1 | INTRODUCTION

The collapse of sedimentary piles under the effect of gravity-driven processes is a common trait of continental margins (Butler & Turner, 2010; Ogata et al., 2019). Collapse commonly takes place across multiple scales and processes resulting in a wide spectrum of products that historically have been classified with a complex variety of terms (Butler & Turner, 2010; Moscardelli & Wood, 2008; Posamentier & Martinse, 2011; Scarselli, 2020 and references therein). Gravity-driven linked systems (also known as megaslides) are commonly observed on regional seismic profiles forming margin-scale collapse systems that can involve sedimentary units several kilometres thick (Butler & Turner, 2010; Rowan, 2020; Scarselli et al., 2016). These large systems display a proximal zone of extension characterised by listric faults that are connected through a basal detachment to a down-dip treatment.
contractional domain, where a fold and thrust belt occurs (Ahmed et al., 2022; Cai et al., 2020; Cobbold et al., 2004; Corredor et al., 2005; Damuth, 1994; de Vera et al., 2010; Lawrence et al., 2002; Mangano et al., 2020; Rowan, 2020; Scarselli et al., 2016; Souza et al., 2020; Trincardi & Argnani, 1990; Zhang et al., 2021). Megaslides contrast with failures displacing near-seabed strata and forming comparably thin, highly deformed products, often referred to in the literature as mass transport complexes (MTCs), slump complexes or submarine landslides (Frey-Martínez et al., 2005; Moscardelli et al., 2006; Moscardelli & Wood, 2008; Posamentier & Martinsen, 2011; Scarselli, 2020; Scarselli et al., 2020). Geophysical and field investigations have shown that these systems can develop an intricate internal architecture as material is transported from an updip zone of evacuation to a down-dip zone of accumulation (Bull et al., 2009; Scarselli, 2020 and references therein).

The rising interest in the products of gravity-driven processes as targets for resource exploration has focused attention on the study of these systems at every scale, from the giant anticlinal structures within the contractional domain of megaslides in the Niger Delta (Corredor et al., 2005) to rotated blocks at the head-scarp of submarine slumps in the Statfjord Field, North Sea (Welbon et al., 2007). Also, gravity-driven sediment deformation forms a significant geohazard for offshore platforms and pipelines (Frey-Martínez et al., 2011; Lu & Shipp, 2011) and is also able to trigger tsunamis that can impact coastal infrastructure and communities (Løvholt et al., 2015; Masson et al., 2006; Urgeles et al., 2019).

A wealth of literature provides analysis of slope failure events based on field (e.g., Alsop et al., 2017, 2020; Alves & Lourenço, 2010; Butler & McCaffrey, 2010; Farrell & Eaton, 1987; Martinsen, 1989; Ogata et al., 2012; Poprawski et al., 2021; Strachan, 2002; Woodcock, 1979) and geophysical data (e.g., Alves, 2015; Bull et al., 2009; Frey-Martínez et al., 2005; Gee et al., 2006; Jackson, 2011; Moscardelli et al., 2006; Nugraha et al., 2020; Sawyer et al., 2009; Smit & Stemmerik, 2022; Wu et al., 2021). Comparably, little research has been carried out at a regional scale to understand how lateral variability of margin characteristics can affect gravity-driven processes (e.g., Moscardelli & Wood, 2016; Naranjo-Vesga et al., 2020; Völker et al., 2012). Previous regional works have used sidescan sonar and multibeam bathymetry to build catalogues of slope failures over large extents of the US continental slope (McAdoo et al., 2000; Twichell et al., 2009) as well as the slopes surrounding the North Atlantic (Huhnerbach & Masson, 2004) and eastern New Zealand (Watson et al., 2020). Discussion of factors controlling the emplacement of these failures was restricted to understanding the recent, near-seabed geology as observed from bathymetric data.

In this research, we analysed an extensive set of seismic profiles that penetrate the Cretaceous and Cenozoic section offshore Namibia, revealing the products of gravity-driven processes along ca. 500 km of the shelf-slope segment of the margin (Figure 1). The results of this research provide a comprehensive catalogue of the location and morphological characteristics of failures on the Namibia margin. General trends in the distribution, size and style of failures are discussed in terms of potential controlling factors. These results may have wide applicability for improving understanding of how gravity-driven deformation is linked to the wider tectono-stratigraphic evolution of passive margins.

**Highlights**

- Margin-scale evaluation of gravity-driven processes of sediment deformation and remobilisation.
- A semiquantitative seismic workflow for the analysis of slope failure products is implemented.
- Regional geologic controls on distribution and style of gravity-driven deformation are proposed.
- Results suggest overpressures, sediment types, pre-existing slope relief and mode of failure as key controls.
- Triggering and preconditioning factors to vary with the evolution of the margin through geological time.

**2 | REGIONAL SETTING**

The Atlantic margin of Namibia and South Africa is a typical rifted passive margin that formed from stretching of Gondwana since the Late Jurassic and subsequent opening of the South Atlantic Ocean between the Nubian and the South American plates (Bauer et al., 2000; Chauvet et al., 2021; Gladczenko et al., 1998; Heine et al., 2013; Koopmann et al., 2014; McDermott et al., 2015; Pérez-Díaz & Eagles, 2014). The margin extends for ca. 1800 km across the coasts of South Africa and Namibia and is defined by two prominent morphotectonic features, the Walvis Ridge to the north and the Aguilhas-Falkland Fracture Zone to the south (Figure 1). The margin hosts two main basins, the Orange and Walvis basins. These basins contain continental and marine sediments up to 8 km thick that record the evolution of the margin from rift to oceanic spreading (Austin & Uchupi, 1982; Bauer et al., 2000; Brown, 1995;
The tectono-stratigraphic evolution of the margin has been previously studied by the analysis of subsurface data consisting of deep reflection profiles and well data mainly acquired during exploration of the Orange Basin (Figure 2; Austin & Uchupi, 1982; Bauer et al., 2000; Brown, 1995; Chauvet et al., 2021; Emery et al., 1975; Gerrard & Smith, 1982; Gladczenko et al., 1998; Koopmann et al., 2014; Light et al., 1993; McDermott et al., 2015; Mohammed et al., 2017; Paton et al., 2008; Planert et al., 2017). These studies have interpreted a series of seismic surfaces that define the stratigraphic packages deposited during key basin-forming phases (Table 1).

