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Sheet-crack cements and early regression in Marinoan (635 Ma) cap dolostones: Regional benchmarks of vanishing ice-sheets?

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1. Introduction

During the younger Cryogenian (Marinoan) glaciation, the continents were huddled between 30°N and 70°S latitude (Fig. 1). The 'Northern Ocean' would have dominated ocean circulation had it not been covered by perennial sea ice. Dynamic ice-sheets shrouded virtually every continent and many drained directly into the ocean, even along the palaeoequator (Evans, 2000, 2003; Evans and Raub, in press; Hambrey and Harland, 1981; Harland, 1964; Hoffman and Li, 2009; Trindade and Macoun, 2007). Whether a dynamic 'sea-glacier' fully covered the ocean, or large areas of seasonally open water persisted through the glacial maxima, is controversial (Goodman and Pierrehumbert, 2003; Hyde et al., 2000; Lewis et al., 2007; Liu and Peltier, 2010; Pollard and Kasting, 2005; Warren et al., 2002). Widespread geological evidence for dynamic (wet base) glaciation on land (Deynoux, 1985; Domack and Hoffman, in press; Dow and Gemuts, 1969; Edwards, 1984; Hambrey, 1982; Hambrey and Spencer, 1987; Hoffman, 2005; Kellerhals and Matter, 2003; McMechan, 2000) suggests that the ice-sheets were in dynamic steady state, broadly comparable to the present East Antarctic Ice Sheet (Donnadieu et al., 2003; Liu and Peltier, 2010; Pollard and Kasting, 2004). It has a mean thickness above sea-level of ~2.0 km (Lythe et al., 2001), and most of its interior area and outlet ice-streams undergo basal melting (Pattyn, 2010). Given the extent and thickness of Marinoan ice-sheets, the glacial ocean was ~25% smaller in volume, with a proportional increase in salinity, than when ice-sheets were absent. This corresponds to a difference in mean sea-level of ~1.0 km, insensitive to the contested differences in marine ice extent.

The abrupt termination of the Marinoan glaciation at 635 Ma (Condon et al., 2005) caused vast coastal flooding. Remarkably, meters to decameters of carbonate sediment, dubbed "cap dolostones", were draped across continental margins and marine platforms worldwide as flooding progressed (Bertrand-Sarfati et al., 1997; Font et al., 2010; Grotzinger and Knoll, 1995; Halverson et al., 2006; Kennedy, 1996; Nogueira et al., 2003; Rose and Maloof, 2010; Shields, 2005). They directly overlie terminal glacial deposits without evident hiatus, and they drape the sub-glacial erosion surface far beyond the confines of glacial deposits. They represent singular events: multiple cap dolostones are not observed, even where multiple "glacial–nonglacial" cycles are present (Allen et al., 2004; Rieu et al., 2007). Cap dolostones were deposited diamorphously, from oldest to youngest with increasing palaeo-elevation. Their mass and extent connote an anomalous flux of alkalinity during deglaciation (Higgins and Schrag, 2003), in addition to strong surface warming and the disproportionate concentration of continents in low latitudes, where carbonate productivity is greatest. Systematic spatial and temporal variations in δ13C (Halverson et al., 2004; Hoffman et al., 2007; James et al., 2001; Jiang et al., 2006; Kennedy, 1996; Nogueira et al., 2003; Rose and Maloof, 2010; Shields, 2005). It has a mean thickness above sea-level of ~2.0 km (Lythe et al., 2001), and most of its interior area and outlet ice-streams undergo basal melting (Pattyn, 2010). Given the extent and thickness of Marinoan ice-sheets, the glacial ocean was ~25% smaller in volume, with a proportional increase in salinity, than when ice-sheets were absent. This corresponds to a difference in mean sea-level of ~1.0 km, insensitive to the contested differences in marine ice extent.

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2007; James et al., 2001; Kennedy et al., 1998) demonstrate that cap dolostones were not detrital in origin, as does their occurrence in areas where sources of detrital carbonate were absent (Jiang et al., 2006; James et al., 2001; Nogueira et al., 2003). Marinoan cap dolostones are typically pale pinkish-grey (weathering to yellowish-grey) in colour and are extremely lean, containing less than 0.001 by weight of total organic carbon. Their conspicuous lamination is defined by normal and reverse graded micro- and macropeloids, deposited as hydraulically-sorted silt-, sand- and gravel-size (3.0 mm diameter) spherical aggregates (Aitken, 1991; Hoffman et al., 2007; James et al., 2001; Kennedy, 1996; Xiao et al., 2004). They were deposited mainly below sea-level and above prevailing wave base; inter- and supratidal indicators are uncommon except in specific regions (Shields et al., 2007; Zhou et al., 2010). Cap dolostones represent the transgressive tract of a global depositional sequence (Hoffman and Schrag, 2002) and, where fully developed, they grade upward into deeper water (below storm wave base) hemipelagic (limestone, marl and/or shale) rhythmites, representing the maximum of the postglacial flood.

