
305 (Fig. 3A). This interpretation is supported by the same observation from the Rapanui MTD (Cardona
306 et al. 2020), where the thickness of the deformed substrate is correlated to higher concentrations of
307 rafted blocks (i.e. megaclasts), and not the thickness of overlying MTC. The higher concentration of

308 megaclasts indicates an increase in flow competence overriding the area of the deformed substrate.
309 In addition, the long-axis orientations of the megaclasts in the lower translation domain are generally
310 oblique to sub-parallel, in contrast to the ones in the upper translation domain that are generally
311 perpendicular, to the transport direction (Fig. 10B). Their long-axis orientations are likely to be
312 controlled by velocity gradients (Mazzanti and De Blasio 2010), where the megaclasts in the lower
313 translation domain are adjacent to the NE lateral margin, and experienced abrupt change of velocity
314 gradient as they moved against the stationary older MTC (Fig. 9B). In contrast, the cluster of
315 megaclasts in the upper translation domain experienced a lower velocity gradient, which then formed
316 the convex-upslope geometry (Fig. 7B).

317 The basal megaclasts have transparent internal facies and display folded tops (Fig. 9D). We suggest
318 they are more deformed compared to the adjacent megaclasts (Fig. 9D). Their transparent internal
319 facies could be related to intense shearing during transport (Alves 2015; Gamboa and Alves 2015), and
320 the folded top may be formed due to impingement of the megaclasts by the slide against the
321 underlying ramp (see Fig. 9D) (Jackson 2011). This impingement is expressed as a positive relief of the
322 seabed (Fig. 9D).

323 **Top surface.**– The top surface shows that the internal shear zone merges with the longitudinal shear
324 zone in the distal part of the lower translation domain (Fig. 9C). These shear zones outline an area
325 covering convex-upslope ridges that narrows and diminishes downflow. Consequently, Area A
326 becomes dominated by the convex-downslope ridges (Fig. 9C). However, we can see that immediately
327 downflow from the merging point of the shear zones, the ridges in Area A have convex-upslope
328 geometries, most notably adjacent to the longitudinal shear zone (Fig. 9C).

329 The top surface supports the interpretation of kinematic indicators within the internal body of the
330 slide (Figs. 9B-C). Here, it is also evident that velocity perturbation induced by the cluster of megaclasts
331 in upper translation domain (Figs. 7B and 8A) had decreased, and diminished downflow, as clearly

332 marked by the merging of the two shear zones (Fig. 9C). However, downflow from the merging point,
333 the presence of convex-upslope ridges within Area A (terminating at the longitudinal shear zone)
334 suggest that internal velocity perturbations continued to occur (Fig. 9C).

335 *Toe domain*

336 **Basal-shear surface.**– The basal-shear surface in this domain serves as the frontal margin of Area A,
337 and swings through 90° to join the lateral margin of Area B that continues downdip, beyond the 3D
338 seismic reflection data area (Fig. 11A). The deformation style of the substrate (Fig. 11A) resembles
339 that of the upper and lower translation domains (see Figs. 7A and 9A). In the SW part of this domain,
340 there is a c. 30 m-high ramp (Figs. 10A and D), which is more profound than that in the lower
341 translation domain (c. 20 m).

342 The geometry of the basal-shear surface indicates that Area B extends further downdip than Area A
343 (Fig. 11A). The deformed substrate and the ramp indicate that substrate deformation and erosion
344 continued to occur beneath the main body of the slide, despite being located the furthest from the
345 headwall.

346 **Internal body.**– SF-1 and SF-3 (i.e. folds and thrusts system) dominate the distal part of Area A and B,
347 respectively (Fig. 11B). The thrusts within Area B dip to the SE, sub-parallel to the transport direction
348 of the slide (see Figs. 5G and 11B and D).

349 Within the older MTC (see Fig. 11B), there is a cluster of megaclasts (c. 2.5 km-wide and c. 5 km-long).
350 The SE and SW margins of this cluster of megaclasts are defined by Area A frontal margin and Area B
351 lateral margin, respectively (Fig. 11B). Within the cluster of megaclasts, there are lineations (NNW-SSE
352 trending) that are broadly perpendicular to the orientation of the Area A frontal margin and the
353 thrusts within Area B (i.e. NE-SW trending). In seismic section, these megaclasts are characterised by
354 medium-amplitude sub-parallel reflections at the base, which become folded towards the top (Fig.

355 11D). The internal reflections are separated by NE-dipping thrusts. The lineations on the time-slice
356 (Fig. 11B) correspond to these thrusts. Thus, we name this cluster of megaclasts as 'thrusted
357 megaclasts'.

358 Abrupt truncation of the thrusted megaclasts by the frontal margin of Area A, and the thrusts within
359 Area B, suggests a cross-cutting relationship (Fig. 11B). Thus, we interpret that the thrusted megaclasts
360 had been emplaced at their present location prior to the emplacement of the Gorgon Slide. Some
361 thrusted megaclasts (i.e. indicated by high RMS amplitude) are observed within the frontal part of
362 Area A (Fig. 11B). However, these thrusted megaclasts are distinctly different from those of the thrust
363 system within Area B (Fig. 11D). This suggests that the thrusted megaclasts were only entrained by
364 the Gorgon Slide in the frontal part of the Area A (Fig. 11B). In contrast, the Area B lateral margin
365 developed along the SW margin of the thrusted megaclasts, without any evidence of entrainment of
366 the thrusted megaclasts by the slide within Area B.

367 The longitudinal shear zone that extends from the upper translation domain (see Figs. 7A and 8A) joins
368 Area B lateral margin in the toe domain (Fig. 11B). This may indicate a relationship between the
369 thrusted megaclasts and inferred intra-MTC velocity perturbation. Specifically, the velocity
370 perturbation could have originated from the SW margin of the thrusted megaclasts (i.e. Area B lateral
371 margin) and propagated upflow (Fig. 11B). Therefore, this velocity perturbation caused by the
372 thrusted megaclasts connected with the downflow-propagating velocity perturbation induced by the
373 cluster of megaclasts in the upper translation domain (Figs. 7B and 8A). Frey-Martínez et al. (2006)
374 also observed similar role of pre-existing blocks (megaclasts), where a single MTC flow bifurcates to
375 form two flows with different transport directions.

376 **Top surface.**– The rugose seabed with c. 30 m-high ridges relative to the flat seabed above the older
377 MTC defines the frontal margin of Area A (see Fig. 11C). The vertical relief of the ridges of Area A is
378 higher than in both Area B (c. 10 m, Fig. 11C) and the lower translation domain (c. 10 m, Fig. 9).

379 The ridges at the Area A frontal margin could indicate a buttressing effect of the slide against the
380 thrusting megaclasts (Fig. 11C), which then formed ridges that decrease in height upflow. In contrast,
381 the ridges in Area B are not as high as those in Area A. Thus, the slide was not buttressed against the
382 thrusting megaclasts and it translated further downdip (Fig. 11C). These differential processes in the
383 toe domain of Areas A and B reflect the merging between Area B lateral margin and the longitudinal
384 shear zone (Fig. 11C).

385 DISCUSSION

386 *Emplacement processes of the Gorgon Slide: a multi-cell flow emplacement mechanism*

387 A multiple flow cell model based on field studies has been proposed by Alsop and Marco (2014), who
388 advance the notions of Farrell (1984) of a large single-cell flow model that controls deformation
389 patterns within an MTC. They suggest that a large (first-order) MTC consists of a number of smaller
390 (second-order) flow cells formed during emplacement of the MTC. These smaller flow cells may
391 interact with each other and cause overprinting on initially formed structures. Interaction between
392 flow cells has also been documented from sonar (e.g. Prior et al. 1984; Masson et al. 1993; Gee et al.
393 2001) and 3D seismic reflection data (e.g. Bull et al. 2009; Steventon et al. 2019), and captured in the
394 form of primary (longitudinal shear zone) and secondary (sinuous shear bands) flow fabrics (*sensu* Bull
395 et al. 2009). These kinematic indicators indicate differential speeds and/or timing of downslope
396 translating material (Masson et al. 1993; Gee et al. 2005).

397 In this study, the Gorgon Slide appears to comprise two intra-MTC (second-order) flow cells. These are
398 represented physically by Areas A and B, and for the purpose of this process-based interpretation are
399 re-named as Cells A and B, respectively. The emplacement processes of the Gorgon Slide are captured
400 in a schematic model that recognises three stages of development (Fig. 12).