2.1 Rift to early post rift: Late Jurassic to Aptian

The earliest evidence of crustal stretching is provided by marginal rift basins and seaward dipping reflectors (SDRs) that form Sequence I (Figures 2 and 3). The marginal rift basins are SE-NW trending half grabens up to 60 km across located in the inner part of the margin that contains up to 2 km thick Upper Jurassic–Lower Cretaceous continental strata (Figure 3; Broad et al., 2006; Broad & Mills, 1993; Coward et al., 1999; Gerrard & Smith, 1982; Jungslager, 1999; Light et al., 1993; McMillan, 2003; Mohammed et al., 2017). The Late Jurassic age limit of these sediments provides constraints for the onset of rifting in the Orange Basin (Figure 2; Gerrard & Smith, 1982; Jungslager, 1999; Koopmann et al., 2014; Light et al., 1993; McMillan, 2003; Séranne & Anka, 2005). Lithospheric stretching and initial break-up coincided with voluminous volcanic activity as testified by the ubiquitous SDRs observed along the margin (Figure 3; Bauer et al., 2000; Chauvet et al., 2021; Elliott et al., 2009; Gladczenko et al., 1998; Intawong et al., 2019; Jackson et al., 2000; Mohammed et al., 2017; Paton et al., 2017). SDRs are wedges of volcaniclastic strata representing the extrusive counterpart of igneous rocks accreted during the late rift stage, possibly associated with the rise of the Tristan de Cunha mantle plume (Bauer et al., 2000; Chauvet et al., 2021; Elliott et al., 2009; Hirsch et al., 2010; Paton et al., 2017).

Along the margin, the transition to a passive margin has been generally associated with the Barremian “break-up” erosional unconformity (Reflector 1 in Figures 2 and 3) that truncates both the SDRs and the infill of the marginal rift basins (Figures 2 and 3; Brown, 1995; Gerrard & Smith, 1982; Jungslager, 1999; Light et al., 1993; Séranne & Anka, 2005).

2.2 Post-rift: Aptian to Maastrichtian

Cooling of the newly accreted crust caused the onset of subsidence and the deposition of shallow marine strata forming the early post-rift, Barremian-Aptian, Sequence II (Figure 2; Baby et al., 2018; Gerrard & Smith, 1982; Intawong et al., 2019; Séranne & Anka, 2005). Continued thermal subsidence brought about the deposition of a fully
marine post-rift Sequence III (Gerrard & Smith, 1982; Light et al., 1993; McMillan, 2003; Séranne & Anka, 2005). In the Orange Basin, the unit contains prograding clinoforms infilling a 5 km thick depocenter, suggesting conspicuous sediment delivery from the Orange River to the margin since the Aptian (Figures 1a and 3; Baby et al., 2018; Brown, 1995; Hirsch et al., 2010; Light et al., 1993; Mohammed et al., 2017). An additional depocenter developed from the early Late Cretaceous 300 km north of the Present-Day Orange River, suggesting an important sediment input was formed at that time (Baby et al., 2018; Mohammed et al., 2017).

Truncations of the uppermost strata of Sequence III below the Base Cenozoic unconformity marked by Reflector 3 have been widely recognized along the Present-Day shelf of the basin margin (Figures 2 and 3). These truncations together with peaks in denudation rates between 80 and 60 Ma reported by apatite

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**FIGURE 2** Tectono-stratigraphic chart for the Orange Basin from Scarselli et al. (2016). The chart delineates the key tectonic events and sedimentary units in the study area.

**TABLE 1** Table showing the key regional seismic surfaces interpreted in this work and their nomenclature relative to previous seismic studies on the Namibian margin

| Reflector | This Study | Gerard and Smith (1982) | Bauer et al. (2000) | Brown (1995) | Light et al. (1993) | Austin and Uchupi (1982) | Gladczenko et al. (1998) |
|-----------|------------|------------------------|-------------------|--------------|-------------------|------------------------|------------------------|
| Reflector 3 | L | L | 22At1 | L | D | D | Base Cenozoic ca. 66 Ma |
| Reflector 2a | N | N | 15At1 (?) | N | – | – | Turonian ca. 90 Ma |
| Reflector 2 | P | – | 13At1 | P | All (?) | – | Aptian ca. 115 Ma |
| Reflector 1 | R | B (?) | 6At1 | Q | – | TB (?) | Early Barremian ca. 129 Ma |
| Reflector 0 | T | – | T | T | – | RUC (?) | Upper Jurassic ca. 150 Ma |
fission track studies (Gallagher & Brown, 1999; Kounov et al., 2009; Tinker et al., 2008a, 2008b), indicate a long-lasting phase of inner margin uplift at the end of the Cretaceous (Figure 2; Paton et al., 2008; Baby et al., 2018).

2.3 | Cenozoic canyons and contourites

Open marine sedimentation continued in the early Cenozoic (Gerrard & Smith, 1982; Light et al., 1993). Basin morphology at that time was controlled by the development of a series of canyons up to 20 km wide and up to 150 km long that incised deeply into the underlying Upper Cretaceous strata (Figure 4a; Bagguley & Prosser, 1999).

Since the Paleocene, the margin has experienced the effect of strong bottom currents, mainly related to the Antarctic Bottom Water current (Bagguley & Prosser, 1999; Scarselli, 2014; Ward et al., 1983). The action of margin-parallel contour currents is displayed by the occurrence of prominent erosive channel complexes on dip-oriented seismic profiles as well as mounded drifts up to 25 km across in the slope strata of the Cenozoic succession (Figure 4b; Bagguley & Prosser, 1999). The abundance of these features indicates that the Cenozoic basinal sedimentation in the margin was largely affected by the onset of bottom currents (Figure 2). The rise of the Benguela current since the Neogene (e.g., Dupont et al., 2005; Stone, 2013; Uenzelmann-Neben et al., 2007) may also be responsible for the widespread occurrence of contourite drifts and associated erosional features observed in the slope strata of Sequence IV.