Cap dolostones display idiosyncratic sedimentary features (Table 1) whose mutual stratigraphic relations are broadly consistent within and between cap dolostones in different regions (Allen and Hoffman, 2005; Hoffman et al., 2007). Here, we focus on one such structure, sheet-crack cements (Kennedy et al., 2001), which we observe to be consistently associated with an early fall in relative sea-level, preceding the major flood. We suggest that it is analogous to the rapid regression of early Holocene age in Greenland, which is attributed to the combined effects of glacial isostatic adjustments (rebound) and the weakening gravitational ‘pull’ on the adjacent ocean by the receding ice sheet (Clark, 1976). The gravitational effect is instantaneous, dictated by the rate and pattern of ice-sheet mass loss (Clark, 1976; Clark et al., 2002; Farrell and Clark, 1976; Tanner and Clark, 1976; Tappert and Clark, 1976). The early Holocene regression in Greenland occurred later, relative to the ‘glacioeustatic’ rise in global mean sea-level, than the early regressions in Marinoan cap dolostones in Namibia. This is because the palaeotropical Congo and Kalahari cratons (Fig. 1) were among the first in the world to lose their ice-sheets (Hoffman and Li, 2009). The idea that regional sea-level must rise and fall, by tens of meters, in response to the waxing and waning of nearby ice-sheets is not new (Penck, 1882, cited in Jameson, 1882).

First, we document the stratigraphic association of sheet-crack cements with early regression in Marinoan cap dolostones on the Congo and Kalahari cratons of Namibia. The sedimentary facies and pattern of $^{13}C_{\text{carb}}$ variation in the Keilberg cap dolostone of the Otavi Group (Congo craton) has been described previously (Hoffman et al., 2007) and only a brief recap is needed. The Dreigratberg (new name) cap dolostone of the Port Nolloth Group in the Gariep Belt (Kalahari craton) is described here for the first time. It was previously correlated with the Bloeddrif Member of the Holgat Formation, which is the cap

Table 1

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<tr>
<th>Sedimentary structure</th>
<th>Description</th>
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<tr>
<td>Sea-floor barite</td>
<td>Roseate clusters of bladed barite crystals in cm-scale digitate masses with internal growth laminae. Interspaces are filled by laminated peloidal ferroan dolomite, intermittently bridged by barite</td>
<td>Sulfate/ferrous boundary in the water column (Hoffman and Schrag, 2002)</td>
<td>Methane cold seepage (Jiang et al., 2006)</td>
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<td>Giant wave ripples</td>
<td>Steep, highly aggradational megaripples with sharp, straight and parallel crestlines, and bidirectional laminae that intersect in the crestal region and coarsen crestward. Relief typically ~35 cm crest to trough; width ~150 cm crest to crest. Ripples set aggrade sigmoidally for up to 140 cm and terminate through onlapping</td>
<td>Growth faulting (Allen and Hoffman, 2005)</td>
<td>Organosedimentary growth form (Corsetti and Grotzinger, 2005)</td>
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<td>Tubestone stromatolite</td>
<td>Confluent, meter-scale, domal or corrugate stromatolites, hosting synsedimentary, cm-scale tubes, oriented palaevorosally (geoplumb), filled by meniscus laminated dolomicrite and/or void-filling cements</td>
<td>Gas or fluid escape (Cloud et al., 1974)</td>
<td>Dissolution and flowage (Kenny, 1996)</td>
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<td>Low-angle crossbedding</td>
<td>Laminated, normal and reverse graded, peloidal grainstone with meter-scale, low-angle toplaps, downlaps and onlaps</td>
<td>Storm waves (James et al., 2001)</td>
<td>Pore-fluid overpressures (Corkeron, 2007)</td>
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<td>Sheet-crack cements</td>
<td>Bedding-parallel (rarely perpendicular) extension cracks, filled by fibrous, isopachous dolosparr and locally late drusy quartz</td>
<td>Pore-fluid overpressures (Corkeron, 2007)</td>
<td>Methane cold seeps (Kennedy et al., 2001)</td>
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limestone above the older (Sturtian) Numees glacigenic diamictite (Frimmel, 2008). The description here complements and extends previous observations and correlations (Macdonald et al., in press). Finally, we speculate on a causal mechanism that links early regression with the development of sheet-crack cements in cap dolostones.

