INTRODUCTION

The Permo-Carboniferous glaciogenic deposits of southern Africa, known as the Dwyka Group, have been the focus of scientific investigations for over a century (Sutherland, 1870; Du Toit, 1921; Gravenor et al., 1984; Visser, 1987, 1990, 1997; von Brunn, 1994, 1996; Bangert et al., 1999; Haldorsen et al., 2001; Bordy and Catuneanu, 2002; Catuneanu, 2004; Isbell et al., 2008; Stollhofen et al., 2008; Andersen et al., 2016). These strata played a pivotal role in developing ideas...
about the Late Palaeozoic Ice Age (LPIA), the longest of all glacial periods experienced by our planet during the Phanerozoic (Eyles, 2008; Fielding et al., 2008a). In addition, the Dwyka Group played a seminal role in the elaboration of the plate tectonic concept (Wegener, 1915; Du Toit, 1937). Since these early studies, many others, encompassing sedimentology, stratigraphy, palaeontology and geochemistry, have been conducted on the Dwyka glaciogenic deposits. This interest derives partly from the fact that these strata are widespread throughout southern Gondwana and, as such, have been used to infer growth and recession of ice sheets or caps, climate changes during Gondwana times, and palaeogeography. It is now widely accepted that the deposition of LPIA-related strata across southern Gondwana resulted from successive phases of growth and decay of ice sheets of somewhat restricted extent rather than a single massive ice sheet covering most of the land masses (Fielding et al., 2008a; Horton and Poulsen, 2009; López-Gamundi and Buatois, 2010; Huuse et al., 2012; Isbell et al., 2012; Montañez and Poulsen, 2013). These successive ice sheet waxing and waning events probably depended on parameters such as the wander of the South Pole (Opdyke et al., 2001), variable concentration of atmospheric CO₂ (Frank et al., 2008; Montañez et al., 2016; Myers, 2016), and/or the erection of mountain belts (Isbell et al., 2012; Goddérís et al., 2017). Investigations on correlative glaciogenic deposits across Gondwana have been carried out in South America (Dykstra et al., 2012; Vesely et al., 2015; Tedesco et al., 2016; Fallgatter and Paim, 2017; Valdez Busto et al., 2017; Assine et al., 2018), the Middle East (Martin et al., 2008; Stephenson et al., 2013), India (Wopfner and Jin, 2009), Madagascar (Wescott and Diggins, 1997), Australia (Fielding et al., 2008b), Antarctica (Isbell, 2010; Cornamusini et al., 2017) and North Africa (Le Heron et al., 2009; Bussert, 2010; Le Heron, 2018), where spectacular, glacially related palaeoreliefs (fjords, ice stream corridors) have been discovered.

In southern Africa, research on the Dwyka Group has predominantly focused on the well-exposed western part of the South African Main Karoo Basin and the Namibian Kalahari-Karoo Basin, where the LPIA is now relatively well-constrained both in time and space (Visser, 1983, 1987, 1990, 1993, 1997; Grill, 1997; Bangert et al., 1999, 2000; Stollhofen et al., 2000,2008; Himmler et al., 2008; Andrews et al., 2019). Here, four phases of ice growth and decay have been proposed on the basis of interstratified glaciogenic (diamictites) and non-glaciogenic (Ice Rafted Debris (IRD)-free mudstones) deposits. Chronological inferences are in addition well-established due to abundant and datable ash layers distributed throughout the deglacial sedimentary successions (Bangert et al., 2000; Stollhofen et al., 2008; Isbell et al., 2008). A smaller number of comparable studies have been conducted in the eastern Karoo Basin where no formal interglacial periods are recognized and where age constraints are essentially lacking (Matthews, 1970; von Brunn, 1994, 1996; Haldorsen et al., 2001; Bangert and von Brunn, 2001). Palaeogeography, bathymetries, sedimentary dynamics and depositional conditions thus remain poorly understood in the eastern Karoo Basin. This part of the South African Karoo Basin is, however, crucial as it may correspond to the central part of the ice sheet (Haldorsen et al., 2001) and to the transition between an emerged domain to the north (the Cargonian Highlands) and a deeper basinal setting to the south (Figure 1; Von Brunn, 1994).

In an attempt to bridge this gap, new sedimentological and palaeo-geomorphological data from the eastern Karoo Basin (Figure 1) are presented in this paper. This paper describes glacial surfaces (GS), provides information on palaeo-landscapes and interprets sedimentological facies in terms of depositional environments in order to provide a deglaciation time framework, including ice-margin fluctuations and associated sedimentation processes. Also discussed are the spatial and temporal correlation of the glacial deposits with coeval strata found elsewhere in the Karoo Basin as well as time inferences on relative sea-level (RSL) changes, most probably forced by glacio-isostatic adjustment (GIA). Although the GIA is a process barely constrained in the deep time record, its unravelling remains crucial to disentangle ice-margin advance-retreat cycles and palaeo-geographic inferences (Dietrich et al., 2018). Particular emphasis will be given to grounding zone wedges (GZWs) discovered here, which are believed to constitute the first outcrop example of this kind of sedimentary depocenter in the deep geological record. A conceptual model that accounts for the characteristics of such depocenters will be proposed.

2 | GEOLOGICAL SETTING

In South Africa, the glaciogenic deposits of the Dwyka Group recording the LPIA constitute the lowermost stratigraphic unit of the world-famous Karoo Supergroup. The evolution of the Karoo Basin, which has a present-day extent of 600,000 km² and a cumulative thickness of up to 6 km, started in the Carboniferous by the deposition of glacially related strata, followed by the deposition of marine, lacustrine and continental successions through
Permo-Triassic time and ended in the early Jurassic with the extrusion of continental flood basalts related to the break-up of Gondwana (Smith et al., 1993; Johnson et al., 2006; Isbell et al., 2008; Tarkand et al., 2009). At its initiation, the Karoo Basin formed an E-W striking depocenter deepening southward. The tectonic setting that permitted the initiation of the Karoo Basin and deposition of the lowermost Dwyka and overlying Ecca groups has either been interpreted as a retro-arc foreland basin tied to the Cape orogeny (Catuneanu et al., 1998, 2005; Catuneanu, 2004) or alternatively, if the onset of this orogeny occurred only later (Linol and De Wit, 2016), as a lithospheric deflection produced by far-field stresses related to subduction (Pysklywec and Mitrovica, 1999; Tarkand et al., 2009). At present, the sedimentary Dwyka, Ecca, Beaufort and Stormberg groups and the volcanic Drakensberg Group that constitute the Karoo Supergroup from base to top crop out concentrically due to basin inversion (Figure 1A).