A recent phase of inner margin uplift is indicated by clear erosion at the seabed of the Base Cenozoic unconformity marked by Reflector 3 and of the overlying Cenozoic strata of Sequence IV (Figures 2 and 3). This phase of uplift during the late stage evolution of the margin may be related to the short-lived phase of uplift between 16 and 14 Ma reported by Hirsch et al. (2010).

3 | METHODS

The research was based on the analysis of a 2D reflection seismic dataset that comprises a total of 220 time-migrated seismic lines part of seven separate surveys (see Appendix S1). The lines cover an area of 119,000 km² between latitudes 24.5°S and 30.5°S offshore Namibia.
The area encompasses the present-day shelf to the outer slope of the Namibia margin, between water depths of 50 m and 3000 m (Figure 1). Spacing between seismic lines is highly variable. In the southern part of the study area, the overlap of a number of surveys provides a line spacing of ca. 4 km (Figure 1b). In the central and northern portions of the area, line spacing ranges from 4 to 17 km. The acquisition parameters along with the estimated seismic resolution for the 2D surveys are summarised in Appendix S1.

Quantitative characterisation of the gravity-driven systems (i.e., the products of gravity-driven processes) was achieved through a series of measurements derived from the analysis of depth-converted seismic data and from gridded depth surfaces. Depth conversion and the gridding workflow are presented in Appendices S2 and S3, respectively.

Gravity-driven systems have been assessed by adapting the methods of McAdoo et al. (2000) and Clare et al. (2019). The systems were characterised according to their temporal and geographical location, general morphometry and style of emplacement.

3.1 Location and age

Mapping of the basal surface was used to assess the location and shape of gravity-driven systems (Figure 5). Systems were located by measuring the X and Y coordinates of the headscarp. To maintain consistency, the measurements were taken at the most external point of the headscarp of each system as shown in Figure 5.

The age of gravity-driven systems has been broadly assigned to that of the sequences in which they were found. Sequences are those defined in Figure 2.

3.2 Position relative to the framework of the margin

Gravity-driven systems were qualitatively and quantitatively characterised with respect to their position relative to the morphological elements of the Namibia margin.

Three main qualitative geographical categories were defined (Figure 6):

- **Shelf**: gravity-driven systems found landward of the coeval shelf-break (Figure 6a).
- **Slope**: gravity-driven systems found seaward of the coeval shelf-break (Figure 6b).
- **Slope within canyons**: gravity-driven system found in the slope, within a canyon-filling sequence (Figure 6c,d).

From a quantitative stand point, three measurements were taken (1) headscarp to shelf-break distance; (2) thickness of the sequence containing a given gravity-driven system and (3) dip of the pre-failure section.

The headscarp to shelf-break distance measures the horizontal distance between the headscarp of a fossil gravity-driven system and the coeval shelf-break (Figure 7). This measurement allows the understanding of the down-dip distribution of gravity-driven systems across the margin.

The thickness of sequence containing a given gravity-driven system measures the mean of the isopach values of the sequence calculated at the headscarp location. To allow comparison across different systems, this measure was then expressed as percentage of the maximum thickness of the sequence hosting the given failure. For example, given a sequence with maximum thickness of 100 m, a measured thickness of 10 m would result in a value of 10%. This measurement allows understanding of the...
distribution of gravity-driven systems with respect to the location of sedimentary depocenters along the margin.

The dip of the pre-failure section is the mean of the dip values of the surface of the first continuous and laterally traceable reflection underneath a given gravity-driven system (Figure 7). The dip values recorded are Present-Day dips that may not represent original depositional dips due to later subsidence. Backstripping to pre-failure dips was not attempted as it was assumed that the effect of subsidence did not vary substantially across the margin. Therefore, the analysis of relative dip values from different gravity-driven systems retains geological significance.

These last two measurements were made by computing values of thickness and dip included in a polygon that covered the headscarp of the gravity-driven system under investigation (Figure 5). This is because the headscarp is considered the location where gravity-driven deformation initiates.

### 3.3 General morphometry

The following measurements were recorded to quantitatively characterise the morphology of gravity-driven systems: (1) runout, width and their ratio, (2) thickness, area and volume, (3) average dip of basal surface and (4) headscarp height. Runout, width, area and volume are considered minimum values as a number of gravity-driven systems were not entirely imaged by the available seismic coverage.

The runout and the width of a given gravity-driven system were calculated once the planimetric morphology was known from the interpretation of the basal surface (Figure 5). The runout was measured as the length of the segment that links the extremities of a gravity-driven system, from the headscarp to the toe (Figure 5). As gravity-driven systems often exhibit irregular shapes, their width was measured along the longest segment that links the opposite lateral flanks of a system (Figure 5).

The ratio between runout and width was calculated and regarded as a shape index. Large values of the shape index indicate long and narrow systems (high aspect ratio). Values around one mean an equant shape, and values smaller than one denote wide and short systems.

Thicknesses were calculated as the mean of the isopach values of a given gravity-driven system defined by the basal and top surfaces. The area was calculated as areal extent of the basal surface. Volumes were computed from the basal and top surfaces bounding a given gravity-driven system. Thickness, area and volume were calculated using standard processing algorithms within the seismic interpretation tool utilised in the research.

The dip of the basal surface was computed as the mean of the dip values of this surface. For similar reasons to those presented above, backstripping of this surface was not attempted and the dip values are Present-Day values.

Headscarp height was measured on a vertical seismic section that best imaged the headscarp of a given system. The headscarp height was measured as the length of a segment that perpendicularly joins the upper tip of the headscarp and the reflection that marks the base of a gravity-driven system (Figure 7).

### 3.4 Styles

Styles of the gravity-driven systems were qualitatively based on (1) typology, (2) degree of vertical and lateral complexity, and (3) morphology of headscarp area. Systems were divided into two types based on diagnostic internal seismic characters (Figure 8a; Moscardelli et al., 2006; Posamentier & Martinsen, 2011; Scarselli, 2020). Slumps are seismically imaged with a typical contorted to chaotic facies, and slides forms seismic bodies of coherent stratigraphy internally displaced by distinct shear surfaces (Figure 8a).

Gravity-driven systems resulting from multiple failures were identified as complex. Diagnostic features of complex gravity-driven systems include the presence of multiple adjacent headscarp salients (Figure 5) as well as failures exhibiting evidence of multiple internal basal surfaces that cannot be linked to distinct headscarps (Figure 8a).

