SEISMIC EXPRESSION, STRUCTURE, AND EVOLUTION OF FLOW CELLS WITHIN A SUBMARINE LANDSLIDE

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ABSTRACT

Submarine landslides are not a singular mass of sediment, but may instead comprise shear zone-bound, intra-flow cells of different sizes, rheological states, and strains. Constraining their evolution is important for assessing risks to subaqueous infrastructure and coastal communities, yet controls on their initiation, translation, and cessation remain unclear. We here use five, high-quality post-stack time-migrated (PSTM) 3D seismic reflection datasets to investigate the structure and evolution of flow cells in the Gorgon Slide, a near-seabed submarine landslide on the Exmouth Plateau, offshore NW Australia. Our data suggest that the slide originated from a c.18 km-long evacuation zone bounded updip by a NE-trending headwall scarp that dips steeply (c.30^{o}) seaward, and travelled northwestward over a strongly erosional basal-shear surface that deepens downdip. The slide is dominated by chaotic seismic reflections, interpreted as a debrite, containing large clasts (up to 1 km-long) derived from the headwall and/or entrained from the layered slope substrate. The morphology of the basal-shear surface focused slide transport, resulting in clustering of megaclasts in the medial part of the still-translating flow. This cluster became an obstacle to subsequent flows, resulting in two flow cells (Cells A and B), separated by a longitudinal shear zone. Interaction between the flow cells is recorded by sinuous flow fabrics within, and pressure ridges on the top surface of, the slide. The intra-flow fabrics and top-surface ridges in Cell A were dragged downdip by the relatively more mobile Cell B, which continued translating downdip without intra-flow obstacles. The transport and emplacement processes inferred for the Gorgon Slide suggest that entrainment and abrasion of megaclasts can induce velocity perturbations during slide emplacement, causing changes in flow rheology, and the initiation and cessation of flow cells. This reconstruction of flow-cell evolution within a submarine landslide may provide geological constraints for future geohazard risk assessments.

INTRODUCTION

Subaqueous landsliding can trigger the emplacement of sediment gravity flows that commonly emplace clast-bearing debrites (e.g. Dott, R., 1963; Nardin, T. R. *et al.*, 1979; Nemec, W., 1991; Moscardelli, L. & Wood, L., 2008; Posamentier, H. W. & Martinsen, O. J., 2011). Subaqueous landslides can evolve temporally and spatially. For example, a relatively large slide (i.e., first-order flow cell; *sensu* Alsop and Marco, 2014) may entrain ambient water, causing it to become more dilute and increasing intra-flow turbulence (Fisher, R. V., 1983; Talling, P. J. *et al.*, 2012; Sun, Q. *et al.*, 2018a). Field studies of metre-scale slides (referred to as 'slumps' by Alsop, G. I. & Marco, S., 2014) show that these first-order cells can also break down into second-order (i.e., intra-flow) cells, between which strain styles, patterns, and magnitudes can abruptly vary (Farrell, S. G., 1984; Alsop, G. I. & Marco, S., 2014). Alsop, G. I. and Marco, S. (2014) propose that these complexities arise due to local (i.e., intra-flow) variations in: (i) the lithology and stratigraphic architecture of the failed and translating sediment mass; (ii) fluid pressure within and below the translating sediment mass; and (iii) the attitude of the basal-shear surface, or *décollement*, underlying the translating sediment mass. Although these studies are undoubtedly informative, it is unclear if the structures observed within, and models developed from, these relatively small-scale two-dimensional exposures are applicable at much larger scales.

Several 3D seismic reflection data-based studies have documented flow cells that are significantly larger (i.e., kilometre-scale) than those identified in the field. These cells occur within relatively long run-out deposits emplaced by one or more flows (Gee, M. et al., 2005; Steventon, M. J. et al., 2019; Abu, C. et al., 2022; Scarselli, N. et al., 2016). For example, Steventon, M. J. *et al.* (2019) show that up to 18 km-wide cells developed in the contractional domain of a large, megaclast-bearing slide from offshore Uruguay. These flow cells are defined by differing magnitudes of contractional strain (i.e., thrusts and related folds), and are bound by up to 35 km-long shear zones. The mechanism for flow cell development appears to record some combination of an abrupt or progressive lateral variation in flow velocity, and/or the total amount of emplacement-related strain (see also Bull, S. *et al.*, 2009). In this study, we use the term '*flow cell*' to describe a kinematically distinct compartment within a larger translating slide mass that records a systematic difference in translation and/or strain relative to adjacent compartments (e.g., Farrell, S. G., 1984; Alsop, G. I. & Marco, S., 2014; Scarselli, N. et al., 2016; Steventon, M. J. et al., 2019; Nugraha, H. D. et al., 2020; Abu, C. et al., 2022), where: (i) adjacent compartments occur within the same mapped slide body and are not separated by a resolvable stratigraphic boundary; (ii) they are separated by a persistent intra-slide kinematic boundary; (iii) structures and flow fabrics across that boundary record systematic differences in strain and relative motion; and (iv) the compartments record distinct translation and/or arrest histories.

Thus, despite a growing body of work on the structure and kinematics of intra-flow cells, the mechanisms responsible for their initiation, translation, and cessation, across a range of scales, remain poorly understood. Progress towards mitigating the geohazards associated with subaqueous landsliding at least partly depends on a thorough understanding of their emplacement processes, including cell formation (Masson, D. *et al.*, 2006). For example, the rheology and emplacement direction of a slide can change during transport (Iverson, R. M., 1997; Dykstra, M. *et al.*, 2011; Joanne, C. *et al.*, 2013; Ortiz-Karpf, A. *et al.*, 2017; Hodgson, D. *et al.*, 2018), which may influence the amount and direction of drag forces exerted on subaqueous pipelines (e.g., Zakeri, A., 2009).

Here, we use five high-quality 3D seismic reflection datasets from the Exmouth Plateau (offshore NW Australia) to study the Gorgon Slide (Fig. 1). The slide is at or just below the seabed and is accordingly well-imaged, enabling: (i) use a range of seismic attributes to map kinematic indicators below, within, and above the slide; (ii) identify and map kinematically distinct intra-slide flow cells; (iii) reconstruct slide emplacement, inferring the impact cell formation and evolution had on overall flow behaviour; and (iv) consider the potential factors controlling cell formation. We therefore test existing models for flow cell formation within sediment gravity-flows and assess how high-quality 3D seismic reflection data allows reconstruction of submarine landslide emplacement processes, and thus, better understand their related geohazard.

DATA AND METHODOLOGY

The five, high-quality post-stack time-migrated (PSTM) 3D seismic reflection datasets image *c.*93% (i.e. 1594 km^2 of 1760 km^2 total area) of the Gorgon Slide, covering all of the evacuation zone and most of the deposition zone (Figs. 1B-C). We used three 2D seismic reflection lines to constrain the downdip limit of the deposition zone (green lines in Fig. 1C). The vertical resolution of the 3D seismic reflection data at the base of the slide (*c.*500 m below sea floor) ranges from 8-11 m, based on near seabed sediment velocity and dominant seismic frequency of 1824 m/s and 40-60 Hz, respectively. Bin spacing of the 3D seismic volumes ranges from 12.5 x 18.75 m to 20 x 25 m (see Appendix 1 for details). Depth conversion of seabed and basal-shear surface time-structure maps was conducted by using average seismic velocities for water (1519 m/s) and weakly compacted, near-seabed sediment (1824 m/s), respectively (Appendix 1). The average water velocity is constrained by ten industry wells (Fig. 1B), and the near-seabed sediment velocity data is available from well ODP 762 (see Figs. 1A and 2). As the depth conversions uses average rather than spatially varying interval velocities, the resulting depth-derived dimensions should be regarded as approximate. This uncertainty affects absolute estimates of slide thickness, basal-shear-surface relief, ramp and megaclast heights, but does not impact the mapped plan-view relationships that underpin the kinematic interpretation. The available velocity data do not permit a more robust numerical estimate of this uncertainty.

We mapped the seabed and basal-shear surface of the Gorgon Slide to define its kinematics as it initiated, translated, and arrested. We also employed an iso-proportional slicing method (Zeng, H. *et al.*, 1998), midway between the seabed and basal-shear surface of the slide, to visualise and map the heterogeneity of its constituent seismic facies. Several seismic attributes were used in this analysis, particularly (see also Appendix 2 for details): (i) *variance*, to better image discontinuities (Chopra, S. & Marfurt, K. J., 2007), including grooves on the basal-shear surface of a slide (e.g., Bull, S. *et al.*, 2009); (ii) *Root Mean Squared (RMS) amplitude*, to better delineate features that have distinct positive or negative amplitudes resulting from an acoustic (velocity and/or density) contrast (Brown, A. R., 2011), such as megaclasts encased within a relatively transparent debritic matrix (e.g., Ortiz-Karpf, A. *et al.*, 2017); (iii) *dip*, to better image rugosity of a surface (Brown, A. R., 2011), including seabed relief (e.g., Bull, S. *et al.*, 2009); and (iv) *spectral decomposition*, to highlight internal stratigraphic (seismic facies) heterogeneities within a geological body, such as a slide (Partyka, G. *et al.*, 1999; Eckersley, A. J. *et al.*, 2018).

GEOLOGICAL SETTING

The Exmouth Plateau is a part of the North Carnarvon Basin, which has experienced multiple rifting events from the Late Jurassic to Early Cretaceous (Fig. 2) (Tindale, K. *et al.*, 1998; Longley, I. M. *et al.*, 2002). Post-rift deposition was initially dominated by fine-grained siliciclastic sediments, becoming increasingly carbonate-rich as the Australian plate drifted northward into more equatorial conditions (e.g., Apthorpe, M., 1988; Hull, J. N. F. & Griffiths, C. M., 2002). A prograding carbonate-dominated margin characterises the Oligocene to the present-day (Fig. 2B) (Cathro, D. L. *et al.*, 2003; Moss, G. D. *et al.*, 2004). Collision between the Australian and Eurasian plates (Miocene to present-day) reactivated some of the rift-related faults, forming inversion structures such as the large NE-trending Exmouth Plateau Arch (Figs. 1 and 2; Keep, M. *et al.*, 1998). It is also likely that inversion-related deformation of the seabed triggered the widespread emplacement of the numerous carbonate-dominated, sediment gravity-driven deposits across the plateau, including the focus of this study, the Gorgon Slide, which extends from the seabed (blue) down to its basal-shear surface (yellow, see Fig. 2B) (Boyd, R. *et al.*, 1993; Hengesh, J. V. *et al.*, 2013; Scarselli, N. *et al.*, 2013; Wu, N. *et al.*, 2026; Scarselli, N. *et al.*, 2026). The headwall of the Gorgon Slide is underlain by a rift-related horst block, which was drilled by Bluebell-1 (Fig. 2B) (McCormack, K. & McClay, K., 2013).

THE GORGON SLIDE

General Characteristics

External geometry and morphological domains.– The Gorgon Slide was sourced from the slope defining the present-day shelf-edge of NW Australia (Fig. 1). It was deposited in the adjacent Kangaroo Syncline, forming a lenticular, NW-trending body that wedges-out (i) to the SE against the continental slope, and (ii) to the NW against the eastern margin of the Exmouth Plateau Arch (see Figs. 1B-C and 2). The slide has a maximum runout distance of *c.*70 km, is up to 500 m thick, and has a total volume of *c.*500 km^3 (Fig. 3A) (Nugraha, H. D. *et al.*, 2022). The slide terminates against the SW and NE lateral margins (Fig. 3A), and is *c.*30 km-wide in the central part and abruptly narrows to *c.*18 km at its frontal end (Fig. 3A). Its downdip termination is defined by the two relatively steeply dipping (18°) eastern and western frontal margins (Fig. 3A).

The central and frontal parts of the slide display changes in seabed rugosity, which defines two distinctive regions: Areas A and B (Fig. 3B). Area A is bound by the NE lateral and eastern frontal margins and is characterized by a highly rugose seabed (Fig. 3B). Area B is bound by the SW lateral and western frontal margins, and the seabed is relatively smoother than that of Area A (Fig. 3B). Areas A and B are separated by a linear, NW-trending feature (see inset image in Fig. 3B) that is subparallel to both lateral margins of the slide and has a seabed relief of *c.*10-20 m (Fig. 3C). This linear feature is narrow (*c.*170-300 m-wide) and extends for *c.*26 km updip, before terminating where Areas A and B merge and become indistinguishable, while downdip it merges with the NE lateral margin of Area B near the slide terminus (see Figs. 3B-C). We interpret this linear seabed feature as a longitudinal shear zone (sensu Bull, S. *et al.*, 2009), which records internal variations of transport velocity within the slide (see also Steventon, M. J. *et al.*, 2019).