In the eastern part of the Karoo Basin (KwaZulu-Natal Province, Figure 1B), the Dwyka Group unconformably overlies Archaean and Proterozoic basement. In the north of the study area (Figure 1B), the basin is formed by the south-eastern margin of the Kaapvaal Craton, which consists of an assemblage of Archaean lithologies. The ~2.9 Ga Pongola Super group, which forms the dominant substrate for the Dwyka Group, is made up of tilted and folded sedimentary and volcanic rocks metamorphosed at greenschist facies grade (Luskin et al., 2019). Locally, in the central part of the study area, fragments of 3.3–3.4 Ga greenstone belts are exposed. These fragments consist mainly of tightly folded mafic to ultramafic volcanic rocks and cherts subjected to amphibolite facies grade (Hofmann et al., 2019). Further north-east, the basin is formed by 3.1 Ga granitoids of the Mpuluzi suite (Kröner et al., 2019). In the southern part of the study area, the Natal Suture Zone (Figure 1B) dated to 1.20–1.08 Ga is characterized by an array of major east-west striking thrust faults that formed during the accretion of terranes to the Kaapvaal Craton (Spencer et al., 2015). Mainly south and east of the Natal Suture Zone, Siluro-Ordovician elastic sedimentary rocks of the Natal Group cover the older basement and underlie the Dwyka Group (Marshall, 2006).

In south-eastern South Africa, Dwyka Group strata are usually divided into three palaeogeomorphic regions according to syn-depositional bathymetries (Figure 1A). The northern region is ascribed to a levelled highland palaeo-plain, the central one being the coastal-platform domain, while the southern region is commonly interpreted as the deep part of the Karoo Basin (Matthews, 1970; von Brunn, 1994, 1996; Haldorsen et al., 2001). The boundary between the central and the southern regions may correspond to the Natal Suture Zone, which had a profound influence on sedimentation modes (Figure 1B; von Brunn, 1994; Tarkand et al., 2009). The present study only focuses on the central region (Figure 1; von Brunn, 1996). Here, the Dwyka Group ranges in thickness between 0 and 200 m as a result of palaeorelief, and predominantly consists of massive to faintly bedded diamictites, interstratified with sandstones and conglomerates. Contrary to glacial deposits of the western Karoo Basin where four deglacial successions have been recognized, only a single deglacial event has been proposed, within which discrete, high-frequency ice fluctuation sequences might be recognized (von Brunn, 1996; Haldorsen et al., 2001). In the study area, although ash layers are occasionally interstratified with the glaciogenic deposits, no formal age inferences have been obtained for the Dwyka Group (Bangert and von Brunn, 2001). Ice flow throughout the eastern Karoo Basin have been confidently assessed on the basis of glacial striae and streamlined landforms carved into the bedrock: while mainly flowing S to SSE in the northern and central regions, the ice flow was directed towards the SW in the southern domain underlain by post-Archaean rocks, indicating a strong basement control on ice flow (Du Toit, 1954; Matthews, 1970; Versfeld, 1988; von Brunn, 1994). In the study area (Figure 1), the contact of the basin with the overlying glaciogenic Dwyka Group is characterized by a highly uneven relief that consists of palaeotopographic lows and highs that differ in elevation by up to 200 m (von Brunn, 1994). Hard-bed polished, striated and grooved surfaces indicating a SSE-directed ice flow are commonly found on the basement rocks (Du Toit, 1954; von Brunn and Talbot, 1986; Versfeld, 1988; von Brunn, 1994, 1996; Haldorsen et al., 2001). This relief was subsequently almost entirely infilled by the glaciogenic deposits that filled local depressions, onlapped on slopes and sealed or fringed most palaeotopographic highs. Directly lying on the glaciogenic Dwyka Group, the lower Ecca Group is subdivided into the lower mudstone-dominated Pietermaritzburg Formation (Bordy et al., 2017), which bears rare dropstones (Tarkand et al., 2009) and the upper sandstone-dominated Vryheid Formation. Interpreted as postglacial sediments deposited after the retreat of the ice margin from the basin, this succession might have been fed at times by the ice sheet that remained on the northern highland, the so-called Cargonian Highland (Figure 1), long after its disappearance from the basin (Visser, 1990, 1993, 1997, 1995; von Brunn, 1994, 1996; Isbell et al., 2008; Buatois et al., 2010).

3 | MATERIAL AND METHODS

Fieldwork was carried out in 2017 and 2018 in KwaZulu-Natal Province of South Africa (Figure 1). While attention
was paid to preglacial landforms throughout the study area (Figure 2), representative high-resolution stratigraphic sections (scale 1:100) were logged at outcrops and facies along these sections were systematically noted and grouped into sedimentary units (Figures 3 and 4). Facies analyses included observation of grain size, sorting, sedimentary structures, stratigraphic architecture and stacking pattern, geometries of the sedimentary bodies and stratigraphic relationships with underlying, overlying and adjacent units wherever feasible (Figure 5). Samples of representative facies were collected and thin sections prepared at the University of Johannesburg for petrographic studies (Figure 6). Characteristic features of both hard and soft-bed GS and landforms, such as polish, striation (direction, shape, size, etc.), asymmetry, nature and lithology, were systematically investigated and mapped (Figures 1 through 10) when encountered.

FIGURE 2  Hard-bed streamlined landforms developed on Archaean basement rocks that underlie the Dwyka Group in the study area. (A) Roches moutonnées developed in Archaean mafic volcanic rocks. Ice flow was towards the south (towards the observer) as indicated by the plucked, lee side of the form. (B) Conspicuously asymmetric Roches moutonnées developed in Archaean quartzites of the Pongola Supergroup. Ice flow towards the south (left) is clearly evidenced by the asymmetry of the form, the up-glacier side (right) being shallow and striated while the down-glacier (left) side is devoid of striae (plucked) and steeper. Note the faintly stratified diamictite covering the striated pavement (white arrows) (C) A U-shaped trough, 800 m wide and 100 m deep, carved into Archaean basement rocks of the Pongola Supergroup, filled with Dwyka sediments (dark strata, indicated by white arrows) and subsequently exhumed by sub-modern erosion. This trough, oriented in a NNW-SSE direction, is thought to have been carved by flowing ice during Dwyka times. Figure 1B for location.

4  |  RESULTS

4.1  |  The pre-Dwyka palaeosurface: a hard-bed GS

A marked palaeorelief characterizes the Archaean basement onto which the Dwyka Group lies in the study area. The mostly gentle slopes can, however, locally dip at up to 35–40° and vertical walls against which glaciogenic strata onlap have also been observed sporadically (fig. 4.5 in von Brunn, 1994, see also Versfeld, 1988). Polished, striated and grooved surfaces as well as trains of medium-scale streamlined, asymmetrical erosional landforms have been recurrently observed carved into the bedrock throughout the study area (Figure 2). Asymmetrical landforms display gently sloping polished and striated NNW-facing sides...
**FIGURE 3** Synthetic stratigraphic log of the Dwyka Group in the study area. The three sedimentary units (SU1 to SU3) delineated by glacial erosion surfaces (GS) are shown. Note that the measured section is located in the southward continuation of the glacially carved U-shaped trough (Figure 2C). See Figures 1B and 6A for location.
FIGURE 4  Detailed sedimentary logs from the study area. See Figure 1B for location.
and steeper SSE-facing sides devoid of polishing or stria-
tion (Figure 2A and B). Of particular interest is an out-
standing exhumed U-shaped trough, 100 m deep, 800 m
wide and at least 2.3 km long, carved into strata of the
Pongola Supergroup that has been discovered in the study
area (Figure 2C). This trough’s long profile is oriented
NNW-SSE (Figure 1) i.e. parallel to inferred ice flow.
Collectively, these glacial erosion features, as they under-
lie the studied glaciogenic sedimentary succession, define
the lowermost GS which is here named GS0.

