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The Cryogenian Ghaub Formation of Namibia – New insights into Neoproterozoic glaciations

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ABSTRACT

The Neoproterozoic Cryogenian ('Marinoan') Ghaub Formation of northwestern Namibia represents an important founding pillar of the Snowball Earth hypothesis and its derivative, the Panglacial Earth hypothesis. These hypotheses assume oceans and continents covered by thick ice, even in the tropics, which caused a very distinct drop in eustatic sea-level. Over time, strongly increased CO_2 contents of the atmosphere led to sudden ice melting, very substantial sea-level rise, and strong weathering on the continents associated with the deposition of cap carbonates in the newly ice-free oceans. The ongoing controversy about Snowball-type glaciations in Namibia and elsewhere is reviewed, and other hypotheses (Slushball Earth, Waterbelt Earth, Jormungand state of the Earth, Thin Ice state of the Earth, Zipper-Rift Earth, High-Obliquity Earth) are discussed. We prefer the term 'Waterbelt Earth' instead of the originally proposed 'Waterbelt state' because of the clearer contrast with 'Snowball Earth'.

Because a great deal of information related to Cryogenian glaciations comes from the Ghaub Formation of northwestern Namibia, these hypotheses should be tested independently based on a time-equivalent depositional system. This analogue was found in the carbonate-dominated successions of the Otavi Mountainland (OML), northeastern Namibia, and is highly comparable with the successions in the well-investigated northwest of the country. An extreme eustatic sea-level drop caused by a global glaciation of oceans and continents and imposed on a carbonate platform or ramp such as the one in the OML would have led either to glacial cover or widespread subaerial exposure and extensive erosion, including deeply incised valleys. The presence of such features would strongly support the Snowball Earth hypotheses if tectonic effects did not play a major role. During the postglacial transgression, distinct reworking of the carbonate platform/ramp surface would have occurred, leaving behind lag deposits, as well as infills of incised valleys with fluvial, reworked glacial, and marine deposits. The main objective of our research was to weigh and investigate the strengths and weaknesses of the proposed Snowball Earth model of glacially induced large-amplitude sea-level changes during Ghaub time and to compare different models to obtain a rough estimate of the amount of glaciation.

The study area in the OML includes two different, age-equivalent facies realms: platform sedimentation in the Southern area without diamictites, and slope deposits, including Ghaub diamictites, in the Northern area. The southern, continuously shallow-marine area shows a shallowing-upward succession from the pre-glacial lower Auros Formation, often varve-like laminated shales formed below wave base, to metre-high columnar stromatolites and microbial mat-related carbonates with intervals of vertical tubes (degassing features) of the upper Auros Formation, overlain by cap carbonates of the Maieberg Formation. The columnar stromatolites and the microbial tubestone lithotypes were clearly deposited in the euphotic zone. Indications for tidal conditions or subaerial exposure were not recorded in this platform succession without unconformities. Neither dropstones, nor incised channels, nor transgressive lag deposits were observed. The facies changes from below storm wave base to the photic zone and finally a shallow subtidal zone is explained by a prolonged, modest sea-level fall, partly counterbalanced by subsidence, followed by a slow transgression.

In contrast, coarse-grained sedimentary rocks (e.g., oolites, debrites) characterise the time-equivalent successions in the Northern area. Starting with laminated shales at the base, similar to the Southern area, the overlying redeposited oolites and breccias of the Auros Formation show distinct lateral and vertical inhomogeneities and thickness changes, which indicate long-lasting synsedimentary tectonism. The same phenomenon is observed in the overlying diamictites of the Ghaub Formation. Their variable clast content indicates erosion of a strongly uplifted local source area covered by a thick carbonate succession, which was downstripped to the crystalline basement. The prograding diamictite succession with repeatedly intercalated silt-stringers is interpreted as periglacial debris flows into a marine environment. Sparse striated clasts in the diamictites and very rare dropstones (much less common than in northwestern Namibia) are indicators of glaciations somewhere in the area. However, compared with other glacial sequences, e.g. Quaternary periglacial sediments at the forefront of continental ice, dropstones and striated clasts would be expected to be much more common and more uniformly distributed if the entire area was covered by melting continental ice, as proposed in the Snowball/Panglacial Earth scenario. In the Southern area, dropstones would be expected to occur on the flooded platforms/ramps as well even when diamictites are absent.

Both the relatively moderate sea-level change and the less common, irregular distribution of locally concentrated glacial rainouts provide strong evidence against the presence of a thick, laterally continuous ice cover over oceans and continents extending to equatorial areas. The oceans possibly corresponded to the scenario of a Waterbelt Earth or High-Obliquity Earth; evidence of open oceanic water exists, which would have enabled the continued evolution of biota. Glacial ice was present on tropical continents, but its occurrences may have been regional in patches, sourced from mountainous areas, and ice streams would have reached the oceans only locally, unrelated to a thick continental ice cover.

Keywords: Snowball Earth; Panglacial Earth; Slushball Earth; Waterbelt Earth; High-Obliquity Earth; Neoproterozoic climatic paradox; glaciation; sea-level change; synsedimentary tectonics; stromatolites; degassing structures; dropstones; diamictites; cap carbonates.

1 Introduction

Harland (1964, 2007) and Harland and Rudwick (1974) were pioneers in proposing low-latitude, late Proterozoic glaciations, but they were not the first authors to assume very widespread glacial cover for specific Proterozoic intervals, including in Namibia (e.g. Gevers, 1931; Le Roex, 1941). Kirschvink (1992) outlined the Snowball Earth hypothesis, which was subsequently refined by Hoffman et al. (e.g. 1998a, b). More recently, a distinction was made between the hypotheses of a hard Snowball (the 'classical' view) and a soft Snowball or Slushball (e.g. Cowen, 2001; Lewis et al., 2007; Yang et al., 2012a, b, c; Liu and Peltier, 2013). According to the (hard) Snowball hypothesis, the Earth's oceans became entirely frozen over, even at tropical latitudes, during long-lasting glacial episodes in the late Proterozoic Cryogenian (pre-Ediacaran) interval. Namibia was located at the margin of the Congo craton in a low latitude position. The equatorial settings of this and other "Snowball successions" were confirmed by plate tectonic reconstructions such as those of Gray et al. (2006), Trindade and Macouin (2007), Li et al. (2008), and Scotese (2009), which show the break-up and subsequent development of the supercontinent Rodinia.

Hoffman (2009) proposed the Panglacial hypothesis, which postulates that during Snowball periods the continents were also covered by thick ice. We will not discuss the possible trigger for the initiation of a Snowball Earth; among the reasons considered are astronomical influence, supervolcano eruption, reduction in the amount of greenhouse gases, or even biological evolution (Tziperman et al., 2011). A wealth of data regarding the assumed glaciations is based on analysis of the Ghaub Formation, especially of the Kaokoveld area and Fransfontein Ridge in northwestern Namibia (Fig. 1) (e.g. Hoffman et al., 1998a, b; Hoffman and Schrag, 2002); recently, the older Chuos Formation in the same area was also investigated in greater detail (e.g. Le Heron et al., 2012; Hoffman et al., 2017). The Chuos (Varianto) diamictites have typically been correlated with the 'Sturtian' glaciation of South Australia, whereas the younger Ghaub diamictites have been correlated with the 'Marinoan' (also called Elatina) glaciation, also from South Australia (Hoffmann and Prave, 1996; Hoffman et al., 1998a, b), although problems exist with this correlation (see 2.4).

Glacigenic sediments commonly consist of poorly sorted pebbles, cobbles, and larger clasts incorporated in a finer-grained matrix. These features are also characteristic of mass- and debris-flow deposits formed with or without glacial influence. Therefore, demonstrating or disproving a glacial origin of such diamictites is difficult, especially when facetted or striated clast are absent (as in the Fransfontein Ridge), or may have been re-deposited after a glacial transport. Even more difficult is to confirm the Snowball Earth hypothesis. Eyles and Januszcak (2004, 2007) and Eyles (2008) are among the authors who doubted the Snowball interpretation and interpreted the Ghaub succession as tectonically induced sediment gravity flows generated by rifting associated with the breakup of the supercontinent Rodinia.

The Snowball Earth hypothesis implies totally frozen oceans, whereas the Panglacial hypothesis refers to ice sheets over all continents. According to the original versions of these hypotheses, glaciation ended because of very strong volcanic CO_2 enrichment in the atmosphere, which built up during glaciation and reached a turning point for the climatic change. This incoming super-greenhouse setting resulted in (i) sudden melting of all ice, (ii) extreme sea-level rise, and (iii) extensive continental weathering, providing strong input of dissolved ions into the oceans. This phenomenon caused the chemical precipitation of 'cap carbonates', deposited on top of diamictites (Hoffman et al., 1998a, b). It has been suggested that deposition of cap carbonates was diachronous as sea-levels rose from the exceptionally low levels of the glacial period (Hoffman et al., 2007). The postulated strong volcanic CO_2 enrichment is necessary to resolve the 'Runaway Snowball Paradox' of Harland and Rudwick (1964); that is, the problem of how a Snowball stage could be terminated. Fairchild et al. (2016) described synglacial carbonates from Svalbard, which were interpreted to represent late stages of a panglaciation, characterised by a build-up of atmospheric CO_2 to high levels. Abbot et al. (2012) suggested that in addition to highly elevated values of CO_2, the presence of clouds and volcanic and continental dust would make it significantly easier to escape a Snowball than was previously thought. For the Namibian Ghaub glacial aftermath, Kasemann et al. (2010) inferred marked ocean acidification and elevated atmospheric CO_2 concentrations from boron isotope measurements. In 2014, Kasemann et al. published Ca and Mg isotope data from the same area in Namibia investigated earlier. The post-Ghaub continental weathering flux transitioned from mixed carbonate and silicate to silicate dominated, together with dolomite formation in the cap dolostones. Thus, acidification ended promptly and ocean chemistry returned to a higher pH.