The Gorgon Slide originated from a *c.*18 km-long evacuation zone, bound on its updip margin by a relatively steeply-dipping (*c.*30°), *c.*350 m-high headwall scarp (Fig. 4A). The scarp has a relatively simple geometry, defined by a single, smooth, seaward-facing plane. The frontal margin of Area A is marked by positive seabed relief (*c.*30 m) relative to the smooth seabed bounding pre-slide slope strata immediately downdip. Pre-slide strata are represented by older slides (defined by chaotic, weakly reflective seismic reflections), and hemipelagic and pelagic slope strata (defined by continuous, reflective, sub-parallel seismic reflections) (Fig. 4A). The juxtaposition of seismically chaotic, interpreted debrite-rich, strata can make establishing stratigraphic relationships challenging, not least because the kinematic development of a slide basal-shear surface can also involve remobilization and deformation of older strata (e.g., Butler and McCaffrey, 2010; Hodgson et al., 2018). We differentiate between the debrite seismic facies of the Gorgon Slide and pre-slide hemipelagic/pelagic seismic facies based on the following: (i) hemipelagic and pelagic slope strata being sharply truncated along the basal-shear surface of the Gorgon Slide (Fig. 4); (ii) older slides capped by hemipelagic and pelagic slope strata (paleo-seabed), imply periods of suspension-dominated deposition separating gravity-driven debrite emplacement events (Fig. 4B); and (iii) a relatively thin interval of the Gorgon Slide material overlies hemipelagic and pelagic slope strata capping older slides (Fig. 4B).

Internal seismic facies.– The Gorgon Slide contains a range of seismic facies that we infer record deposition by a range of sediment transport processes and variable strain styles (Fig. 5) (e.g., Alves, T. M. *et al.*, 2014). Our seismic facies classification is supported by calibrated seismic reflection data with lithology from well data (e.g., Sawyer, D. E. *et al.*, 2009), including field data used to construct seismic reflection forward models (Dykstra, M. *et al.*, 2011). We define five seismic facies (SF) (Fig. 5A) based on variations in reflection strength and internal deformation (see Figs. 5B-G): (i) SF-1 – low-to-variable amplitude, chaotic reflections - these are interpreted as debrites (cf. Posamentier, H. W. & Kolla, V., 2003; Posamentier, H. W. & Martinsen, O. J., 2011; Ortiz-Karpf, A. *et al.*, 2017); (ii) SF-2 – low-to-medium amplitude, discontinuous reflections that are folded in cross-section, and that are associated with very well-developed sinuous lineations within and at the top of the slide in plan-view – these are interpreted as slumps to debrites containing seismic-scale clasts and pressure ridges (Bull, S. *et al.*, 2009); (iii) SF-3 – high-amplitude, relatively continuous, folded reflections that are offset by thrusts – these are interpreted as fold and thrust systems developed within relatively cohesive portions of the slide; (iv) SF-4 – packages of relatively continuous, sub-parallel, variably deformed reflections that are surrounded by SF-1 and -2 and that extend vertical through the entire thickness of the deposit – these are interpreted as megaclasts embedded within the debritic matrix (cf. McGilvery, T. A. H., Geoffrey, 2004; Bull, S. *et al.*, 2009; Jackson, C. A., 2011; Ortiz-Karpf, A. *et al.*, 2017; Hodgson, D. *et al.*, 2018); and (v) SF-5 – variable amplitude, continuous, sub-parallel reflections – these are interpreted as pelagic and hemipelagic slope strata (see Fig. 4A) (e.g., Prélat, A. *et al.*, 2015). The *c.*26 km-long, NW-trending longitudinal shear zone described above is generally characterised by SF-1 (Fig. 3C) (i.e. debrite) (cf. Ogata, K. *et al.*, 2014; Bull, S. & Cartwright, J. A., 2019; Omeru, T. & Cartwright, J. A., 2019), although its relatively narrow width means it can be difficult to differentiate it from adjacent deposits within Area A and B where more continuous reflections (i.e. megaclasts; SF-4) are not juxtaposed across it (Figs. 4B, and 5B-E).

The Gorgon Slide's four principal domains are discussed in relation to their primary kinematic indicators and related emplacement processes, as follows : (i) the headwall domain; (ii) the upper translational domain (UTD); (iii) the lower translational domain (LTD); and (iv) the toe domain (Fig. 3B).

Headwall domain

Description.– The largest feature in the headwall domain is a large, NW-dipping scarp (Fig. 6). Immediately updip of this feature are: (i) a small, NNE-trending scarp, next to which are (ii) numerous circular depressions that have diameters of *c.*100-300 m, and (iii) at least four, *c.*3-5 km-long, *c.*15 m-deep, linear depressions (see zoomed-in image in Fig. 6A). Downdip of the headwall scarp, within the source area of the Gorgon Slide (i.e. the region between the headwall scarp and evacuation-deposition zone boundary), there are numerous *c.*5-16 km-long, broadly NW-trending, elongate features (Figs. 6A-B) that have a v-shaped geometry in cross-section, which are (*c.*150-300 m-wide and *c.*10-25 m-deep (see zoomed-in image in Fig. 6B).

Interpretation.– The main headwall scarp cross-cuts a smaller scarp (*c.*10 m-high) (Fig. 6B). The smaller, older scarp may have been associated with earlier slope failure responsible its deposition, and that of other older, thinner slump deposits, which pre-date the Gorgon Slide (Fig. 4). The circular depressions are interpreted as pockmarks, suggesting active vertical fluid expulsion (e.g., Hengesh, J. V. *et al.*, 2013; Scarselli, N. *et al.*, 2013). The linear depressions are interpreted as crown cracks, possibly marking the location of future slope failure events (Varnes, D. J., 1978; Frey-Martinez, J. *et al.*, 2005). We interpret the elongate v-shaped features as grooves (sensu Bull, S. *et al.*, 2009) formed due to tooling by megaclasts into the substrate during slide emplacement (Posamentier, H. W. & Martinsen, O. J., 2011; Ortiz-Karpf, A. *et al.*, 2017; Hodgson, D. *et al.*, 2018; Sobiesiak, M. S. *et al.*, 2018; Scarselli, N. *et al.*, 2026). The orthogonal relationship with the headwall scarp enables the trend of these grooves to define reliable translation pathways of the slide through the evacuation zone.

Upper translational domain

Description.– Grooves in the updip part of the upper translational domain (Fig. 7A) are the downdip continuation of those within the evacuation zone (see Fig. 6), displaying similar dimensions and geometries (see Headwall domain section). However, grooves in this domain converge downdip towards the NE lateral margin (Fig. 7A), which contrasts to the more commonly described downdip-diverging grooves (e.g. Posamentier, H. W. & Kolla, V., 2003; McGilvery, T. A. H., Geoffrey, 2004; Ortiz-Karpf, A. *et al.*, 2017). In the central part of the basal-shear surface is a pair of broadly NW-trending, slightly curved lineations that bound an area slightly elevated (*c.*10 m) compared to its surrounding area, and which mark subtle changes in the depth of the basal-shear surface (medium grey-shaded grey in Fig. 7A). Adjacent to the NE lateral margin is an area of highly discontinuous reflections, as captured on the variance map (Fig. 7A). Within this area lineations display an oblique orientation relative to the NE lateral margin. This area is also characterised by low-to-medium amplitude, discontinuous reflections at, and immediately beneath, the basal-shear surface (Fig. 7D). On top of the older slide, there are several 0.5-1.5 km-long lineations that connect to, and trend at *c.*45 $^{\circ}$ to, the NE lateral margin (Figs. 7A and D).

The proximal part of the upper translational domain is dominated by debrite (SF-1) that surrounds scattered megaclasts (SF-4) (Fig. 7B). These megaclasts have elliptical to rectangular planview geometries, with long-axis lengths ranging from *c.*0.18 to 1 km and thicknesses of *c.*70 to 140 m (Fig. 7B). Seismic sections show that these megaclasts are sometimes internally folded and faulted (Fig. 7D). Megaclasts are concentrated in the central part of this domain, where they form a *c.*15 km-long and 3 km-wide, convex-updip cluster. This cluster is bound by a gradational boundary with SF-1 in the E, and an abrupt boundary in the W with Area B. The latter is defined by the longitudinal shear zone across which they abruptly change their orientation from convex-updip to -downdip (Fig. 7B). Most of the megaclast cluster occurs within Area A, apart from a smaller cluster located *c.*5 km downdip to the N within Area B (Fig. 7B), which are separated by the longitudinal shear zone. Immediately downdip of the Area A megaclast cluster are several convex-updip bands developed within the slides debritic matrix (Fig. 7B). These extend upwards through the slide and are expressed on its top surface, where they form positive relief at the seabed (Fig. 7C). These bands are sub-parallel to the overall, convex-updip geometry of the cluster and to the long-axis of elongated clasts (Figs. 7B and 8A). In contrast, downdip from the eastern margin of the cluster, the bands show a convex-downdip geometry, terminating at the NE lateral margin (Figs. 7B and 8A). A NW-trending, narrow band (*c.*500 m-wide and *c.*10 km-long) of debrite (SF-1), defines the boundary between these two sets of bands (Fig. 7B).

Interpretation.– The converging-downdip geometry of the grooves implies that the pathway of the slide was focused towards the steep, NE lateral margin (Fig. 7A). As a result, the slide is thickest adjacent to this margin (see Fig. 3A). The pathway was likely controlled by the morphology of the basal-shear surface that broadly follows the morphology of the underlying substrate (see Fig. 4B). This supports the observations of Ortiz-Karpf, A. *et al.* (2017), who provide evidence for the impact of seabed morphology on the emplacement of submarine landslides. We interpret the pair of curved lineations as 'ramps' bounding an area called a 'flat' (Fig. 7A) (Trincardi, F. & Argnani, A., 1990; Lucente, C. C. & Pini, G. A., 2003; Frey-Martinez, J. *et al.*, 2005; Bull, S. *et al.*, 2009). The ramps record basal erosion by the overlying slide that are commonly expressed by truncated reflections of underlying substrate by a basal-shear surface (e.g. Bull, S. *et al.*, 2009). However, as the ramps in this domain represent relatively small steps (i.e. 10 m), the basal-shear surface does not truncate more than one reflection. The lineations oriented oblique to the NE lateral margin on, and immediately beneath, the basal-shear surface (Fig. 7A) implies a substrate that has undergone contractional deformation due to stress exerted by the passing slide, forming a 'basal-shear zone' (Butler, R. & McCaffrey, W., 2010; Hodgson, D. *et al.*, 2018; Cardona, S. *et al.*, 2020). Lineations on top of the older slide (Fig. 7A) are interpreted as shear fractures (i.e. Riedel shears) that developed due to dextral strike-slip movement along the NE lateral margin as the Gorgon Slide translated north-westward (e.g. Fleming, R. W. & Johnson, A. M., 1989; Martinsen, O., 1994; Fossen, H., 2016) (Fig. 7A). Fleming, R. W. and Johnson, A. M. (1989) suggest that this type of fracture is developed during an early stage of strike-slip faulting along the lateral margin of slides, prior to the formation of through-going lateral margins. They recorded fractures oriented at 45^{o} clockwise from the trend of a dextral lateral margin, similar to the shear fractures observed here.

The scattered megaclasts in the proximal part of this domain are clustered (see Fig. 7B), possibly due to downdip convergence of material within the Gorgon Slide (see Fig. 7A). We suggest that the clusters of megaclasts in Areas A and B were initially emplaced as a single cluster that was subsequently cross cut by the longitudinal shear zone (Figs. 7B and 8), with clustering of megaclasts possibly inducing intra-slide velocity perturbations and/or variable total strains. These perturbations are evidenced by across-strike variations in the attitude of flow fabrics, which we infer are formed due to intra-flow compression (cf. 'pressure ridges' or 'secondary flow fabrics' of Bull, S. *et al.* (2009)); i.e. convex-downdip flow fabrics in Areas A and B are located further downflow of the convex-updip flow fabrics preserved immediately downdip, and perhaps in the strain shadow zone of the megaclast cluster in Area A (Figs. 7B and 8). Another indicator of internal variations in intra-flow velocity and/or total strain is the narrow area within Area A that separates the convex-downdip and -updip flow fabrics (Fig. 7B). This area is interpreted as an 'internal shear zone' (cf. Ogata, K. *et al.*, 2014; Bull, S. & Cartwright, J. A., 2019; Omeru, T. & Cartwright, J. A., 2019) that contains disaggregated material due to intense shearing. Other studies have also discussed how the entrainment and abrasion of megaclasts during slide emplacement could affect flow rheology (e.g. Joanne, C. *et al.*, 2013; Ortiz-Karpf, A. *et al.*, 2017; Hodgson, D. *et al.*, 2018; Sobiesiak, M. S. *et al.*, 2019), and therefore spatial variations in intra-slide flow velocity.

Lower translational domain

Description.– On the basal-shear surface the majority of kinematic indicators observed in the upper translational domain extend downdip into the lower translational domain (i.e. the basal-shear surface ramp and its associated shear fractures; Fig. 9A). Grooves on the basal-shear surface are, however, are not apparent, plus the ramps are deeper (*c.*20 m-deep, Fig. 9D) than in the upper translational domain, and lineations within the deformed substrate area are more apparent (Fig. 9A). Downdip from the deformed substrate are several SE-facing ramps (i.e. perpendicular to transport direction) that merge updip with the ramp extending downdip from the upper translational domain (Fig. 9A). The area NE of the deformed substrate, beyond the lateral margin and on top of the older slide, displays the downdip continuation of the N-trending shear fractures seen in the upper translational domain. These fractures die-out downdip to the N.