2. Sheet-crack cements and associated intrastratal folds

Sheet-cracks are planar openings, commonly buckled, that accommodate extension perpendicular, rarely parallel, to bedding. Their typical aspect ratio of ~0.02 (width/length) is supported by isopachous cement, precipitated as linings of constant-thickness, which are composed of fibrous dolomite oriented normal to the crack walls and local, late-stage, drusy quartz (Figs. 2a, b, 3a). Sheet-crack cements are characteristic of but not unique to Marinoan cap dolostones (Bertrand-Sarfati et al., 1997; Corkeron, 2007; Edwards, 1984; Hoffman et al., 2007; Jiang et al., 2006; Kennedy, 1996; Macdonald et al., in press; McCoy et al., 2006; Nédélec et al., 2007; Plummer, 1978; Shields, 2005; Sumner, 2002), where they are typically confined to a meter-thick zone near the base of the cap dolostone (Corkeron, 2007; Kennedy, 1996; Kennedy et al., 2001). Volumetric expansion, both normal and parallel to bedding, is required to accommodate densely-developed sheet-crack cements. This is expressed in the form of variably-oriented intrastratal buckles and folds (Figs. 2a, b, 3a), the severity of which increases directly with the volumetric fraction of cement. The combination of intrastratal folds and isopachous cements invites comparison with peritidal ‘tepee’ structures (Assereto and Kendall, 1977), but sheet-crack cements in cap dolostones formed

Fig. 2. Lithofacies of Marinoan (635 Ma) cap dolostones on the distal foreslope of the Otavi platform in northern Namibia (a–c) and on the western margin of the Kalahari craton in southern Namibia (d–f). (a) Sheet-crack cements composed of fibrous isopachous dolomite (white) and late drusy quartz (orange stain), with associated intrastratal folds, basal Keilberg cap dolostone, Garettes pos (section P6540). (b) Micropeloidal dolostone (yellow) with isopachous sheet-crack cements (white) at the crest of an intrastratal anticline, Fransfontein (section P7002). (c) Terminal Bethanis Member of the glacigenic Ghaub Formation, conformably overlain by basal turbidites and sheet-crack cements of the Keilberg cap dolostone, Narachaams se pos (section A, P1607, in Fig. 4). (d) Namaskluff diamictite (ND) overlain by channelized white limestone (DC) and buff micropeloidal dolostone (DCd) of the Dreigratberg member, looking west from the top of the escarpment above Namaskluff Camp. (e) Clasts of Palaeoproterozoic basement rocks (arrows) at base of Dreigratberg member above Namaskluff Camp. Note also the sheet-crack cement and buckling of beds developed to the left of the hammer. (f) Giant wave ripples and elongate stromatolite (S) above Namaskluff Camp.
the sea-shelf was not well insulated by an ice sheet. One dif

Kennedy et al., 2001). The existence of permafrost assumes that the
during the glacial sea-level lowstand (Jiang et al., 2003, 2006;
hydrates presumed to have developed in organic-rich shelf sediments

oscillatory traction across the ripple crest (Allen and Hoffman, 2005).

laminae interdigitate in the crestal region (Fig. 3b) as a result of
ripples (Table 1), which lack void-filling cement and in which the
laminae interdigitate in the crestal region (Fig. 3b) as a result of
oscillatory traction across the ripple crest (Allen and Hoffman, 2005).

We are aware of only two explanations for sheet-crack cements in
Marinoan cap dolostones. The first is that they represent cold seeps on
the sea-floor, caused by flood-induced destabilization of permafrost
hydrates presumed to have developed in organic-rich shelf sediments
during the glacial sea-level lowstand (Jiang et al., 2003, 2006;
Kennedy et al., 2001). The existence of permafrost assumes that the
shelf was not well insulated by an ice sheet. One difficulty with this
model is the extreme paucity of cement with very depleted $\delta^{13}C$
values (Jiang et al., 2003; Wang et al., 2008), which are diagnostic of
modern and ancient cold seep carbonate cements (Kauffman et al.,
1996; Michaelis et al., 2002). In the Keilberg cap dolostone of the
Congo craton, the $\delta^{13}C$ and $\delta^{18}O$ values for isopachous cements and
eocoeexisting sediments are statistically indistinguishable (Supplemen-
tary Fig. S1). A second difficulty is the lateral continuity of the meter-

thick zone of sheet-crack cements. We observed no discrete center in
tens of kilometers of continuous outcrop section, nor are the under-
lying strata or the base of the cap dolostone disturbed as would be
expected if methane had erupted from below (Hoffman et al., 2007).

Taking a different approach, Corkeron (2007) attributes sheet-

cracking and brecciation to pore-fluid overpressures, caused by
rapid carbonate sedimentation and differential burial compaction of
stratigraphically underlying mud, now shale. We are attracted to
pore-fluid overpressures as a means of jacking open sheet-cracks, but
place the ultimate cause of the overpressures elsewhere.