Regarding the duration of the ice ages and deposition of cap carbonates, in addition to still-sparse radiometric age data (see 2.4), there are estimates available from the multitude of observed magnetic reversals (with rates of the recent past applied to the late Proterozoic). The Elatina ('Marinoan') glaciation was estimated to have lasted several tens of thousands to hundreds of thousands of years (Schmidt and Williams, 1995; Sohl et al., 1999); the duration of deposition of the cap dolostone was assumed to be > 10^5 yr (Trindade et al., 2003; Kilner et al., 2005; Schmidt et al., 2009). The Sturtian lasted much longer and persisted for around 100 Myr (Kendall et al., 2009).

Because of the many arguments against the occurrence of a hard Snowball Earth, not the least being the continued evolution of life (see 2.1; also Christie-Blick et al., 1999; Kerr, 2000; Allen and Etienne, 2008), other conceptual models have also been suggested that modify this hypothesis or contrast with it (Slushball Earth, Waterbelt Earth, Jormungand state of the Earth, Thin Ice state of the Earth, Zipper-Rift Earth, High-Obliquity Earth) to explain the enigmatic diamictite deposits and associated facies (e.g. Williams, 1975; McKay, 2000; Runnegar, 2000; Hyde et al., 2000; Cowen, 2001; Eyles and Januszczak, 2004, 2007; Pollard and Kasting, 2005; Harland, 2007; Allen 2007; Fairchild and Kennedy, 2007; Williams, 2008; Williams et al., 2016). Schmidt and Williams (1995) coined the more neutral term 'Neoproterozoic climatic paradox' to address the different diamictite intervals intercalated within apparently warm-water carbonates at tropical latitudes without specifying the preferred genetic hypothesis already in the terminology.

The Slushball Earth hypothesis, developed by Richard Cowen (2001) and modified by many authors thereafter (e.g. Lewis et al., 2007; Yang et al., 2012a, b, c; Liu and Peltier, 2013), proposes that oceans were not completely frozen during extreme glaciation. No statement was made on the conditions on land. According to this hypothesis, in addition to thicker ice covering most oceans, a thin, watery layer of ice (slush) draped the equatorial oceanic areas, or was even absent. This distribution would have enabled photosynthetic organisms to survive and evolve. It also would have allowed the hydrological cycle to continue in many areas, which is corroborated by observations discussed in chapter 2.5, and indicates rapidly moving ice, cyclicity, and interglacial periods. McKay (2000) discussed, and Pollard and Kasting (2005) modelled, a 'Thin Ice solution' for Snowball Earth, which would have allowed the survival of biota and would be much easier to reconcile with the existing phylogenetic and fossil records. This Thin Ice hypothesis (critically discussed by Warren and Brandt, 2006) still explains the presence of cap carbonates and implies much more modest sea-level change. With the assumption of a Thin-Ice state (Pollard and Kasting, 2005) or Slushball-like scenarios for the Cryogenian glaciations (Hyde et al., 2000; Peltier et al., 2007; Abbot et al., 2011; Yang et al., 2012a, b, c), these models for Neoproterozoic glaciations could also deglaciate at sufficiently low CO_2 concentrations (Abbot et al., 2012). Pierrehumbert et al. (2011) coined the more appropriate term 'Waterbelt state' instead of Slushball for an Earth 'where sea ice and active land ice sheets reach the tropics, but in which substantial areas of open tropical ocean remain'. We prefer the term 'Waterbelt Earth' instead of 'Waterbelt state' because of the clearer contrast with the term 'Snowball Earth'. Rose (2015) modelled such a scenario with a stable 'Waterbelt' in the equatorial realm caused by convergence of ocean heat transport at tropical ice edges.

Abbot et al. (2011) proposed the term 'Jormungand' state, complementary to the hard Snowball, soft Snowball/Slushball and Thin-Ice hypotheses. This term denotes a stable, ice-free equatorial fringe, which is achieved in modelling when the marine ice rim first propagates toward the equator but then stabilises at a certain position in the subtropical zone. This halt in propagation is the result of the relatively low top-of-atmosphere albedo of the low-latitude, snow-free (bare) sea ice and relatively lower cloud cover. The term 'Jormungand' state is used in modelling because it generally cannot be verified in field studies.

A completely different proposal is the Zipper Rift Earth hypothesis (Eyles and Januszczak, 2004), which suggests a relationship between regional (not global) glaciations and pulses of supercontinental breakup. The rift-related tectonic uplift formed high plateaus where glaciers may have developed, as well as new oceans with shallow volcanic ridges. A link would therefore exist between glaciations, the amounts of eroded, redistributed, and deposited sediment, and the timing of rifting related to the break-up of Rodinia and Laurentia. Continental rifting and associated subsidence could have produced the accommodation necessary for deposition of the often fairly thick sedimentary successions without involving a large eustatic sea-level rise caused by extensive and rapid melting of ice. Banded iron formations (BIFs) are locally associated with some Sturtian glacial records and were interpreted by earlier authors as evidence for global ice cover. BIF's are, however, absent from many

"Snowball" successions, but occur either stratigraphically below such deposits or are post-depositional (Eyles and Januszczak, 2004). The limited record and extent of these Neoproterozoic banded iron deposits (in contrast to the massive and extensive Palaeoproterozoic BIFs) may indicate that they were deposited in restricted rift basins, where heavy input of Fe ions and anoxic bottom water would have favoured precipitation of iron (Eyles and Januszczak, 2004).

Williams (1975, and thereafter) and Williams et al. (2016) listed observations of different facies from Australia and elsewhere that indicate strong seasonal oscillations of the climate, which is not possible in equatorial regions today, although the equatorial position of South Australia in particular is well established by palaeomagnetism. The authors describe glacier-free continental regions with large, permafrost sand wedges stacked atop each other, which indicate a mean monthly air temperature range of 40ºC. The specific climate is further corroborated by non-tidal seasonal (annual) oscillations of sea-level showing a sawtooth pattern of the 'monthly inequality' of spring-tidal height, explained by an eccentric lunar orbit. These and other observations are interpreted to represent a high-obliquity position of the Earth during the Proterozoic (including the Palaeoproterozoic) low-latitude glaciations (Williams et al., 2016). This scenario was also discussed and modelled by other authors, including Jenkins (2000, 2004). In contrast to large seasonal changes in temperature over continents, high obliquity would cause only moderate seasonal changes in ocean temperatures. At high latitudes, the limited snow that forms during the arid winter melts during the very hot summer. High obliquity alone would not have been the primary cause of glaciation, but would have influenced the latitudinal distribution of glaciers (Williams et al., 2016).

The evidence for these changes of temperatures at low palaeolatitudes presents a major difficulty for the Snowball Earth, Slushball Earth and related hypotheses. The initiation of the high-obliquity position is assumed to be the result of a single, immense extraterrestrial impact at 4.5 Ga, the widely favoured mechanism to explain the origin of the moon. The specific high-obliquity position would then have lasted until after the Cryogenian, when it slowly changed to an obliquity approaching the present-day value. According to Williams et al. (2016), this change would have yielded a more habitable globe with greatly reduced seasonal stresses, which likely contributed to the evolutionary progress of life after the Cryogenian. The assumed high obliquity can explain specifics of the glaciations and the emplacement of glacial ice in equatorial regions, but it cannot explain either the onset of glaciations or the relatively sudden deglaciations. The assumed long-lasting change of the high obliquity position back to "normal" values poses problems. Although an early high obliquity is a plausible outcome of a major impact on the early Earth, no convincing mechanism has been identified to date to reduce high obliquity to the present setting (Williams et al., 2016). However, when Wegener (1912) proposed continental drift, it needed about 40 years until a proper mechanism was found to explain plate movements, and even longer until the theory was fully accepted. Williams (2008) proposed possible mechanisms for a change of obliquity between end of Cryognian and early Paleozoic, to permit Phanerozoic circum-polar glaciations, but the debate is ongoing. At the moment it might be more appropriate to search for arguments pro or contra high obliquity, instead of rejecting the theory because no mechanism is accepted to change axial tilt to the present position.

The thick ice cover of a Snowball and/or Panglacial Earth implies a very substantial eustatic sea-level fall. The magnitude of this change was estimated for the Ghaub Formation of western Namibia (Kaokoland, Fransfontein Ridge) as 500–800 m (Hoffman, 2005; Hoffman et al., 2007), or at least 1,250 m (Hoffman et al., 2006). Hoffman (2011a, b) and Domack and Hoffman (2011) gave a more extreme number and assumed a sea-level drop of more than 1,500 m. Two years later, Hoffman (2013) pointed at erroneous assumptions in the 2011 calculations and reduced the amount of sea-level fall to ≥ 500 m. Can this assumed sea-level fluctuation be corroborated elsewhere? We concentrated our sedimentological research on the Ghaub Formation and stratigraphic equivalents in the Otavi Mountainland (OML), about 300 km to the east of Fransfontein Ridge (Fig. 1). This area was fairly well mapped because of economically important occurrences of metallic ores at Tsumeb and elsewhere (Geological Survey of Namibia, 1999). Detailed field-work, including mapping and tracing of horizons, was carried out (e.g. Bechstädt et al., 2009) with the intent to confirm or disprove current theories on sea-level change during Neoproterozoic glaciations. Our work was focused on obtaining better data to evaluate this postulated strong sea-level change during the Ghaub interval, because this data may serve to validate or challenge the Snowball hypothesis.

Therefore, the emphasis of the present paper is to: (i) Evaluate the magnitude of sea-level changes by comparing data from the OML with those from the Fransfontein Ridge in particular. Carbonate platforms and/or ramps were widely developed in the OML, and the Ghaub Formation is present there, including the type locality, but diamictites do not occur everywhere. A major, eustatic sea-level drop associated with a worldwide Snowball glaciation would have caused a regional erosional unconformity, especially on top of the former shallow-water platform/ramp sediments, even where diamictites were not present. In addition, extensive dropstone horizons would be expected to have formed at the onset of melting. (ii) Investigate the eventual impact of synsedimentary tectonism on the strata. If the amount of relative sea-level change can be constrained, and if rift tectonics were minor, clues on the volume of ice and therefore the magnitude of worldwide glaciation could be drawn. However, if rift tectonism had been strong and caused unsteady subsidence, it would be more difficult to corroborate or challenge the Snowball/Panglacial Earth scenario because in such a case, the eustatic component of sea-level change can be difficult to separate from the tectonic component.