The convex-updip intra-flow fabric extends vertically through the deposit, it is bound between the longitudinal and internal shear zones within Area A, and its top surface dies-out downdip (Fig. 9B). In contrast, the convex-downdip flow fabrics within Area B continue and are more prominent in this lower translational domain (Fig. 9B). Another cluster of megaclasts occurs adjacent to the NE lateral margin (Fig. 9B). In cross-section (Fig. 9D), this cluster contains megaclasts that have similar seismic expression to those in the upper translational domain (Figs. 7D and 8B-C). However, these megaclasts are smaller (length of long axes *c.*0.05 to 0.54 km) and the unit thinner (*c.*73 to 220 m) compared to the upper translational domain (long axes *c.*0.17 to 0.98 km and unit thickness *c.*70 to 137 m) (see Fig. 10A). The long-axes trends of megaclasts also differ: oriented NE-SW in the lower translation domain and NNW-SSE in the upper translation domain (Fig. 10B).

The internal shear zone within Area A merges with the longitudinal shear zone in the distal part of the lower translational domain, and defines the boundary between areas A and B (Fig. 9C). These shear zones outline a downdip-narrowing area defined by broadly convex-updip ridges. Consequently, Area A becomes dominated by the convex-downdip ridges (Fig. 9C). However, immediately downdip from the point where the shear zones merge, the ridges in Area A have slightly convex-updip geometries, most notably adjacent to the longitudinal shear zone (Fig. 9C).

Interpretation.– The ramps, deformed substrate, and shear fractures indicate that seabed erosion and deformation also occurred in this lower translational domain (Fig. 9A). The continuation of the longitudinal and internal shear zones suggests that the related intra-flow velocity variation identified in the upper translational domain also characterized the lower translational domain (Fig. 9B). Critically, however, the gradual downflow disappearance of the convex-updip flow fabrics between these shear zones suggests a downflow decrease in the internal velocity perturbations and/or tota induced by the cluster of megaclasts in the upper translational domain (Fig. 7B).

The cluster of megaclasts adjacent to the NE lateral margin (Figs. 9B and D) is located immediately downflow from, and has a similar width (2.5 km) to, the deformed substrate area (Fig. 9A). Similar to the observation from the Rapanui MTD (Cardona, S. *et al.*, 2020), where the presence of deformed substrate could be correlated to higher concentrations of rafted blocks (i.e. megaclasts). The higher concentration of megaclasts indicates an increase in flow competence overriding the area of the deformed substrate. In addition, the long-axis orientations of the megaclasts in the lower translational domain are generally oblique-to-sub-parallel to the overall north-westerly transport direction, which contrasts to those in the upper translational domain that are generally perpendicular to the transport direction (Fig. 10B). Their long-axis orientations are likely to be controlled by velocity gradients (Mazzanti, P. & De Blasio, F., 2010), where the megaclasts in the lower translational domain, adjacent to the NE lateral margin, were dragged against the stationary lateral wall (Fig. 9B). In contrast, the cluster of megaclasts in the upper translational domain, further away from the lateral margin, experienced a lower across-strike velocity gradient, which resulted in their long axes displaying an overall convex-updip geometry (Fig. 7B).

The top surface supports the interpretation of kinematic indicators within the internal body of the slide (Figs. 9B-C). Here, it is also evident that velocity perturbations induced by the cluster of megaclasts in upper translational domain (Figs. 7B and 8A) decreased downdip, and caused the two shear zones to merging (Fig. 9C). However, further downdip from the merger point, convex-updip ridges within Area A, which terminate at the longitudinal shear zone, suggests continued internal velocity perturbations (Fig. 9C).

Toe domain

Description.– The basal-shear surface in this domain serves as the frontal margin of Area A, and swings through 90^o to join the lateral margin of Area B, which that continues downdip, beyond the area imaged by 3D seismic reflection data (Fig. 11A). The seismic attribute expression of the deformed substrate (Fig. 11A) resembles that of the upper and lower translational domains, especially the high variance values (see Figs. 7A and 9A). In the SW part of this domain, there is a *c.*30 m-high ramp (Figs. 10A and D), which is of higher-relief than the one developed updip in the lower translational domain (*c.*20 m).

Debrite (SF-1) and a fold-and-thrusts system (SF-3) dominate the distal part of the toe domains in Area A and B, respectively (Fig. 11B). The thrusts within Area B dip to the SE, sub-parallel to the transport direction of the slide (Figs. 5G, 11B and D). Within the older slide, there is a cluster of relatively large megaclasts (*c.*2.5 km wide and *c.*5 km long) that are deformed by NNW-SSE-striking, NE-dipping thrusts; these trend broadly perpendicular to the frontal margin of Area A and to the NW-SE-striking thrusts within Area B (Fig. 11B). There is a relatively small number of thrusted megaclasts (i.e., indicated by high RMS amplitude and brown colour in Fig. 11B) within the frontal part of Area A, where they are encased in debrite (green colour in Fig. 11B). The thrusted megaclasts in Area A (NW-SE-to-E-W) have more variable strikes than those in the older slide (NW-SE) (Fig. 11B). Additionally, thrusts within the older slide and Area A are unidirectional, whereas those in Area B form pop-up-bounding conjugate pairs (Fig. 11B and D). The toe region of Area A is defined by a rugose seabed characterized by *c.*30 m-high ridges that are elevated above the flat seabed capping the older slide (see Fig. 11C). The vertical relief of the ridges in Area A is higher than in Area B (*c.*10 m, Fig. 11C) and in the lower translational domain (*c.*10 m, Fig. 9).

Interpretation.– The presence of a ramp along the basal-shear surface, which caps deformed substrate material, indicate that substrate deformation and erosion continued beneath the main body of the slide, even in this relatively distal location. Critically, evidence suggests that most of the thrusted megaclasts belong pre-date emplacement of the Gorgon Slide: (i) the abrupt truncation of most of the thrusted megaclasts in the older slide by the frontal margin of Area A; (ii) the strike difference between thrusts within areas A and B and those in the older slide (Fig. 11B); and (iii) the structural style difference between the thrusts in Area B, and Area A and the older slide. However, the small number of thrusted megaclasts in the frontal part of Area A (Fig. 11B) also suggests erosion and cannibalization by the Gorgon Slide.

In addition, Area B extends further downdip than Area A (Fig. 11A) and the longitudinal shear zone extends from the upper translational domain (see Figs. 7A and 8A) to join the lateral margin of Area B in the toe domain (Fig. 11B). This suggests that the frontal margins of the cells may influence the inferred intra-slide velocity perturbations. Specifically, the velocity perturbation could have originated due to the passage of the Gorgon Slide onto (i.e. Area A), and partly around (i.e. Area B), frontal confinement of Cell A, and causing the subsequent arrest of the flow in Area A with some upflow propagation (Fig. 11B). Therefore, this velocity perturbation may have connected with the downflow-propagating velocity perturbation induced by the cluster of megaclasts in the upper translational domain (Figs. 7B and 8A). The frontal confinement of Cell A is likely to be controlled by the depth of basal-shear surface, which is deeper than that of the Cell B. Moernaut, J. and De Batist, M. (2011) suggest that the depth of the basal shear surface is a key control on frontal emplacement style (frontally confined versus emergence, sensu Frey-Martínez, J. *et al.*, 2006). A deeper basal shear surface implies a higher potential-energy threshold that must be exceeded for the moving mass to abandon the basal-shear surface and emerge onto the coeval basin floor, thereby favouring frontal confinement. In contrast, a shallower basal-shear surface lowers this threshold, making frontal emergence more likely. In addition to basal-shear-surface depth, the architecture of the pre-existing slide (i.e., the thrusted megaclasts) downdip from the toe of Cell A may also have contributed locally to buttressing and arrest of the flow cell (Figs. 11B).

DISCUSSION

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

Single versus multiple failure events.– *Was the Gorgon Slide deposited during a single event, or does it comprise multiple, stacked deposits, emplaced by several, temporally discrete events?* Establishing this is key to developing a kinematic model for slide emplacement, and for assessing the geohazard potential of the related flows. Such stacked deposits might be anticipated in the Cenozoic stratigraphy of the NW Shelf of Australia, given its history of repeated slope failures (Hengesh, J. V. *et al.*, 2013; Nugraha, H. D. *et al.*, 2019; Wu, N. *et al.*, 2026). However, answering these questions is not straightforward, given our lack of borehole data and the still limited resolution (8-11 m) of even these high-quality 3D seismic reflection data. For example, individual gravity-flow deposits, and/or thin pelagic or hemipelagic intervals separating them, could be thinner than the resolution of our seismic reflection data.

These uncertainties are considered in more detail in the following five discussion points:

(i) *Does the slide source area contain evidence for multiple slope failure events?* We note that the headwall scarp of the slide is relatively simple, being defined by a single, relatively smooth, downdip-facing plane. This supports a single slope failure event, compared to a punctuated, retrogressive failure, which would result in a composite, more complex staircase-like geometry (e.g., Sawyer, D. E. *et al.*, 2009). We observe crown cracks immediately updip of the main scarp; these possibly mark the location of future slope failure events, which ultimately would give rise to a staircase-like scarp geometry (Varnes, D. J., 1978; Frey-Martinez, J. *et al.*, 2005).

(ii) *Does the basal-shear surface contain evidence for temporally separate, but spatially related flows?* We observe downdip converging grooves on the basal-shear surface immediately downdip of the slide headwall (Fig. 6) and in the upper translational domain (Fig. 7A). These grooves do not cross-cut one another (cf. Wu, N. *et al.*, 2026), suggesting they formed during a single event (e.g., Scarselli, N. *et al.*, 2013). There is also no evidence for intra-slide basal-shear surfaces decorated by grooves.

(iii) *Is there evidence for stratigraphically discrete deposits defined by their own distinct structural style?* Within the upper and lower translational domains, and the toe domain, the shear zones (e.g. Figs. 5D-E), thrusts (e.g. Fig. 5G), flow fabrics (e.g., Figs. 7B-C and 9B-C), and megaclasts (Fig. 8B) extend over the entire height of the slide. This is observed on broadly stratigraphically concordant maps generated within the deposits, as well as at the top of the deposit, which implies that these features are through-going rather than being stratigraphically decoupled.

(iv) *Does the lateral margin contain evidence for formation by multiple slope failure events?* The lateral margin is geometrically similar to the headwall and is defined by a structurally simple, sub-vertical plane that extends continuously for c.30 km), suggesting emplacement during a single event (e.g. Fig. 3B).

(v) *Is there any evidence for relatively continuous, through-going, intra-slide fine-grained (e.g., pelagic, hemipelagic) deposits separating individual sediment gravity-flow deposits (e.g., mass-transport deposits)?* Across the deposit there is no seismically resolvable evidence for intra-deposit hiatuses that would support multiple catastrophic, slide-related depositional events. However, the hemipelagic/pelagic deposits capping the older slide, and the evidence of erosion at the base of the Gorgon Slide, supports a significant break between the deposition of the these two units (e.g. Fig. 4B). Nevertheless, the *c.*8-11 m vertical resolution of the seismic reflection data means that closely spaced sub-events separated by thin deposits or short depositional hiatuses cannot be entirely excluded. We therefore treat emplacement during a single event as the interpretation most consistent with the resolved stratigraphic and structural relationships.

The available evidence supports the Gorgon Slide emplacement during a single event. Deposition of such a large slide (i.e., total volume of *c.*500 km^3 and thickness up to 500 m) by a single event is not unusual, with similar features described from offshore New Zealand (Joanne, C. *et al.*, 2013), South China Sea (Sun, Q. *et al.*, 2018b), and offshore Brunei (Gee, M. *et al.*, 2007).

A multi-cell emplacement model.– A multiple flow cell model for subaqueous slides is proposed by Alsop, G. I. and Marco, S. (2014) based on field data, developing a model originally proposed by Farrell, S. G. (1984). They suggest that a large (first-order) slide consists of several smaller (second-order) flow cells formed during the transport and ultimate emplacement of a sediment mass. These smaller flow cells may interact with each other and cause overprinting on earlier formed structures. Similar kinematic interactions between intra-flow cells are documented from sonar (e.g., Prior, D. B. *et al.*, 1984; Masson, D. *et al.*, 1993; Gee, M. J. *et al.*, 2001) and 3D seismic reflection data (e.g., Bull, S. *et al.*, 2009; Steventon, M. J. *et al.*, 2019; Nugraha, H. D. *et al.*, 2020; Abu, C. *et al.*, 2022; Scarselli, N. et al., 2016), where primary (longitudinal shear zone) and secondary (sinuous) flow fabrics form between and define evolving cells (sensu Bull, S. *et al.*, 2009). These kinematic indicators suggest the portions (i.e. cells) of the translating mass were travelling at different speeds and/or travelled at slightly different times (Masson, D. *et al.*, 1993; Gee, M. *et al.*, 2005).