3. Otavi platform, Congo craton, northern Namibia

The Otavi Group is a 770–590 Ma carbonate platform exposed in the
Damaran (590–530 Ma) Otavi fold belt, which rims the (present)

southwestern promontory of the Congo craton. Two Cryogenian
 glaciations are represented in the Otavi Group, the older Chuos
Formation and the ca 635 Ma Ghaub Formation (Halverson et al.,
2005; Hoffman and Halverson, 2008; Hoffman et al., 1998; Hoffmann
and Prave, 1996; Hoffmann et al., 2004;). The Ghaub Formation forms
a marine ice grounding-zone wedge situated on the distal part of the
south-facing foreslope of the Otavi platform (Domack and Hoffman, in
press; Hoffman, 2005, 2010). Glacial deposits are absent on the upper
foreslope and outer platform, 5 km seaward and 50 km landward of
the slope break respectively, but a discontinuous veneer of lodgement
tillite is found on the erosionally-deepened inner platform. A
syndeglacial cap dolostone, the Keilberg Member of the Maieberg
Formation (Hoffmann and Prave, 1996), conformably overlies the
Ghaub Formation or the contiguous glacial erosion surface (Hoffman
et al., 2007). The Keilberg Member (Fig. 4) is thicker on the upper

![](Fig. 3. Contrast between the crestal regions of (a) an intrastratal anticline associated with sheet-crack cements and (b) a giant wave ripple. Note the fibrous radial
dolosparite (clear) filling the folded sheet-cracks (a); and the interdigitation of
opposing laminae, crestward-coarsening of peloids and strongly aggradational
character of the wave ripple (b).

Fig. 4. Representative columnar sections with $\delta^{13}C_{\text{carb}}$ data of the Keilberg cap dolostone on the Otavi platform and its foreslope (modified after Hoffman et al., 2007). Columns are
arranged vertically to fit a sigmoidal $\delta^{13}C$ profile with time progressing upward on the Y-axis. The steep initial decline in $\delta^{13}C$ is captured in sections a–c, the long slow rise in d–f, and
the steep final decline in e–g (arrows at correlative inflections). Cap dolostone thickness variation controlled by tubestone stromatolite thickness. Greater thickness on the raised
outer platform (e), compared with the depressed inner platform, indicates higher accumulation rates, not greater accommodation space. $\delta^{13}C$ and $\delta^{18}O$ values for sheet-crack
cements and coexisting dolomite in sections a–c are plotted in Supplementary Fig. S1. 

S Distal Slope

Upper Slope

Outer Platform

Inner Platform

LEGEND

Maieberg Formation

pink marly limestone rhythmite

with sea-floor crystal fans

Keilberg Member cap dolostone

marly micropeloidal dololite turbidites

peloidal dolarenite, giant wave ripples

as above, swaley low-angle crossbedding

tubestone stromatolite dolostone

peloidal dolarenite, sheet-crack cements

marly micropeloidal dolostone turbidites

Bethanis Member of the Ghaub Formation

stratified proglacial detrital carbonate with ice-rafted debris

Ombaatjie Formation cycle b7-b8

peritidal cycles of dolostone ribbonite, stromatolite, grainstone
and microbialaminite)

$\delta^{13}C_{\text{carb}}$ sediment

$\delta^{13}C_{\text{carb}}$ sheet-crack cement

foreslope (~100 m) and outer platform (~75 m) where the Ghaub Formation is absent, and thinner on the inner platform (~25 m) and distal foreslope (~10 m). It forms the transgressive tract of the postglacial depositional sequence, the Maieberg Formation, and is everywhere conformably overlain by deeper water, marly limestone rhythmite of the maximum postglacial flood (Hoffman and Halverson, 2008; Hoffman and Schrag, 2002).

There are five basic carbonate lithofacies within the Keilberg cap dolostone (Fig. 4). All were deposited above prevailing wave base except for the carbonate turbidites, which mark the upper transition nearly everywhere and the base of the cap dolostone exclusively on the distal foreslope (Fig. 4). The regression (shoaling-upward transition) from basal turbidites into peloidal grainstone with low-angle cross-bedding is exceptional; it goes against the overall deepening-upward trend. The basal turbidites (Fig. 2c) are discrete, parallel-sided, graded units of dololutite, separated by shale partings. They thicken (0.5 to 2.0 cm), coarsen and amalgamate upward. The zone with sheet-crack cements consistently occurs directly above the turbidite-to-grainstone transition (Figs. 2c, 5). It is ~1.0 m thick and continues laterally for tens of kilometers parallel to depositional strike. Although spatially variable in the intensity of development, the cements do not form centered complexes, nor are the turbidites or the basal contact disturbed. The turbidites pinch out at the base of the upper foreslope, coincident with the disappearance sheet-crack cements (Fig. 4).