4.2 | Facies analysis, soft-bed GS and
stratigraphic architecture

In the study area, the Dwyka Group ranges in thickness
from 0 to 200 m, which is mainly controlled by the pre-gla-
cial relief, the greatest sedimentary thicknesses being found
in palaeo-depressions (Figures 3 through 5). In palaeo-de-
pressions, ~ 80–90% of the total thickness of the Dwyka is
made up of massive to crudely stratified diamictite, with
the remainder consisting of conglomerate and sandstone.
Within each of these three fundamental lithofacies (diam-
ictite, conglomerate, sandstone), variations in grain size
and sedimentary structures were observed (Figures 3 and
4, see also von Brunn, 1994). Photomicrographs of some
peculiar facies are displayed in Figure 6. Beside the above-
described basal hard-bed glacial erosion surface character-
izing the base of the sedimentary succession (GS0), three
other soft-bed GS (GS1–GS3) have been unravelled at dif-
ferent stratigraphic levels in the Dwyka succession where
the thickness is greatest and hence the sedimentary record
the most complete. In areas of reduced thickness, only one
or two GS are commonly found. These soft-bed GS are
associated with striated and grooved pavements, glacial
lineations and fluting or glaciotectonic deformation, as de-
tailed below. These GS were used to delineate three super-
imposed sedimentary units (SU1–SU3) characterized by
particular assemblages of diamictites, conglomerates and
sandstones. These three sedimentary units are interpreted
as ice fluctuation sequences. Note that other subordinate
GS observed throughout the sedimentary succession are interpreted as within-trend, lower-rank fluctuation events, as they are not distributed regionally and/or do not delineate fundamentally different facies assemblages. Interestingly, the log displayed in Figure 3, flanked by basement slopes, and representing one of the most complete sedimentary section found in the study area, lies in the direct projected south-eastward continuation of the cross-shelf trough described above (Figure 1).

4.2.1 Sedimentary unit 1

The lowermost sedimentary unit ranges from 0 to 70 m in thickness and exclusively occurs in topographic depressions of the palaeo relief – and notably in the southward continuation of the above-described cross-shelf trough (Figures 5A, 6A and 7), the slopes of which are onlapped or conformably draped by glaciogenic deposits (Figure 7A). This sequence is made up almost entirely of a greyish to bluish diamictite bearing abundant clasts wrapped in a muddy to sandy matrix (Figures 6A and 7A,B,C). The diamictite may either be massive or crudely bedded, exhibiting both normal and inverse grading (Figure 7C). The massive variety under the microscope reveals a fabric devoid either of clear stratification or deformation patterns (Figure 6A). The clasts occur in a muddy matrix, range in size from sand to boulder, show the whole range of degrees of roundness and consist predominantly of granitoids, but also of komatiite, gneiss, basalt, conglomerate, quartzite, banded iron formation and (banded) chert, representing both local (Archaean basement) and exotic lithologies brought from at least 250 km away (Du Toit, 1954). Clasts are occasionally striated and/or faceted. Some diamictite intervals are conspicuously devoid of small clasts (Figure 7D), and locally interstratified by well-sorted, hummocky cross-stratified, fine-grained sandstones (Figure 7E). In some rare occasions, limestones are found puncturing and downwarping the faint bedding developed in some finer-grained diamictite. Laminated, decimetre-thick and very poorly sorted polymictic conglomeratic sandstone beds and lenses that generally wedge out over a few metres are occasionally incorporated within the diamictite (Figure 7F). Clasts incorporated in these conglomeratic lenses are generally well-rounded and are wrapped in a sandy matrix exhibiting faint horizontal to undulating bedding.

The uppermost 10 m of SU1 is formed by the same greyish to bluish diamictite bearing abundant clasts but bedding is significantly better developed (Figures 3 and 8). The planar to undulating top surface of this bedded diamictite, which also forms the top surface of this first sedimentary unit, is prominently marked by soft-bed glacial erosion forms such as striae, grooves and flutes, corresponding to glacial surface 1 (GS1, Figures 5A,B and 7G). These
Glacial erosion features are commonly reworked by wave and interference ripples (Figure 7H) or are occasionally covered by pebble and cobble lags. Many basement highs protrude from this GS against which the wave-rippled unit reworking the GS wedges out (Figure 5A). As sealed by finer-grained deposits (see below, sedimentary unit 2), GS1 capping this first fluctuation sequence hence appears in the landscape as a horizontal surface highlighted by recent erosion (Figure 5A and B).

In some places, a more complex facies assemblage up to 10 m thick characterizes the top of SU1, displaying abrupt vertical and lateral facies change (log C in Figures 4 and 5).
This assemblage stacks from base to top, or laterally juxtaposes: (a) a fine-grained, dark grey, fissile argillaceous and weakly laminated diamictite with clasts up to boulder size (Figure 8A and E); (b) a coarse-grained, massive, arenaceous and clast-bearing diamictite (Figure 8A); (c) trough cross-stratified, poorly sorted and occasionally conglomeratic sandstones (Figure 8B) that rework the underlying coarse-grained diamictite and incorporate some of its biggest clasts; (d) normally graded, decimetre-thick and laminated sandstone beds interstratified with laminated siltstones (Figure 8C); and (e) a 1 m thick sheet of massive diamictite bearing angular rock fragments conspicuously characterized by soft sediment deformation (Figure 8D). All these facies are disrupted, sometimes intensely, by clastic dykes and sheet intrusions, small-scale downstepping extensional fractures, conventional and intraformational striae and evidence of horizontal shearing (Figure 8A and E), features which collectively indicate the presence of a composite GS (see also von Brunn and Talbot, 1986). At one locality, large-scale elongated ridges, 5–10 m in width, with a relief of up to 1 m and at least 100 m in length, and moulded in a greenish, massive and fine-grained diamictic material characterized by abundant angular pebble-sized clasts have been observed on top of this particular assemblage of facies (Figure 8F). Small-scale soft sediment deformation such as folds and shear bands is frequently observed within this diamictite (Figure 8D).