401 **Stage 1.**— Prior to slope degradation, a surface rupture might have been triggered by two main factors
402 (Fig. 12A). First, the normal faults bounding the horst could have been inverted due to compression

403 to destabilise the slope (Keep et al. 1998; Nugraha et al. 2019b). Second, the existence of pockmarks
404 observed on the seabed (see Fig. 6) implies that there has been active fluid venting in the headwall
405 area (Hengesh et al. 2013), most likely originating from the deeper gas-bearing horst block of the
406 Gorgon Field (Fig. 2). Gas leakage into shallower sediments could have lowered the shear strength of
407 these sediments, and primed the slope for subsequent failure (Scarselli et al. 2013). However, the
408 Gorgon Slide was not an isolated occurrence, but rather the most recent. Previous collapse of the
409 continental margin is reflected in the older (pre-Gorgon) MTC, which would have left a remnant
410 topography both on the slope but especially in the area of subsequent Gorgon Slide deposition. Most
411 notably, thrust megaclasts had already been emplaced in the vicinity of the future toe domain of
412 the Gorgon Slide (Fig. 12A).

413 **Stage 2.**— The arcuate geometry of the main headwall scarp indicates that the failed sediments were
414 evacuated during a single mass-transport event (see Fig. 6). The evacuated sediments might include
415 megaclasts derived from either the headwall and/or entrained from the layered slope substrate (Figs.
416 11B and 4A). During translation, the megaclasts could be deformed and fragmented (e.g. Gee et al.
417 2005; Alves 2015).

418 The downslope-converging grooves within the headwall and upper translation domains suggest a
419 convergent pathway of the slide, resulting in the clustering of the megaclasts (Fig. 12B). In the lee-side
420 of the cluster of megaclasts, the following features were formed: (1) convex-upslope shear bands
421 within the slide, and (2) convex-upslope ridges on top of the slide. These features indicate slower
422 transport velocity in and around the area of concentrated megaclasts (Fig. 12B). Higher transport
423 velocities of flows moving around the megaclast-rich area led to the formation of the longitudinal
424 shear zone, and the initiation of Cells A and B. The cluster of megaclasts effectively acted as an obstacle
425 to the initial, single-cell flow. Other studies have also documented such mechanism, where the
426 geometry of shear bands and ridges of MTCs downslope from translating megaclasts suggest slower-

427 moving flows than surrounding materials (e.g. Masson et al. 1993; Lastras et al. 2005; Gee et al. 2006;
428 Bull et al. 2009).

429 **Stage 3.**– The downslope propagation of the basal-shear surface was coupled with the evolution of
430 the internal body and top surface of the slide. The area covering the convex-upslope shear bands and
431 ridges narrowed downslope (Fig. 12C), which suggests a reduction in the influence of the cluster of
432 megaclasts on slowing down the flow of material in its lee-side. Thus, we interpret this area as a
433 'shadow zone'. The shadow zone is bound by the longitudinal shear zone separating the two cells, and
434 the internal shear zone within Cell A (Fig. 12C). The shadow zone is an example of how megaclasts
435 influence flow processes of an MTC (e.g. Masson et al. 1993; Lucente and Pini 2003; Jackson 2011;
436 Hodgson et al. 2018).

437 Downflow from the shadow zone, ridges of Cell A show convex-upslope geometries adjacent to the
438 longitudinal shear zone. In contrast, ridges of Cell B consistently exhibit convex-downslope geometries
439 (Fig. 12C). These geometries indicate that Cell A resisted the downslope translation of Cell B. As a
440 result, the ridges of Cell A were dragged downslope, and the ridges of Cell B were dragged upslope
441 (Fig. 12C). Therefore, we suggest that Cell A was travelling more slowly than Cell B. Furthermore, the
442 high seabed relief of Cell A at the frontal margin suggest that it was forced to stop its translation by
443 the pre-existing thrusting megaclasts (Fig. 12C), and, thus, can be considered as "stopping structures"
444 that were formed during cessation (Masson et al. 1993; Gee et al. 2006). In contrast, the position of
445 the thrusting megaclasts allowed Cell B to translate further downdip than Cell A. Thus, Cell B was not
446 only travelled faster, but also further, than Cell A.

447 *Impact of flow cells formation on MTCs flow behaviour*

448 Submarine debris flow can travel for tens to hundreds of km across low gradient (c. <1°) continental
449 slope, despite its cohesive nature (Gee et al. 1999; Lastras et al. 2005). This mobility can be explained
450 by, for instance, sustained pore-fluid pressure within the flow during transport (Major and Iverson

451 1999; McArdell et al. 2007), and also the presence of a thin lubricating layer of fluid at the base of the
452 frontal part of the flow (i.e. hydroplaning, Mohrig et al. 1998). Ultimately, a debris is formed by *en*
453 *masse* freezing of the debris flow (e.g. Talling et al. 2012), where materials at flow margins (i.e. frontal
454 and lateral) cease moving first, followed by materials in the main body of the flow (Iverson 1997).

455 The Gorgon Slide provides evidence of how a mass flow split into two smaller flow cells (Cell A and B,
456 Fig. 12). The relationship between the two cells suggests that Cell A ceased movement, while Cell B
457 was still in motion. This suggests that *en masse* freezing did not occur across the entire body of the
458 flow synchronously. Instead, individual flow cells underwent differential timing of freezing, resulting
459 in different runout distance of the cells. We propose that lateral friction and related pore-fluid
460 pressure played an important role in controlling the runout distance of the two cells, in addition to
461 the presence of pre-existing thrusting megaclasts. The longitudinal shear zone may have sustained
462 excess pore-fluid pressure between the two cells, such that low friction between the two cells could
463 be maintained, allowing Cell B to keep in translation despite the impediment by Cell A. In contrast,
464 pore-fluid pressure was likely to dissipate at the lateral margins during translation (e.g. NE lateral
465 margin, Fig. 3). The lack of excess pore-fluid pressure would have resulted in high friction between the
466 moving slide (e.g. Cell A) and stationary lateral substrate (i.e. the older MTC). This high friction at the
467 lateral margin was likely to reduce runout distance more significantly than the friction at the
468 longitudinal shear zone. Such mechanisms are also observed from experimental studies (Major and
469 Iverson 1999; De Haas et al. 2015).

470 Our results suggest that a debris flow comprising smaller flow cells could experience a 'punctuated'
471 freezing, where a flow cell can have shorter runout distance than the others, resulting from differential
472 friction and pore-pressure dissipation at flow cells margins. The flow behaviour documented in our
473 study may be considered for modelling the potential impact of MTCs on subsea infrastructures.

474

Controls on flow cell formation

475 Flow cell formation within an MTC depends on internal velocity perturbations, which are controlled
476 by variations in at least three local factors (Farrell 1984; Alsop and Marco 2011; Alsop and Marco
477 2014): (i) lithology and/or geometry of stratigraphic element (e.g. MTCs, channels and lobes); (ii) fluid
478 pressures within the MTC and/or substrate; and (iii) slope and/or geometry of basal-shear surface
479 underlying the MTC.

480 In the Gorgon Slide, a cluster of megaclasts within a debritic matrix initiated flow cell formation. This
481 implies that lithology, in particular variations of the degree of disaggregation within the slide, play a
482 key role in forming the two seismic-scale flow cells. In addition, the geometry of the basal-shear
483 surface was also important in converging the flow, clustering the megaclasts, initiating velocity
484 perturbation, and terminating the flow cells. Those three local variations (i.e. lithology, fluid pressures,
485 and basal-shear surface geometry) may have been influential prior to emplacement, but their
486 properties could also evolve during translation and cessation of the parent flow (Iverson 1997; Dykstra
487 et al. 2011; Joanne et al. 2013; Alsop and Marco 2014; Ortiz-Karpf et al. 2017; Hodgson et al. 2018).

488

Origin of the pre-existing thrust megacasts

489 We have established that the thrust megacasts are encased by the older MTC, and, thus, had
490 existed in their present position prior to the Gorgon Slide emplacement. Here, we discuss possible
491 origins of the thrust megacasts. Deformations within, and the present location of, the thrust
492 megacasts might indicate that they are either *in-situ* (remnant) or translated megacasts.