The question that emerges is, do sheet-crack cements systematically accompany regressive changes from basal turbidites to low-angle crossbedded grainstones? Kennedy (1996) documented a similar relationship in the lower part of the Marinoan cap dolostone in the Amadeus Basin of central Australia, but attributed the low-angle crossbedding (his Lithofacies II) to differential dissolution and semi-plastic slumping in a deepwater setting. We suggest that his graded, well-sorted macropeloids (Kennedy, 1996, Fig. 4b) accumulated above prevailing wave base. To further illustrate the coincidence of early regression and sheet-crack cements in cap dolostones, we next describe a new cap dolostone from a different craton in Namibia.

4. Gariep belt, Kalahari craton, southern Namibia

The Gariep belt is a Pan-African orogenic belt exposed on the western margin of the Kalahari craton in southwestern Namibia and northwestern South Africa (Stowe et al., 1984; Tankard et al., 1982). Folded strata in this late Ediacaran to Terraneuvian transpressional orogen (Davies and Coward, 1982) include the Port Nolloth Group (PNG), which formed as a consequence of rifting on the Kalahari margin of the Adamastor palaeocean (Frimmel, 2008), and the Nama Group, which was deposited in a foreland basin developed in response to collisions between the Kalahari, Congo and Rio de la Plata Cratons (Gresse and Germs, 1991; Grotzinger and Miller, 2008; Grotzinger et al., 1995). The PNG contains a pre-Sturtian diamictite (Kaigas Formation) of uncertain origin, and discrete Sturtian (Numees Formation) and Marinoan (Nama Group diamictite) glacigenic deposits (Macdonald et al., in press). The Nama Group diamictite is capped by the Dreigratberg member of the Holgat Formation, which contains geochemical and sedimentological features characteristic of Marinoan (basal Ediacaran) cap dolostones. Below we compare proximal sections of the Dreigratberg member that show evidence for upward-shallowing, with more distal sections that contain sheet-crack cements.

4.1. Inner shelf sections at Namaskluft Camp

On the escarpment above Namaskluft Camp (Fig. 5), the PNG fills a ~12-km-wide and 1-km-deep palaeo-valley that is incised into
crystalline basement (Fig. 6). This panel is separated from more internal, parautochthonous sections by crystalline basement (Fig. 5).

The palaeo-valley at Namaskluft Camp was cut into an uplifted rift-shoulder during the Sturtian glaciation (Macdonald et al., in press). It is lined by Sturtian-age glacial deposits (Numees Formation) and largely filled by allodapic carbonate and silicilastic turbidites (Wallekraal Formation), consisting predominantly of Bouma sequences Ta-b. During the Marinoan glaciation, a valley developed on the south-side of the palaeo-canyon, preferentially eroding the underlying sedimentary rocks (Fig. 6). The upper palaeo-valley is at least 3 km-wide and 200 m deep, with a cross-sectional morphology similar to Pleistocene sub-glacial tunnel valleys (Boyd et al., 1988; Van Dijke and Veldkamp, 1996). Tunnel valleys form under high glaciostatic pressures during rapid deglaciation (Van Dijke and Veldkamp, 1996). The lower stratified diamictic facies of the Namaskluft diamictite include plowed clasts, laminated muds with soft-sediment deformation of glaciectonic origin, graded sandstone with rare limestones, and planar bedding passing upwards to aggradational ripples, characteristic of upper flow regime (Macdonald et al., in press). This facies assemblage is characteristic of Pleistocene tunnel valleys (Eyles and McCabe, 1989). The massive diamictic facies at the top of the Namaskluft diamictite are interpreted to reflect a rainout till formed at the termination of glaciation. The overlying Dreigratberg cap carbonate is also channelized with a <30-m thick package of limestone turbidite (Fig. 2d) present between the Namaskluft diamictite and the cap dolostone (Fig. 7). This limestone body pinches out laterally and is succeeded by the buff-coloured dolostones of the Dreigratberg member, which contains giant wave ripples and tubestone stromatolites.

Fifteen detailed stratigraphic sections were measured through the Dreigratberg cap carbonate along the escarpment above Namaskluft Camp to track facies changes in a three dimensional framework (Fig. 7). Deposition begins with a channelized body of white to violet allodapic limestone with reduction spots (Fig. 2d) and shallows up to a fine-laminated micropeloidal dolomite with low-angle crossbedding. The limestone beds are graded and also contain thin green marl partings. Within 5 km to the northwest of its greatest thickness, the limestone turbidite interval thins to less than a meter. Cobble-sized limestones of metamorphic basement and carbonate are present in these thin turbidite beds on the margin of the channel (Fig. 2e), but were not found in the thickest bodies of the basal Dreigratberg limestone.