4.2.2 | Sedimentary Unit 2

The second sedimentary unit (SU2), floored by GS1, also shows thickness variations (10–50 m), partly owing to underlying basement highs not completely sealed by SU1 and protruding from GS1 (Figures 3 and 5). The facies association (Figure 9) characterizing this unit differs from the
underlying one as it consists of an interbedding of somewhat finer-grained diamictic material and sandstone beds that overall show an upward increase in the sandstone proportion along with a general grain size coarsening and bed thickening trend (Figure 3). The base of the sequence consists of a faintly bedded yellowish-greenish diamictite made up of faintly laminated argillaceous siltstones with granule-sized lonestones (Figure 9A) and showing occasional arthropod trackways (von Brunn and Talbot, 1986). This facies coarsens upward and becomes incrementally interstratified by thin, well-laminated, well-sorted and normally graded yellowish fine-grained sandstone beds (Figure 9B and C) that themselves coarsen and thicken upward. Interestingly, on rare occasions, rhythmites formed by supercritical to in-phase climbing current ripples (sensu Hunter, 1977) and developed in fine-grained sandstones have been observed (Figures 6B,C and 9D). Microfacies analysis reveals conspicuous inverse grading and the presence of sand-sized clasts puncturing the bedding, and against which further laminae onlap (Figure 6B and C). Above, the sandstone beds thicken and coarsen. Some reddish scour-based conglomeratic sandstone beds display normal and inverse grading as well as soft sediment deformations (Figure 9E and F). In addition, the above-described interstratified diamictite grades into a greyish to bluish massive to crudely stratified, sandy diamictite similar to the one making up the bulk of SU1, and hosting abundant lonestones up to boulder-sized. The upper 5 m of the sedimentary sequence is made up of stacked conglomeratic sandstone beds that commonly display horizontal, trough and planar cross-bedding (Figure 9G), climbing current and wave ripples, gravel

**FIGURE 9** Facies that make up the second sedimentary unit (SU2) (A) Faintly laminated argillaceous siltstone with granule-sized lonestones. (B) Alternation between facies shown in (A) and planar-laminated, and occasionally normally graded sandstone beds. (C) Close-up view of planar-laminated sandstone beds. (D) Supercritical to in-phase climbing current ripples displaying rhythmic modulation of the angle of climb, see Figure 6B and C for photomicrographs. (E) Alternation between reddish coarse-grained, normally graded sandstone and diamictite bearing abundant clasts. (F) Close-up view of the coarse-grained, sometimes conglomeratic, normally graded sandstone beds. (G) Trough cross-bedded, poorly sorted, occasionally conglomeratic sandstones. (H) Boulder pavement with planar upper surfaces observed on top of SU2.
lags, soft sediment deformations and fluid escape structures. Some segments of the uppermost sandstone beds are folded at a large wavelength of hundreds of metres forming large depressions, while grain-band structures (sensu Busfield and Le Heron, 2018) have been observed microscopically (Figure 6D). A 3 m thick, boulder-bearing, clast-supported, massive to faintly laminated conglomeratic sheet capped by a boulder pavement with planar upper surfaces has been observed at the top of this coarsening-upward succession at one locality (Figure 9H and log A in Figure 4).

4.2.3 | Sedimentary Unit 3

The third and uppermost sedimentary unit (SU3) consists of an alternation of diamictite and sandy-conglomeratic deposits and is commonly 60–70 m thick but reaches 100 m in places (Figures 4, 5C and 10). The base of the sequence ubiquitously consists of the greyish to bluish massive to crudely stratified diamictite that bears abundant clasts up to boulder size, similar to the one making up the first sedimentary unit. Further upward, interlayered with diamictite beds, thick (1 m), poorly sorted, faintly laminated and mostly clast-supported conglomeratic layers occur in association with deformed sandstone rafts, clastic dykes and occasional sandstone beds with climbing ripple cross-lamination. On a few occasions, elongated meso-scale ridges 5–10 m in width and 1 m in depth and moulded in a greenish massive diamictite that show soft sediment deformations and pebble-sized clasts (Figure 10A) have been observed. The top surface of these meso-scale streamlined features bears striae and grooves.
oriented parallel to the long axis of these ridges, and occasional rill marks (Figure 10B). Rare boulder pavements with faceted upper faces occur in association with the elongated ridges and are sealed by finely laminated, poorly sorted siltstones with abundant IRD up to cobble size. The top of this sedimentary unit is either made up of a conspicuously stratified diamictite, which differs from the basal diamictite in possessing a coarser-grained matrix, or in a similar way to the lowermost fluctuation sequence, or by a stack of poorly sorted conglomeratic sandstone beds displaying planar and undulating laminations, planar, trough and climbing-dune cross laminations and climbing current ripples (Figure 10C and D).

A striated pavement displaying striae and small-scale grooves was observed at the top of this sandstone succession at one locality (Figure 10F), where it was reworked by small-scale, curve-crested dunes. Alternatively, sandstones and diamictites are capped by a thin (1–3 m) layer of well-laminated IRD-bearing greenish mudstones (Figure 10E), intensively deformed by pervasive soft sediment deformation such as folding and faulting (downstepping extensional fractures; Figure 10G and H). The top surface of this unit, which also constitutes the top of the Dwyka Group in this region, is highly undulating and uneven, forming a relief that can attain 40 m over kilometre-scale wavelengths. The contact with the overlying black shales of the lower Ecca Group (Pietermaritzburg Fm) is either abrupt above deformed IRD-bearing mudstones or transitional where undeformed mudstones grade upward into the black shales (Figure 3 and logs E and F in Figure 4).

5 | INTERPRETATIONS: DEPOSITIONAL ENVIRONMENTS, ICE-MARGIN FLUCTUATIONS AND ASSOCIATED MODE OF SEDIMENTATION

5.1 | The pre-Dwyka palaeosurface: an exhumed glacial landscape

Polished and striated surfaces undoubtedly indicate that flowing ice carved the bedrock prior to deposition of the Dwyka Group in this region. Asymmetrical bedforms are interpreted as Roches moutonnées and confirm that ice was flowing towards the SSE. Oriented parallel to the inferred deepening axis of the basin, and eroded into the central palaeobathymetric domain interpreted as a coastal-platform domain (see above and Figure 1A), the U-shaped trough might be regarded as a small-scale cross-shelf trough carved by subglacial processes probably linked to flowing ice, as also evidenced by streamlined forms and polished pavements (Krabbendam et al., 2016; Newton et al., 2018; see also Andrews et al., 2019).

The palaeosurface onto which the Dwyka Group lies thus corresponds to a glacial landscape scoured by the SSE-ward moving glaciers and subsequently sealed by glaciogenic sediments. In recent times, preferential erosion of the Dwyka Group over the more resistant Archaean lithologies of the basement has exhumed this pre-Dwyka palaeosurface. The path of the modern Tugela and Buffalo rivers (Figures 1 and 2C) closely follows the network of pre-Dwyka fluvial and/or glacial valleys that had been completely filled up with glaciogenic sediments during the glacial period and subsequently re-excavated during Cenozoic and Recent times. Hence, in KwaZulu-Natal Province, in areas where glaciogenic deposits have been entirely eroded away, the modern topography exhibits Dwyka-aged polished and striated surfaces as well as Roches moutonnées carved into the ancient basement and hence corresponds to the exhumed Permo-Carboniferous glacial landscape that prevailed at the onset of the deposition of the Dwyka sediments (von Brunn, 1994, 1996; see also Guillocheau et al., 2018). Such an exhumed LPIA glacial landscape echoes those of the same age found in Namibia (Andrews et al., 2019), Chad (Le Heron, 2018) and in South America (Assine et al., 2018).