For more thorough reviews of previous research on the Neoproterozoic climatic paradox, recent results, and thought-provoking points of debate, the reader is referred to Fairchild and Kennedy (2007), Etienne et al. (2008), Allen and Etienne (2008), Pierrehumbert et al. (2011), Young (2013), Spence et al. (2016), Williams et al. (2016), and Brocks et al. (2017).

2 The Neoproterozoic climatic paradox

2.1 Biological evolution during Cryogenian glaciations

Different authors established a continuous record of microfossil life since the Palaeoproterozoic. By the end of the Mesoproterozoic, all general characteristics of prokaryotic and eukaryotic life had developed (Vidal and Knoll, 1982; Vidal and Moczydlowska-Vidal, 1997; Huntley et al., 2006; Waldbauer et al., 2009; Nagy et al., 2009; Brocks et al., 2017). The survival of these eukaryotes, particularly photosynthetic ones, poses a major challenge for the Snowball Earth hypothesis. Several million years of global ice-cover, as proposed by this hypothesis, would suggest the termination, or at least an extreme extinction, of life in the late Neoproterozoic (Hoffman and Schrag, 2002). The reported extreme decline in global microfossil diversity during the glacial intervals of the Cryogenian seems to support such a severe limitation of life (Zang and Walter, 1989; Knoll, 1994; Vidal and Moczydlowska-Vidal, 1997; Huntley et al., 2006; Brocks et al., 2017). This evolutionary 'bottleneck' is proposed to have triggered distinct biotic turnover during the Cryogenian leading to the evolution of Ediacaran biota, and even influencing the Cambrian Explosion (Hoffman et al., 1998b). Short-lived rapid diversification after the Cryogenian glaciation events seems to support this 'bottleneck' scenario (Huntley et al., 2006). Also of interest are the significant increases of steroid diversity and abundance in the interval between the 'Sturtian' and 'Marinoan' glaciations, which indicate the rapid rise of marine planktonic algae in the aftermath of the Sturtian because of the surge of nutrients supplied by deglaciation (Reinhard et al., 2017; Brocks et al., 2017).

However, palaeontological data from the pre-Sturtian Chuar Group in Arizona indicate that this biotic turnover had already occurred before the first low-latitude glaciation of the Cryogenian (Nagy et al., 2009; Porter and Riedmann, 2016), which clearly brings the 'bottleneck scenario' into question. Biogeochemical studies of the early evolution of life have shown continuous evolutionary lineages throughout the Proterozoic, with distinct diversification before the Cryogenian glaciations, and no abrupt rise in diversification thereafter (Javaux, 2007; Sánchez-Baracaldo, 2015). Several major photosynthetic clades, including green, red, and chromophyte algae, evolved prior to the first Cryogenian glaciation and survived these glacial intervals without major changes (Runnegar, 2000). These findings provide evidence of at least partially ice-free equatorial oceans during glaciation, as proposed in the Slushball Earth/Thin Ice Earth/Waterbelt Earth/High-Obliquity Earth scenarios, but clearly present a challenge to the inference of extreme extinctions during Cryogenian glaciations, as proposed in the (hard) Snowball Earth hypothesis.

According to data presented by Kendall et al. (2009), the Sturtian interval lasted from approx. 750–643 Ma, which implies that the glacial period was either long-lasting but diachronous, or that there were multiple episodes of glaciation (see 2.4). Despite this long interval, the microfossil record is sparse, and no detailed studies across the glacial interval(s) are available. In Namibia, early sponge-like fossils were reported from different stratigraphic levels, the oldest being pre-Sturtian at about 760 Ma (Brain et al., 2012). Well-preserved eukaryotes were found in the Chuar Group in Arizona (Nagy et al., 2009; Porter and Riedmann, 2016), as mentioned above, in deposits from before the Sturtian (Dehler et al., 2017). In addition, the Death Valley microbiota were documented from carbonates between glacial deposits (Corsetti et al., 2006), which conflicts with a proposed extreme extinction during the Sturtian glaciation. Other eukaryote remains were recorded from the aftermath of this glaciation (Nagy et al., 2009). Despite the Neoproterozoic low-latitude glacial events, the fossil record reveals little fundamental change between pre-, syn-, and post-glacial microbiota. If the Sturtian glaciation was diachronous (see above), affecting different continents at different times, survival of the biota poses even fewer problems.

For the Marinoan, detailed microfossil data were reported from pre- to postglacial successions in Australia and China. In Australia, low-diversity acritarch assemblages occur continuously from the Meso- to Neoproterozoic, but 'with some large gaps through the time of the Sturtian and Marinoan glaciations' (Grey and Calver, 2007), including a barren interval in the upper Marinoan diamictites and the cap dolomite. More recently, well-preserved eukaryotes were reported from siltstones intercalated with Marinoan glacial deposits in southern Australia (Le Heron et al., 2016). An important record was found in China, where prokaryotic and eukaryotic microbiota occur in the preglacial Datangpo Formation and the basal Nantuo tillites of the Marinoan glaciation (Zang, 1992). These deposits are followed by an interval with no microfossil record up to the postglacial Doushantuo Formation. A few metres above the cap dolomite, low-diversity assemblages first appear, followed by robust diversification of acritarchs tens of metres higher in the succession (Yin et al., 2007; Yin and Yuan, 2007; Zhou et al., 2007).

The diversity minima in the global occurrences of microfossils in the Cryogenian were partially interpreted as gaps in the fossil record with 'substantially different biota' before and after these gaps (Knoll et al., 2006), which would appear to support the (hard) Snowball Earth hypothesis. To date, many overview studies have based their interpretations of the Neoproterozoic fossil record on the Snowball Earth scenario (Huntley et al., 2006; Nagy et al., 2009; Brocks et al. 2017). Other studies, focused on detailed microfossil records, have clearly proposed a continuous evolution of microbial life in the Neoproterozoic, and denied any gaps or extreme extinctions. The strong similarities between pre- and postglacial microfossil assemblages provides evidence that life continuously developed throughout these glaciations, although few microfossils have been observed from within the glacial intervals thus far (Grey et al., 2003; Grey and Calver, 2007; Javaux, 2007; Moczydlowska, 2008). There is strong evidence that the concept of a glacial interval assumed to be microfossil-free represents an artefact of biased sampling and not a real gap of life. Most research has avoided lithologies thought to be unsuitable for palynology (e.g. synglacial diamictites and dolomites) and has been focused on sedimentary records that appear to be more suitable, such as fine-grained shales and micritic marlstones of pre- and postglacial times.

Another challenge for the Snowball Earth hypothesis is the survival of eukaryotic life for several million years in a globally ice-covered world. It was proposed that life survived Snowball glaciations in various refugia, such as brine channels, cracks in ice induced by tides, and meltwater ponds, as known from recent glacial environments (Hoffman and Schrag, 2002; Hoffman, 2009). However, the duration of recent glacial environments, and even more importantly, the interaction between glacial and non-glacial areas in the recent world, differ entirely from the Snowball Earth scenario. This makes use of the recent glacial system as template for assumed Snowball refugia a dubious exercise. Furthermore, the palaeontological record contradicts the survival of microbiota in such refugia for several million years. The well-known metabolic processes and life modes for the recorded prokaryotic and eukaryotic microbiota apply constraints that the environmental conditions must have been relatively stable and robust throughout the Cryogenian. The fossil evidence also requires that there were habitats with oxygenated marine waters in the photic zone, open oceans with an active hydrologic cycle, ice-free shelves, and access to the sea floor (Moczydlowska, 2008). The presence of planktonic eukaryotes (acritarchs and prasinophycean algae) in intercalations within the diamictites of the Sturtian and Marinoan glaciations (Corsetti et al., 2006; Zang, 1992; Grey et al., 2003, Le Heron et al., 2016) provides clear evidence for an active hydrological cycle and open oceans during the glaciation. All of this evidence strongly contradicts the (hard) Snowball Earth scenario, but is consistent with a ‘Waterbelt’ setting (Pierrehumbert et al., 2011) that contained ice-free oceans along the equator during the glaciations (Runnegar, 2000), related or unrelated with a High-Obliquity Earth (Williams, 1975; Jenkins, 2000, 2004; Williams et al., 2016). In contrast to large seasonal changes in temperature over continents, high obliquity would cause only moderate seasonal changes in ocean temperatures. Life was clearly affected by Neoproterozoic glaciations, but not severely damaged or extinct. During the glacial intervals of the Cryogenian, the oceans were at least partially ice-free, and biological evolution continued. Evolution was most likely stimulated in the warmer interglacial or post-glacial periods when the oceans became more nutrient-rich; eutrophication also occurred regionally before the Sturtian glaciation (Nagy et al., 2009) and long after the Cryogenian.

2.2 Modelling Cryogenian glacial scenarios

Computer modellers of Cryogenian glacial settings have employed a variety of models, including general circulation models (GCMs) developed to understand and predict climate evolution. The input variables of these models were deduced from geochemical proxies, and others from assumptions such as the nature of the frozen surface, its albedo (influenced by different factors), the amount of clouds, sea-ice dynamics, and oceanic heat transport. Pierrehumbert et al. (2011) wrote a word of caution in this respect, mentioning that the GCMs used are outside their intended ranges. In our opinion, geological factors have not been adequately considered in many of these models, which span a long time interval. This is especially the likely relationship between Cryogenian glaciations and plate tectonics, such as the fragmentation of Rodinia (Eyles and Januszczak, 2004), which caused increased production of oceanic crust, changes in subduction volcanism (increasing CO_2), ash content, and mountain building (affecting climate and the amount of erosion). The Sturtian interval is thought to have lasted more than 100 Myr (see 2.4); during such an extended interval, plate tectonic processes would have modified the Earth substantially, including ocean chemistry.

Most models assume the present obliquity of the ecliptic (present-day value 23.44º, ranging between approx. 22º and 24º over a cycle of 41 kyr), which controls the strength of the global seasonal cycle, climatic zonation, and Milankovitch cyclicity. With an obliquity of 65º, seasonal temperature ranges would exceed 50ºC for continental regions and 10ºC for the ocean at 30º latitude (Jenkins, 2000). A high obliquity would therefore cause only moderate seasonal changes in ocean temperatures, in contrast to large temperature changes over continents, because of the higher heat capacity of the ocean compared to that of land masses. With a high obliquity, evidence for large seasonal changes of temperature could be expected in the (limited) continental geological record, whereas indications of strong seasonality may not be prominent in the marine record (Jenkins, 2000; Williams et al., 2016). According to the latter authors, facies observations, especially for stacked m-scale primary sand wedges, formed in perennially frozen ground, point to such large temperature differences affecting terrestrial environments because of high obliquity.