The limestone turbidites are succeeded gradationally by fine-laminated, buff-coloured, micropeloidal dolostone with low-angle crossbedding. Sheet-crack cements occur near the base of the dolostone (Fig. 2e), but are poorly developed and laterally discontinuous. Within 2 m of the base of the dolostone, giant wave ripples form off of planar surfaces. Wave ripples near the base of the dolomite have <30 cm of synoptic relief and wavelengths >2 m (Fig. 2f). Elongate stromatolites appear ~2 m higher in the section, preferentially nucleated on the crests of the giant wave ripples. Microbialites become more dominant upwards and modify the size of the wave ripples, decreasing their wavelength and regularity (Fig. 2f). Throughout the cap dolostone, both the wave ripples and the elongate stromatolites maintain a consistent orientation ~100° azimuth. Some of the wave ripples are refolded, but the orientation of these later tectonic folds is ~158° azimuth, distinct from the orientation of the elongate stromatolites and wave ripples. Higher in the section, the elongate stromatolites coalesce to form a massive bioherms with tube-stone structures (Corsetti and Grotzinger, 2005; Macdonald et al., in press). The bioherms are flooded by <300 m of pink to light grey allodapic limestone and siltstone with hummocky cross-stratification (upper Holgat Formation), variably truncated by the sub-Nama unconformity (Fig. 6).
4.2. Upper slope sections at Namaskluft Farm

The sections at Namaskluft Camp are separated from those at Namaskluft Farm by a basement high (Fig. 5). Along the southernmost exposures at Namaskluft Camp, the cap dolostone rests on basement, suggesting that this basement high persisted during deposition of the Dreigratberg member. At Namaskluft Farm, the Namaskluft diamictite is succeeded by less than a meter of green marl, a couple of meters of thin-laminated peloidal dolomite, then 

\[ N \sim 30 \text{ m of massive stromatolite bioherm with irregular cements} \]  

(Fig. 8). This is overlain by an ~50 m thick transgressive sequence of folded pink limestone rhythmite and an additional ~50 m of mixed allodapic carbonate and siliciclastic rocks of the upper Holgat Formation.

4.3. Distal sections at Dreigratberg

In deeper water sections ~20 km to the southwest of Namaskluft Camp at Dreigratberg, the Neoproterozoic stratigraphy is very condensed and consists predominantly of allodapic carbonate and argillite (Fig. 8). The Namaskluft diamictite is stratified and contains rare, subrounded carbonate and sandstone cobble dropstones that pierce the siltstone matrix lamination. The diamictite is sharply capped by 1.6 m of graded beds of allodapic limestone, which weathers white but is blue when fresh. These beds are interpreted as turbidites. They are succeeded by 0.4 m of dolostone with cements that are parallel to sedimentary bedding and buckled upwards into pseudo-tepees. These sheet-crack cements are isopachous, fibrous, and consist of dolospar. Unlike the giant wave ripples, the pseudo-teepee structures associated with the sheet-crack cements show no preferred orientation. The Dreigratberg member continues upwards with an additional ~50 m of pink limestone, characteristic of the upper portion of the Dreigratberg cap carbonate.

Sheet-crack cements are also present in distal sections of the Dreigratberg member in the Dolomite Peaks area of South Africa (Macdonald et al., in press).

4.4. Chemostratigraphy

Samples were collected from measured stratigraphic sections for \( \delta^{13}C \) and \( \delta^{18}O \) analyses. They were processed and analyzed using standard laboratory procedures (described in detail in Halverson et al., 2004).

The Dreigratberg member at Namaskluft Camp, Namaskluft Farm, and Dreigratberg all display \( \delta^{13}C \) profiles with values beginning near \(-1\%\), decreasing sharply in the lower couple of meters, and then increasing upwards through the most of the cap dolostone before plunging to the most depleted values at the top (Fig. 8). Values of \( \delta^{13}C \) at Namaskluft Farm are consistently offset by \(+1\%\) from those at Namaskluft Camp (Fig. 8).

4.5. Base-level fall

The Neoproterozoic palaeotopography on the escarpment above Namaskluft Camp affords a unique window into the relationship...
between base-level and facies change. Typically, the sedimentary facies of the Marinoan cap dolostone are in a characteristic order, progressing in a transgressive sequence up-section from a micro-peloidal dolomite with low-angle crossbedding, to tubestone stromatolites, to giant wave ripples, and culminating with crystal fans precipitated at a dolomite–limestone transition (Hoffman et al., 2007), although one or more of these features are commonly missing from any individual outcrop. However, unlike cap carbonates elsewhere, along the escarpment, a channelized body of algalic limestone is at the base, and is succeeded by a micropeloidal dolostone. The limestone hosts rare ice-rafted debris, marking the retreat of the ice-line. Within the overlying dolostone, giant wave ripples are always below or interbedded with elongate gutter stromatolites that progress upwards to tubestone stromatolites. Elongate stromatolites grow preferentially on the crests of giant wave ripples and coalesce upwards to form tubestone stromatolites. These facies patterns describe a high-stand tract prior to a transgression, and the characteristic sequence of sedimentary structures is inverted.