Although it appears evident that glacial erosion was at least partly responsible for carving and shaping the pre-Dwyka palaeosurface (striated and polished surfaces, Roches moutonnées), it seems, however, probable that other processes also contributed to the arrangement of this surface prior to the deposition of the Dwyka Group. Indeed, as stated by von Brunn (1994, 1996), the complex structural grain of the Archaean basement rocks probably contributed significantly to the uneven configuration of the palaeosurface which was later modified and accentuated by glacial scouring that exploited existing weaknesses (Newton et al., 2018). The presence of vertical walls undoubtedly indicates that the structural heritage had a preponderant influence in the shaping of the basement. Moreover, a pre-existing network of fluvial valleys was potentially reworked and amplified by direct glacial abrasion (von Brunn, 1994, 1996), as is the case for some Quaternary high-latitude valleys that originated from the combination of both pre-glacial fluvial and glacial erosion processes (Lajeunesse, 2014; Livingstone et al., 2017). Indeed, the major hiatus that separates the basement from the Dwyka Group suggests that the study area constituted an emerged domain during pre-Dwyka times as a result of crustal uplift preceding the inception of the Karoo Basin and the onset of glaciation (Visser, 1990; Tankard et al., 2009) and onto which a fluvial system may well have developed.
5.2  Depositional environments

5.2.1  SU1: Grounding zone wedges

Von Brunn (1994, 1996) and Haldorsen et al. (2001) interpreted the diamictite-dominated unit SU1 as having resulted from the settling of suspended material and dropstones dumped from drifting icebergs close to the ice front as well as resedimentation of glaciogenic material through debris flows (see also Visser, 1997). Indeed, the large array of predominantly diamictic facies supplemented by intervening conglomerate and sandstone beds observed throughout this sequence indicates a variety of highly energetic depositional processes. Indeed, resedimentation of (subglacial) glaciogenic material through sediment gravity flows (mass and debris flow, turbidities) is thought to have originated in the deposition of massive to faintly bedded diamictite, conglomerate lenses as well as normally graded sandstone beds (Talling et al., 2012; Le Heron et al., 2017 and references therein), while massive rain-out of debris and dumping of debris from either a floating ice shelf or abundant drifting icebergs are being expressed by IRD-bearing diamictite (e.g. Figure 6D) possibly supplied by a meltwater plume (Dowdeswell et al., 2015). Hummocky cross-stratified deposits (Figure 6E) are interpreted as originating from storm events. So the build-up of this thick sequence thought to have occurred in a subaqueous, ice-proximal setting continuously supplied by glaciogenic materials, as evidenced by exotic lithologies, and the presence of intervening GS. Furthermore, the occurrence of the fine-grained, greenish diamictite material bearing soft sediment deformation (Figure 8D) and interpreted as subglacial till (Evans et al., 2006), striae and grooves forming a composite GS (log C in Figure 4 and Figures 7 and 8) capping SU1 points towards the presence of a nearby grounded, fluctuating ice front prone to override its own proximal sedimentary products. Finally, elongated ridges moulded into subglacial till represent mesoscale streamlined landforms interpreted as flutes (Ely et al., 2016) may indicate they were implemented in a subglacial environment by overriding flowing ice (Schoof and Clarke, 2008). Collectively, the association of these deposits (Figures 7 and 8) capped by a composite GS (Figure 8) enables this succession to be interpreted as a GZW (see also Visser, 1997). The GZW correspond to asymmetrical, wedge-shaped depocenters emplaced at the submarine grounding zone of fast-flowing or streaming ice by the continuous delivery and resedimentation through a variety of sedimentary processes of subglacial material (till) from up-glacier (von Brunn and Talbot, 1986; Powell and Domack, 2002; Dowdeswell and Fugelli, 2012; Bell et al., 2016; Dowdeswell et al., 2016a, 2016b; Lajeunesse, Bart et al., 2017; Prothro et al., 2018; Lajeunesse et al., 2018; Batchelor et al., 2018; see also Demet et al., 2019 for some facies). Although generally interpreted as being deposited at the grounding line of a floating ice shelf (Batchelor and Dowdeswell, 2015), GZWs may also form at tidewater termini, depending on the availability of meltwater and/or deformability of subglacial till (Powell and Alley, 1997). These authors, however, indicate that a tidewater terminus would promote the deposition of ice-contact fan and hemipelagic sediments in proximal and distal positions, respectively, in which case IRD would be exhausted within a few hundred metres of the grounding zone. Hence, the virtual absence of hemipelagics as well as the wealth of clasts (dropstones) throughout this unit suggests that a floating ice shelf probably existed during the deposition of this lowermost unit.

Such an interpretation would account for the diversity of facies and associated processes observed in this sedimentary unit as well as its significant thickness throughout the study area. In such a context, the complex facies assemblage at the top of SU1 and displaying abrupt vertical and lateral facies change (Figure 8), clastic dykes and sheets and superimposed striated surfaces (composite GS) might correspond to the effective grounding zone of the GZW prone to multiple ice-marginal fluctuation and episodic grounding (von Brunn and Talbot, 1986). Wave and interference ripples reworking in place the GS indicate that the ice retreated after the construction of GZW. It is thought that the GZW was then abandoned and subsequently winnowed and washed-out of the fine fraction of the diamictite by nearshore processes, either immediately or after a fall of RSL that exposed the summit of the GZW – the effective GS – to wave action (Dietrich et al., 2017, 2018; Demet et al., 2019). Furthermore, the wedging of wave-ripped sediments or the pebble/cobble lag against the basement slopes indicate that by the time basement highs situated above this horizon emerged only palaeotopographic lows were inundated. An outstanding modern analogue of such a GZW reworked in a shallow-marine domain and which correlativeshore-related deposits onlap basement slopes is given in Lajeunesse et al. (2018, their fig. 12).

5.2.2  SU2: Ice-contact fan

On the one hand, the upward coarsening and thickening of the sandstone beds is indicative of a progradational sequence of a subaqueous sedimentary system over which high-density sediment gravity flows such as turbidites and debris flows expressed by normally graded sandstone beds and conglomeratic layers (Talling et al., 2012) spread. On the other hand, the diamictite interstratified in between these sandstone and conglomeratic beds indicates the proximity of an ice margin. The general coarsening-upward trend, as well as the increasing proportion of IRD, suggests that the ice margin was advancing towards the depositional area. The coarse grain size of the uppermost sandstone beds, the rapid coarsening-upward as well as soft sediment deformation and sedimentary structures witnessing traction flows indicate that the deposition was rapid.
and sustained in a dynamic, proximal setting. In such a glacial context, this sedimentary sequence is therefore interpreted as a glaciolfluvial or ice-contact delta or fan (sensu Lønne, 1995, see also Lønne et al., 2001; Dowdeswell et al., 2015). Delta topset beds might be represented by the uppermost cross-bedded and rippled sandstone beds. The boulder conglomerate capping this sedimentary sequence is interpreted as the most dynamic deposit, most probably deposited in an ice-contact environment by the deconfinement of meltwater flows exiting subglacial tunnels (Russell et al., 2006; Alexander and Cooker, 2016; Aquino et al., 2016). Altogether, the boulder pavement, the undulating geometry of the sandbads making up the top surface of SU2, as well as the microscopic grain bands that could be interpreted as boudins are evidence of overriding ice (Visser and Hall, 1985; Buechi et al., 2017; Busfield and Le Heron, 2018 and references therein). The rhythmic climbing ripples seem to indicate a tidal influence on deposition and might have resulted from the interaction of tides with glaciofluvial inputs to generate and/or support tide-influenced sediment gravity flows (Smith et al., 1990; Cowan et al., 1998; Dietrich et al., 2017). Observed inverse grading (Figure 6B and C) may represent waxing flow of hyperpycnal events (Mulder et al., 2001), whereas sand clasts puncturing the laminae are interpreted as dropstones derived from floating ice (sea ice, drifting icebergs or ice shelf; Dowdeswell et al., 2015).