Poulsen pointed out in 2003 that in the Snowball hypothesis, ice–albedo feedback is too simple and cannot be considered realistic because it does not account for enough parameters. Pollard and Kasting (2005) modelled a 'thin-ice' setting, which would allow the survival of biota. Lewis et al. (2007) modelled hard versus soft Snowball scenarios and found the latter only stable with a purely thermodynamic component, whereas it was not stable with dynamic and thermodynamic sea-ice components. Le Hir et al. (2008) used a numerical model of the carbon-alkalinity cycle. Even a very limited area of open water (10^3 km^2) allows efficient atmospheric CO_2 diffusion into the ocean, blocking any exit from Snowball Earth. In their modelling, even after the maximum estimated duration of the glaciation (30 Myr), the atmospheric CO_2 was far from reaching the minimum deglaciation threshold. If there is open water, the dissolution of the basaltic oceanic crust (see also Gernon et al., 2016) delays the rise in atmospheric CO_2 required by the Snowball hypothesis to reach the melting threshold to unacceptable time spans. The hard Snowball can be ended in the model only if no areas of open water are left. Additional plausible processes, for instance through an albedo decrease, are needed to lower the CO_2 threshold sufficiently to achieve deglaciation. Modelling by Pierrehumbert (2005) and Pierrehumbert et al. (2011) resulted in stable, long-lasting conditions, but a Snowball/Non-Snowball Earth bifurcation was more stable than a state with an ice-free 'Waterbelt' in the tropical oceans. The authors also mentioned that for initiation and deglaciation, the nature of the ice and snow surface must be understood in terms of its albedo and the hydrology of the surface meltwater. Williams et al. (2016) critically discussed some of the boundary conditions of this modelling. In the models of Voigt and Abbott (2012), sea-ice dynamics have a far larger effect on the CO_2 values critical for deglaciation than bare sea-ice albedo and ocean-heat transport. Hadley circulation leads to three mechanisms that stabilise and one that destabilises tropical sea-ice margins. Convergence of ocean heat transport at tropical ice edges stabilises, according to the model of Rose (2015), a Waterbelt Earth in the equatorial realm. The GCM modelled has four stable equilibria ranging from 0% to 100% ice cover, including a 'Waterbelt' state with tropical sea ice. This Waterbelt is stabilised against albedo feedback by intense but narrow wind-driven ocean overturning cells that deliver heat to the ice edges.

Le Hir et al. (2009) investigated what would occur in the glacial aftermath of a hard Snowball. If the CO_2 partial pressures were several hundred times greater than at present, the Snowball Earth hypothesis assumes extremely warm temperatures and a vigorous hydrological cycle. However, different scenarios modelled by the above-mentioned authors indicate that the hydrological cycle intensifies, but only moderately. Although in this limited amount of fluvial water the maximum discharge of dissolved elements was approximately 10 times greater than the modern flux, silicate weathering would take several Myr to reduce CO_2 levels to pre-glacial settings. In addition, continental weathering alone would not supply enough cations to account for the observed volumes of cap carbonates. Mills et al. (2011) revealed that over long time-scales, the availability of fresh rock becomes a limiting factor for silicate weathering rates. When this transport-determined limitation is incorporated into the model, which includes a simple ice–albedo feedback, the stabilisation time is tens of millions of years, and the model produces greenhouse–icehouse oscillations. According to Mills et al. (2011), the long gaps between Snowball glaciations may be explained by the mentioned limitations on silicate weathering rates. Fabre and Berger (2012) obtained different results. They modelled the formation of cap carbonates in glacial aftermath and postulated that the dissolution of weathered tillites (in contrast to solid rock) may contribute significantly to riverine ionic input, and may promote the deposition of thick cap carbonates over a short period of time. However, one problem to solve is whether these widespread, thick tillites were present at all. It is a puzzling question, why so few tillites are reported from the shelves and low-lying land areas.

Gernon et al. (2016) discussed an important additional factor. According to these authors, the alkalinity changes of the ocean necessary to form cap carbonates cannot be explained solely by increased terrestrial runoff associated with increased CO_2. An important factor is the supercontinental breakup of Rodinia. This tectonic process created submarine shallow volcanic ridges with large amounts of glassy hyaloclastites, which were altered to palagonite (see also the model of Le Hir et al., 2008, mentioned above). Gernon et al. (2016) estimated the associated fluxes of calcium, magnesium, phosphorus and silica, and argued that this process made an important contribution to changes in ocean chemistry during Snowball Earth glaciations. This alteration would lead to Ca^{2+} and Mg^{2+} supersaturation, which is thought to be sufficient to explain the volume of cap carbonates deposited. The model of Gernon et al. (2016) assumes a ‘hard’ Snowball with a thick ice-cover over the oceans. However, according to the comments of Fairchild (2016) in response to that publication, there is considerable evidence that larger areas of open water existed that were sufficient to permit exchange of CO_2 between the atmosphere and the oceans (see also Le Hir et al., 2008). Moreover, the continuation of biological evolution (see 2.1) invokes areas of open water. Nevertheless, the alteration of ocean chemistry described by Gernon et al. (2016) is an important factor because distinctly higher concentrations of Ca^{2+} and Mg^{2+} than those of the present may have occurred in Neoproterozoic intervals of rifting, even without glaciation. Another modelling attempt was performed by Yang et al. (2017) to deal with changes in ocean chemistry in the aftermath of a ‘hard’ Snowball. These authors modelled the time-scale of mixing oceanic water with meltwater, giving a best estimate of $\sim 5 \times 10^4$ yr, up to 100 times longer than that of the modern ocean, where it is in the range of 10^3 yr, a value used by Shields (2005) in his scenario. Liu et al. (2014) assumed a 10^4 yr mixing time from another geochemical model; Font et al. (2010) and Kasemann et al. (2014) supposed a duration of 10^5 yr based on geochemical modelling, sediment thicknesses and palaeomagnetic reversal frequencies. According to Yang et al. (2017), during this long interval, a meltwater plume floated on the oceanic water. Their model suggests that the cap dolostones formed predominantly in a freshwater environment (also postulated by Liu et al., 2014), not well-mixed with saline deep water, and deposition took a long time. Of course, these results would only be valid in the aftermath of a (hard) Snowball Earth with thick ice-cover; Yang et al. (2017) did not model other scenarios such as a Waterbelt Earth.

2.3 Evolution of the continental margin of the Congo Craton

Extensional tectonic activities are well known in the Neoproterozoic of Namibia, reported by workers such as Miller (1983, 2008), Martin et al. (1985), and Eyles and Januszczak (2007), who described local intrabasinal controls on sedimentation. According to an overview given by Miller (2008), continental rifting occurred in the late Neoproterozoic at about 750 Ma and was associated with distinct volcanism. More recently, based on correlations with Zaire and the Congo, the initiation of rifting was placed at around 880 Ma (Miller, 2013). Hoffman et al. (1996) gave a minimum age of 756 ± 2 Ma (obtained from zircon dating) for the Nosib Group terrigenous rift sedimentation. This long-lasting rifting dissected Rodinia and opened the Damara Ocean in a roughly NE–SW direction, and the Adamastor Ocean, a part of it represented by the Kaokoveld, aligned N–S (all directions are with respect to the present continental orientation) (Fig. 1). A former part of Rodinia, the Congo Craton, came into existence, north of the Damara and east of the Adamastor Ocean, whereas the Kalahari Craton formed on the southeastern side of the Damara Ocean. Gray et al. (2006), Trindade and Macouin (2007), Li et al. (2008), and Scotese (2009) are among the authors who gave general reconstructions of the arrangement of continents in the late Proterozoic.

The successions at the margin of the Congo Craton to the Damara and Adamastor oceans consist of crystalline basement overlain by very thick upper Proterozoic sediments of the Damara Supergroup (Miller, 1983, 2008). In the west, in the Kaokoveld area, these continent-marginal successions strike coast-parallel, almost N–S. Eastward, the rocks of interest are well exposed in a roughly ENE–WSW-trending belt, mainly outcropping in the Fransfontein Ridge (in the area of Khorixas) and the OML around Otavi, Tsumeb and Grootfontein (Fig. 1). This Proterozoic stratigraphic interval continues in the subsurface to the north (Owambo Basin) and east, where outcrops are found in Botswana and Zambia (Lufilian arch).

In the OML, on top of a crystalline basement (Grootfontein metamorphic complex), the Damara Supergroup (Fig. 2) starts with pebbly, continental clastics of the Nosib Group. At its top, the glacially influenced Chuos (Varianto) Formation follows. The succession is overlain by carbonates of the Otavi Group, which were deposited at low latitudes, represents a time span of roughly 200 Myr, and reaches a maximum thickness of about 5,000 m (after Söhnge, 1957; Beukes, 1986; Miller, 2013). These carbonates were said to represent a 'Bahama-type' relatively shallow-water tropical platform system (Hoffman et al., 1998a, b; Hoffman, 2005, 2011). However, it is unknown if this platform was detached from or attached to the Congo Craton in the north. The Otavi Group in the OML is subdivided into the Abenab Subgroup (Berg Aukas, Gauss, and Auros formations), and the overlying Tsumeb Subgroup (Ghaub, Maieberg, Elandshoek, and Hüttenberg formations) (Fig. 2). The clastic Mulden Group follows unconformably to paraconformably, with a great break, but without any conspicuous angular unconformity; it is interpreted to represent the foredeep or molasse stage (e.g. Martin and Porada, 1977; Halverson et al., 2002; Miller, 2013).