493 Relatively continuous, sub-parallel reflections at the base of the thrust megacasts could indicate
494 that they had not been translated (Fig. 11D). This might support the interpretation that they are *in-*
495 *situ* megacasts. However, the NE-dipping thrusts originating from the base, and folded reflections
496 toward the top of the megacasts, record contractional strain as a result of broadly NE-SW trending σ_1

497 stress (Fig. 11D). This stress was unlikely exerted by the Gorgon Slide onto the *in-situ* megaclasts, as
498 the Gorgon Slide was transported towards the NW. Likewise, it is unlikely that the NE-dipping thrusts
499 were formed by an older MTC that translated to the SW. This is because there is no possible source of
500 MTCs towards the NE (see location of the NW Australian shelf, Fig. 1A).

501 If the thrusting megaclasts were to be deformed or translated by an MTC, the MTC should be sourced
502 either from the Exmouth Plateau Arch (i.e. to the SW from the megaclasts) or from the NW Shelf of
503 Australia (see Fig. 1A). As the thrusts of the megaclasts are NE-dipping (Fig. 11D), the thrusting
504 megaclasts are unlikely to be deformed or translated by a NE-flowing MTC originated from the arch.
505 This MTC should produce SW-dipping thrusts. Thus, the MTC forming the thrusting megaclasts was
506 more likely sourced from the NW Shelf, similar to the source of the Gorgon Slide. However, a NW-
507 flowing MTC should generate SE-dipping thrusts, instead of NE-dipping thrusts. This fact suggests that
508 the thrusting megaclasts are unlikely to be deformed *in-situ*. Thus, thrusting megaclasts might be
509 deformed during translation within the NW-flowing MTC, rotated counter-clockwise (*c.* 70-80°) and
510 then rested at their present location.

511 The thrusting megaclasts have similar dimensions (Figs. 13A-B) and seismic facies (Figs. 13C-D), to the
512 basal megaclasts (see Lower translation domain section, Fig. 9D). Thus, it is possible that the thrusting
513 megaclasts were deformed and translated by the NW-flowing MTC. Deformation and rotation of
514 megaclasts during MTC translation have been documented in other studies, such as in Storegga Slide
515 (Bull et al. 2009). Here, the megaclasts were re-oriented from perpendicular to become sub-parallel
516 to transport direction with increasing distance from headwall scarp.

517 **CONCLUSIONS**

518 We use 3D seismic reflection data covering a recent mass-transport complex (MTC), the Gorgon Slide,
519 from the Exmouth Plateau, offshore NW Australia, to investigate how flow cells within an MTC was
520 formed, translated, and finally ceased. This study concludes that:

- 521 1. The Gorgon Slide was evacuated from a steep, NE-SW trending, c. 350 m-high headwall scarp
522 and transported towards the NW. Layered slope strata in this headwall domain are the likely
523 source of megaclasts that are subsequently transported downslope.
- 524 2. In the proximal part of the translation domain, downslope-converging grooves on the basal-
525 shear surface indicate that the pathway of the slide was focused towards its lateral margin in
526 the NE. The convergent pathway of the flow results to the clustering of the megaclasts, whose
527 long-axes are generally trending NE-SW, perpendicular to the transport direction. This cluster
528 of megaclasts became an obstacle to flow, causing velocity perturbation within the slide. The
529 velocity perturbation is recoded within the internal body by convex-upslope shear bands, and
530 on the seabed by convex-upslope ridges. These features indicate slower transport velocity of
531 the cluster of megaclasts and materials in its lee-side. The area of the slower-moving materials
532 narrows downslope, indicating that velocity perturbation caused by the cluster of megaclasts
533 gradually diminished downflow, forming a 'shadow zone'. Transport velocities of flows were
534 higher around the megaclasts, resulting in the formation of longitudinal shear zone and the
535 initiation of two flow cells, namely Cells A and B.
- 536 3. In the distal part of the translation domain, kinematic indicators recorded on the basal-shear
537 surface indicate that erosional and deformational processes occurred. Erosional processes are
538 evidenced by ramp, and deformational processes are evidenced by deformed substrate or
539 basal-shear zone and shear fractures adjacent to the NE lateral margin. The deformed
540 substrate and shear fractures are closely related to the thickest part of the slide, comprising a
541 cluster of megaclasts, with individual megaclasts generally trending NNW-SSE, oblique to sub-
542 parallel to the transport direction. Shear bands within the slide and ridges on the seabed of
543 Cell A were dragged downslope, while those of Cell B were dragged upslope. This points to
544 Cell A acting as an impediment to the movement of faster-moving Cell B.
- 545 4. In the toe domain, the frontal margin of Cell A is marked by positive seabed relief (c. 30 m-
546 high) that gradually decreases upflow, which is significantly higher than the relief of Cell B (c.

547 10 m-high). This suggests that Cell A was buttressed against a pre-existing cluster of
548 megaclasts (i.e. encased by older MTC), while Cell B was not. Therefore, as there were no flow
549 obstacles, Cell B was able to travel further than Cell A.

550 5. The morphology of the basal-shear surface and the degree of disaggregation within the slide,
551 especially the megaclasts, played important roles in flow cell evolution. The basal-shear
552 surface controlled the pathway of the slide, and, the clustering of the megaclasts. The
553 megaclast clusters then induced internal velocity perturbation that could result in the
554 initiation and cessation of intra-MTC flow cells.

555 6. *En masse* freezing was unlikely to occur throughout the body of the Gorgon Slide at the same
556 time. Instead, 'punctuated' freezing, where Cell A has halted while Cell B was still in motion,
557 occurred due to differential friction and pore-fluid pressure dissipation at flow cells margins.
558 For instance, excess pore-fluid can be maintained within the longitudinal shear zone, so that
559 Cell B only experienced minimal friction despite Cell A impeding its movement. In contrast,
560 excess pore-fluid pressure was likely to dissipate at lateral margins (e.g. the NE lateral margin
561 separating Cell A and stationary substrate). Thus, Cell A experienced higher lateral friction
562 than that of Cell B, resulting in reduced runout distance. This punctuated freezing mechanism
563 may be considered for modelling the impact of MTCs on submarine infrastructures.

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569 **CONFLICT OF INTEREST**

570 No conflict of interest declared.

571

REFERENCES

572 ALSOP, G.I., AND MARCO, S., 2011, Soft-sediment deformation within seismogenic slumps of the Dead
573 Sea Basin: *Journal of Structural Geology*, v. 33, p. 433-457.

574 ALSOP, G.I., AND MARCO, S., 2014, Fold and fabric relationships in temporally and spatially evolving
575 slump systems: A multi-cell flow model: *Journal of Structural Geology*, v. 63, p. 27-49.

576 ALVES, T.M., KURTEV, K., MOORE, G.F., AND STRASSER, M., 2014, Assessing the internal character, reservoir
577 potential, and seal competence of mass-transport deposits using seismic texture: A
578 geophysical and petrophysical approach: *AAPG Bulletin*, v. 98, p. 793-824.

579 ALVES, T.M., 2015, Submarine slide blocks and associated soft-sediment deformation in deep-water
580 basins: A review: *Marine and Petroleum Geology*, v. 67, p. 262-285.

581 APTHORPE, M., 1988, Cainozoic depositional history of the North West Shelf: The North West Shelf,
582 Australia: *Petroleum Exploration Society of Australia*, p. 55-84.

583 BOYD, R., WILLIAMSON, P., AND HAQ, B., 1993, Seismic Stratigraphy and Passive-Margin Evolution of the
584 Southern Exmouth Plateau, *in* Posamentier, H.W., Summerhayes, C.P., Haq, B.U., and Allen,
585 G.P., eds., *Sequence Stratigraphy and Facies Associations*: Oxford, Blackwell Scientific
586 Publications, p. 579-603.

587 BROWN, A.R., 2011, Interpretation of three-dimensional seismic data: *AAPG Memoir 42*, SEG
588 Investigations in Geophysics No. 9: Tulsa, The American Association of Petroleum Geologists
589 and the Society of Exploration Geophysicists.

590 BULL, S., CARTWRIGHT, J., AND HUUSE, M., 2009, A review of kinematic indicators from mass-transport
591 complexes using 3D seismic data: *Marine and Petroleum Geology*, v. 26, p. 1132-1151.