The morphology of the headscarp area was characterised as filled or evacuated (Figure 8b). This attribute was relevant for several systems found in the study area. Filled headscarps show the head region occupied by failed material. Systems with evacuated headscarps show a draping section that fills the section between the headscarp scar and the failed sediments further down dip. The horizontal distance between the headscarp scar and the failed material was measured to get insight into the mobility of the gravity-driven systems.
RESULTS

A total of 39 gravity-driven systems were identified in the study area (Figures 9a and 10a). The systems have been found in the Upper Cretaceous Sequence III and in the Cenozoic Sequence IV (Figure 2) and affected ca. 60,000 km$^3$ of post-rift strata (Figure 9b). This is approximately one-third of the volume of the entire post-rift succession in the study area.

Slumping was the most common type of failure observed in the post-rift strata and it was the dominant sedimentary deformation process in the Cenozoic in terms of number of failures and volume displaced (Figure 9a,c). Although slides are substantially less numerous than the slumps, they displaced a greater amount of sediment during the Cretaceous (Figure 9a,c). As shown in the following sections, this is due to the fact that slides are able to affect sections of strata up to 2 km thick.

4.1 Distribution

Figure 10 shows the distribution of the gravity-driven systems mapped in this research. Overall, they are evenly distributed along the whole segment of the margin under examination. Virtually all systems occurred in slope settings, and very few (ca. 9% of the total) on the shelves (Figures 10 and 11a). The Cenozoic gravity-driven systems
have been developed along shelves, slopes and in clusters within Cenozoic canyons (Figures 4a, 10b, and 11a). The Late Cretaceous systems occurred almost entirely on open slopes (Figure 11a).

In general, slides were found at distances from the shelf-break ranging from a few kilometres up to 50 km (Figure 11b). A cluster of slides was found at or immediately down dip of the shelf-break (Figure 11b). These slides are here named ‘shelf-break block slides’ for their particular distribution and are presented in detail in the following paragraphs. Slumps instead were commonly seen at distances of ca. 50 km up to ca. 100 km from the shelf-break (Figure 11b). When taken as a whole, there is a frequency peak in the total number of slumps and slides at ca. 30 km from the shelf-break (Figure 11c). Strikingly, slumps located at increasing distance from the shelf-break exhibit a decrease in thickness (Figure 11c). Slides seem to follow an opposite trend, with the thinner systems found closer to the shelf-breaks (Figure 11c).

Critically, slides are preferentially found where the host section is the thickest (Figure 12a). Most of the slides, in fact, were constrained within the two main sediment depocenters that developed during the Late Cretaceous (Figure 10c). In contrast, the occurrence of slumps seems unrelated to the areal variation of thickness of their host sequence (Figure 12a).

As a whole, gravity-driven systems occur above a substratum with slopes ranging from 0.5° up to 9°, with a peak in frequency at slopes of ca. 2° (Figure 12b). There is a weak negative correlation between the volume of gravity-driven systems and dip of the pre-failure section (Figure 12c), meaning that large failures tend to occur along gentler slope dips. Overall, there is a positive correlation between the dip of the intact substratum and the dip of the basal surface of gravity-driven systems, but strikingly slumps consistently have gentler basal surfaces than the undeformed slopes above which they form (Figure 12d).
FIGURE 9  Histograms showing the impact of gravity-driven deformation along the Namibian margin in terms of occurrence of failures and volume of sediments that these have displaced according to the stratigraphic sequences in which they are found. The sequences are described in Figure 2. (a) Occurrence of slides and slumps is described by their total number. (b) Volume of sediments involved in gravity-driven systems (i.e., slides and slumps) and total volume of the host sequences. (c) Relative percentages of volumes of sediments involved by slides and slumps. See text for full discussion. ‘Post rift’ refers to the total of Sequence III and IV.

FIGURE 10  Maps of the distribution of gravity-driven systems in the study area. (a) Distribution of all systems mapped in this research. Systems are colour coded according to the host sequence where these were found. (b) Distribution of the Cenozoic systems overlain on the depth structure map of the Base Cenozoic Reflector 3. (c) Distribution of Late Cretaceous systems overlain on the isopach map of the Upper Cretaceous Sequence III. See text for full discussion.
FIGURE 11 Charts detailing the occurrence of gravity-driven systems within the geomorphic framework of the Namibian margin. (a) Occurrence (in percentage) of systems divided by the environments in which they have been found. ‘Post rift’ refers to the total of Sequence III and IV. (b) Frequency of failures at increasing distance of headscarp to shelf-break. (c) Semi-log plot of distance of the headscarp to shelf-break versus thickness of a system. Trend lines are power-law fit. $R^2$ values are 0.49 and 0.03 for slides and slumps, respectively. See text for full discussion of the charts.

FIGURE 12 Charts detailing key morphometries of gravity-driven systems in relation to local geological characteristics of the margin. (a) Histogram showing the ratio (in percentage) between thickness of the host sequence at the headscarp location and the maximum thickness of the host sequence. Where this ratio is highest is an indication of systems that are preferentially located in sediment depocenters. (b) Frequency of failures at increasing dips of the substratum. (c) Semi-log plot of volume of gravity-driven systems against dip of the pre-failure section. Trend lines are power-law fit. $R^2$ values are 0.53 and 0.03 for slides and slumps, respectively. (d) Log-log plot of dip of the detachment of gravity-driven systems versus the dip of the intact substratum. Trend lines are power-law fit. $R^2$ values are 0.83 and 0.19 for slides and slumps, respectively. $R^2$ value for slumps and slides considered together is 0.49. See text for full discussion of the charts.
4.2 | General morphometry of slides and slumps

The main morphometric parameters for slides and slumps mapped in the study area are reported in Figure 13 with a series of frequency histograms. Typical dimensions of these parameters are given as median values.

The range of values for runout, area and volume can be vast, with few extreme values up to two orders of magnitude greater than typical values (e.g., Figure 13a,e,f). This is due to the fact that a few systems are complex being composed of amalgamated failures (e.g., Figure 8a). The occurrence of complex events is marked with “c” in the frequency charts shown in Figure 13.