Carbon isotope chemostatigraphy suggests that the high-stand tract on the escarpment can be correlated with the condensed section at Dreigratberg. That is, the progression from channelized limestone turbidite to giant wave ripples and tubestone stromatolites on the escarpment is equivalent to the condensed turbidite to sheet-cement succession at Dreigratberg, both recording a fall in relative sea-level prior to the maximum flood at the top of the Dreigratberg member.

5. Discussion

5.1. Variable expressions of early regression

There are significant differences as well as obvious similarities between the Dreigratberg cap dolostone and its equivalent on the Otavi platform, the Keilberg Member (Figs. 4 and 8). In both areas, the cap dolostone is thinnest on the distal slope, thickest on the upper slope, and intermediate in thickness on the platform. In both areas, it features an overall sigmoidal $\delta^{13}C$ profile composed of three stages—a steep early decline, a long gradual rise, and a steep final descent. In both areas, $\delta^{13}C$ values through the gradual rise are consistently higher on the slope than on the inner shelf/platform. And in both areas, rifting had been active before, during and after the Sturtian glaciation, but the Marinoan glaciation encountered young passive margins undergoing regional subsidence of presumed thermal origin (Halverson et al., 2002; Macdonald et al., in press).

Hoffman et al. (2007) demonstrated that the Keilberg Member was strictly deposited within the flooding stage associated with global ice-sheet melt-down. On the assumption that this occurred rapidly because of positive climate feedbacks (e.g., ice-elevation, ice-albedo and greenhouse-gas feedbacks), they concluded that the observed change in $\delta^{13}C$ of $\sim 4\%$ was far too large to reflect change in the isotopic composition of seawater, giving the long residence time of C in seawater (100 s of kyr) with elevated atmospheric $pCO_2$ (Bao et al., 2008). They postulated that the changes in $\delta^{13}C$ reflect a dominant role for temperature-dependent $CO_2$ (gas)–$CO_2$ isotopic fractionation at low pH ($<7.2$). Accordingly, the overall upward decline in $\delta^{13}C$ reflects strong warming as ice-sheets receded, lowering the planetary albedo by 0.3 (snowball Earth deglaciation). With insolation reduced by 6% relative to present, the albedo change amounts to a radiative forcing of nearly 100 W m$^{-2}$, of which $\sim 11$ W m$^2$ would be taken up in melting the global ice-sheets (Wallace and Hobbs, 1977, p. 320). Lower $\delta^{13}C$ values on the inner platform reflect warmer waters there during the gradual rise, compared to the slope. This reasoning may also apply to the Gariep belt, although the estimated horizontal temperature gradient across the Otavi platform of 0.1 $\degree$ C km$^{-1}$ (Hoffman et al., 2007) must have been $10 \times$ steeper to account for the telescoping of distance (Fig. 8). Alternatively, the lateral gradient could be a product of an isotopically-light carbon flux from the continent. Either way, we postulate that this gradient was maintained in part by a palaeotopographic basement high between the inner shelf and upper slope, above which the tubestone stromatolite of the Dreigratberg member was best developed, restricting circulation between the inner shelf and the open ocean.

Of the differences, four stand out. First, stratigraphic and isotopic variability is more localized in the Gariep belt. Changes that take place in 15 km across strike in the Gariep belt occur in 150 km across the Otavi platform (Figs. 4 and 8). This reflects more localized palaeotopography as discussed above. Second, compared with the Keilberg Member, the Dreigratberg sections are displaced ‘downward’, relative
to the sigmoidal $\delta^{13}C$ profile. The early steep decline is captured in all three sections in the Gariarp bel but only on the distal foreslope of the Otavi platform (Figs. 4 and 8). The final steep descent is recorded in dolostone in all sections on the Otavi platform, but in the Gariarp belt it only occurs in the limestone above the Dreigratberg member. This means that the Dreigratberg member began and ended earlier than the Keilberg Member, relative to the warming trend and the glacioeustatic flood. Third, the early regression recorded in the Dreigratberg member extends over a thicker stratigraphic interval compared with the Keilberg Member. The regressive segment is <1.0 m thick in the Keilberg (Fig. 2c) but up to 40 m thick in the Dreigratberg member (Fig. 8). This is consistent with the Dreigratberg cap being slightly older than the Keilberg, which only caught the tail end of the early regression in its oldest sections (distal foreslope). And fourth, the early regression occupies only part of the initial steep decline in $\delta^{13}C$ in the Keilberg Member, but outlasts it in the Dreigratberg member (Figs. 4 and 8). This implies that relative sea-level continued to fall in the Gariarp belt after it had stopped falling around the Otavi platform, assuming that the warming of surface waters (governing the $\delta^{13}C$ trajectory according to Hoffman et al., 2007) occurred simultaneously in both areas.