In this study, the Gorgon Slide appears to comprise at least two intra-slide (second-order) flow cells, which are Areas A and B. Hereafter, we interpret both areas as Cells A and B because they satisfy the diagnostic criteria (see the Introduction). First, both occur within the Gorgon Slide body and are not separated by a resolvable stratigraphic boundary. Second, they are separated by a persistent longitudinal shear zone located within, rather than at the external margin of, the slide. Third, the two areas contain systematically contrasting flow fabrics across the longitudinal shear zone. Finally, the two areas record different translation and arrest histories: Cell A is frontally confined and ultimately ceased translating, whereas Cell B extends farther downdip and remained mobile after Cell A had arrested. Collectively, these relationships indicate lateral kinematic partitioning within the translating Gorgon Slide. Such partitioning into flow cells is comparable to other intra-slide shear zones. For instance, in offshore Mozambique, Abu, C. et al., (2022) document transport-parallel shear zones c.90-

120 m wide and c.12 km long, characterised by chaotic seismic facies and c.4-6 m differences in top-surface relief. Similar kinematic partitioning occurs in the Haya Slide, Makassar Strait (c.150 km²; Nugraha, H.D., *et al.,* 2020), and in the Namibia Slump Complex (c.1200 km²; Scarselli N., *et al.,* 2016). Furthermore, the shear zone in the Gorgon Slide differs from a lateral margin because slide material occurs on both sides of the shear zone, while an inherited lineament is not favoured because the zone is confined to the slide and is spatially associated with contrasting emplacement-related flow fabrics. The emplacement processes of the Gorgon Slide are captured in a schematic model that recognises three stages of development (Fig. 12).

Stage 1.– Prior to slope degradation, a surface rupture might have been triggered by two main mechanisms (Fig. 12A). First, the normal faults bounding the horst could have been inverted due to regional compression, which then destabilised the slope (Keep, M. *et al.*, 1998; Nugraha, H. D. *et al.*, 2019). Second, the existence of pockmarks observed on the seabed (see Fig. 6) implies that there has been active fluid venting in the headwall area (Hengesh, J. V. *et al.*, 2013), most likely originating from the underlying, gas-bearing horst block (Fig. 2). Gas leakage into shallower sediments could have lowered the shear strength of these sediments, and primed the slope for subsequent failure (Scarselli, N. *et al.*, 2013). However, the Gorgon Slide was not an isolated occurrence, but rather the most recent. Previous collapse of the continental margin is recorded in the older (pre-Gorgon) slide (see Fig. 4).

Stage 2.– The arcuate geometry of the main headwall scarp is consistent with our preferred interpretation that the failed sediments were evacuated during a single landslide event (see Fig. 6). The evacuated sediments might include megaclasts derived from either the headwall and/or megaclasts entrained from the layered slope substrate. During translation, the megaclasts were deformed and fragmented (e.g., Nwoko, J. *et al.*, 2020). The downdip-converging grooves within the headwall and upper translational domains suggest a convergent pathway of the slide, resulting in the clustering of the megaclasts (Fig. 12B). In the leeside of the cluster of megaclasts, the following features formed: (i) convex-updip flow fabrics within the slide, and (ii) convex-updip ridges on top of the slide. These features indicate slower transport velocity in and around the area of concentrated megaclasts (Fig. 12B). Higher transport velocities of flows moving around the megaclast-rich area led to the formation of the longitudinal shear zone, and the initiation of Cells A and B. The cluster of megaclasts effectively acted as an obstacle to the initial, single-cell flow. Other studies have also documented such mechanism, where the geometry of flow fabrics and ridges downdip from translating megaclasts suggest slower-moving flows than surrounding materials (e.g., Masson, D. *et al.*, 1993; Lastras, G. *et al.*, 2005; Gee, M. *et al.*, 2006; Bull, S. *et al.*, 2009).

Stage 3.– The downdip propagation of the basal-shear surface was coupled with the evolution of the internal body and top surface of the slide. The area covering the convex-updip flow fabrics and ridges narrowed downdip (Fig. 12C), which suggests a reduction in the influence of the cluster of megaclasts on slowing down the flow of material in its leeside. Thus, we interpret this area as a 'shadow zone'. The shadow zone is bound by the longitudinal shear zone separating the two cells, and the internal shear zone within Cell A (Fig. 12C). The formation of this shadow zone and related bounding structures illustrates how megaclasts influence flow processes within a slide (e.g. Masson, D. *et al.*, 1993; Lucente, C. C. & Pini, G. A., 2003; Jackson, C. A., 2011; Hodgson, D. *et al.*, 2018).

Downdip from the shadow zone, ridges within Cell A show convex-updip geometries adjacent to the longitudinal shear zone. In contrast, ridges within Cell B consistently exhibit convex-downdip geometries (Fig. 12C). These geometries indicate that Cell A resisted the downdip translation of Cell B, meaning its internal ridges were dragged downdip whereas those in Cell B were dragged updip (Fig. 12C). These relationships indicate that Cell A had lower mobility than Cell B, either because it translated more slowly and/or because it ceased translating earlier. Cell B consequently remained mobile farther downdip and achieved a greater runout. Furthermore, the prominent seabed relief characterizing the frontal margin of Cell A suggest a shortening and thickening effect driven by this part of the flow being frontally confined (Fig. 12C) (Masson, D. *et al.*, 1993; Gee, M. *et al.*, 2006). Cell

B, however, was able to translate further downdip than Cell A, meaning the former had relatively higher mobility than the latter.

Impact of flow cell formation on submarine landslide kinematics and structure

Submarine debris flows can travel for tens to hundreds of km across low gradient (*c.*<1°) continental slopes, despite their cohesive nature (Gee, M. *et al.*, 1999; Lastras, G. *et al.*, 2005). This mobility can be explained by sustained pore-fluid pressure within the flow during transport (Major, J. J. & Iverson, R. M., 1999; McArdell, B. W. *et al.*, 2007), and the presence of a thin lubricating layer of fluid at the base of the frontal part of the flow (i.e. hydroplaning, Mohrig, D. *et al.*, 1998). Ultimately, a debrite is formed by *en masse* freezing of the debris flow (e.g. Talling, P. J. *et al.*, 2012), where materials at flow margins (i.e. frontal and lateral) cease moving first, followed by materials in the main body of the flow (Iverson, R. M., 1997).

The Gorgon Slide provides evidence of a mass flow splitting into two smaller flow cells (Cell A and B, Fig. 12). The relationship between the two cells suggests that Cell A ceased movement, while Cell B was still in motion. This suggests that *en masse* freezing did not occur across the entire body of the flow synchronously. Instead, individual flow cells froze at different times, resulting in different runout duration and distance of the cells. We propose that lateral friction and related pore-fluid pressure played an important role in controlling the runout distance of the two cells, in addition to the frontal confinement. The longitudinal shear zone may have sustained excess pore-fluid pressure between the two cells, such that low friction between the two cells may be maintained, allowing continued translation of Cell B despite being partly impeded by Cell A. In contrast, pore-fluid pressure may likely be dissipated at the lateral margins of the flow during translation (e.g. NE lateral margin, Fig. 3), which could have promoted relatively high friction between the moving slide (e.g. Cell A) and stationary lateral substrate (i.e. the older slide and locally other slope strata). This high friction at the lateral margin may reduce runout distance more substantially than the friction at the longitudinal shear zone.

Such mechanisms are also observed from experimental studies (Major, J. J. & Iverson, R. M., 1999; De Haas, T. *et al.*, 2015). Our results suggest that a multi-cell debris flow could undergo 'punctuated' freezing, where one cell may have a shorter runout distance than the others due to spatial differences in the pore pressure and related friction between bounding cells.

Controls on flow cell formation

Flow cell formation within a slide depends on internal velocity perturbations, which are controlled by variations in at least three local factors (Farrell, S. G., 1984; Alsop, G. I. & Marco, S., 2011; 2014): (i) the lithology and/or geometry of stratigraphic elements overridden by the slide (e.g. older slides, channels and lobes); (ii) fluid pressures within the slide and/or its substrate and lateral margins; and (iii) the slope gradient and/or geometry of basal-shear surface underlying the slide.

In the Gorgon Slide, a cluster of megaclasts within a debritic matrix-initiated flow cell formation. This implies that lithology, in particular variations of the degree of disaggregation within the slide, play a key role in forming the two seismic-scale flow cells. Variability of local fluid pressures within the slide could be represented by the interactions between the slide and clustered megaclasts in the upper translational domain (e.g., Figs. 7B and 12B). As an obstacle to flow, the clustered megaclasts may increase local fluid pressure as they resisted the translating slide, and thus induce the formation of the cells. In addition, the geometry of the basal-shear surface was also important, given it caused the flow to converge, clustering the megaclasts, initiating velocity perturbation, and terminating the flow cells. Most notably, frontal confinement of Cell A, along with the high-relief top surface, coincides with deeper basal-shear surface as compared to those of Cell B that has shallower basal-shear surface (see Fig. 4B). As demonstrated by Moernaut, J. & De Batist, M., (2011), the depth of the basal shear surface exerts a key control on the frontal margin of a slide: a deeper basal-shear surface may be associated with frontal confinement, whereas a shallower basal-shear surface may be associated with frontal emergence. By having laterally different depth of basal-shear surface, a single slide event could therefore display both frontal confinement and emergence along its toewall, for example, along the toewall of the Haya Slide in the Makassar Strait, Indonesia (Nugraha, H. D. *et al.*, 2020). In addition, megaclasts incorporated within an older slide may locally promote frontal confinement along the toewall (Fig. 12C).

Local variations of lithology, fluid pressure and basal-shear surface geometry may have been influential prior to emplacement, but their properties could also evolve during translation and cessation of the parent flow (Iverson, R. M., 1997; Dykstra, M. *et al.*, 2011; Joanne, C. *et al.*, 2013; Alsop, G. I. & Marco, S., 2014; Ortiz-Karpf, A. *et al.*, 2017; Hodgson, D. *et al.*, 2018).

Implications for geohazard risk assessment

Submarine landslide emplacement can severely damage seabed infrastructure, including submarine pipelines (Parker, E. J. *et al.*, 2008; Zakeri, A., 2009; Randolph, M. F. & White, D. J., 2012; Vanneste, M. *et al.*, 2013). The Gorgon Slide study shows that high-quality 3D seismic reflection data allows the internal partitioning, differential translation and sequential arrest of a submarine landslide to be reconstructed from its preserved kinematic indicators, which otherwise require direct monitoring of an active landslide. The seismic reflection data do not directly constrain flow velocity, rheology, drag force or pipeline response, and therefore cannot provide quantitative estimates of infrastructure loading. However, recognition of two kinematically distinct cells within a single emplacement event provides a geological framework and conceptual constraints. For example, a key control on pipeline survivability is the drag imposed by the moving mass flow, which can determine whether a pipeline remains in place or is torn out and transported downslope (Parker, E. J. *et al.*, 2008; Zakeri, A., 2009). Quantifying this drag is therefore a critical requirement in the design of seafloor installations (e.g., Zakeri, A., 2009). The translation of distinct flow cells, such as Cells A and B, suggests that pipelines may be subjected to two kinematically distinct cells within a single landslide event. This may potentially produce spatially variable drag forces and, along the longitudinal shear zone, intensified shearing may further elevate the likelihood of pipeline failure. Accordingly, the flow behaviour recorded in the Gorgon Slide provides useful conceptual constraints for experimental (e.g., Zakeri, A. *et al.*, 2008) and numerical (e.g., Zakeri, A., 2009; Randolph, M. F. & White, D. J., 2012) modelling.

CONCLUSIONS

This analysis of the emplacement of the Gorgon Slide concludes the following:

1. The Gorgon Slide was evacuated from a steep, NE-SW trending, *c.*350 m-high headwall scarp and transported towards the NW. The megaclasts were likely derived from the headwall and/or entrained from layered slope substrate during translation.
2. The proximal part of the translation domain contains downdip-converging grooves on the basal-shear surface, which indicate that the pathway of the slide was focused on its lateral margin in the NE. The convergent pathway of the flow results in the clustering of the megaclasts, whose long-axes are generally trending NE-SW, perpendicular to the transport direction. This clustering of megaclasts obstructed flow, causing velocity perturbation within the slide, as recorded by convex-updip flow fabrics within the internal body, and by convex-updip ridges on the seabed. These features indicate a slower transport velocity of the cluster of megaclasts and materials in its leeside. The area of the slower-moving material narrows downdip, indicating that velocity perturbation caused by the cluster of megaclasts gradually diminished downflow, forming a ‘shadow zone’. Relative mobility was greater around the megaclast cluster, resulting in the formation of longitudinal shear zones and the initiation of two flow cells, namely Cells A and B.
3. The distal part of the translation domain contains kinematic indicators recording erosional and deformational processes on the basal-shear surface. Erosional processes are evidenced by a ramp, and deformational processes are evidenced by deformed substrate or basal-shear zone and shear fractures adjacent to the NE lateral margin. The deformed substrate and shear fractures are closely related to the thickest part of the slide, comprising a cluster of

megaclasts, which individually trend NNW-SSE, oblique to sub-parallel to the transport direction. Flow fabrics within the slide and ridges on the seabed of Cell A were dragged downdip, while those of Cell B were dragged updip. This points to Cell A acting as an impediment to the movement of the more mobile Cell B.