Slumps have runout distances from a few kilometres up to 120 km, with typical distances of ca. 50 km (Figure 13a). Single events with runouts in excess of 100 km are slumps that originated in the slope and plunged into Cenozoic canyons (Figure 10b). Slides have shorter runouts, with typical values of ca. 35 km (Figure 13a). Slides have shape index values (runout over width) ranging from 0.2 to 3 with typical values of ca. 1 (Figure 13g). Slides with shape index <1 are the shelf-break block slides which are wide features (ca. 60 km wide) with short runouts (ca. 10 km).

Shape index values for slumps are higher and with a wider range that goes from ca. 1 to ca. 5 (Figure 13g). Slumps and slides with thickness of 300 m, and can be up to ca. 2 km thick (Figure 13c). Slumps and slides with thickness <20 m are likely to be present on the margin, but would not be visible as they are below seismic resolution.

The frequency histogram of headscarp height shows a similar pattern, with slumps having typical headscarp heights of ca. 100 m whereas slides have taller headscarps, commonly in excess of 300 m (Figure 13c).

The typical dip of detachments of slumps is ca. 1°, whereas slides commonly reach dip values in excess of ca. 2° and up to 9° (Figure 13d). These uncommon, steep basal surfaces are distinctive of shelf-break block slides.

Slumps vary greatly in areal extent from 50 to 4000 km² and can commonly cover areas in excess of 1000 km² (Figure 13e). Complex events can be up to an order of magnitude more extensive than the largest single event (Figure 13e). Slides are less extensive and typically cover areas of ca. 800 km² (Figure 13e).

The volume of sediments that slumps contain ranges between 10 and 500 km³ (Figure 13f). Slides show a wider range in size that includes relatively small failures with volumes of ca. 50 km³ and larger features with volumes up ca. 5000 km³ (Figure 13f). These larger slides, here indicated as megaslides, are up to an order of magnitude larger than the largest slumps (Figure 13f). Megaslide complexes are able to displace volume of sediments in excess of 50,000 km³ (Figure 13f).

4.3 | Styles of deformation

Megaslides and shelf-break block slides are the two main types of slides observed along the margin (Figure 14). Shelf-break block slides show an up-dip extensional domain characterised by sets of listric normal faults that displace the shelf-edge into distinct rotated blocks (Figure 14a). Blocks are commonly up to 2 km across, and overall these failures extend laterally for up to ca. 60 km and have short runouts, ranging from 5 to 10 km. Late Cretaceous and Cenozoic shelf-break block slides have been mapped in the northern end of the study area (Figures 10a and 14a). Note that these systems lack a contractional domain as observed in megaslides.

Megaslides form extensional-contractional linked systems that extended for several tens of kilometres along the margin and can displace thousands of cubic kilometres of strata (Figure 14b,c). Megaslides observed within the large Cretaceous depocenter to the south are the largest, affecting the whole shelf/slope segment of the margin (Figure 14c). Generally, these systems lack the undeformed translational domain observed in other megaslides along the margin (Figure 14b), but instead show an overlap of the extensional and contractional domains, with cannibalization of older thrust ramps by younger extensional structures (Figure 14c).

The observations of headscarp morphologies indicate that slumps exhibit filled and evacuated styles in similar proportion and virtually all slides show filled headscarps (Figure 15a,b). Evacuated slumps have a distance from headscarp to failed materials that range from ca. 2 km up to 18 km with typical values of ca. 6 km (Figure 15b). Interestingly, evacuated systems don’t exhibit longer runout distances (Figure 15b) and show a very tenuous positive correlation between runouts and distance between headscarp and failed material (Figure 15c).

5 | Discussion—Control of margin architecture on slope processes and their distribution

The analysis has highlighted trends regarding the distribution, morphology and style of gravity-driven systems. The factors that might control such trends are as follows.
One of the most striking trends that emerges is that slides are commonly found in the proximal part of the margin and constrained within the margin depocenters, whereas slumps seem to preferentially occur in the more distal segments of the slope with examples of slumps observed at distances in excess of 100 km from the shelf-break (Figures 11b and 12a). This may indicate that distinct geological processes acting in these different segments of the margin promote different types of gravity-driven deformation.

Sedimentation rates are commonly highest within sediment depocenters. Decay of sediment transport capacity with increasing distance from the shelf edge is responsible for the highest sedimentation rates at the shelf and upper slope and confers typical sigmoidal profiles to offshore margins (Adams et al., 1998; Adams & Schlager, 2000; Brothers et al., 2013; Schlager & Adams, 2001).

### 5.1 Overpressures and sediment strength

Figure 13 Frequency charts illustrating the variability of the main morphometric parameters of slides and slumps. (a) runout; (b) thickness; (c) headscarp hight; (d) dip of basal surface; (e) area; (f) volume; (g) shape index, defined as the ratio of runout over width. Extreme values are marked with “c” if failures are complex. (h) diagram reporting shape index values of gravity-driven systems grouped according to the environment in which they were found. See text for detailed discussion.
Sedimentation rates decrease by up to an order of magnitude from the inner slope towards the lower slope (e.g., Ruddiman & Sarnthein, 1989). The sediments deposited in margin depocenters are mainly a mixture of sands, silt and muds discharged from rivers and transported across the shelf by surface currents (e.g., Nittrouer et al., 2009; Stow & Mayall, 2000) with the coarse-grained component carried by gravity flows that are able to transport large amounts of fluvial-derived sediments across the margins (e.g., Mulder et al., 2003; Mulder & Syvitski, 1995; Nittroeur et al., 2009; Stow & Mayall, 2000). Fine-grained, muddy material in suspension from turbid layers (nepheloid layer) brings sediment to the distal slopes (e.g., Nittroeur et al., 2009; Stow & Mayall, 2000).

**FIGURE 14** Representative seismic sections showing the key types of slides. (a) Shelf-break block slide. (b) Megaslide with typical tripartition into extensional, translational and contractional domains. (c) Example of a margin-scale megaslide affecting the shelf and slope components of the margin. These large systems are mainly restricted to the central areas of the margin’s depocenters. Section locations are shown in Figure 10c.