In contrast to the underlying sedimentary succession either emplaced at the grounding zone of a floating ice shelf or at a tidewater terminus, ice-contact fans are exclusively associated with tidewater ice margins (Powell and Alley, 1997; Batchelor and Dowdeswell, 2015). The presence of a floating ice shelf nearby cannot, however, be conclusively ruled out, as both the diamictite and sandstone horizons bear IRD. Lonstones could also originate from drifting icebergs or sea ice or, alternatively, from reedimentation of till material through debris flow (Vesely et al., 2018). Hence, the upward disappearance of the diamictite in favour of the sandstones might either be interpreted as the increasing influence of the progradation of the ice-contact fan on sedimentation or, alternatively, as the emergence of an initially subaqueous ice-contact fan that would have grown up into an ice-contact or glaciofluvial delta. Such a transition would have promoted the demise of a glaciomarine system and permitted the transition from a marine to a continental ice margin (Powell, 1990; Lønne et al., 2001; Dowdeswell et al., 2015).

### 5.2.3 SU3: Mixed-influenced GZW-ice-contact fan

This third and uppermost fluctuation sequence appears as an amalgamation of deposits resulting from GZW sedimentation (diamictite) and from ice-contact fans and deltas (coarse-grained sandstones). It is therefore suggested that SU3 corresponds to a glaciomarine setting comprising and interdigitating ice-contact fans and GZWs, similar to the setting shown in fig. 20b of Dowdeswell et al. (2016b). Such a combination arguably results from the availability of basal meltwater emerging at the grounding zone and permitting the deposition of ice-contact fans in a setting otherwise dominated by the continuous delivery of subglacial material along a line source (GZW). The meso-scale elongated ridges made up of greenish diamictic material showing soft sediment deformation are thus also interpreted as flutes and hence suggest the presence of flowing ice. Finally, the intensively deformed and folded horizons of IRD-bearing mudstones, sometimes associated with downstepping extensional fractures, and lying in the same stratigraphic position as striated and grooved pavement (Figure 9F) are interpreted as the effect of subglacial glaciotectonism (Evans et al., 2006).

### 6 DISCUSSION

#### 6.1 Ice margin fluctuations and grounding-line sedimentation

As each sedimentary unit is bounded at its bottom and top by GS, they are interpreted as recording ice-margin fluctuations. In such a context, the lower part of each sequence records the retreat of the ice margin from the position it occupied at the implementation of the underlying GS while the upper part corresponds to the glacial advance culminating in the overriding of ice (development of the overlying GS). Note, however, that the turnaround between the ice-margin retreat and the subsequent advance may be largely undecipherable in the sedimentary record; each fluctuation sequence can be highly asymmetric and facies dislocation does not necessarily represent any ice-margin retreat-advance trend reversal but rather reflect progradation of related sedimentary systems (Normandeau et al., in press; Powell and Cooper, 2002; Normandeau et al., 2017; Dietrich et al., 2018).

Glacial maxima conditions during Dwyka times are thought to be represented by the lowermost GS carved into the bedrock (GS0) when the ice margin was located in southwest South Africa and KwaZulu-Natal was entirely covered by ice (Visser, 1997; Haldorsen et al., 2001; Tankard et al., 2009). The ice-margin then retreated throughout the Karoo Basin and stabilized over the study area. This stabilization was probably permitted by the grounding of the ice margin on the coastal-platform domain of the eastern Karoo Basin (Figure 1A), where abundant bedrock pinning points may have acted as a zone of reduced water depth that permitted the ice anchorage. The first fluctuation sequence (SU1) was deposited during ice-margin stabilization in front of the grounding line by the deposition of the GZW (SU1;
Figure 11. Synthetic depositional model for the Dwyka Group in the study area, depicting the three fluctuations of the ice-margin and their associated deposits. The last cartoon shows the modern-day situation where Cenozoic (Guilloucheau et al., 2018) and sub-modern erosion exhumed glacial landscapes carved by the Dwyka ice sheets.

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The GS (GS1) capping this first sedimentary sequence is interpreted as a subsequent re-advace of the ice margin over its own proglacial deposits. This glacial advance may either have been climate-triggered (cooling) or, alternatively, originated from a reduction of the water depth. Such a shallowing could have either been due to a RSL fall forced by a GIA consecutive to the ice-margin retreat and/or to the accumulation of sediments in front of the grounding zone and possibly underneath the floating ice shelf that would have permitted the anchoring of the ice and then its advance (Alley et al., 2007; Anandakrishnan et al., 2007; Brinkerhoff et al., 2017; Batchelor et al., 2018). It is here thought that positive feedback existed between pre-glacial topography, ice-margin stabilization and deposition of sediments: ice anchorage and position of the grounding zone immediately after the initial ice-margin retreat were at least partly controlled by bedrock pinning points, which in turn also controlled the position of the GZW themselves, acting to further stabilize the ice margin and even permitting its autogenic re-advance. This first fluctuation sequence was hence mainly deposited in troughs of the preglacial topography (Figure 11; see also Fallgatter and Paim, 2017), while highs may have pinned and maintained a more or less perennial ice shelf.

The onset of deposition of the second ice-margin fluctuation sequence (SU2) corresponds to the retreat of the ice margin from its stillstand position (GS1) and its reworking by shallow-water currents, giving rise to wave and interference ripples. Hence, the initial ice-margin retreat may be only virtually marked by the abandonment and reworking of GS1 (Figure 11); the overlying coarsening-upward succession (SU2) marking the subsequent ice-contact fan and/or deltaic progradation possibly tied to the glacial advance culminating in the overriding of its own proglacial deposits (GS2, see also Powell and Cooper, 2002). As may have been the case for the underlying GZW, the progradation to aggradation of the ice-contact depositional system could have permitted an autogenic glacial advance (Boulton, 1990). A similar scenario is envisioned for the third, uppermost fluctuation sequence (SU3). It should be noted, however, that although amplitudes of ice-margin advance-retreat fluctuations are not constrained here, the virtual absence of prominent deposits lacking glacial features, like the interglacial black shales in the western Karoo Basin (Visser, 1997; Isbell et al., 2008) suggests that...
retreat phases of the three fluctuations were probably of restricted amplitude and of short term. The final retreat of the ice-margin from the study area, marked by the facies dislocation characterizing the transition from the Dwyka to the Ecca, may have abandoned the subaqueous glaciogenic depocenters (GZW, ice-contact fan) over which the Pietermaritzburg and Vryheid delta subsequently prograded (Figure 11; see discussion in Dietrich et al., 2018 as well as their fig. 7). Such a deltaic progradation that would have sealed abandoned glaciogenic depocenters may have permitted the preservation of the described sedimentary succession.