4. In the toe domain, the frontal margin of Cell A is marked by positive seabed relief (*c.*30 m-high) that gradually decreases updip, which is significantly higher than the relief of Cell B (*c.*10 m-high). This suggests that Cell A was buttressed against its frontally-confined toewall that coincides with a pre-existing cluster of megaclasts (i.e. encased by older slide) and deeper basal-shear surface, while Cell B was not. Therefore, as there were no flow obstacles, Cell B was able to travel further than Cell A.
5. The morphology of the basal-shear surface and the degree of disaggregation within the slide, especially the megaclasts, played important roles in flow cell evolution. The basal-shear surface controlled the pathway of the slide, and the clustering of the megaclasts. The megaclast clusters induced internal velocity perturbations that controlled the initiation and cessation of intra-slide flow cells.
6. Simultaneous *en masse* freezing was unlikely to have occurred throughout the body of the Gorgon Slide. Instead, 'punctuated' freezing, where motion in Cell A had ceased while Cell B was still active, occurred potentially due to differential friction and pore-fluid pressure dissipation at flow cells margins. For instance, excess pore-fluid may have been maintained within the longitudinal shear zone, so that Cell B likely only experienced minimal friction against the Cell A. In contrast, excess pore-fluid pressure may have been dissipated at lateral margins, such as at the NE lateral margin separating Cell A and stationary substrate. Thus, Cell A may have experienced higher lateral friction than that of Cell B, resulting in reduced runout distance. This punctuated freezing mechanism may be considered for modelling the impact of submarine landslides on submarine infrastructures.

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CONFLICT OF INTEREST

No conflict of interest declared.

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FIGURE CAPTIONS

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

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

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

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

Fig. 5.--- Seismic facies classification used in this study. **A)** Seismic facies description and interpretation. **B)** Variance attributes extraction between the basal-shear surface and an iso-proportional surface (50% between the basal-shear surface and the seabed). **C-E)** Seismic sections showing seismic facies within the translation domain. **F)** A time-slice of variance attribute extraction (see G for position) showing seismic facies in the toe domain. **G)** A seismic section showing seismic facies in the toe domain. Vertical exaggeration of all seismic sections is 15.

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

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

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

Fig. 9.--- Lower translational domain of the Gorgon Slide. **A)** Basal-shear surface variance map (top) and its interpretation (bottom). **B)** Internal body RMS amplitude map (extracted 50 ms above and below isoproportional horizon, orange) (top) and its interpretation (bottom). **C)** Top surface dip map

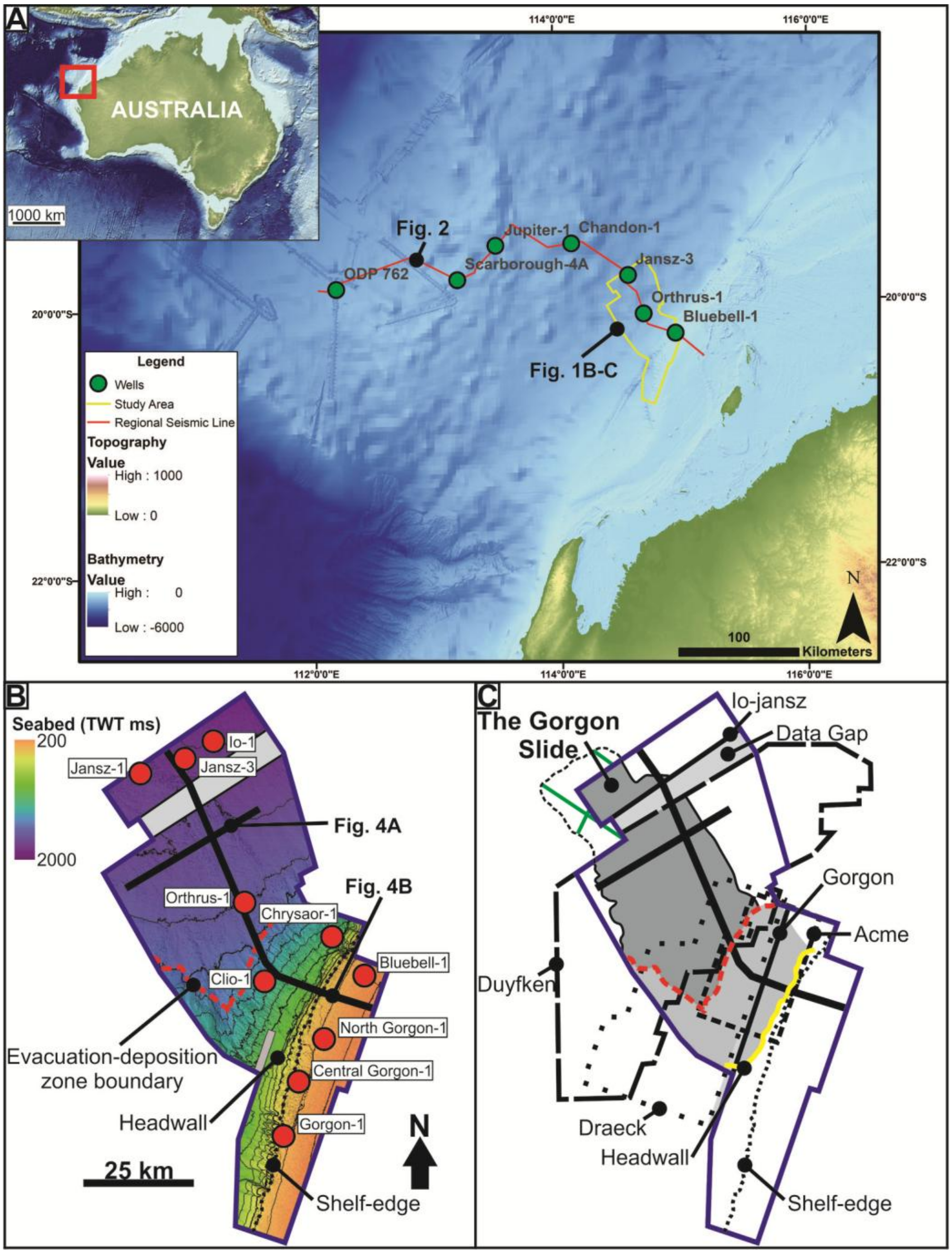
(top) and its interpretation (bottom). **D)** Seismic sections, uninterpreted (above) and interpreted (bottom), showing seismic facies across the upper translational domain. See text for discussion.

Fig. 10.--- Dimensions and orientation of the megaclasts in the upper and lower translation domains. **A)** Megaclasts in the upper translational domain are generally thinner with longer long-axes, as compared to the ones in the lower translational domain that are thicker with shorter long-axes. **B)** Megaclasts in the upper translational domain are generally oriented perpendicular, and the ones in the lower translational domain are oblique to sub-parallel, to the transport direction.

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

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

Fig. 1



A
AUSTRALIA
1000 km
114°0'0"E
116°0'0"E
112°0'0"E
20°0'0"S
22°0'0"S
Fig. 2
ODP 762
Scarborough-4A
Jupiter-1
Chandon-1
Jansz-3
Orthrus-1
Bluebell-1
Fig. 1B-C
Legend
Wells
Study Area
Regional Seismic Line
Topography
Value
High : 1000
Low : 0
Bathymetry
Value
High : 0
Low : -6000
100
Kilometers
N
B
Seabed (TWT ms)
200
2000
Io-1
Jansz-1
Jansz-3
Fig. 4A
Fig. 4B
Orthrus-1
Chrysaor-1
Bluebell-1
Clio-1
North Gorgon-1
Central Gorgon-1
Gorgon-1
Evacuation-deposition zone boundary
Headwall
Shelf-edge
25 km
N
C
The Gorgon Slide
Io-jansz
Data Gap
Gorgon
Acme
Duyfken
Draeck
Headwall
Shelf-edge

Fig. 2

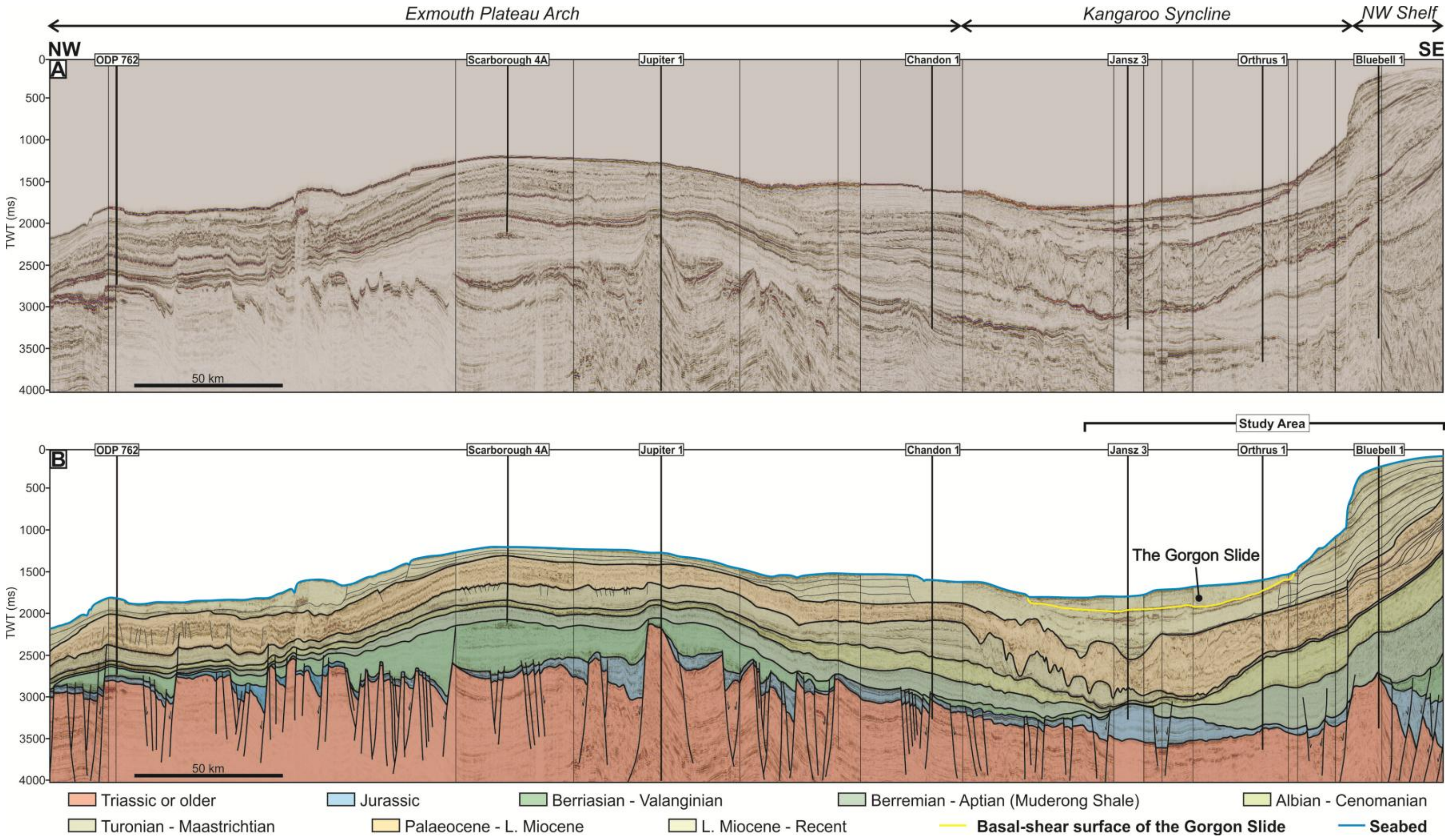
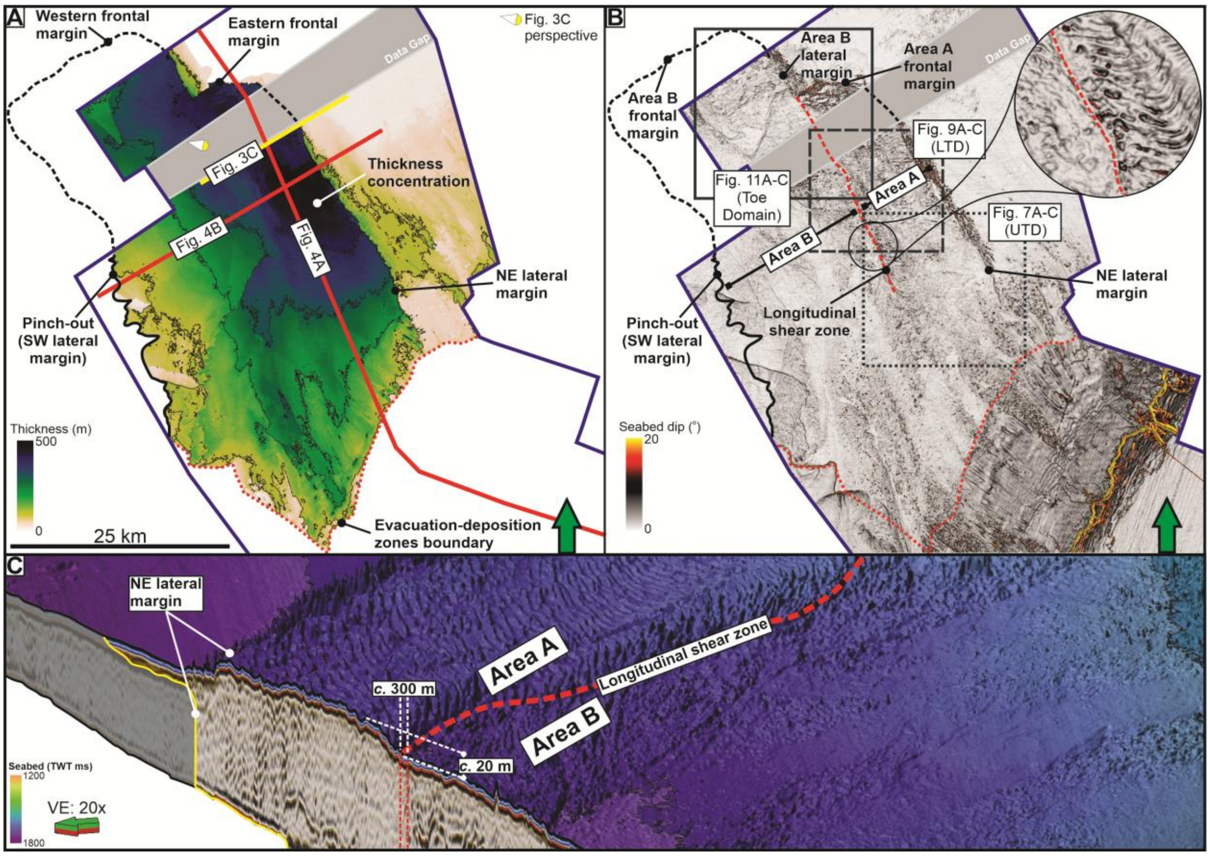