**FIGURE 15** Charts detailing the characteristics of headwall styles of the gravity-driven systems. (a) Frequency of distance between the headscarp and the failed sediment mass for systems with evacuated scarps (see example of evacuated systems in Figure 8). (b) Frequency of runout distances for evacuated and non-evacuated systems. (c) Plot of runout distance versus distance between headscarp and failed sediment mass for evacuated systems. See text for discussion.
margin setting is applicable to the Orange Basin, which has been largely dominated by siliciclastic input from the Orange River since the Albian (Brown et al., 1995; Garzanti et al., 2014; Gerrard & Smith, 1982; Kirkpatrick et al., 2019; Light et al., 1993; Paton et al., 2008). In this framework, high sedimentation rates and burial on the upper slopes of basin depocenters are likely to create distinct, weak layers at depth due to disequilibrium compaction and fluid expansion overpressures (Figure 16; Osborne & Swarbrick, 1997; Swarbrick & Osborne, 1998; Tingay et al., 2009). The common occurrence of fluid escape pipes and associated pockmarks in the Orange Basin are regarded as direct evidence for widespread overpressures in the inner part of the margin (Boyd et al., 2011; Hartwig et al., 2012; Kuhlmann et al., 2010; Moss & Cartwright, 2010). It is proposed that disequilibrium compaction of Uppermost Cretaceous units and fluid expansion related to maturation of Cretaceous source rocks are the likely mechanisms for the creation of overpressure in the Namibian margin (Hartwig et al., 2012).

External loading of the more coarse and likely high-strength sediments in this part of the margin will therefore result in sliding where failure focuses along these weak layers without exceeding the shear strength of the sediment above, producing coherent slides (Figure 16). By contrast, in the distal slope, fine-grained and weak sediments deposited at low sedimentation rates would not develop distinct, deep, overpressured weak layers. Therefore, when critically loaded by external factors, the shear strength of the sedimentary section is overcome throughout resulting in fully disruptive slump failures (Figure 16).

Studies have documented an inverse correlation between sedimentation rate and strength of sediments as slow rates of deposition may allow sediment strength to build up relatively soon after burial commences (Chassefière et al., 1985; Keller, 1974; Moore, 1961; Morelock, 1969). This might imply that under the same loading conditions the thickness of the unconsolidated and weak sediments prone to failure decreases towards the distal part of the slope where sedimentation rates are expected to be low. This may explain the trend observed in this research that thinner slumps are found in the more distal segments of the slope (Figure 11c).

5.2 | Control of failure processes on morphometry and style of products

The contrasting nature of disruptive slumping and coherent sliding has a direct control on the morphometry and style of the products of these two processes. This is supported by the fact that slumps are areally more extensive and can attain greater runouts than slides (Figure 13) due to pervasive internal disruption and spreading of the failing mass. The striking fact that slump thickness is 50% of their headscarp height (Figure 13) may be an additional indication of the dispersive nature of slumping. This could also be the expression of substantial volume loss due, for example, to release of interstitial fluids during or after failure. Fluid escape structures are commonly observed in exposed examples of submarine failures (e.g., Martens, 1989; Ogata et al., 2012; Strachan, 2002). Disruption and fluid release might be intimately linked as fluid mobilization would likely minimise internal friction by lubrication, promoting further disaggregation of the failing mass. Direct measurement of shear strength within slumps has indicated that these are stronger than the encasing sediments likely due to shearing of the failing mass and the resulting loss of water during emplacement (Sawyer et al., 2009).

Full evacuation of headscarsps is a distinctive and common trait of slumps (Figure 15) and suggests a marked mobility of slumps with respect to slides. In evacuated slumps, distances from slump masses to their relative headscarsps are typically in the order of ca. 5 km with extreme cases reaching distances of ca. 20 km (Figure 15c). Counter-intuitively evacuated slumps do not show extreme runouts, and higher degrees of evacuation do not correspond to longer runouts (Figure 15b,c). This indicates that the degree of headscarp evacuation should only be used as a proxy for the overall mobility of slumps with caution, and that other geological constraints, such as lateral confinement (see below) and possibly sediment strength, are the main controlling factors.

Margin uplift and sediment loading of the shelf are both key factors initiating megaslides along a margin through processes of sliding and spreading (Rowan et al., 2004). The variability of products related to sliding observed in the Namibia margin is remarkable (Figure 14). It is proposed that different styles of sliding can be related to progressive stages of sediment input and margin uplift which form early shelf-break block slides evolving subsequently into megaslides as internal deformation increases (Figure 17). In this view, initial margin tilting and shelf loading would drive limited extension which is entirely balanced by down-dip lateral compaction and layer parallel shortening with no contractional domain formed at this stage (Figure 17a). The occurrence of such an initial stage in the evolution of megaslides would explain the paradox of extension exceeding contraction commonly observed in these systems (Butler & Paton, 2010; de Vera et al., 2010). In an evolving margin, shelf-break block slides would be preserved at the periphery of a basin or in minor basins where sediment input
is limited. This is what is observed in this research, with shelf-break block slides mainly mapped to the north of the study area where no major sediment depocenters are observed (Figure 10). This is also what is observed in the Amazonas Megaslide Complex where numerous basinward dipping listric extensional faults occur at the periphery of the Amazon Fan, whereas a series of laterally segmented megaslides affect the inner part of the system (Figure 18a; Reis et al., 2010).

Continued margin uplift together with increasing sediment delivery and progradation across the shelf-break further drive extension. After significant lateral volume loss, further up-dip extension is balanced by down-slope contraction with the nucleation and growth of thrust faults and the formation of full slides characterised by distinct extensional, translational and contractional domains (Figure 17b).

Sustained margin uplift together with high rates of sediment delivery and progradation drives gravity-driven deformation further down slope. Activation of extensional faults at the headwall further down dip leads to progressive shortening of the undeformed translational domain. Extension superimposed on contraction and cannibalization of older thrust systems is likely to occur when expanded megaslide headwalls overlap early structures in the contractional domain. These structurally complex megaslides are expected to occur in the axial part of basins where sediment delivery and loading are expected to be the highest. In this study, these ‘mature’ megaslide systems are in fact observed in the large southern Cretaceous depocenter (Figure 10). A similar evolutionary pattern is observed in the Amazonas Megaslide Complex where towards the central part of the Amazon Fan widening of the contractional and extensional domains (Figure 18a) and
Cannibalization of early thrusts are observed (Figure 18b). In the Namibian margin, the area where the southern depocenter occurs is also where the margin seems to have experienced the highest late Cretaceous uplift as shown by extensive erosion of the Base Cenozoic surface (Figure 10b). Vigorous uplift would have further enhanced down dip migration of gravity-driven deformation, further explaining why such structurally complex megaslides are commonly observed in that part of the basin.