5.2. Early regression and sheet-crack cements

Why are sheet-crack cements closely associated with early regression? Before they were buckled, sheet-cracks opened vertically, parallel to the fibres of the cement, and against the force of gravity. This implies that pore-fluid pressure (pore pressure) exceeded the lithostatic pressure. When sedimentation occurs, the sediment acquires a pore pressure equal to the hydrostatic pressure. With burial, pore pressures increase on a trajectory intermediate between the hydrostatic and lithostatic pressure gradients (Fig. 9). This is referred to as pore-fluid overpressure. Overpressure allows sheet-cracks to form within the sediment. We propose that regionalized sea-level falls associated with the disappearance of ice-sheets lowered lithostatic pressures below the ambient pore pressure at shallow depths beneath the sediment–water interface (Fig. 9). Hydrostatic and lithostatic pressures drop by ~1 bar for every 10 m fall in sea-level. For this mechanism to be viable, the rate of sea-level fall must be rapid relative to the rate at which the sediment can depressurize by pore-fluid escape. Rapid cementation, which lowers permeability while increasing pore pressure, is an essential part of the proposed mechanism. The isopachous cements themselves provide undeniable evidence for rapid carbonate precipitation from pore waters synchronously with incremental sheet-crack opening. We discount remineralization of organic matter or methane hydrate as a driving force for overpressure because the $\delta^{13}C$ of the sheet-crack cement provides no support for an isotopically-light source of alkalinity (Supplementary Fig. S1). Our proposal provides a causal mechanism for the observed correlation of sheet-crack cements and early regression within cap dolostones.

5.3. Early regression and ice-sheet mass loss

What caused the early fall in relative sea-level? When a grounded ice-sheet forms near a coast, sea-level in its vicinity rises due to mutual gravitational attraction between the ice-sheet and the adjacent waters. At a glacial termination, sea-level in the vicinity of a melting ice-sheet will fall at a rate determined by the rate and distribution of ice-sheet mass loss (Clark, 1976; Farrell and Clark, 1976). The fall will be fastest where mass loss is concentrated at the periphery of an ice-sheet, causing the ice edge to retreat. By raising the land surface, glacioisostatic adjustments (GIA) also contribute to early regression but, as the time-scale for GIA is more prolonged than the recession of the Laurentide and Scandinavian ice-sheets, Marinoan GIA has been identified with ‘late regression’, at the stratigraphic tops of cap dolostones (Bertrand-Sarfati et al., 1997; James et al., 2001; Nogueira et al., 2003; Shields et al., 2007; Zhou et al., 2010). We propose that early regressions in Marinoan cap dolostones in central Australia (Kennedy, 1996) and on the Congo and Kalahari cratons, are primarily related to the instantaneous gravitational effect on regional sea-level of ice sheet mass loss, complemented by GIA.

How fast did Marinoan ice-sheets vanish? From a climate physics perspective, it is generally assumed that Marinoan deglaciation was very rapid (a few kys) because of the preponderance of ice-sheets at low palaeolatitudes (Fig. 1), combined with extreme atmospheric pCO$_2$ (Bao et al., 2008, 2009). In contrast, the existence of apparent geomagnetic excursions and reversals in cap dolostones implies a time-scale of tens to hundreds of kys if Marinoan geomagnetic field behavior was similar to the Cenozoic (Font et al., 2010; Raub and Evans, 2006; Trindade et al., 2003). This conflict remains unresolved, but we find the shorter time-scale more compatible with the sedimentology of cap dolostones (Table 1). If Marinoan ice-sheets were buttressed by ice-shelves or a continuous ‘sea-glacier’ (Goodman and Pierrehumbert, 2003; Pollard and Kasting, 2005; Warren et al., 2002), their removal would have triggered rapid ice-sheet drainage, as observed in modern outlet glaciers when confined ice-shelves are lost (De Angelis and Skvarca, 2003). The sea-glacier would be lost before the ice-sheet because it is thinner and has a lower surface elevation. Because the rise in global mean sea-level due to the melting of Marinoan ice-sheets globally was roughly ten times larger than the regional fall in relative sea-level due to the gravitational effect of local
ice-sheet mass loss, ice-sheets must have vanished in sequence, not in
unison, for the gravitational effect to reverse, ... in Namibia. GSA Today 8, 1–9.
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