592 BULL, S., AND CARTWRIGHT, J.A., 2019, Line length balancing to evaluate multi-phase submarine landslide
593 development: an example from the Storegga Slide, Norway: *Geological Society, London*,
594 Special Publications, v. 500.

595 BUTLER, R., AND MCCAFFREY, W., 2010, Structural evolution and sediment entrainment in mass-transport
596 complexes: outcrop studies from Italy: *Journal of the Geological Society*, v. 167, p. 617-631.

597 CARDONA, S., WOOD, L.J., DUGAN, B., JOBE, Z., AND STRACHAN, L.J., 2020, Characterization of the Rapanui
598 mass-transport deposit and the basal shear zone: Mount Messenger Formation, Taranaki
599 Basin, New Zealand: *Sedimentology*.

600 CATHRO, D.L., AUSTIN JR, J.A., AND MOSS, G.D., 2003, Progradation along a deeply submerged
601 OligoceneMiocene heterozoan carbonate shelf: How sensitive are clinoforms to sea level
602 variations?: *AAPG bulletin*, v. 87, p. 1547-1574.

603 CHOPRA, S., AND MARFURT, K.J., 2007, Seismic attributes for prospect identification and reservoir
604 characterization, v. 11, Society of Exploration Geophysicists Tulsa, Oklahoma.

605 CLEGG, L., SAYERS, M., AND TAIT, A., 1992, The Gorgon Gas Field: Chapter 32.

606 DE HAAS, T., BRAAT, L., LEUVEN, J.R., LOKHORST, I.R., AND KLEINHANS, M.G., 2015, Effects of debris flow
607 composition on runout, depositional mechanisms, and deposit morphology in laboratory
608 experiments: *Journal of Geophysical Research: Earth Surface*, v. 120, p. 1949-1972.

609 DOTT, R., 1963, Dynamics of subaqueous gravity depositional processes: *AAPG Bulletin*, v. 47, p. 104-
610 128.

611 DYKSTRA, M., GARYFALOU, K., KERTZNUS, V., KNELLER, B., MILANA, J.P., MOLINARO, M., SZUMAN, M., AND
612 THOMPSON, P., 2011, Mass-transport deposits: Combining outcrop studies and seismic forward
613 modeling to understand lithofacies distributions, deformations, and their seismic stratigraphic
614 expression: *SEPM Special Publication*, v. 96, p. 293-310.

615 ECKERSLEY, A.J., LOWELL, J., AND SZAFIAN, P., 2018, High-definition frequency decomposition: *Geophysical*
616 *Prospecting*, v. 66, p. 1138-1143.

617 FARRELL, S.G., 1984, A dislocation model applied to slump structures, Ainsa Basin, South Central
618 Pyrenees: *Journal of Structural Geology*, v. 6, p. 727-736.

619 FISHER, R.V., 1983, Flow transformations in sediment gravity flows: *Geology*, v. 11, p. 273-274.

620 FLEMING, R.W., AND JOHNSON, A.M., 1989, Structures associated with strike-slip faults that bound
621 landslide elements: *Engineering Geology*, v. 27, p. 39-114.

622 FOSSEN, H., 2016, *Structural geology*, Cambridge University Press.

623 FREY-MARTINEZ, J., CARTWRIGHT, J., AND HALL, B., 2005, 3D seismic interpretation of slump complexes:
624 examples from the continental margin of Israel: *Basin Research*, v. 17, p. 83-108.

625 FREY-MARTÍNEZ, J., CARTWRIGHT, J., AND JAMES, D., 2006, Frontally confined versus frontally emergent
626 submarine landslides: A 3D seismic characterisation: *Marine and Petroleum Geology*, v. 23, p.
627 585-604.

628 GAMBOA, D., ALVES, T., CARTWRIGHT, J., AND TERRINHA, P., 2010, MTD distribution on a 'passive' continental
629 margin: The Espírito Santo Basin (SE Brazil) during the Palaeogene: *Marine and Petroleum*
630 *Geology*, v. 27, p. 1311-1324.

631 GAMBOA, D., AND ALVES, T.M., 2015, Three-dimensional fault meshes and multi-layer shear in mass-
632 transport blocks: Implications for fluid flow on continental margins: *Tectonophysics*, v. 647, p.
633 21-32.

634 GEE, M., MASSON, D., WATTS, A., AND ALLEN, P., 1999, The Saharan debris flow: an insight into the
635 mechanics of long runout submarine debris flows: *Sedimentology*, v. 46, p. 317-335.

636 GEE, M., GAWTHORPE, R., AND FRIEDMANN, J., 2005, Giant striations at the base of a submarine landslide:
637 *Marine Geology*, v. 214, p. 287-294.

638 GEE, M., GAWTHORPE, R., AND FRIEDMANN, S., 2006, Triggering and evolution of a giant submarine
639 landslide, offshore Angola, revealed by 3D seismic stratigraphy and geomorphology: *Journal*
640 *of Sedimentary Research*, v. 76, p. 9-19.

641 GEE, M.J., MASSON, D.G., WATTS, A.B., AND MITCHELL, N.C., 2001, Passage of debris flows and turbidity
642 currents through a topographic constriction: seafloor erosion and deflection of flow
643 pathways: *Sedimentology*, v. 48, p. 1389-1409.

644 GOLOVCHENKO, X., BORELLA, P.E., AND O'CONNELL, S.B., 1992, Sedimentary cycles on the Exmouth Plateau,
645 *in* Von Rad, U., Haq, B.U., Kidd, R.B., and O'Connell, S.B., eds., *Proceedings of the Ocean Drilling*
646 *Program, Scientific Results: College Station, TX, Ocean Drilling Program*, p. 279-291.

647 HENGESH, J.V., DIRSTEIN, J.K., AND STANLEY, A.J., 2013, Landslide geomorphology along the Exmouth
648 Plateau continental margin, North West Shelf, Australia: Australian Geomechanics, v. 48, p.
649 71-92.

650 HODGSON, D., BROOKS, H., ORTIZ-KARPF, A., SPYCHALA, Y., LEE, D., AND JACKSON, C.-L., 2018, Entrainment and
651 abrasion of megaclasts during submarine landsliding and their impact on flow behaviour:
652 Geological Society, London, Special Publications, v. 477, p. SP477. 26.

653 HULL, J.N.F., AND GRIFFITHS, C.M., 2002, Sequence stratigraphic evolution of the Albian to Recent section
654 of the Dampier Sub-basin, NorthWest Shelf, Australia: The Sedimentary Basins of Western
655 Australia 3: Proceedings of the Petroleum Exploration Society of Australia Symposium, p. 617-
656 639.

657 IVERSON, R.M., 1997, The physics of debris flows: Reviews of geophysics, v. 35, p. 245-296.

658 JACKSON, C.A., 2011, Three-dimensional seismic analysis of megaclast deformation within a mass
659 transport deposit; implications for debris flow kinematics: Geology, v. 39, p. 203-206.

660 JOANNE, C., LAMARCHE, G., AND COLLOT, J.Y., 2013, Dynamics of giant mass transport in deep submarine
661 environments: the Matakaoa Debris Flow, New Zealand: Basin Research, v. 25, p. 471-488.

662 KEEP, M., POWELL, C., AND BAILLIE, P., 1998, Neogene deformation of the North West Shelf, Australia: The
663 sedimentary basins of Western Australia, v. 2, p. 81-91.

664 LASTRAS, G., DE BLASIO, F.V., CANALS, M., AND ELVERHØI, A., 2005, Conceptual and numerical modeling of
665 the BIG'95 debris flow, western Mediterranean Sea: Journal of Sedimentary Research, v. 75,
666 p. 784-797.

667 LONGLEY, I.M., BUESSENSCHUETT, C., CLYDSDALE, L., CUBITT, C.J., DAVIS, R.C., JOHNSON, M.K., MARSHALL, N.M.,
668 MURRAY, A.P., SOMERVILLE, R., AND SPRY, T.B., 2002, The North West Shelf of Australia - a
669 Woodside Perspective, in Keep, M., and Moss, S.J., eds., The Sedimentary Basins of Western
670 Australia 3: Petroleum Exploration Society of Australia Symposium: Perth, p. 28-88.

671 LUCENTE, C.C., AND PINI, G.A., 2003, Anatomy and emplacement mechanism of a large submarine slide
672 within a Miocene foredeep in the northern Apennines, Italy: A field perspective: *American*
673 *Journal of Science*, v. 303, p. 565-602.