5.3 | Control of pre-existing slope relief

Clusters of slumps observed within canyons and their draping sequences form a considerable number of the failures mapped along the Namibian margin (Figure 11a), clearly indicating that these negative relief features provide control on the distribution of gravity-driven deformation.

Plunging of slumps into canyons has been noted in a number of cases (Figures 6 and 10b) indicating these areas acted as sinks into which failures were funnelled. Although less numerous, slumps have been mapped entirely within canyons. The limited spatial resolution of 2D data may have hindered a wider recognition of these failures. Accurate mapping of canyon floors with 3D seismic data and high-resolution bathymetry reveals that these failures are common (e.g., Frey-Martinez et al., 2011; McAdoo et al., 2000; Mountjoy et al., 2009; Paull et al., 2013; Puga-Bernabéu et al., 2011; Twichell et al., 2009). Destabilisation of these canyon-contained failures is related to canyon incision and flank undercut which induce slope steepening and collapse (Frey-Martinez et al., 2011; McAdoo et al., 2000; McArthur & McCaffrey, 2019; Mountjoy et al., 2009).

Typical characteristics of the slumps found in these lows are their high aspect ratios and long runouts, in some cases in excess of 100 km (Figure 13h). A canyon-confined pathway of collapse minimizes lateral spreading of a failure and hence the high aspect ratio of the related deposits, but it doesn’t account for the long runouts. A possible explanation is that confinement limits dissipation of inertial energy of failures allowing them to attain the longest runouts and high-aspect-ratios. Other studies have found that confinement due to folds and salt diapirs at the seafloor was crucial for failures developing over long distances (e.g., Moscardelli et al., 2006; Trincardi & Field, 1992; Trincardi & Normark, 1989). By contrast, Twichell et al. (2009) have reported that inter-canyon and canyon-confined failures in the U.S. Atlantic margin have similar runouts suggesting that, aside from failure confinement, the physical properties of the sediments in the source area might play an equally important role in the runout distance of slope collapses.

5.4 | Dip of substratum

Measurements of the dip of the intact substratum and the dip of the basal surface of the overlying slope failures
show a direct positive correlation (Figure 12d). This suggests that failures develop above basal surfaces that are generally conformable with the dip of the strata forming the substratum. This might explain why slides, which are typically located in the inner part of the margin where strata are generally steeply dipping with respect to flatter basinial strata, exhibit steep basal surfaces (Figure 13d). Slumps follow this general trend but are commonly characterised by basal surfaces shallower than the underlying substrate (Figure 12d). This might be related to enhanced mobility of slumps which make them able to reach the distal, and flatter, segments of the margin. Shallow basal surfaces of slumps might be also related to a preferential frontal emergence of slumps along the Namibia margin. This is because, in frontally emergent slumps, the failing mass tends to ‘ramp up’ stratigraphy and spread above the seafloor (Frey-Martínez et al., 2005; Moernaut & De Batist, 2011), making the basal surface flatter overall.

While the dip of the substratum seems to play a role in controlling the dip of the basal surface of failures, steeply
dipping substrata do not seem to have a major impact on the occurrence of failures. This is shown by the fact that although failures have been observed across slope gradients between 0.5° and 9°, most initiated above substrata with dips ca. 2° (Figure 12b). Previous statistical analyses of slope failures along the North American (Booth et al., 1993; Mcdoo et al., 2000), Africa and northern Europe margins (Huhnerbach & Masson, 2004) have come to the same conclusion. Similar to those studies, this research also shows a tenuous negative correlation between volume of slope failures and dip of the intact substratum above which they occur (Figure 12c). This confirms that slope steepness is a poor indicator of susceptibility to slope collapse as increasing slope angles do not drive the size and occurrence of failures (Huhnerbach & Masson, 2004; Masson et al., 2006; Mcdoo et al., 2000).

5.5 Triggering and preconditioning factors of gravity-driven deformation

Considerations of the geological history of the Orange Basin together with the characteristics of gravity-driven deformation allow potential triggering and preconditioning factors that controlled the initiation of gravity-driven deformation to be highlighted.

Earthquakes are effective triggers of slope failure as they induce shear stress in the sediment column as well as shear strength degradation due to transitory excess pore pressure during seismic shaking (Biscontin et al., 2004; Lee et al., 2007). Earthquakes with magnitude up to 5 have been reported onshore and offshore Namibia (Figure 19, Mangololo & Hutchins, 2008) suggesting that seismic activity might be a key triggering factor of recent slope failure in the region. Seismic activity is widespread onshore, with epicentres mainly concentrated along the Great Escarpment, whereas few earthquakes with magnitude up to 5 have been registered offshore, where near-seafloor slumps were also mapped (Figure 19). Mangololo and Hutchins (2008) have suggested that ridge push from the Mid-Atlantic Ridge may provide a main driving mechanism for the numerous earthquakes observed in Namibia, but the mechanisms and regional stresses behind seismicity in Namibia are not well understood. Although the occurrence of seismic activity in the geological past is uncertain, earthquakes are plausible triggering factors for the deep water Cenozoic slumps observed in the Orange Basin.

Wave loading during tsunamis, storms and hurricanes (Canals et al., 2004; Lee et al., 2007; Masson et al., 2010), and overpressure generation due to dissociation of gas hydrates (Canals et al., 2004; Horozal et al., 2017; Li et al., 2016; Masson et al., 2010) are considered to be very effective triggering mechanisms of slope failure. Wave loading is able to induce shear stress at the seafloor in water depths up to 100 m and gas hydrates located at water depth in excess of >500 m are thought to be stable even under significant variations of pressure and temperature (Kvenvolden, 1993; Milkov et al., 2000; Reagan & Moridis, 2007; Xu & Germanovich, 2006). The range of water depths in which these two triggers are thought to operate suggests that they may not have played an important role in generating the failures analysed in this research (Table 1). This is because most of the failures in the Namibian margin are located seaward of the shelf-slope transition (Figure 11a), hence likely occurred at water depths in excess of 500 m.