In the Fransfontein Ridge, the overall stratigraphic succession is lithologically comparable, but with important differences in the details (Fig. 3). The prominent Huab and Makalani ridges at the margin of the Fransfontein shelf succession to the open marine Outjo Basin (part of the Damara Basin) in the south have counterparts in the Grootfontein area of the OML (Fig. 4). In the Damara Basin, the Swakop Group occurs, which is the basinal, often turbiditic equivalent of the Otavi Group (e.g. Miller, 1983, 2008; Miller et al., 1983, 2010; Hoffmann, 1983). The Otavi succession was interpreted as a 'passive' continental margin (Dürr and Dingeldey, 1997; Hoffman et al., 1998a, b); the former authors emphasized, however, repeated phases of tensional tectonic activity during its evolution, the latter authors assumed a stable margin after late Abenab time (see 5.2). Halverson et al. (2002) and Hoffman et al. (2017) assigned the Otavi Group to the rift–drift transition and the following thermal-subsidence stage. According to these authors, a stable southern margin of the Congo craton was permanently established from middle Abenab time onward, which lasted until the Damara collision. According to Miller (2013), continental collision occurred at about 542 Ma and the post-tectonic peak of regional metamorphism was at about 535–530 Ma.

Prave (1996, 1997) proposed a much earlier start (middle Otavi time) of an active foreland basin setting. According to this author, evidence for this interpretation includes (i) an intra-Otavi angular unconformity, (ii) concomitant development of increased accommodation in the deep-basin and slope settings, which shows basin segmentation and backstepping depositional systems composed mostly of hemipelagic and gravity-flow deposits, and (iii) a near reversal in palaeocurrents. Furthermore, Nascimento et al. (2016) reported an unconformity within the Swakop Group of the Outjo basin, which records basin margin onlap and slope instability.

Despite their contrasting assumptions of a tectonically stable margin in Tsumeb time undergoing thermal subsidence only, Halverson et al. (2002) mentioned a tensional phase during the (glacial) Ghaub Formation affecting the Makalani Ridge, when block rotation also occurred in the basinal Outjo area farther to the south, which corresponded to the above-mentioned intra-Otavi/Swakop unconformity of Prave (1996, 1997) and Nascimento et al. (2016). Hoffman et al. (2017) described from the Vrede Dome area (Northern Zone) rift-related Chuos diamictites truncated by a thin Ghaub diamictite and followed by cap carbonates. Despite the unconformity at the base, these authors again assumed passive-margin subsidence for the Ghaub interval.

2.4 Stratigraphy of the diamictites

Early on, two separate diamictite horizons were differentiated in the Damara succession (Kröner and Rankama, 1972; Guj, 1974). Subsequently, it was assumed that there was only one glacigenic diamictite interval (e.g. Hedberg, 1979; SACS, 1980; Miller, 1997), called the Chuos Formation. It was presumed to represent a single, time-stratigraphic marker separating the Abenab from the Tsumeb Subgroup. The South African Committee for Stratigraphy (SACS, 1980) adopted this interpretation, which served as a key element for subsequent stratigraphic correlations. However, Prave and Hoffmann (1995), Hoffmann and Prave (1996) and Hoffman et al. (1998a, b) emphasised that two diamictite horizons existed: an older one, the Chuos Formation at the base of the Abenab Subgroup, and a younger one, the Ghaub Formation at the base of the Tsumeb Subgroup. The Chuos Formation at the type locality in the Chuos Mountains of the central Damara belt is equivalent to the Varianto Formation of SACS (1980), and is clearly older than the 'Otavi Tillite' in the OML, which Hoffman and Prave (1996) re-named Ghaub Formation. The type locality is at Ghaub Farm in the OML. When studying older literature, it is often unclear which of these two glacial intervals was referred to as Chuos.

The two intervals of Cryogenian climate perturbations in Namibia were also correlated with other continents. To overcome the paucity of radiometric age dating, chemostratigraphy (carbon and strontium isotopes) was employed for correlation (e.g. Kaufman et al., 1991, 1997; Halverson et al., 2005; MacDonald et al., 2010). The authors assumed isochronous and global glaciations based on correlatable chemostratigraphic trends, especially carbon isotopes. However, the $\delta^{13}C$ trends described by Sato et al. (2016) from South China confirmed that such trends differ regionally and are distinct from other areas in the world. Furthermore, Kasemann et al. (2014) found lateral differences in isotopic changes across the southern Congo shelf, which indicate local influences modifying and amplifying the global signal. Differences in carbon isotope values are also shown when the sections of Domack and Hoffman (2011) and of Hoffman et al. (2014) are compared with each other. Such differences must be considered, not only in modelling chemical weathering flux, but also in intercontinental chemostratigraphic correlations, which are often not robust because different rift basins may not have possessed uniform water chemistries, as discussed by Alan and Etienne (2008).

The Chuos (Varianto) diamictites have typically been correlated with the 'Sturtian', and the younger Ghaub diamictites with the 'Marinoan' (Hoffmann and Prave, 1996; Hoffman et al., 1998a, b). The type areas for the Sturtian and Marinoan (Elatina) are both located in South Australia, but there is discussion on the local stratigraphy and exact age assignments (Williams et al., 2008). A study by Kendall et al. (2009) highlights the problems of correlation when radiometric age dates are scarce or missing. Hoffman et al. (1996) published ages of 746 ± 2 Ma and 747 ± 2 Ma for the Naauwport rhyolites, which are said to represent a maximum age for the Chuos Formation. Nascimento et al. (2016) obtained a somewhat older minimum age of 757 ± 5 Ma for the base of the

Chuos Formation. Their measurements employed an ion microprobe using SHRIMP II and RG on a sample from a Naauwport dacite sill intercalated with the basal Chuos Formation. This age is very close to the minimum age of 756 ± 2 Ma given by Hoffman et al. (1996) for the lower Nosib Group terrigenous rift sedimentation. Therefore, the age of the Chuos Formation apparently is not synchronous everywhere, and differs greatly from the Sturtian ages for Australia (see below). Rifting in the area may have begun as early as about 880 Ma, if the correlations with Zambia and the Congo are robust (Miller, 2013).

In addition to zircon dating (e.g. Hoffmann et al., 2004; Condon et al., 2005), correlations based on Re–Os ages have been published. Kendall et al. (2006, 2009) discussed different age data, including data from Corsetti and Lorentz (2006), and measured Re–Os ages of 657.2 ± 5.4 Ma for Sturtian glacial deposits in South Australia, and 643.0 ± 2.4 Ma for the post-Sturtian lower Tindelpina shales. For Tasmania, Kendall et al. (2009) obtained an age of 640.7 ± 4.7 Ma from black shales immediately overlying 'Sturtian' diamictite-bearing successions. Both post-Sturtian Re–Os age data are statistically identical (within 2σ uncertainties). Rooney et al. (2015) assumed globally synchronous glaciations and presented four new Re–Os ages from different parts of the world. For the onset and demise of the long-lasting 'Sturtian' glacial epoch, a duration of almost 60 Myr, from approx. 717–660 Ma was suggested by these authors. According to the age data of Kendall et al. (2009), an even longer Sturtian interval may have existed, lasting more than 100 Myr, from approx. 750 to approx. 645 Ma. The above-mentioned ages of the Chuos Formation in Namibia (Hoffman et al., 1996; Nascimento et al., 2016) are at the beginning of this long time span; most of the ages reported from Australia closer to the end. Therefore, Kendall et al. (2009) discussed that the 'Sturtian' glacial period was either long-lasting, but diachronous, or that there were multiple episodes of glaciation. Le Heron (2012) and Le Heron et al. (2011a, b, 2012) found clear evidence in Namibia and South Australia for 'Sturtian' sea-ice-free sedimentation, including a Sturtian interglacial intercalated within glacial sediments. However, the Chuos successions in Namibia may be almost 100 Myr older than parts of the Sturtian in South Australia. Much of the Sturtian interval sensu Kendall et al. (2009) is represented in Namibia by shallow-water sediments of the Abenab Subgroup, deposited at a time when Sturtian diamictites were deposited elsewhere. If one assumes more than one ''Sturtian'' glaciation of different age in different parts of the world, these deposits may represent the products of the protracted breakup of the supercontinent Rodinia (sensu Eyles and Januszczak, 2004). If the 'Sturtian' glacial interval was diachronous across different continents and if 'Sturtian'-type diamictite–cap carbonate pairs cannot be used as chronostratigraphic marker horizons (Kendall et al., 2009), many global correlation problems would be solved. These problems exist because of the questionable assignment of many late Proterozoic glaciations to the Sturtian or the Marinoan group (if such groups exist at all).

For the 'Marinoan' (or Elatina) of the type area, Williams et al. (2008, 2011) gave maximum and minimum ages of approx. 640–580 Ma. The Namibian Ghaub Formation could be dated precisely. U–Pb zircon dating of ash layers within dropstone-bearing glaciomarine sediments in central Namibia gave an age of 635.5 ± 1.2 Ma (Hoffmann et al., 2004), which is largely the same as ages obtained in China from the base of the Doushantuo Formation overlying the Nantuo diamictites (632.5 ± 0.5 Ma, 635.5 ± 0.6 Ma: Condon et al., 2005). Other authors reported ages with larger uncertainties for the Doushantuo Formation (628.3 ± 5.8 Ma: Yin et al., 2005; 621 ± 7 Ma: Zhang et al., 2005). The correlation of the Namibian Ghaub Formation with the Marinoan of Australia has been corroborated by data from Tasmania, although robust ages of the Marinoan (Elatina) glaciation at the type localities are still missing. Calver et al. (2013) dated the topmost metre of the Cottons Breccia in Tasmania (at the transition to “Marinoan” cap carbonates) and received a predominant age of 636.41 ± 0.45 Ma from slightly recycled detrital zircons (older ages of stronger recycled zircons have been measured as well). Rooney et al. (2015) provided a Re–Os age for the termination of the Cryogenian glaciation in the Mackenzie Mountains of northwestern Canada with an age of 632.3 ± 5.9 Ma. Within uncertainty, this age is identical to the U–Pb zircon ages from China, Namibia and Australia mentioned above (Rooney et al., 2015). The Marinoan glaciation appears to have ended almost synchronously; however, the initiation time of the Ghaub–Marinoan–Nantuan glaciation remains unknown. Rooney et al. (2015) assumed only a short interlude between the two glaciations. This interval would be even shorter if one considers the above-mentioned Re–Os age directly at the top of a ‘Sturtian’ glacial deposit from Tasmania (640.7 ± 4.7 Ma: Kendall et al., 2009). If the Sturtian would be terminated at about 640 Ma and the Marinoan at approx. 635 Ma, the nonglacial *and* glacial period together would have lasted roughly 5 Myr, compared with the 10 to 20 times longer duration of the entire Sturtian, from which glacial and interglacial intervals were reported (Le Heron et al., 2011a, b, 2012).