It has been shown above that depositional environments evolved throughout the Dwyka Group from GZW through ice-contact fan to a bimodal setting. The occurrence of these environments is thought to be controlled by the availability of basal meltwater possibly reflecting palaeoclimate rather than the morphology of the ice-margin terminus (ice shelf vs. tidewater glacier, Powell and Alley, 1997; Dowdeswell et al., 2016b). Hence, it is suggested that the Dwyka ice sheet responsible for the deposition of the lower fluctuation sequence was characteristic of a high-latitude, arguably cold polar setting which permitted the implementation of GZWs owing to the lack of flowing basal meltwater. Above, the occurrence of ice-contact fans (SU2) and mixed-influenced fan-GZW (SU3) indicates an increase in basal meltwater availability possibly tied to climate amelioration (Bjarnadóttir and Andreassen, 2016; Dowdeswell et al., 2016b). The total lack of eskers and subglacial channels as well as the predominance of streamlined features (flutes, striated pavements, U-shaped trough) throughout the glaciogenic succession indicate, however, that sedimentary processes continued to be dominated by flowing ice, subglacial deformation and erosion and delivery of plastered material (subglacial till) to the ice-contact systems.

6.2 | A Quaternary-style deglaciation timing for the Dwyka Group

Although sparse tuffaceous beds have been described from the Dwyka Group in the vicinity of the study area, no age inference has been extracted from them (Bangert and von Brunn, 2001). Hence, the precise age of Dwyka sedimentation in the eastern Karoo Basin remains unknown. Modes of sedimentation and volume of the sedimentary units can, however, provide valuable insights into the duration of their deposition as they are thought to correlate with the duration of the stillstanding ice margin that permitted their deposition (Bart et al., 2017; Protho et al., 2018). Despite their significant extent, several tens to hundreds of cubic kilometres, GZWs are typically deposited in a few hundreds to thousands of years, while duration of deposition of grounding-line fans and deltas ranges from years to hundreds of years (Dowdeswell et al., 2016b, fig. 13; see also Demet et al., 2019). In fact, one of the largest GZW of the outer Antarctica shelf (eastern Ross Sea) emplaced during the Last Glacial Maximum (LGM) being more than 100 km in length and 200 m in thickness and representing a volume of ~ 500 km$^3$ was deposited in less than 4,000 years (Bart et al., 2017). Although no volume estimates were deduced in the present study considering that the spatial extent of the depocenters is unknown, thicknesses occupied by these GZWs may be viewed as comparable or even less, indicating that the timing of deposition was in the same order of magnitude as the sub-modern examples.

Then, each fluctuation sequence taken separately only represents a few thousands to tens of thousands of years at most. The virtual absence of hemipelagic sediments, as well as condensed or highly bioturbated horizons separating the fluctuation sequences, furthermore indicates rapid deposition of, and no significant time gaps within each fluctuation sequence. The time span encompassed within each glacial erosion surfaces is, however, not constrained but as ice-margin retreat phases and the deposition of associated sedimentary suites were rapid, correlative advances were arguably of the same frequency. In Quaternary series, GS stacked in a sedimentary succession ascribed to a deglacial trend and recording glacial stillstand or re-advance represent short intervals of time (lower-rank glacial retreat surface sensu Zecchin et al., 2015, see also Occhietti, 2007; Lajeunesse et al., 2018; Dietrich et al., 2018). Hence, the duration of the deposition of the whole Dwyka Group in this area may be viewed as having taken place in a time range similar to Quaternary deglacial sequences, as already proposed by Haldorsen et al. (2001), and being thus comparable to other ancient deglacial sequences (Girard et al., 2015; Dietrich et al., 2018).

6.3 | RSL change and associated GIA

As outlined above, the duration of the deposition of the whole Dwyka Group in the study area may have been short, on the order of a few thousands to tens of thousands of years. In a stable epicratonic tectonic setting, RSL variations would have only been controlled, in such a deglacial context, by glacio-eustatic and glacio-isostatic fluctuations; crustal subsidence being arguably too slow to have a notable influence over such short periods of time (Boulton, 1990; Girard et al., 2015; Dietrich et al., 2017, 2018). Because of minor amplitudes, ice-margin advance-retreat phases recorded in the study area are thought to have been largely uncoupled to the pattern of continental-scale ice sheet fluctuations (Isbell et al., 2008, and see below), and hence unaccompanied by important glacio-eustatic sea-level change. The pattern of RSL change throughout the studied sedimentary section is hence thought to solely reflect glacio-isostatic processes tied to ice-margin fluctuations. In terms of vertical glacio-isostatic movement, glacial retreat must be accompanied by RSL fall forced by the GIA (Boulton, 1990; Dietrich et al., 2018) while glacial
advances are preceded by RSL rises induced by glacio-isostatic deflection (Storms et al., 2012).

Although arguably (glacio)marine in origin, depositional bathymetries of SU1 are largely unknown. The reworking of the overlying GS (GS1) by nearshore processes, however, undoubtedly indicates that deposition took place in a very shallow to sub-emerged domain. This emersion most probably implies that RSL fell after the retreat of the ice-margin responsible for the implementation of the GS due to the GIA. Furthermore, if the lowermost GS carved into the bedrock (GS0) corresponds to glacial maxima conditions, then the accompanying GIA must have been the most important one of the whole Dwyka sequence. The second and third fluctuation sequences bear no formal diagnostic criteria to constrain any RSL changes; general trends have then to be inferred from the presence of GS: RSL rise being inferred from glacial advance culminating in the implementation of a GS while RSL falls follow the GS (glacial retreat).

6.4 Relationship with coeval glaciogenic deposits

The presence of GZWs and associated ice-contact fans and/or deltas have been inferred based on facies associations and associated features. As similar deposits cover a large part of the intermediate platform domain (Du Toit, 1954; von Brunn, 1994, 1996; Haldorsen et al., 2001), it is suggested here that they might belong to a compound GZW or an assemblage of backstepping/overlapping GZWs, a setting commonly encountered on formerly glaciated margins (Dowdeswell et al., 2016a; Bjarnadóttir and Andreassen, 2016; O’Brien et al., 2016; Bart et al., 2017). Their overall architecture would, however, in the absence of any regional data (outcrops extending over tens of kilometres, onland seismic reflection surveys, Decalf et al., 2016), remain largely undecipherable. Glacial deposits from KwaZulu-Natal interpreted as lacustrine varvites (Savage, 1971; von Brunn, 1994; Haldorsen et al., 2001) may have formed in depositional environments confined to topographic depressions of the pre-Dwyka GS (glacial valleys) and/or inherited from the ice-proximal glaciogenic features (von Brunn, 1994; Dietrich et al., 2017).