Bottom currents are able to generate erosion at the seabed causing an increase in steepness of the slope and hence favouring failure (Sayago-Gil et al., 2010; Tournadour et al., 2015). This triggering factor cannot be discounted in the Namibian margin where contourites and associated channel complexes are ubiquitous in the Cenozoic slopes (Figure 4b). Also, research suggests that deposition from steady bottom currents imparts good sorting to contourite deposits, which implies higher water content in the sediments, driving their weakening and hence facilitating slope collapse (Laberg & Camerlenghi, 2008). Therefore, the occurrence of contourites may also represent an important preconditioning factor for the Cenozoic slope failures of the Namibian margin.

High sedimentation rate and hydrocarbon generation are effective mechanisms for the creation of pore fluid overpressure that can form discrete, weak sedimentary units (Cobbold et al., 2004, 2009; Rowan et al., 2004; Scarselli, 2020). These mechanisms were likely to be important for the collapse of the Late Cretaceous gravity-driven systems in the Namibian margin. In fact, these failures occurred at a time when two depocenters, likely actively fed by rivers, developed within the margin (Figure 10c). The Turonian marine shale at the base of the Cretaceous megaslides entered the oil window at ca. 85 Ma (Hirsch et al., 2010; Kuhlmann et al., 2011). This suggests that the increase in pore fluid pressure due to hydrocarbon generation created an efficient weak detachment that assisted the emplacement of the megaslides (Figure 16a).

Seepage of fluids through sediments can lead to their weakening and failure (Locat & Lee, 2002). Gee et al. (2006) reported that upward fluid migration and release along fault planes can concur during fault slip and initiate slope collapse in the overlying strata. This mechanism may be relevant for the initiation of slumps in the Orange basin as underlying megaslide structures, such as those in the extensional and contractual domains (Figure 14), may form important conduits for the upward migration of potential fluids sourced in deeper sections. This in turn may promote
weakening and collapse of the section overlaying megaslides. In the case of the Namibian megaslides, the source of fluids may have been in part from the overpressured Turonian marine shales that form the basal surface of the slides (Scarselli et al., 2016). Along the margin, a number of slumps have been found degrading the upper part of megaslides and fluid seepage through faults forming megaslides might have been a prime factor for the initiation of those failures (Mahlalela et al., 2021; Scarselli et al., 2016).

Slope steepening is considered a key trigger for slope collapse (Huhnerbach & Masson, 2004; Masson et al., 2010; van Weering et al., 1998). In the Namibia margin, three main phases of margin uplift might have assisted slope steepening and failure. A widespread and long-lived phase of inner margin uplift occurred in the Late Cretaceous between 80 and 60 Ma (Figure 2). Truncations below the Base Cenozoic unconformity of growth strata associated with the extensional domain of megaslides have been
commonly observed in the study area (Figure 14b,c). This provides evidence that the Late Cretaceous phase of margin uplift broadly coincided with the emplacement of the Cretaceous megaslides and it may therefore represent an important triggering factor for their emplacement.

A late phase of margin uplift occurred during the Cenozoic between 16 and 14 Ma (Figure 2). The short-lived nature of this uplift may have promoted the emplacement of a few coeval failures. This implies that the collapse of the more numerous slumps observed throughout the Cenozoic was likely controlled by other triggering and preconditioning factors discussed above, such as earthquakes and the action of bottom currents.

6 | CONCLUSIONS

The conclusions of this research are:

- Slides are seismically imaged as up to 2 km thick portions of coherent strata displaced by distinct faults commonly found less than 50 km from the shelf-break. They are typically constrained within the margin depocenters where sedimentation rates and burial are likely to be high compared to other locations in the study area.
- Slumps in contrast are characterised by a disrupted seismic facies and form thin failures (ca. 50 m) that commonly occur in the distal section of the slope, at distances from the shelf-break in excess of 50 km. Distal slumps have originated 100 km away from the shelf-break and potentially at water depths in excess of 2 km.
- High sedimentation and burial rates on the upper slope of the margin are likely to produce overpressures, augmented by hydrocarbon generation along discrete horizons at depth. Creation of these distinct and weak layers together with the likely high strength of coarse sediments in this proximal part of the margin promotes the formation of coherent slides in which deformation occurs along distinct shear surfaces.
- By contrast, in the distal slope, fine-grained and weak sediments deposited at low sedimentation rates are not likely to result in the development of distinct deep overpressured weak layers. Therefore, when critically loaded by external forces, the shear resistance of the sedimentary section is less than that of the basal surface and the section is mobilized throughout, resulting in a fully disruptive slump failure.

The contrasting nature of internal disintegration of slumping and coherent sliding has a direct control on the different morphometry and style of the products of these two processes.

- Slumps commonly have thicknesses which are 50% of their headscarp heights. They are typically characterised by evacuated headscars and can extend over large areas in excess of 1000 km².
- By contrast, slides have comparable average thickness and headscarp heights. Slides are less aerially extensive than slumps covering areas rarely in excess of 1000 km².
- Slope relief such as canyons and pre-existing landslide scars have a major role in controlling the distribution and evolution of slumps. Clusters of slumps in these high relief areas indicate that they act as sinks into which failures are funnelled. Also, confinement and hence limited dissipation of inertial energy, allowed these failures to attain the longest runouts and high aspect ratios.
- Steeply dipping substrata do not seem to have a major control on the occurrence and size of slope failures as most gravity-driven systems are initiated on slopes of ca. 2°, with the large systems occurring preferentially above gentle rather than steep slopes.

The emplacement of the gravity-driven systems offshore Namibia is likely related to the combination of key triggering and preconditioning factors that changed through time in response to the tectono-sedimentary evolution of the margin:

- Formation of overpressured layers due to high sedimentation rates and hydrocarbon generation coupled with phases of margin uplift favoured the initiation of Late Cretaceous gravity-driven systems.
- Co-occurrence of weak contouritic strata and seismic shaking aided the initiation of Cenozoic gravity-driven systems.

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SUPPORTING INFORMATION
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