Hofmann et al. (2014) (not to be confused with P.F. Hoffman) described four Neoproterozoic glacial records from southern Namibia and investigated these deposits based on extended single zircon grain analysis. The youngest obtained zircons indicate an age of approx. 1.0 Ga. The different levels of diamictites interpreted as glaciomarine can be separated according to the relative abundance of concordant Mesoproterozoic to Palaeoproterozoic zircons. None of the measured ages corresponds to the Cryogenian rifting events in southern Namibia. Therefore, the source area for these “Sturtian” glacial records was likely located some distance away, in the east of the studied area (Hofmann et al., 2014). These data are of special interest if we compare the assumed four glaciations with the long duration of the Sturtian reported from elsewhere.

According to Eyles (2008), the extensive Cryogenian glacial records represent either diachronous glaciations or discrete pulses of cooling between ∼750 and ∼580 Ma. Allen and Etienne (2008) critically discussed the problems of correlations from one rift basin to another. Furthermore, according to these authors, likely more than two glaciations, not global but regional, occurred in the Neoproterozoic. These events were characterised by a number of different glacial advances and retreats. Therefore, we ask: is the existence of two global glaciations (Sturtian and Marinoan) only wishful thinking? Of importance, however, is the largely synchronous end of the Ghaub–Marinoan–Nantuan glaciation on different cratons, followed by a long non-glacial period that persisted until the glaciations in the late Ediacaran Period.

2.5 Chuos and Ghaub diamictites: Debris flows or glacigenic deposits?

Glacigenic sediments and debris flows commonly consist of differently sized, unsorted clasts incorporated in a finer-grained matrix. Facetted and striated clasts indicate a glacial origin, but later, such clasts may have been reworked and incorporated into fluvioglacial and marine sediments. To assign a depositional process is even more problematic when such rocks are overprinted metamorphically, as is the case in many parts of the Damara Orogen, including greenschist metamorphosis in the Fransfontein area. Miller (1983) and Miller et al. (1983) interpreted thick successions in central Namibia as almost entirely turbiditic. However, according to these authors, the occurrence of rift-related sediments does not rule out glaciations. For instance, ice rafting of some glacial debris was suggested for the occurrence of a 1.5 m boulder encased in distal turbidites. These thick successions indicate a long-lasting turbiditic environment with some glacial influence. Badenhorst (1988) reported evidence for mass flows or reworked glacial origin for the bulk of coarser-grained diamictite deposits in the Karibib/Usakos area in central Namibia. Busfield and Le Heron (2013) describe from the Chuos Formation of the Otavi Mountainland macro- and micro-scaled soft-sediment deformation structures, e.g. pervasive glacitectonic lamination, which supports subglacial deformation under high and sustained porewater pressures. In contrast, soft-sediment structures indicative of mass movements are not present.

The Chuos diamictites are more strongly affected by rifting processes than are the Ghaub diamictites higher in the section; both can only be compared with caution. Because of the difficulty of interpreting such sediments as products of one process or the other, most authors belong to one of two camps, interpreting the diamictites differently:

(i) The diamictites, especially in northern Namibia (Chuos/Varianto and Ghaub Fms.), are tectonically induced mass and debris flows (e.g. Frets, 1969; Guj, 1970, 1974; Schermerhorn, 1974, 1983; Hedberg, 1975, 1979; Porada and Wittig, 1983a, b; Martin et al., 1985; Eyles and Januszczak, 2004, 2007; Bechstädt et al., 2009; Nascimento et al., 2016). Some of these authors allow additional glacial influence (e.g. Badenhorst 1988).
(ii) The successions are mainly glacigenic, although some authors emphasise additional tectonic influence (e.g. Gevers, 1931; Le Roex, 1941; Söhnge, 1957; Martin, 1964, 1965; Kröner and Rankama, 1972; Hoffmann, 1983, 1990; Henry et al., 1986; Kaufman et al., 1991, 1997; Hoffmann and Prave, 1996; Kennedy et al., 1998; Hoffman et al., 1998a, b; Hoffman and Schrag, 1999, 2000, 2002; Condon et al., 2002; Busfield and Le Heron, 2013). In addition, there is discussion of whether the Cryogenian glaciers were dry- or wet-based (see 2.5.3).

2.5.1 Are the diamictites mainly mass and debris flows?

Frets (1969) and Guj (1970, 1974) inferred a non-glacial, sediment gravity-flow origin for the diamictites. For Fransfontein Ridge, Frets (1969) favoured submarine landsliding caused by a fall in base level. In addition, Schermerhorn (1974, 1983) disagreed with a glacigenic origin for these diamictites, but also for other successions in different parts of the world interpreted by others as glacial in origin. He pointed out the absence of dropstone strata in central Namibia, and the association with carbonates in northern Namibia. Badenhorst (1988) pointed out, however, that this absence of dropstones, mentioned also by Martin et al. (1985), is based on incorrect correlations. Perhaps a more important point is the observed close relationship of the diamictites with tectonically active depositional environments, not only in Namibia but also in many other regions (Schermerhorn, 1983). The author also put special emphasis on the types of clasts; these are not exotic but locally derived intrabasinal clasts that mirror the underlying stratigraphy, atypical for glacigenic sediments. The clastic source points at uplift and erosion of the basin edge; the diamictites pinch out toward the stable borderland, source of the clastics, where locally restricted, tectonically induced glaciations might have occurred (Schermerhorn, 1983).

Martin (1964, 1965) initially argued in favour of a glaciomarine origin of the diamictites, but several years later, Martin et al. (1985) reinvestigated the successions, corroborated the interpretation of Frets (1969) and Schermerhorn (1974, 1983), and concluded that the diamictites were synrift deposits, submarine mass flows with no evidence of extensive glaciation. A major eustatic sea-level drop (possibly glacially induced because of a glaciation elsewhere) was assumed to have triggered submarine landslides in different parts of Namibia. Martin et al. (1985) argued that large clasts of carbonate could have been freighted over short distances by mass flows and not necessarily by icebergs, also because of the undoubted mass and debris flow origin of over- and underlying facies. Hedberg (1975, 1979) corroborated the importance of rift tectonics for this type of deposit. Porada and Wittig (1983a, b) investigated the northern rift of the Outjo Basin, located south of the Otavi platform. Here, sandstone successions, originally interpreted by Gevers (1931) as glacio-lacustrine varves, were now explained as graded calcareous turbidites (the formation processes of turbidites were not known in the 1930s); elsewhere, debris flows and siliciclastic turbidites occur in this evolving rift basin. Henry et al. (1986) described evidence for mass flow processes as well as glacial activity from the central zone of the Damara Orogen. Badenhorst (1988) measured different sections of the Chuos Formation in central Namibia and evaluated the lithology of clasts and matrix in diamictites and other interbedded clastic sediments. He interpreted the diamictites as mass flows, but also confirmed glacial activity and assumed a reworked glacial origin for the bulk of the clasts.

Eyles and Januszczak (2004, 2007) investigated the Ghaub and related facies successions in the Fransfontein Ridge in sedimentological detail for the first time. They regarded most of the Ghaub successions of the Fransfontein Ridge and the rocks underneath as water-laid debris flow deposits in a rift basin, and not directly caused by glaciation. The authors mentioned that breccia facies occurs repeatedly within the Abenab ‘preglacial’ and Ghaub ‘glacial’ rocks. These breccias are a key diagnostic indicator of the overall strong tectonic control on the basinal setting. Hoffman (2007) commented, not in the form of a discussion in the journal but on his website, on the publication of Eyles and Januszczak (2007) and mentioned some errors according to his personal opinion. Eyles (2008) discussed the relation between supercontinent cycles and glacial epochs. The extensive

Cryogenian glacial records may represent diachronous glaciations or discrete pulses of cooling. The deposits correspond in the overwhelming majority of cases to glacially influenced marine strata of substantial thicknesses (1 km and more) deposited in rift basins. Bechstädt et al. (2009) highlighted tectonic activity in the OML throughout the Ghaub Formation, as well as before and after the deposition of this formation. Different sub-basins were created within the shelfal area. Only weak indications for glacial activity at some distance, not in the area of Ghaub deposition itself, were found. No basal grooved or striated erosion surfaces occur, striated clasts are rare, and intercalated water-laid sedimentation has been observed (corresponding to the observations of Condon et al., 2002, in the area to the west). The changing clast content in the Ghaub diamictites of the OML indicates uplift and erosion on the scale of possibly thousands of metres, stripping the uplifted local source area down to the crystalline basement. No exotic clasts occur in the Ghaub Formation. After a probable glaciation in the hinterland, the relief was flooded.

Nascimento et al. (2016) interpreted the Nosib and Otavi/Swakop groups of the Outjo Basin as gravitational deposits related to rifting. The sediments of the latter two groups show a deepening upward trend from slope and proximal submarine fan successions to medium to distal fan deposits, divided by the above-mentioned unconformity. Locally, the whole succession incorporates large olistholithic blocks, explained as slide blocks caused by slope failure. The Chuos and Ghaub formations within this succession represent gravitational, subaqueous mass flows. Nascimento et al. (2016) mentioned that rudstone beds similar to those of the Chuos and Ghaub formations also occur in the Berg Aukas and Karibib formations (see also Eyles and Januszczak, 2007). The authors concluded that these might be glacial as well, if the Chuos and Ghaub formations would be glacial.