In the absence of any absolute ages from the study area, correlation of Dwyka strata with those from the western Karoo Basin remains speculative. As the ice-margin fluctuations inferred here are most probably of high frequency, they must be discrete at the scale of sequences deciphered in the Western Karoo Basin (Visser, 1990; Isbell et al., 2008). This suggests that the deposition of the entire deglacial succession described here and formed by three higher-frequency fluctuation sequences corresponds to one of the single glacial–deglacial cycles observed in the western Karoo Basin, and most probably the last one—at least in term of lithostratigraphic unit as the deglaciation of the Karoo Basin may

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**FIGURE 12** Outline for a conceptual model defining identifying criteria for grounding zone wedges (GZW) including the four indicators (geomorphological, architectural, sedimentological and biological). The three inner arrowed circles represent the likelihood that the different indicators will be deciphered by different investigative methods (outcrop and core studies, seismic and bathymetric surveys.)
have been diachronous (Griffis et al., 2019)—since they are both covered by Ecca Group strata. In such a perspective, the three lower glacial–deglacial cycles present in the western Karoo would be virtually not recorded in the KwaZulu-Natal Province of South Africa that was still covered by the ice mass at that time.

6.5 Criteria to identify GZWs

Although extensively imaged on modern high to mid-latitude shelves (Dowdeswell et al., 2016a), this paper reports on what is thought to be the first described outcrop of GZW in the pre-Quaternary record (see Demet et al., 2019 for a post-LGM example), even though Ordovician buried fossil GZWs have been imaged onland by seismic methods (Decalf et al., 2016). This lack of known ancient examples partly arises from the fact that this type of depocenter has been defined from the study of Quaternary glaciogenic successions, the term ‘grounding zone wedge’ itself referring to a geometry (asymmetric sedimentary depocenters; Bell et al., 2016; Batchelor et al., 2018) largely missing in the ancient geological record, erased by post-depositional processes. In addition, the dimensions (hundreds to tens of thousands of metres in length and several metres to hundreds of metres in thickness) and aspect ratio of GZWs may far exceed the size of available outcrops which makes their identification even more problematic. Similarly, even if GZWs are characteristically wedge-shaped and asymmetrical, with a steep up-glacier face and a smoother down-glacier face, slopes generally do not exceed a few degrees for the steepest face and less than 1° for the shallowest face and are thus hardly decipherable at outcrop or in the landscape, especially if affected by regional tectonic deformation. Besides, facies models for GZWs are scarce, which may explain how few have been reported in the stratigraphic record (Prothro et al., 2018; Demet et al., 2019). Indeed, although the definition of GZW implies depositional processes (‘till emerging from beneath the glacier along a line source is redistributed by subaqueous debris flows’: Bell et al., 2016), associated facies are not necessarily straightforward.

Advantage is therefore taken of the described GZW outcrops in the eastern Karoo Basin of South Africa combined with parameters inferred from bathymetric, seismic surveys, outcrop and core studies to propose an outline for a conceptual model defining criteria to identify this type of depocenter. Even if the original morphology of the depocenters, as well as depositional processes often directly observed or inferred in modern glacimarine settings, are intrinsically lacking in ancient successions, the latter offer access to facies and stratigraphic architecture of subglacial to ice-marginal settings unreachable in modern context, or at a spatial resolution that cannot be attained for Quaternary depocenters which are hardly penetrated by high-frequency seismic waves and piston cores (Dowdeswell et al., 2019).

Four indicators (geomorphological, architectural, sedimentological and biological) are proposed to account for the diversity of depositional processes and stratigraphic architecture and geomorphologic expression characterizing GZW depositional environments, as detailed below and summarized in Figure 12. It is highly unlikely that all these indicators would be ticked for most of the studies performed on either modern, Quaternary or ancient settings, yet it is envisioned that the presence of only some of these parameters would permit the confident identification of a GZW.

- Geomorphological indicators encompass the archetypical asymmetric, wedge shape of the GZWs, ranging from 2 to 200 m in height and from 0.2 to 200 km in length (see fig. 8 in Demet et al., 2019) as well as the presence of characteristic glacial features and landforms superimposed on or encompassed within the GZW (megaflutes, iceberg ploughmarks, mega-scale glacial lineations, moraines, etc.). The association or juxtaposition of recessional curve-crested GZWs may strengthen such an interpretation. Bathymetric and seismic surveys are particularly well-suited for the recognition of the geomorphological indicators (Dowdeswell et al., 2016a).

- Architectural indicators, which can be unravelled by both seismic and outcrop studies, comprise the identification of the stacking pattern and the stratigraphic architecture of the GZW pointing to the active progradation of the GZWs (internal architecture showing seaward-dipping clinofolms topped by horizontal sheets, truncation of horizons by erosion surfaces, landward-dipping backsets; Dowdeswell and Fugelli, 2012; Lajeunesse et al., 2018).

- Sedimentological indicators are best emphasized by outcrop investigations and consist, on the one hand, of bedding and lamination patterns that show a predominance of thick, massive to crudely stratified strata, interstratified with cross-stratified conglomeratic sandstones in which the presence of contorted beds and glaciotectonically deformed horizons is common (Demet et al., 2019). On the other hand, a wide range of grain sizes is supposed to characterize the largely unsorted material making up the bulk of GZWs (diamictite) while the abundance of exotic, outsized, faceted and/or striated clasts should be regarded as characteristic.

- Finally, biological indicators can aid in the interpretation of GZW as specific diatom and foraminifera assemblages have been found in the diamictic material forming GZWs (Prothro et al., 2018). It is also envisioned that after their abandonment subsequent to glacial retreat, ice-contact deposits such as GZWs may be colonized by marine fauna and flora that can leave behind bioturbation (O’Brien et al., 2016).
7 | CONCLUSIONS

The glaciogenic Dwyka Group of the eastern Karoo Basin in South Africa consists of an up to 200 m thick succession of diamictite facies interstratified with sandstone and conglomerate. Three distinct sedimentary units have been recognized. The lower one, lying directly on a highly uneven surface on basement rock carved by glacial action, is thought to represent GZW deposits. Such rocks are common to modern high-latitude continental shelves but have so far remained unrecognized in the ancient geological record. In that sense, the present study advances the comprehension of glacial depositional environments. The second and third units are represented by ice-contact delta and mixed-influenced GZW deposits, respectively. The three depositional units are separated by GS (stratified pavements, glacirotectonic complex) and represent ice-margin fluctuation sequences. Their deposit is interpreted, by analogy with Quaternary depositional systems, to have been very rapid (tens to hundreds of thousand years) during retreat of the Gondwana ice sheet. The RSL variations are thought to only represent glacio-isostatic signals, forced by the demise of the ice sheets. Deposition of glaciogenic Dwyka strata was probably very rapid on a geological time scale.

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