Exmouth Plateau Arch
Kangaroo Syncline
NW Shelf
NW
SE
A
B
ODP 762
Scarborough 4A
Jupiter 1
Chandon 1
Jansz 3
Orthrus 1
Bluebell 1
TWT (ms)
0
500
1000
1500
2000
2500
3000
3500
4000
50 km
Study Area
The Gorgon Slide
Triassic or older
Jurassic
Berriasian - Valanginian
Berremian - Aptian (Muderong Shale)
Albian - Cenomanian
Turonian - Maastrichtian
Palaeocene - L. Miocene
L. Miocene - Recent
Basal-shear surface of the Gorgon Slide
Seabed

Fig. 3



A
Western frontal margin
Eastern frontal margin
Fig. 3C perspective
Data Gap
Fig. 3C
Thickness concentration
Fig. 4B
Fig. 4A
NE lateral margin
Pinch-out (SW lateral margin)
Thickness (m)
500
0
25 km
Evacuation-deposition zones boundary
B
Area B lateral margin
Area A frontal margin
Data Gap
Area B frontal margin
Fig. 9A-C (LTD)
Fig. 11A-C (Toe Domain)
Area A
Area B
Fig. 7A-C (UTD)
NE lateral margin
Pinch-out (SW lateral margin)
Longitudinal shear zone
Seabed dip (°)
20
0
C
NE lateral margin
Area A
Longitudinal shear zone
c. 300 m
Area B
c. 20 m
Seabed (TWT ms)
1200
1800
VE: 20x

Fig. 4

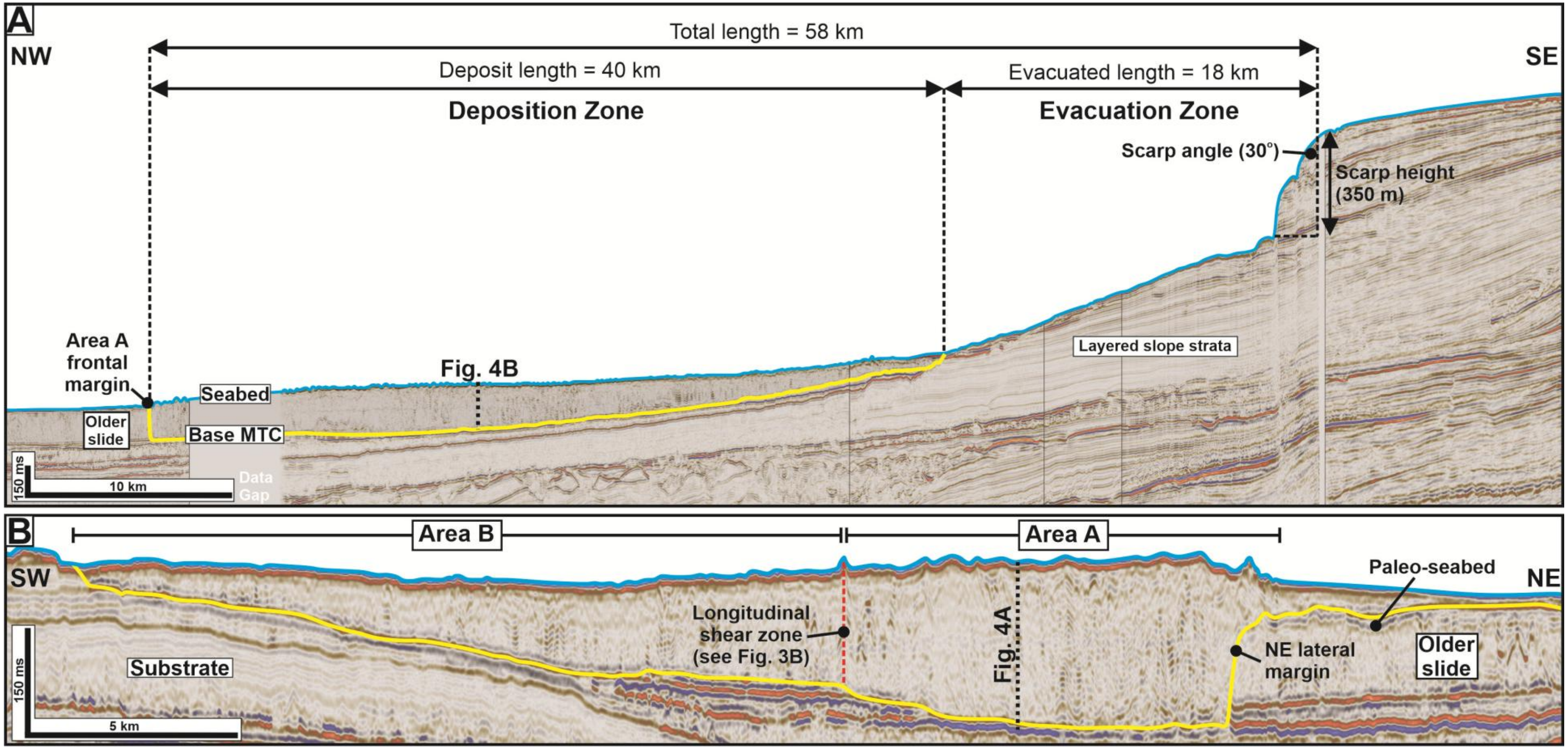

A
NW
SE
Total length = 58 km
Deposit length = 40 km
Evacuated length = 18 km
Deposition Zone
Evacuation Zone
Scarp angle (30°)
Scarp height (350 m)
Layered slope strata
Area A frontal margin
Fig. 4B
Seabed
Older slide
Base MTC
Data Gap
150 ms
10 km
B
Area B
Area A
SW
NE
Paleo-seabed
Longitudinal shear zone (see Fig. 3B)
Fig. 4A
NE lateral margin
Older slide
Substrate
150 ms
5 km

Fig. 5

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).	Slumps to 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).

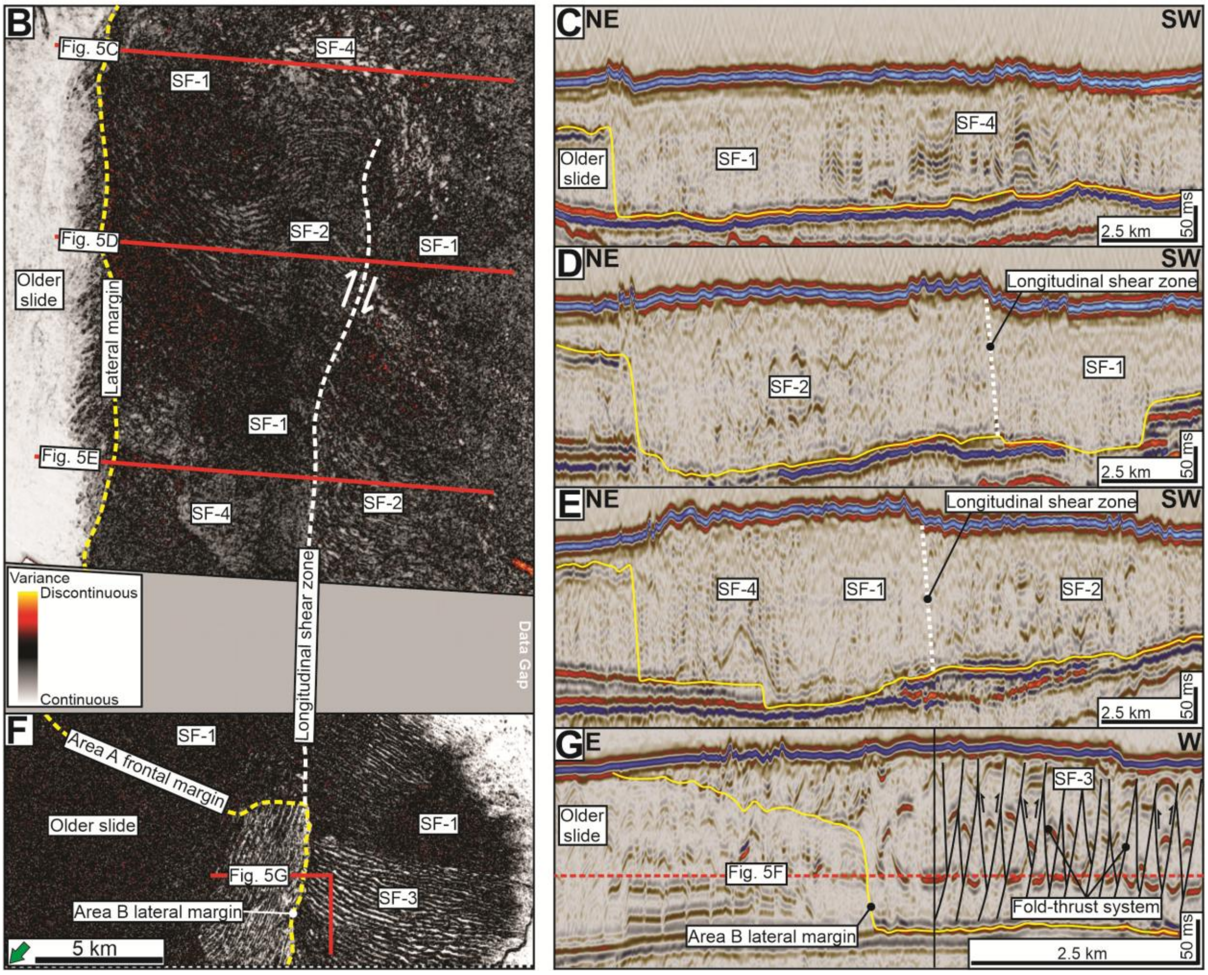
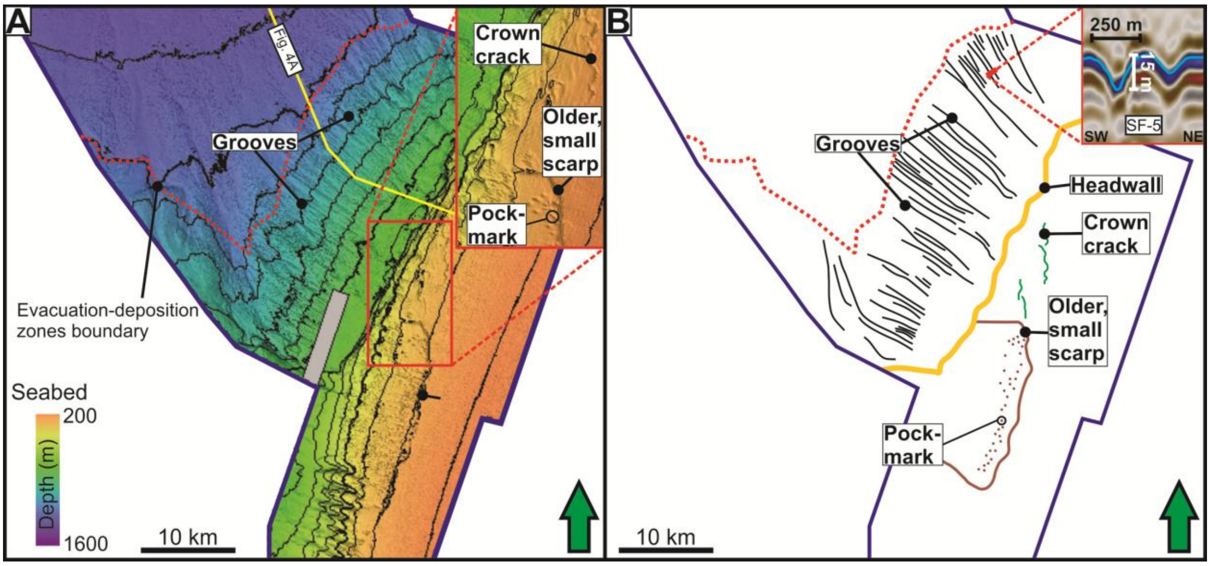
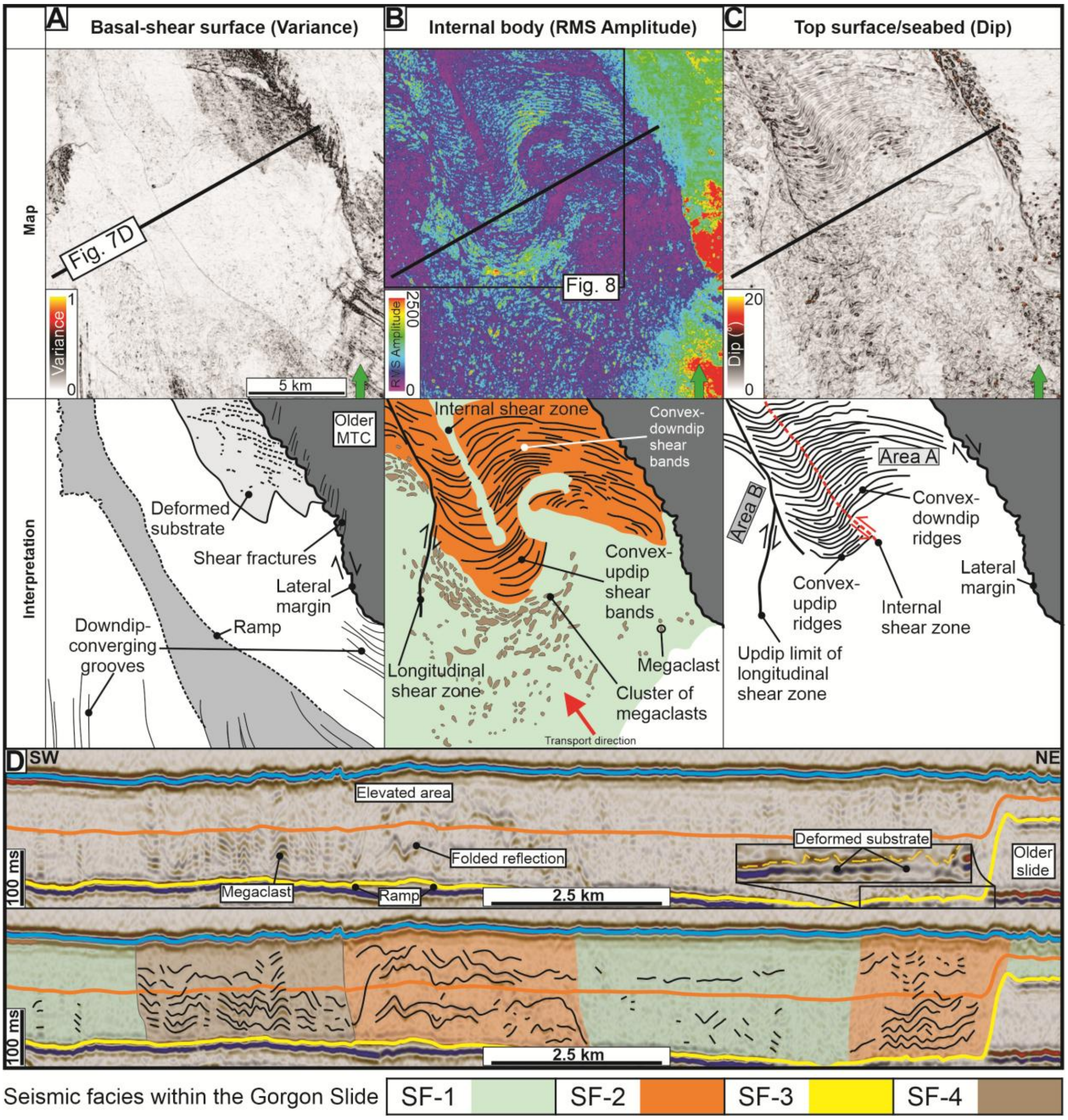