2.5.2 Are the diamictites mainly glacial deposits?

In the early history of research, the Namibian diamictites were declared as glacigenic, partly because sedimentology could not explain turbidites and mass flow/debris flow deposits properly. Their blanket-like nature, massive bedding, lack of sorting, and abundant matrix were emphasised as diagnostic for glacial deposits. Gevers (1931) was the first to describe diamictites from the Chuos Mountains in central Namibia as glacial products. In 1932, de Kock and Gevers used these glacigenic 'pebbly schists' as a correlation tool. Le Roex (1941) described diamictites from the area of Ghaub Farm and interpreted them as glacial ('Otavi Tillite'). The author pointed out the stratigraphic importance of the different occurrences of glacial diamictites. If the correlation of widely separated occurrences with one another would be correct, ice movements on a gigantic scale are indicated. Söhnge (1957) also took the poor sorting and variability within the 'Chuos Formation' as evidence for a glacial origin. Martin (1964, 1965) assumed that the Chuos and the Otavi tillite were contemporaneous and represented probable glaciomarine drifts. The author had some doubts, however, and stated, that a glacial origin cannot be proven with the same degree of conclusiveness as for other diamictites. Twenty years later, Martin et al. (1985) declared these diamictites as synrift deposits (see above). Kröner and Rankama (1972) interpreted the Chuos diamictites once more as glacigenic, chronostratigraphic markers.

Hoffmann (1983, 1990) suggested a continental to shallow-marine, glacial origin for the diamictites in the southern Damara belt, but strongly emphasised the overall tectonic sedimentary environment. Kaufman et al. (1991, 1997) mentioned separate glacial episodes in Namibia and elsewhere, and correlated the records of carbon and strontium isotope fluctuations on different continents with one another. Based on highly negative carbon isotope values, the authors suggested that unusually high rates of organic carbon burial facilitated glaciation by reducing atmospheric greenhouse capacity. Hoffmann and Prave (1996) revised the stratigraphy and described the differences between the strongly mixed, glacial Chuos diamictites and the mainly carbonate-clasts bearing glacial Ghaub diamictites. The authors also pointed out the presence of cap carbonates, which form a good regional marker horizon because they are present, even when no diamictite occurs. When Hoffman et al. (1998a, b) and Hoffman and Schrag (1999, 2000, 2002) wrote their classic, high-impact publications on ‘Snowball Earth’, as mentioned in the introduction, they also interpreted the Ghaub diamictites as glacial tillites. Kennedy et al. (1998) confirmed the existence of two glacial intervals in Namibia. These intervals were correlated chemostratigraphically (C and Sr isotopes) with the Sturtian and Marinoan of Australia and of other continents. More recently, Sato et al. (2016) were able to show distinct differences in carbon isotopy in the Three Gorges area of South China and other parts of the world, and discussed the problems of regional chemostratigraphic correlations, as highlighted earlier by other authors (e.g. Allan and Etienne, 2008).

2.5.3 Dry- or wet-based glaciers, glacial waxing and waning?

In the Cryogenian, continental glaciers may have reached the local sea at low latitudes, in Namibia (e.g. Hoffman, 2011b; Domack and Hoffman, 2011) and elsewhere (e.g. Schmidt and Williams, 1995; Sohl et al., 1999; Evans, 2000; Macdonald et al., 2010). Le Heron et al. (2012) described the infill of a large incised valley in northwestern Namibia, related to the Chuos Formation. Hoffman et al. (2017) reported a palaeovalley and erosional reliefs from the same region. According to Le Heron et al. (2012), Chuos diamictites with dropstones and intensely sheared zones of interpreted subglacial origin occur at the base of the outcrop. Dropstone-free mudstones follow, interpreted as interglacial. Chuos glacial deposits reappear above, and indicate glacial re-advance, followed by another, but more gradual retreat. The authors compared these sediments with Sturtian strata in South Australia (Le Heron et al., 2011 a, b), where evidence for sea-ice-free sedimentation had been established. The findings suggest that a Sturtian interglacial may have been extensive, which implies waxing and waning of ice sheets during these glacial events; the oceans were likely only partly frozen over in these areas during the early Cryogenian. However, the correlation of the Chuos Formation in Namibia with Sturtian successions in Australia is questionable; a strong difference in age is quite likely (see 2.4).

In Namibia and elsewhere, Condon et al. (2002) observed glacial rainout sediments, intervals of well-stratified detrital carbonates, which occur in various positions within the non-stratified diamictites of the Ghaub Formation. According to the authors, this finding indicates ice-line instability and glacial calving, which are incompatible with a Snowball Earth. Hoffman (2011b) and Domack and Hoffman (2011) interpreted the Ghaub sediments as late glacial, whereas the main "Snowball" glaciation was correlated with the base of the Ghaub formation (see below). Non-diamictic sediments were only rarely described from the neighbourhood of 'Marinoan' glaciations. Xiao et al. (2004) reported a probable Marinoan equivalent from the Tianschan, the Tereeken Formation, which consists of at least five diamictite units, tens to hundreds of metres thick, containing clear glaciogenic features. The diamictites are separated by siltstones with thin carbonate laminae, which consist of vertically oriented, upward-growing calcite crystals, 'inundated by overlying siltstone'. These observations were interpreted as another example of an active hydrological system, but more detailed data are not available. Fairchild et al. (2016) described, in terms of facies and geochemistry, an excellent example from Svalbard, assumed to be time-equivalent to the Ghaub Formation. The succession contains synglacial carbonates, which were compared with sediments of dry valleys in Antarctica and interpreted to represent late stages of a panglaciation, characterised by a build-up of atmospheric CO_2 to high levels. Benn et al. (2015) reported glacial–interglacial cycles, interpreted as Milankovitch cyclicity, at the top of glacial sediments from Svalbard, assumed to correlate with the Marinoan.

For clastic sediments, research methods are more limited than for carbonates, and facies analysis alone does often not allow unambiguous interpretation of the depositional setting. Past conditions of physical and chemical weathering can be inferred if comprehensive facies-analysis is combined with the Chemical Index of Alteration (CIA) (Young and Nesbitt, 1999; Young, 2001, 2002; Scheffler et al., 2003). CIA data are therefore instrumental not only in documenting changes between icehouse and greenhouse climates, but also in recognising shorter-term climate oscillations between glacial and warm-humid conditions. Sedimentological and CIA data sets from different 'Marinoan' successions of Namibia, Oman and China (Rieu et al., 2007; Bahlburg and Dobrzinski, 2011) give strong evidence of a functioning hydrological cycle, operative sediment routing systems, and variable climate conditions oscillating between dry-cool/glacial, and warm-humid/interglacial. The cold climate modes of the Cryogenian were cyclical, punctuated with well-defined warm-humid interglacial periods. The hydrological cycle and the routing of sediment were active throughout the glacial epoch, which requires substantial open ocean water. The prominent influence of physical weathering on the detrital silicate matrix of the diamictites, in addition to sedimentological features like the presence of dropstones in laminated facies in the Fransfontein Ridge (Domack and Hoffman, 2011), and to a much smaller degree also in the OML, strengthen the interpretation of a glacially connected origin of the Ghaub diamictites, affected by different episodes of glacial waxing and waning (Bahlburg and Dobrzinski, 2011).

Williams et al. (2008) reported different facies extending over a large area of South Australia, ranging from permafrost regolith and periglacial sand sheets on the Stuart Shelf in the west, through fluvial, deltaic and inner-marine shelf deposits in the western and central parts of the 'Adelaide

Geosyncline', to outer-marine shelf sediments in the north and southeast. The widespread and persistent rainout of fine-grained sediment and ice-rafted debris indicates that in the type area, the sea was not frozen over during the Marinoan (Elatina) glaciation. Sea-ice-free sedimentation and global-scale waxing and waning of ice sheets therefore existed during the early and late Cryogenian (Williams et al., 2008; Le Heron et al., 2011 a, b). Rose et al. (2013) described the Elatina Formation in the Adelaide Rift complex, which consists of pre-, syn-, and postglacial sediments, and incorporates ice-contact tillites, fluvio-glacial sandstones, dropstone intervals, and tidal rhythmites with combined-flow ripples and turbidites. Sedimentological and chemostratigraphic observations reveal the onset of the glaciation and advance of the ice sheet from land to sea, which created a heavily glaciated terrain that was incised down into underlying sediments.

In northern Africa and Australia, 'Marinoan' periglacial, metre-scale, stacked primary sandstone wedges have been reported that are identical to those forming today by subaerial contraction and cracking of permafrozen ground. These findings indicate high seasonal variations in temperature (e.g. Deynoux et al., 1989; Williams, 1986, 1994), which are not possible in low latitudes of the present Earth. A mean monthly air temperature range of 40°C and mean annual air temperatures of −20°C or lower is assumed (Williams et al., 2016). Ewing et al. (2014) present a somewhat different interpretation of the sandstone wedges (critically discussed by Williams et al., 2016), constrained the local mean annual surface temperature to within a few degrees of freezing, and speculated on whether these temperatures represent peak Marinoan glacial conditions (not supporting a persistent deep freeze of a hard Snowball Earth), or were due to glacial onset or deglaciation.

Data of Allen and Etienne (2008) from Oman indicate extended non-glacial, fluvial, deltaic and shallow-marine deposits alternating with thick (< 200 m) diamictites, interpreted as glacially transported debris dropping down near the ice grounding line. Such large-scale alternations require repeated advance and recession of ice and open marine water (Leather et al., 2002; Allen et al., 2004; Rieu et al., 2007; Allen, 2007; Allen and Etienne, 2008). As mentioned earlier for South Australia, the latter authors noted that in Oman, despite the assumed severity of the glaciations, the rocks deposited during the cold intervals indicate that some parts of the ocean must have remained ice-free, because dynamic glaciers and ice streams continued to deliver large amounts of sediment to the open ocean, which would be incompatible with the hypothesis of a totally ice-covered Snowball Earth.