674 MAJOR, J.J., AND IVERSON, R.M., 1999, Debris-flow deposition: Effects of pore-fluid pressure and friction
675 concentrated at flow margins: *Geological Society of America Bulletin*, v. 111, p. 1424-1434.

676 MARTINSEN, O., 1994, Mass movements, *The geological deformation of sediments*, Springer, p. 127-
677 165.

678 MASSON, D., HUGGETT, Q., AND BRUNSDEN, D., 1993, The surface texture of the Saharan debris flow
679 deposit and some speculations on submarine debris flow processes: *Sedimentology*, v. 40, p.
680 583-598.

681 MASSON, D., HARBITZ, C., WYNN, R., PEDERSEN, G., AND LØVHOLT, F., 2006, Submarine landslides: processes,
682 triggers and hazard prediction: *Philosophical Transactions of the Royal Society A:*
683 *Mathematical, Physical and Engineering Sciences*, v. 364, p. 2009-2039.

684 MAZZANTI, P., AND DE BLASIO, F., 2010, Peculiar morphologies of subaqueous landslide deposits and their
685 relationship to flow dynamics, *Submarine mass movements and their consequences*, Springer,
686 p. 141-151.

687 MCARDELL, B.W., BARTELT, P., AND KOWALSKI, J., 2007, Field observations of basal forces and fluid pore
688 pressure in a debris flow: *Geophysical research letters*, v. 34.

689 MCCORMACK, K., AND MCCLAY, K., 2013, Structural Architecture of the Gorgon Platform, North West
690 Shelf, Australia: *The Sedimentary Basins of Western Australia IV: Proceedings of the*
691 *Petroleum Exploration Society of Australia Symposium*, Perth, WA, p. 24.

692 MCGILVERY, T.A.H., GEOFFREY, 2004, Seafloor and shallow subsurface examples of mass transport
693 complexes, *Offshore Brunei, Offshore Technology Conference: Houston*.

694 MOHRIG, D., ELLIS, C., PARKER, G., WHIPPLE, K.X., AND HONDZO, M., 1998, Hydroplaning of subaqueous
695 debris flows: *Geological Society of America Bulletin*, v. 110, p. 387-394.

696 MOSCARDELLI, L., AND WOOD, L., 2008, New classification system for mass transport complexes in
697 offshore Trinidad: *Basin Research*, v. 20, p. 73-98.

698 MOSS, G.D., CATHRO, D.L., AND AUSTIN, J.A., 2004, Sequence biostratigraphy of prograding clinoforms,
699 northern Carnarvon Basin, Western Australia: a proxy for variations in Oligocene to Pliocene
700 global sea level?: *Palaios*, v. 19, p. 206-226.

701 NARDIN, T.R., HEIN, F., GORSLINE, D.S., AND EDWARDS, B., 1979, A review of mass movement processes
702 sediment and acoustic characteristics, and contrasts in slope and base-of-slope systems versus
703 canyon-fan-basin floor systems.

704 NEMEC, W., 1991, Aspects of sediment movement on steep delta slopes, *in* Colella, A., and Prior, D.B.,
705 eds., *Coarsed-Grained Deltas*, International Association of Sedimentologists, p. 29-73.

706 NUGRAHA, H.D., JACKSON, C.A.-L., JOHNSON, H.D., HODGSON, D.M., AND CLARE, M., 2019a, How erosive are
707 submarine landslides?, *EarthArXiv*.

708 NUGRAHA, H.D., JACKSON, C.A., JOHNSON, H.D., HODGSON, D.M., AND REEVE, M.T., 2019b, Tectonic and
709 oceanographic process interactions archived in Late Cretaceous to Present deep-marine
710 stratigraphy on the Exmouth Plateau, offshore NW Australia: *Basin Research*, v. 31, p. 405-
711 430.

712 OGATA, K., MOUNTJOY, J., PINI, G.A., FESTA, A., AND TINTERRI, R., 2014, Shear zone liquefaction in mass
713 transport deposit emplacement: a multi-scale integration of seismic reflection and outcrop
714 data: *Marine Geology*, v. 356, p. 50-64.

715 OMERU, T., AND CARTWRIGHT, J.A., 2019, The efficacy of kinematic indicators in a complexly deformed
716 Mass Transport Deposit: Insights from the deepwater Taranaki Basin, New Zealand: *Marine
717 and Petroleum Geology*, v. 106, p. 74-87.

718 ORTIZ-KARPF, A., HODGSON, D.M., JACKSON, C.A.-L., AND MCCAFFREY, W.D., 2017, Influence of Seabed
719 Morphology and Substrate Composition On Mass-Transport Flow Processes and Pathways:
720 Insights From the Magdalena Fan, Offshore Colombia: *Journal of Sedimentary Research*, v. 87,
721 p. 189-209.

722 PARKER, E.J., TRAVERSO, C.M., MOORE, R., EVANS, T., AND USHER, N., 2008, Evaluation of landslide impact
723 on deepwater submarine pipelines: Offshore technology conference.

724 PARTYKA, G., GRIDLEY, J., AND LOPEZ, J., 1999, Interpretational applications of spectral decomposition in
725 reservoir characterization: *The Leading Edge*, v. 18, p. 353-360.

726 POSAMENTIER, H.W., AND KOLLA, V., 2003, Seismic geomorphology and stratigraphy of depositional
727 elements in deep-water settings: *Journal of Sedimentary Research*, v. 73, p. 367-388.

728 POSAMENTIER, H.W., AND MARTINSEN, O.J., 2011, The character and genesis of submarine mass-transport
729 deposits: insights from outcrop and 3D seismic data: *Mass-transport deposits in deepwater*
730 *settings: Society for Sedimentary Geology (SEPM) Special Publication 96*, p. 7-38.

731 PRÉLAT, A., PANKHANIA, S.S., JACKSON, C.A.-L., AND HODGSON, D.M., 2015, Slope gradient and lithology as
732 controls on the initiation of submarine slope gullies; Insights from the North Carnarvon Basin,
733 Offshore NW Australia: *Sedimentary Geology*, v. 329, p. 12-17.

734 PRIOR, D.B., BORNHOLD, B., AND JOHNS, M., 1984, Depositional characteristics of a submarine debris flow:
735 *The Journal of Geology*, v. 92, p. 707-727.

736 RANDOLPH, M.F., AND WHITE, D.J., 2012, Interaction forces between pipelines and submarine slides—A
737 geotechnical viewpoint: *Ocean Engineering*, v. 48, p. 32-37.

738 SAWYER, D.E., FLEMINGS, P.B., DUGAN, B., AND GERMAINE, J.T., 2009, Retrogressive failures recorded in
739 mass transport deposits in the Ursa Basin, Northern Gulf of Mexico: *Journal of Geophysical*
740 *Research: Solid Earth*, v. 114.

741 SCARSELLI, N., MCCLAY, K., AND ELDERS, C., 2013, Submarine slide and slump complexes, Exmouth Plateau,
742 NW Shelf of Australia: *The Sedimentary Basins of Western Australia IV: Proceedings of the*
743 *Petroleum Exploration Society of Australia Symposium*.

744 SOBIESIAK, M.S., KNELLER, B., ALSOP, G.I., AND MILANA, J.P., 2018, Styles of basal interaction beneath mass
745 transport deposits: *Marine and Petroleum Geology*, v. 98, p. 629-639.

746 SOBIESIAK, M.S., BUSO, V.V., KNELLER, B., ALSOP, G.I., AND MILANA, J.P., 2019, Block Generation,
747 Deformation, and Interaction of Mass-Transport Deposits With the Seafloor: An Outcrop-

748 Based Study of the Carboniferous Paganzo Basin (Cerro Bola, NW Argentina): Submarine
749 Landslides: Subaqueous Mass Transport Deposits from Outcrops to Seismic Profiles, p. 91-104.

750 STEVENTON, M.J., JACKSON, C.A., HODGSON, D.M., AND JOHNSON, H.D., 2019, Strain analysis of a seismically
751 imaged mass-transport complex, offshore Uruguay: *Basin Research*, v. 31, p. 600-620.

752 SUN, Q., ALVES, T., LU, X., CHEN, C., AND XIE, X., 2018, True volumes of slope failure estimated from a
753 Quaternary mass-transport deposit in the northern South China Sea: *Geophysical Research*
754 *Letters*.