Fig. 6



A
Fig. 4A
Crown crack
Older, small scarp
Grooves
Pock-mark
Evacuation-deposition zones boundary
Seabed
200
Depth (m)
1600
10 km
B
250 m
5 m
SF-5
SW
NE
Grooves
Headwall
Crown crack
Older, small scarp
Pock-mark
10 km

Fig. 7



A Basal-shear surface (Variance)
B Internal body (RMS Amplitude)
C Top surface/seabed (Dip)
Map
Interpretation
Fig. 7D
Variance
1
0
5 km
Fig. 8
RMS Amplitude
2500
0
Dip (°)
20
0
Older MTC
Deformed substrate
Shear fractures
Lateral margin
Ramp
Downdip-converging grooves
Internal shear zone
Convex-downdip shear bands
Convex-updip shear bands
Megaclast
Longitudinal shear zone
Cluster of megaclasts
Transport direction
Area A
Area B
Convex-downdip ridges
Convex-updip ridges
Internal shear zone
Lateral margin
Updip limit of longitudinal shear zone
D SW
NE
Elevated area
Deformed substrate
Folded reflection
Older slide
100 ms
Megaclast
Ramp
2.5 km
100 ms
2.5 km
Seismic facies within the Gorgon Slide
SF-1
SF-2
SF-3
SF-4

Fig. 8

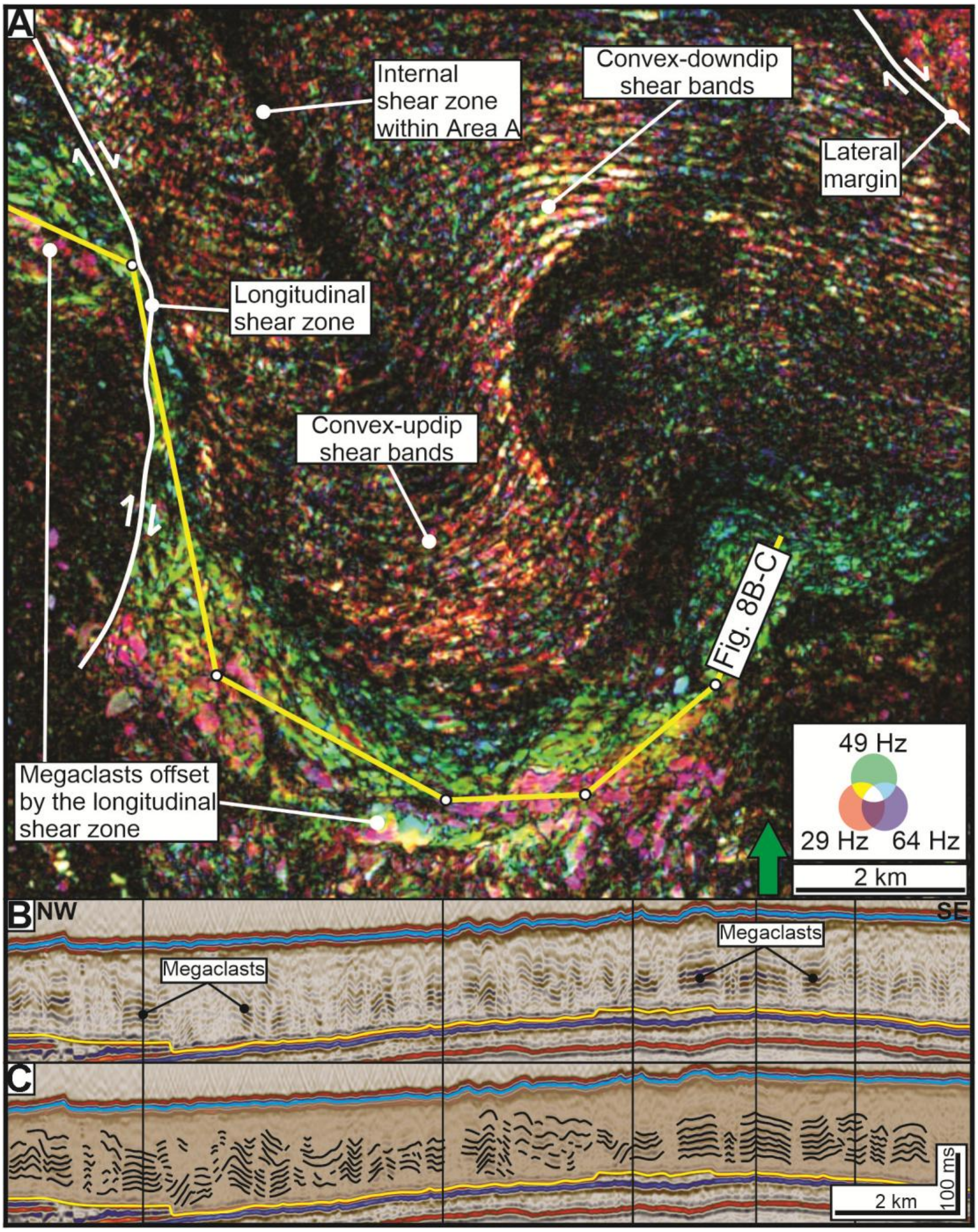
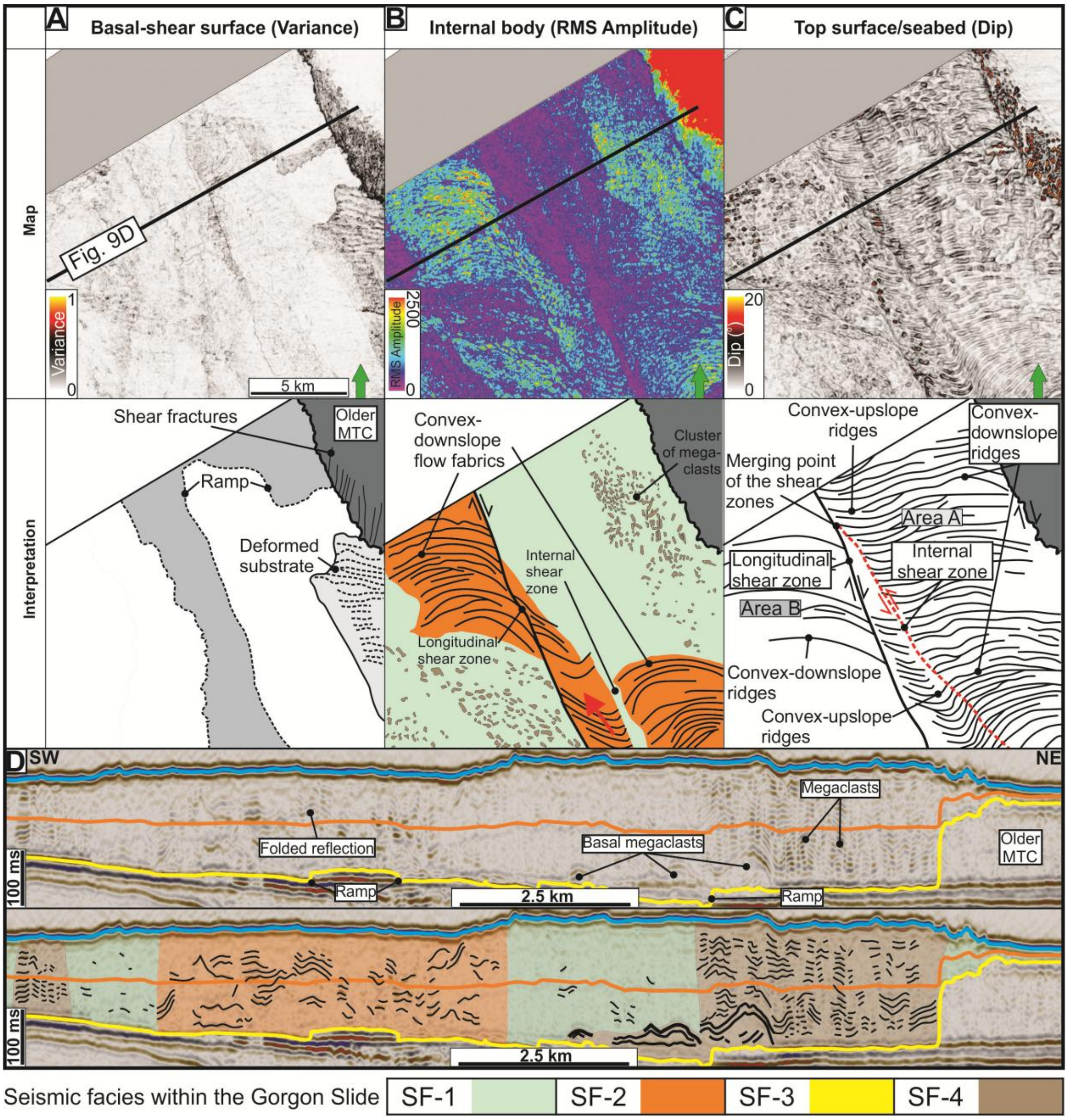

A
Internal
shear zone
within Area A
Convex-downdip
shear bands
Lateral
margin
Longitudinal
shear zone
Convex-updip
shear bands
Fig. 8B-C
Megaclasts offset
by the longitudinal
shear zone
49 Hz
29 Hz
64 Hz
2 km
B
NW
SE
Megaclasts
Megaclasts
C
2 km
100 ms

Fig. 9



A
Basal-shear surface (Variance)
B
Internal body (RMS Amplitude)
C
Top surface/seabed (Dip)
Map
Interpretation
Fig. 9D
Variance
1
0
5 km
RMS Amplitude
2500
0
Dip (°)
20
0
Shear fractures
Older MTC
Ramp
Deformed substrate
Convex-downslope flow fabrics
Cluster of mega-clasts
Internal shear zone
Longitudinal shear zone
Convex-upslope ridges
Convex-downslope ridges
Merging point of the shear zones
Area A
Longitudinal shear zone
Internal shear zone
Area B
Convex-downslope ridges
Convex-upslope ridges
D
SW
NE
Megaclasts
Folded reflection
Basal megaclasts
Older MTC
100 ms
Ramp
2.5 km
Ramp
100 ms
2.5 km
Seismic facies within the Gorgon Slide
SF-1
SF-2
SF-3
SF-4

Fig. 10

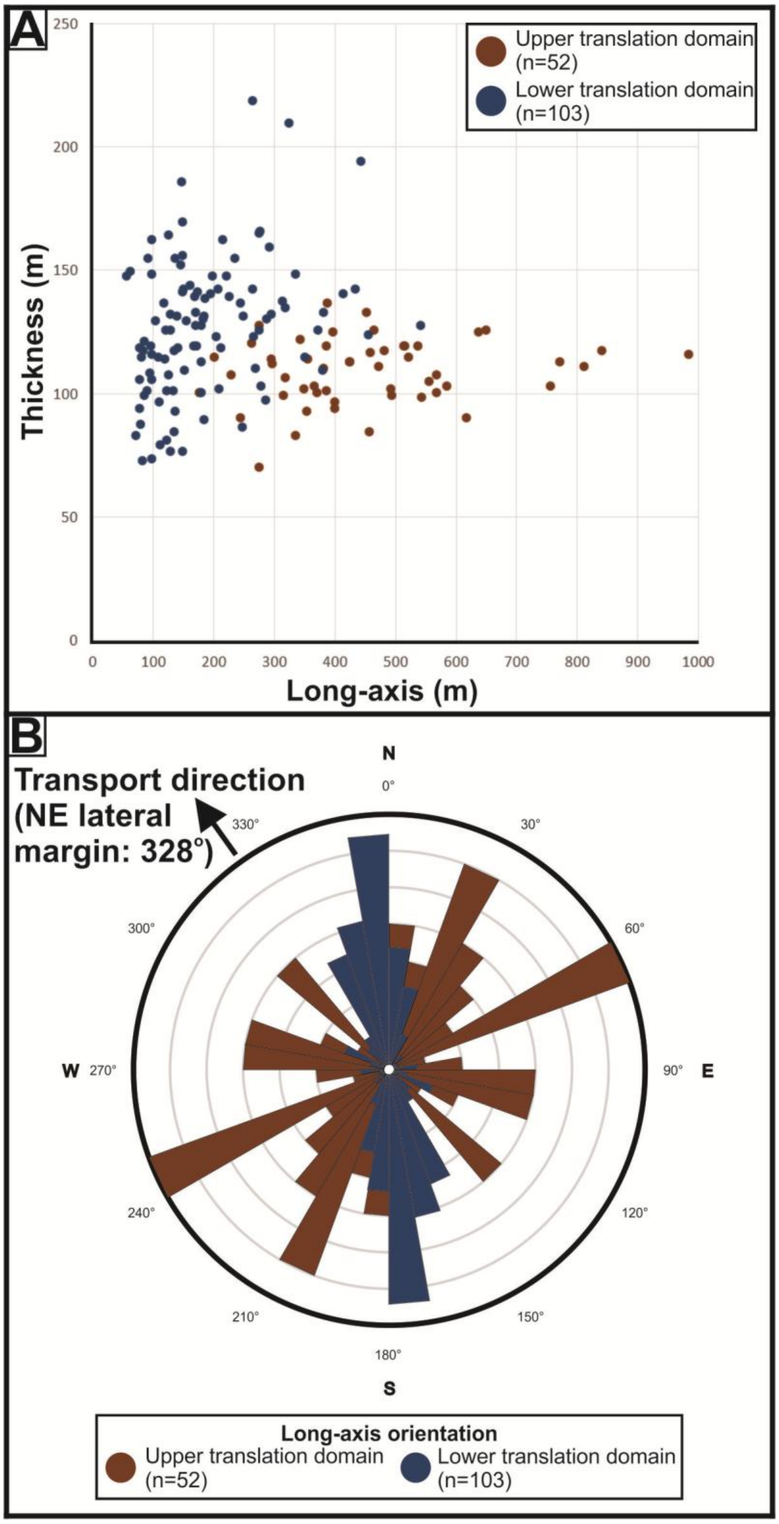

A
Upper translation domain (n=52)
Lower translation domain (n=103)
Thickness (m)
Long-axis (m)
0
50
100
150
200
250
0
100
200
300
400
500
600
700
800
900
1000
B
Transport direction (NE lateral margin: 328°)
N
0°
30°
60°
90°
E
120°
150°
180°
S
210°
240°
270°
W
300°
330°
Long-axis orientation
Upper translation domain (n=52)
Lower translation domain (n=103)

Fig. 11

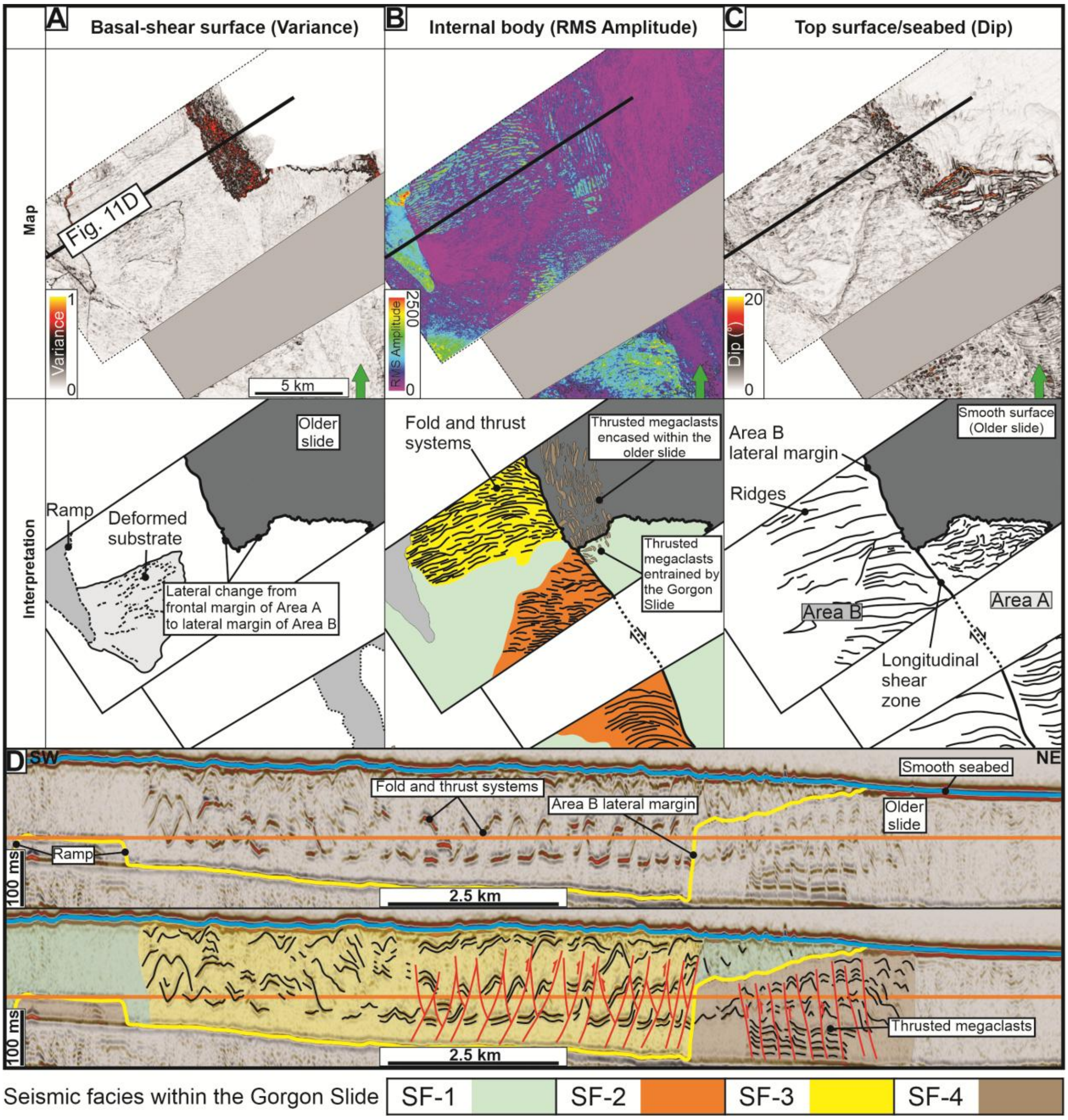
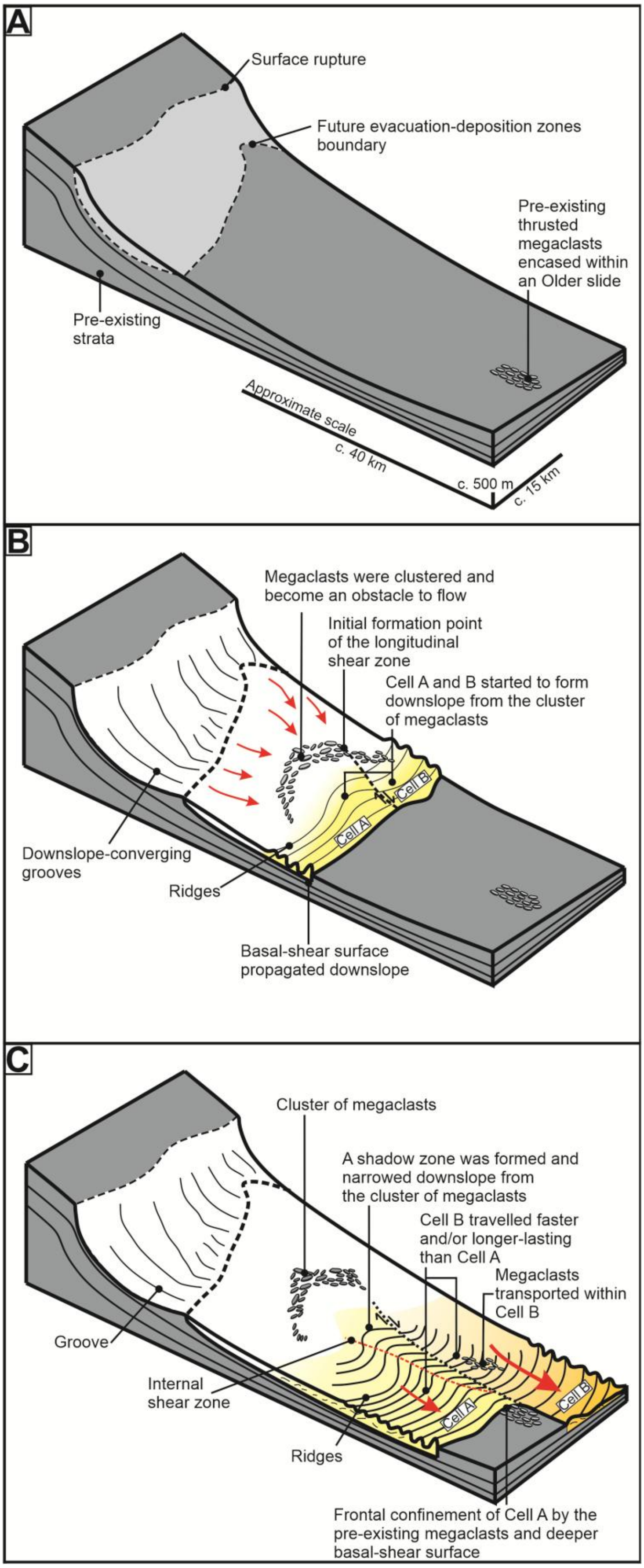

A
Basal-shear surface (Variance)
B
Internal body (RMS Amplitude)
C
Top surface/seabed (Dip)
Map
Interpretation
Fig. 11D
Variance
1
0
RMS Amplitude
2500
0
Dip (°)
20
0
5 km
Older slide
Ramp
Deformed substrate
Lateral change from frontal margin of Area A to lateral margin of Area B
Fold and thrust systems
Thrusted megaclasts encased within the older slide
Thrusted megaclasts entrained by the Gorgon Slide
Smooth surface (Older slide)
Area B lateral margin
Ridges
Area B
Area A
Longitudinal shear zone
D
SW
NE
Smooth seabed
Fold and thrust systems
Area B lateral margin
Older slide
Ramp
100 ms
2.5 km
Thrusted megaclasts
100 ms
2.5 km
Seismic facies within the Gorgon Slide
SF-1
SF-2
SF-3
SF-4

Fig. 12


A
Surface rupture
Future evacuation-deposition zones boundary
Pre-existing thrusted megaclasts encased within an Older slide
Pre-existing strata
Approximate scale
c. 40 km
c. 500 m
c. 15 km
B
Megaclasts were clustered and become an obstacle to flow
Initial formation point of the longitudinal shear zone
Cell A and B started to form downslope from the cluster of megaclasts
Cell A
Cell B
Downslope-converging grooves
Ridges
Basal-shear surface propagated downslope
C
Cluster of megaclasts
A shadow zone was formed and narrowed downslope from the cluster of megaclasts
Cell B travelled faster and/or longer-lasting than Cell A
Megaclasts transported within Cell B
Groove
Internal shear zone
Ridges
Cell A
Cell B
Frontal confinement of Cell A by the pre-existing megaclasts and deeper basal-shear surface