755 TALLING, P.J., MASSON, D.G., SUMNER, E.J., AND MALGESINI, G., 2012, Subaqueous sediment density flows:
756 Depositional processes and deposit types: *Sedimentology*, v. 59, p. 1937-2003.

757 TINDALE, K., NEWELL, N., KEALL, J., AND SMITH, N., 1998, Structural evolution and charge history of the
758 Exmouth Sub-basin, Northern Carnarvon Basin, Western Australia, *in* Purcell, P.G., and Purcell,
759 R.R., eds., *The Sedimentary Basins of Western Australia 2: Proceedings of the Petroleum*
760 *Exploration Society of Australia: Perth*, p. 473-490.

761 TRINCARDI, F., AND ARGNANI, A., 1990, Gela submarine slide: a major basin-wide event in the Plio-
762 Quaternary foredeep of Sicily: *Geo-Marine Letters*, v. 10, p. 13.

763 VANNESTE, M., FORSBERG, C.F., GLIMSDAL, S., HARBITZ, C.B., ISSLER, D., KVALSTAD, T.J., LØVHOLT, F., AND NADIM,
764 F., 2013, *Submarine landslides and their consequences: what do we know, what can we do?*,
765 *Landslide science and practice*, Springer, p. 5-17.

766 VARNES, D.J., 1978, *Slope movement types and processes: Special report*, v. 176, p. 11-33.

767 WEIMER, P., AND SHIPP, C., 2004, *Mass Transport Complexes: Musing on past uses and suggestions for*
768 *future directions*, Offshore Technology Conference: Houston.

769 ZAKERI, A., 2009, Review of state-of-the-art: Drag forces on submarine pipelines and piles caused by
770 landslide or debris flow impact: *Journal of offshore mechanics and Arctic engineering*, v. 131.

771 ZENG, H., HENRY, S.C., AND RIOLA, J.P., 1998, Stratal slicing, Part II: Real 3-D seismic data: *Geophysics*, v.
772 63, p. 514-522.

773 **FIGURE CAPTIONS**

774 **Fig. 1.**--- **A)** Location of the study area. Regional seismic line (orange) across several wells (see Fig. 2).
775 **B)** Seabed map of the Gorgon Slide, and industry well data (red dots) available for this study. The
776 Gorgon Slide is expressed as rugose relief on the seabed. Both evacuation and most of deposition
777 zones are imaged within the 3D seismic reflection data. **C)** Outline of the deposits of the Gorgon Slide
778 (dark grey), where a minor area (c. 7%) of the total slide area in the NW (dashed line) is not imaged
779 within the 3D seismic reflection data. This minor part is delineated using 2D seismic lines (green). Five
780 3D seismic reflection datasets (Gorgon, Acme, Draeck, Duyfken, and Io-Jansz) were used in this study.
781 Bathymetry and topography data are from Geoscience Australia.

782 **Fig. 2.**--- A regional seismic section across the Exmouth Plateau (see Fig. 1 for location). **A)**
783 Uninterpreted. **B)** Interpreted. The Gorgon Slide is bound by a basal-shear surface (yellow) at the base
784 and seabed (blue) at the top. Modified from Nugraha et al. (2019b).

785 **Fig. 3.**--- **A)** Thickness map of the Gorgon Slide showing lateral boundaries of the slide (i.e. NE lateral
786 margin and pinch-out in the SW), with thickness concentration adjacent to the NE lateral margin. We
787 divide rugged geometry of the frontal margin into eastern and western frontal margins. **B)** Seabed dip
788 map showing two distinct sub-bodies (namely Area A and B) within the slide. The two areas are
789 separated by a zone of longitudinal shear. The depositional zone of the slide comprises upper (UTD)
790 and lower (LTD) translation and toe domains. **C)** A 3D perspective of seabed structure map in the LTD
791 showing the geometry of the longitudinal shear zone.

792 **Fig. 4.**--- **A)** Dip-oriented seismic section across the Gorgon Slide showing the headwall scarp,
793 evacuation and deposition zones. **B)** Strike-oriented seismic section showing the asymmetric
794 geometry of the slide, with erosional lateral margin in the NE and pinch-out in the SW.

795 **Fig. 5.**--- Seismic facies classification used in this study. **A)** Seismic facies description and
796 interpretation. **B)** Variance attributes extraction between the basal-shear surface and an iso-

797 proportional surface (50% between the basal-shear surface and the seabed). **C-E**) Seismic sections
798 showing seismic facies within the translation domain. **F**) A time-slice of variance attribute extraction
799 (see G for position) showing seismic facies in the toe domain. **G**) A seismic section showing seismic
800 facies in the toe domain.

801 **Fig. 6.**--- Seabed map showing kinematic indicators in the headwall domain, which include the main
802 headwall of the Gorgon Slide, grooves, crown cracks, pockmarks, and a small scarp. **A**) Uninterpreted.
803 **B**) Interpreted.

804 **Fig. 7.**--- Upper translation domain of the Gorgon Slide. **A**) Basal-shear surface variance map (top) and
805 its interpretation (bottom). **B**) Internal body RMS amplitude map (extracted 50 ms above and below
806 isoproportional horizon, orange) (top) and its interpretation (bottom). **C**) Top surface dip map (top)
807 and its interpretation (bottom). **D**) Seismic sections, uninterpreted (above) and interpreted (bottom),
808 showing seismic facies across the upper translation domain. See text for discussions.

809 **Fig. 8.**--- **A**) Spectral decomposition map within the slide (50% between basal-shear and top surfaces)
810 showing features within upper translation domain in detail. **B**) Uninterpreted, and **C**) interpreted,
811 seismic section along megaclasts (SF-4) across Area A and B. See text for discussion.

812 **Fig. 9.**--- Lower translation domain of the Gorgon Slide. **A**) Basal-shear surface variance map (top) and
813 its interpretation (bottom). **B**) Internal body RMS amplitude map (extracted 50 ms above and below
814 isoproportional horizon, orange) (top) and its interpretation (bottom). **C**) Top surface dip map (top)
815 and its interpretation (bottom). **D**) Seismic sections, uninterpreted (above) and interpreted (bottom),
816 showing seismic facies across the upper translation domain. See text for discussion.

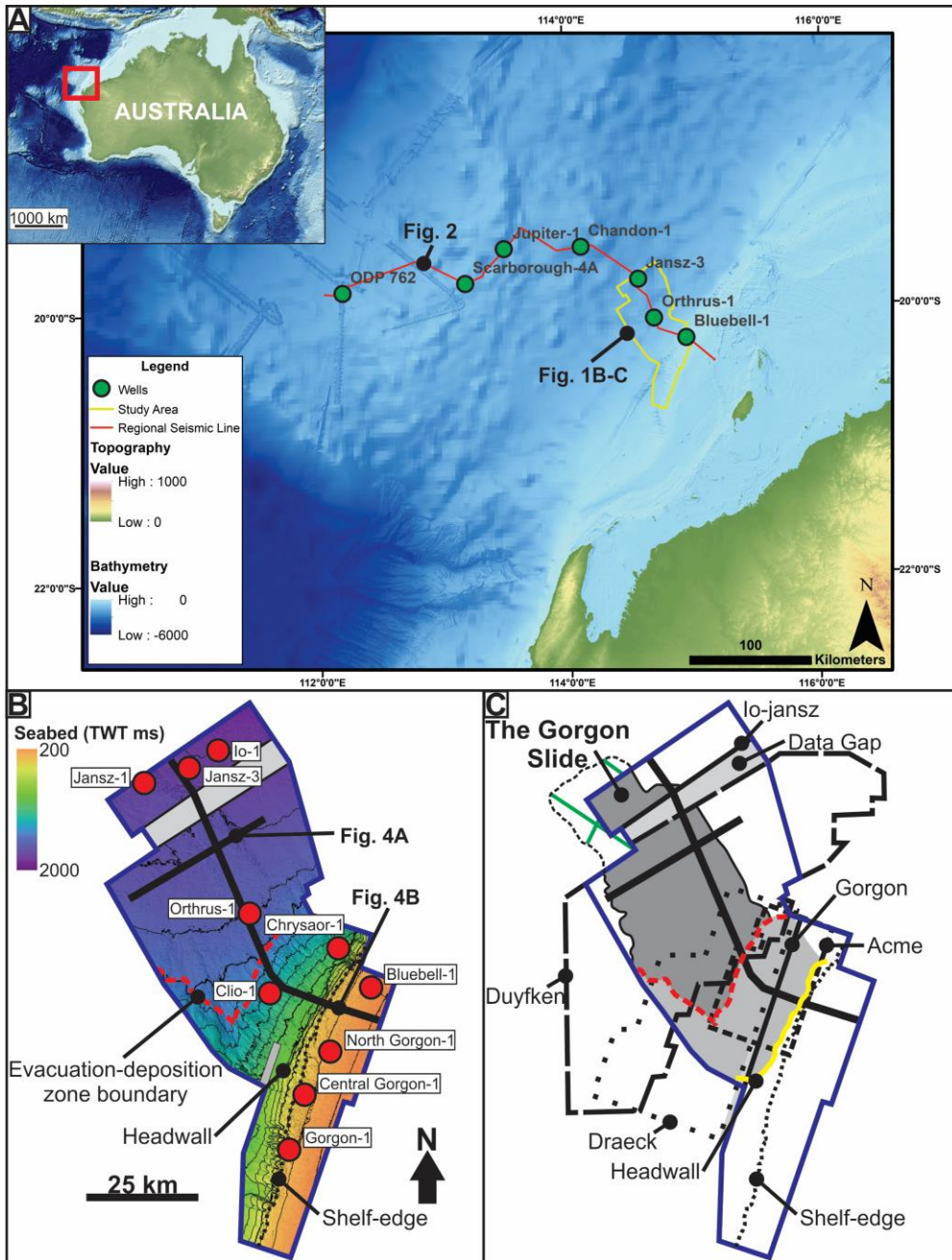
817 **Fig. 10.**--- Dimensions and orientation of the megaclasts in the upper and translation domains. **A**)
818 Megaclasts in the upper translation domain are generally thinner with longer long-axes, as compared
819 to the ones in the lower translation domain that are thicker with shorter long-axes. **B**) Megaclasts in

820 the upper translation domain are generally oriented perpendicular, and the ones in the lower
821 translation domain are oblique to sub-parallel, to the transport direction.

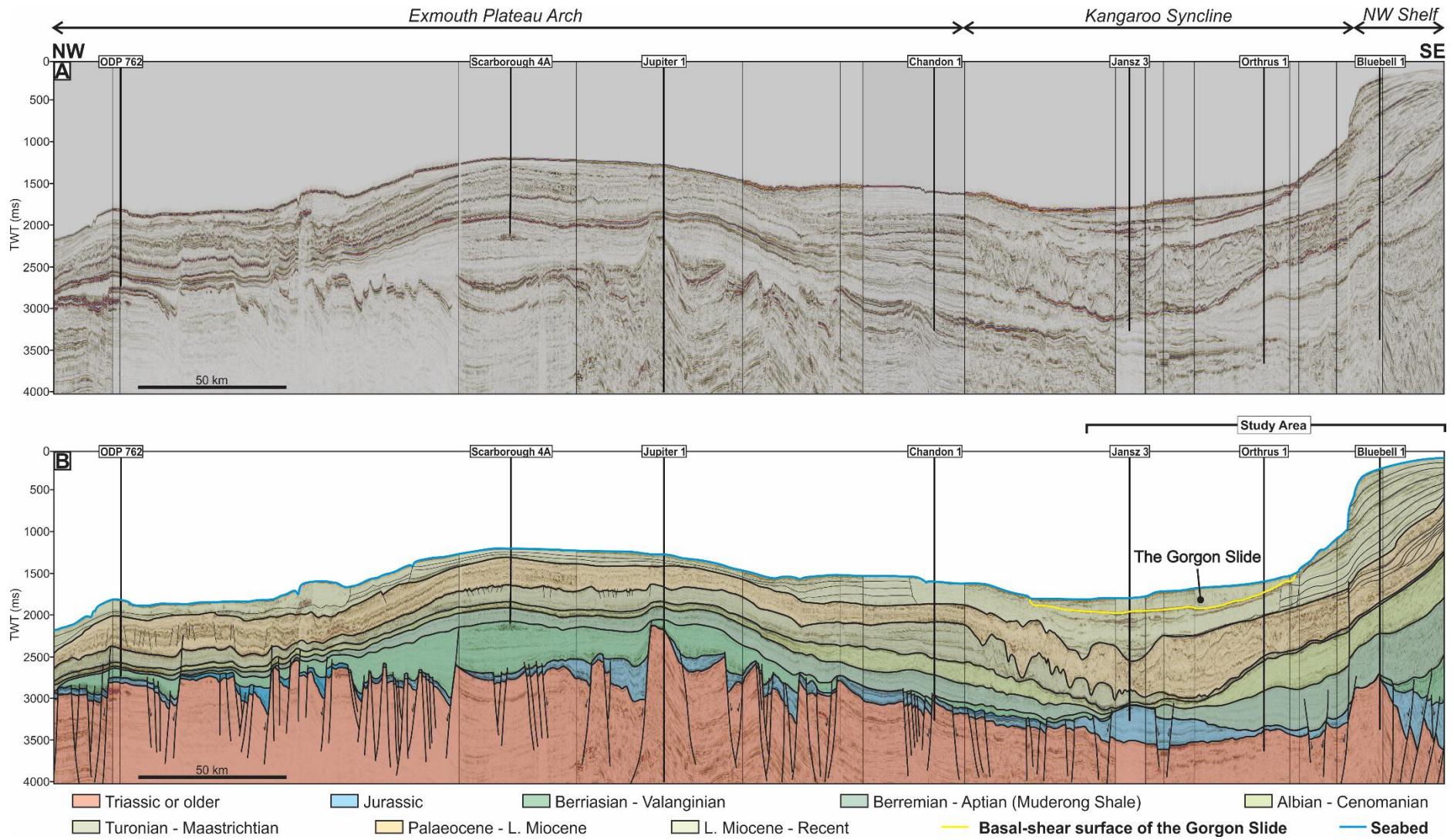
822 **Fig. 11.**--- Toe domain of the Gorgon Slide. **A)** Basal-shear surface variance map (top) and its
823 interpretation (bottom). **B)** Internal body RMS amplitude map (time-slice at the orange horizon in D)
824 (top) and its interpretation (bottom). **C)** Top surface dip map (top) and its interpretation (bottom). **D)**
825 Seismic sections, uninterpreted (above) and interpreted (bottom), showing seismic facies across the
826 toe domain. See text for discussion.

827 **Fig. 12.**--- Schematic diagram of Gorgon Slide depicting three stages of emplacement processes. **A)** A
828 failure event occurred. **B)** The slide split into two flow cells, Cell A and B, due to a cluster of megaclasts
829 derived from the headwall and/or slope strata that acted as a flow obstacle. **C)** Cell A ceased, and its
830 frontal margin is expressed on the seabed, while Cell B flowed beyond the limit of the dataset. See
831 text for discussion.

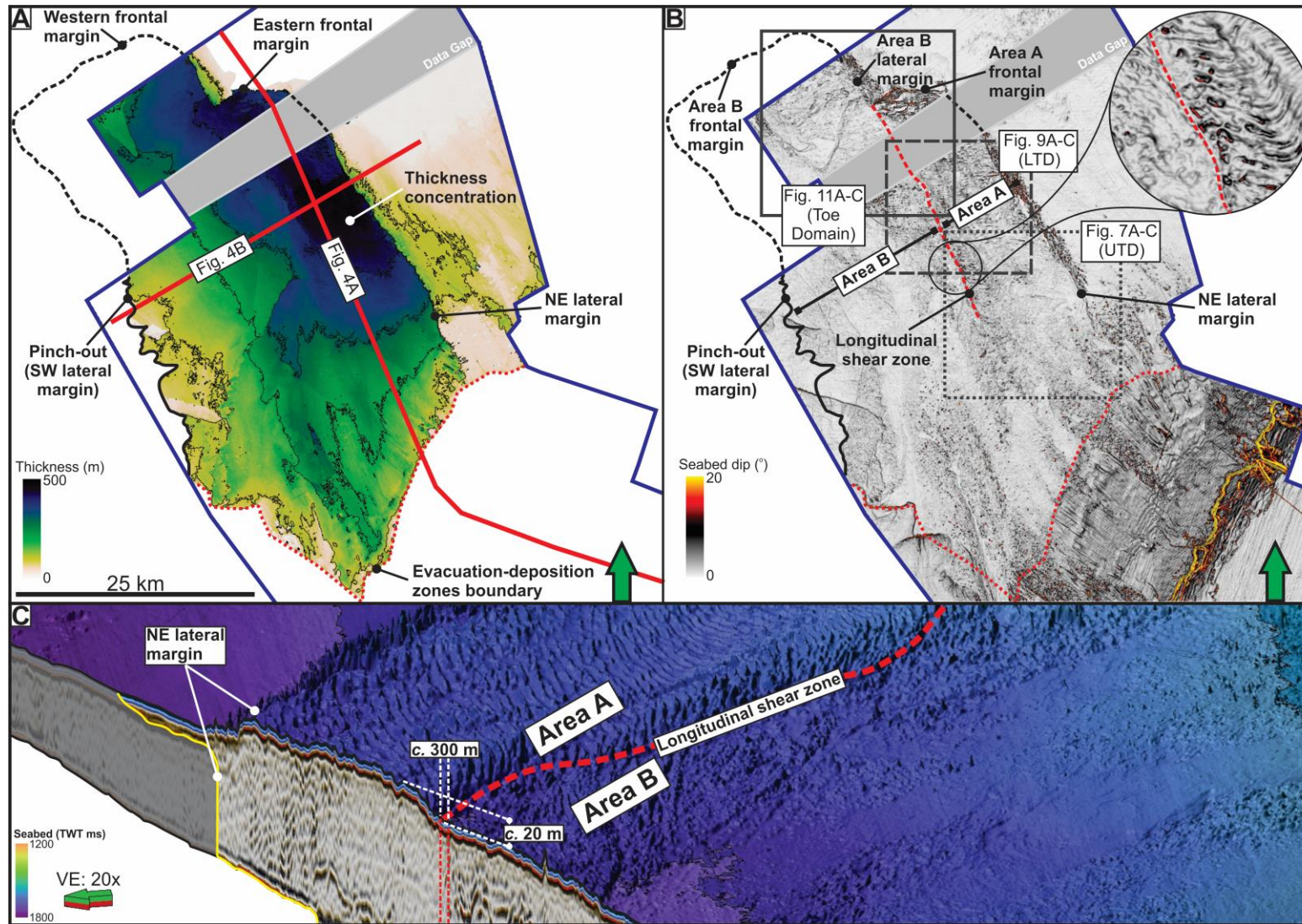
832 **Fig. 13.**--- **A)** Variance map extracted along the orange horizon in C-D, overlaid by time structure map
833 of thrusting megaclasts (see the red horizon in D, left). The thrusting megaclasts define the frontal
834 margin of Cell A and lateral margin of Cell B. **B)** Variance map extracted along the basal-shear surface
835 (yellow) in C-D, overlaid by time structure map of the basal megaclasts (see the red horizon in D, right).
836 **C)** Uninterpreted, and **D)** interpreted seismic section across the thrusting and basal megaclasts. These
837 megaclasts have similar dimension and seismic facies, thus, likely to have a similar origin.



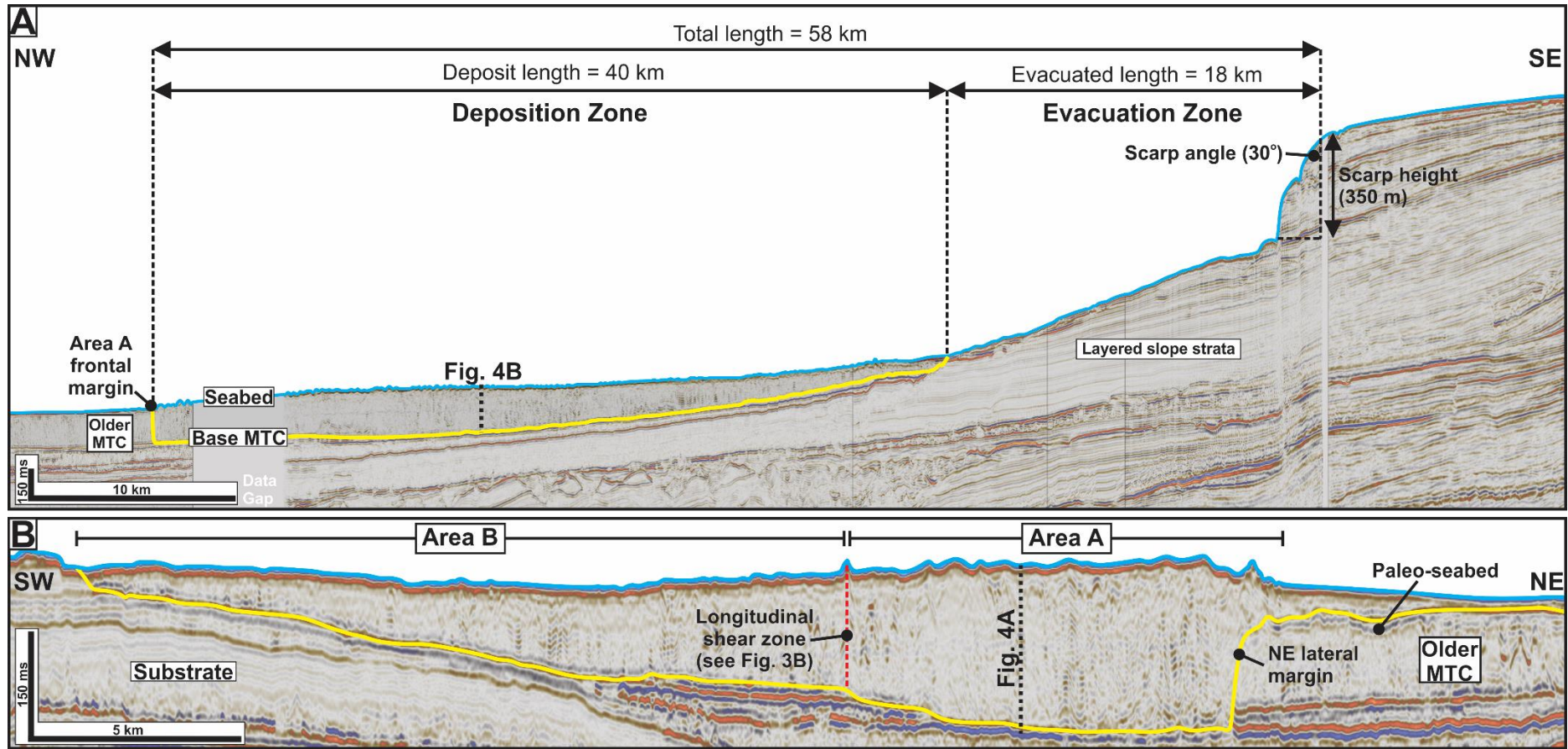
840 Figure 2



841



844 Figure 4



845

A Facies	Description	Interpretation
SF-1	Chaotic and transparent both in cross-section and map-view (Fig. 5B-C).	Debrites containing disaggregated materials (cf. Posamentier and Kolla 2003).
SF-2	Low-to-medium amplitude, discontinuous folded reflections that occasionally form sinuous lineations in map-view (Fig. 5B, D-E).	Debrites containing partially disaggregated materials (cf. Ortiz-Karpf et al. 2017).
SF-3	High amplitude, discontinuous folded reflections that are separated by imbricate thrusts (Fig. 5F-G).	Fold and thrust systems formed by compressional deformation within MTCs, flow direction is generally perpendicular to the strike of the thrusts (Bull et al. 2009).
SF-4	Isolated packages of coherent, sub-parallel reflections within a matrix composed of SF-1 or 2 (Fig. 5B-C, E). In most cases, the reflections are disrupted, e.g. faulted and folded.	Megaclasts transported within debritic matrix (cf. Bull et al. 2009; Jackson 2011; Hodgson et al. 2018).
SF-5	Medium-to-high amplitude, continuous, sub-parallel, downslope-dipping reflections beneath the shelf and within the evacuation zone (Fig. 4A).	Non-MTC deposits, i.e. carbonate progradation and layered slope deposits that were the source of, and eroded by, the Gorgon Slide (Hengesh et al. 2013; Nugraha et al. 2019b).