Hoffman (2005) described a major moraine near a palaeo-ice stream (see also Hoffman et al., 2014), a corridor of fast-flowing wet-based ice. In the present Antarctic and Greenland ice sheets, ice streams are said to be responsible for up to 90% of the total ice drainage. According to Hoffman (2005), climate modelling suggests that annual mean surface temperatures on tropical continents with frozen oceans in the Cryogenian would have been similar to present-day Antarctica. Therefore, the existence of Cryogenian ice streams is not incompatible with a frozen ocean (Hoffman, 2005). However, the author did not discuss the difference between the present-day situation with an ocean frozen over only around Antarctica and the assumed Cryogenian "Snowball" situation with totally frozen oceans. Where should a wet-base ice stream flow to when the ocean is totally frozen over? Such a setting can only exist during deglaciation, and this is likely the reason for a more recently revised interpretation of the Fransfontein outcrops by the Hoffman group. Hoffman (2011a, b) and Domack and Hoffman (2011) interpreted the Ghaub Formation as deposits of an ice sheet grounded on the distal foreslope of the Fransfontein Ridge, down to about 1,500 m below the platform margin at the time of maximum ice cover. This interpretation was associated with extreme sea-level drop (the value was later revised, Hoffman, 2013; see 1 and 5.3). Domack and Hoffman (2011) now assumed that the Ghaub sediments were wet-based late glacial and contain sediments transported by different means and during eventual climatic oscillations, whereas the sub-Ghaub disconformity represents the glacial 'Snowball' maximum. According to these authors, melting of the tropical sea-ice buttress triggered ice-sheet attenuation, and retreat of grounding ice in line with accelerated drainage and calving of icebergs. Most of the Ghaub deposits were regarded as basal and deformation till, but also undermelt deposits and suspension products from meltwater, iceberg-rafted debris, and some sediment gravity flows. According to these authors, the main transgressive flooding event is represented by a fining-upward succession of assumed sediment gravity flow deposits, choked with dropstones (Bethanis Member), and in turn overlain by the cap carbonates. The whole succession was said to represent a very strong sea-level rise, which flooded the platform and caused below-wave-base sedimentation of the cap carbonates on the platform (Domack and Hoffman, 2011), despite the isostatic rise of the crust unloaded from a thick ice-cover. However, this modified interpretation is an argument against the classical Snowball hypothesis, with the assumption of sudden melting of thick ice and concomitant excessive weathering on the continents. If the 'Snowball' period would be represented by the basal unconformity, without sediments present, how could one prove the (hard) Snowball hypothesis at all? Allan and Etienne (2008) already noted that observations and interpretations of this type damage the Snowball Earth hypothesis. According to Hoffman (2011b), 'counter arguments [against Snowball Earth] based on temperate-type glacial sedimentology fail to grasp, that the preserved glacial sedimentary record reflects the end of the Snowball Earth, when melting was bound to emerge triumphant'. However, if (almost) all the recorded sediments are formed because of melting, why is the record of the encroaching glaciers so rare? This modified interpretation also devalues chemostratigraphic correlations, because deglaciation with advances and retreats is hardly correlatable from one continent to another. If the Ghaub and other 'Marinoan' sediments record only the end of a Snowball period, without sedimentary rocks representing the main glaciation or glacial advance, the estimation of the amount and the rate of sea-level change before, and after the assumed main glaciation becomes crucial.

3 Methods

This paper is mainly based on detailed field work, mapping the area and tracing horizons, measuring stratigraphic sections, and examining rock slabs and thin sections. Because of the presence of relatively fine crystalline dolomites, rock slabs are often more diagnostic than thin sections.

Our studies were focused on the successions outcropping on Jakkals Omuramba Farm (incorporating also the former Keilberg Farm and the Maieberg Farm) in the central OML, as well as on Auros and Gauss farms to the south (Figs. 4, 5). The succession from the crystalline basement to the Maieberg and Elandshoek formations crops out, including the Ghaub Formation. The above-mentioned area contains a large anticline, which was mapped, and more than ten longer sections were measured with different levels of detail, with horizons traced laterally (for details see Werner, 2005; Rittersbacher, 2008; Schweisfurth, 2009; Bechstädt et al., 2009). In addition, we investigated cores from the former Khusib Springs mine and from the OKS 1 well (Otjikoto gold mine) in the Outjo Basin, about 50 km south of the OML; these cores contain different levels of diamictites.

4 Stratigraphic succession

In the central OML, there are distinct, persistent facies differences between a 'Northern area' (northern part of Keilberg Farm) and a 'Southern area' (southern Keilberg Farm, as well as Auros and Gauss farms farther south). Both areas contain mainly carbonate sediments, but in the north diamictites also occur, which are not present in the south (Figs. 5–7).

4.1 Gauss and Auros formations

Here, we address only briefly the several-hundred-metre-thick Gauss Formation to better understand the facies development underneath the Auros Formation, which is of greater interest in the context of this study. In the Southern area, the Gauss Formation is dolomitic throughout, often rich in carbonate micrite, and (i) frequently contains 'colloform' structures in its lower parts. Above this basal unit, (ii) an oolitic interval follows, often with large ooids several millimetres in diameter, and then (iii) an upper unit about ten metres thick with different types of well-developed stromatolites and oolites. The stromatolites are characterised by a gradual transition from stromatolitic mats into silicified, mound-shaped forms. In places, these forms were redeposited, as indicated by their broken nature and chaotic superposition. Silicification generally occurred in both the Southern and Northern areas. In the latter, dolomitic wackestones are abundant, stromatolitic intercalations scarce to absent; several intervals with large ooids occur, and some shale intercalations are found at the base and top.

The lower Auros Formation follows conformably. It consists in the Southern area of finely laminated basal shales, which typically do not outcrop well because of their clay content, but are readily recognisable because of the sparse bush cover. The shales contain high amounts of silt. In the Northern area, the Auros shale interval includes intercalated dark, often organic-rich dolomites and many thinner dolomite interbeds. Some of these beds contain microlaminites.

4.1.1 Upper Auros Formation in the Northern area

The upper Auros Formation thickens from east to west, from about 130 to 350 m, and shows distinct lateral and vertical facies differences, with no unconformities. In the western (Fig. 6) and central parts of the working area, the formation consists partly of dark, finely crystalline (formerly micritic), locally varve-like laminated carbonates, similar to intercalations in the Auros shales underneath. The laminated carbonates are mainly limestones, but limestone/dolomite intercalations occur locally. These interbeds are mostly centimetres thick and coarsely crystalline.

Oolite intervals (Fig. 8) are very common, and are centimetres to several metres in thickness. The ooids are often well sorted and graded. Locally, especially above the Auros shales, solid hydrocarbons fill the interstices between ooids. The oolites are partially to strongly silicified, usually in the form of irregular streaks and bands. In the dolomites and within the silicified zones, the ooids maintain their original spherical form, whereas in local calcitic parts, the ooids are mostly strongly elongated and obliterated, affected by the stress field of the Damara Orogen. The obliteration is mainly caused by grain-boundary migration that results in an oblique deformation pattern (Laukamp, 2006).

Eastward, at a kilometre scale, these laminite–oolite successions partially change to thickly bedded carbonates that contain several coarse, matrix-rich breccia intervals with angular carbonate clasts, often a decimetre or greater in size. The clastic successions have no clear boundary with the overlying Ghaub diamictites, and the amount of breccia beds increases further near the Keilberg Farm boundary, until covered by recent debris.

4.1.2 Upper Auros Formation in the Southern area

Columnar stromatolites

The upper Auros Formation above the Auros shales is dolomitic and about 50 to 100 m thick. No unconformity occurs in between the Auros shales and the overlying Auros carbonates, or within the upper Auros carbonates themselves. The dolomites on top of the shales start with a roughly parallel to slightly wavy laminated succession less than 1 m thick, with interlayers 2–10 cm thick of silicified microbial mats. A dolomitic interval up to 40 m thick follows immediately above, and contains often strongly silicified, columnar stromatolites (Fig. 9). The columns extend nearly orthogonal to bedding, have an average diameter of 10 to 30 cm, commonly reach heights of 1–2 m, and are circular and densely packed (cm to dm distances). Krüger (1969) reported heights of more than 7 m even, although it is difficult to follow the same column in two dimensions over such a distance. Joint hulls occasionally attach neighbouring, thinner columns before these columns may separate again. Because of the common irregular silicification, the columns and their circular growth bands typically weather out well; when not silicified, the stromatolitic growth fabrics become less apparent. Very rarely, debris from stromatolitic columns was found between upright columns. Parts broken off from comparable stromatolites (Auros Formation?), were only very scarcely found reworked in the Ghaub diamictite of the Northern area. Less than one-metre-thick lenses of siltstone to fine sandstones occur occasionally, but cannot be followed laterally. At the top of the unit, the columns become not only smaller in height and diameter but are also less common, and the fabrics then slowly change to wavy, laminated microbialites, the so-called tubestones.

Tubestone facies

This overlying interval of microbial laminated dolomites is 10 to about 40 m thick. The transition from the underlying columns to the laminated fabrics is conformable, and no erosional or other unconformity has been found during our fieldwork, neither at the base, nor at the top of, nor within the laminated tubestone interval. The microbial laminations are sub-parallel to bedding, mm- to cm-scale, and are partially silicified. In a fresh cut, the rock shows different shades of grey, in contrast to the typically dark grey to black columnar stromatolites underneath. Interstices between the laminae are 2–10 mm thick. In outcrop the laminations are only rudimentarily observable if the appearance is not enhanced by silicification. Sheet cracks are common (Fig. 10A, B); they are a few to more than ten centimetres in length, and are mostly filled by drusy quartz and/or chert. Some smaller, open or partly calcite-filled vugs of 1–3 mm in length also occur in the otherwise homogeneous, medium-grained, recrystallised microbial bindstone, which internally is mostly packstone.

The most peculiar feature are irregularly silicified tube-like fabrics (Figs. 10, 11), which give the name to this facies. The former cavities inside the tubes are filled with coarse, blocky dolomite or, more often, quartz or chert, which also partially replace the microbial fabrics. The tubes contain many small, weathered iron minerals; the content of the latter diminishes toward the top of the tubestone interval. No fibrous, radial carbonate crystals have been observed on the walls of the tubes, in contrast to the Gauss and Abenab formations, where such cements are very prominent. On the farm Abenab 707 in the eastern OML, galena, sphalerite and sparry calcite occasionally form the infill of tubes (Miller, 2008: p. 170). Because of strong silicification, the tubestone rocks often form steep walls several metres in height. This interval was called 'quartz cluster dolomite' (Söhnge, 1957) and has been used as a regional marker horizon at the top of the Auros Formation (Beukes, 1986; King, 1994). Hoffmann and Prave (1996) and subsequent authors integrated the tubestone facies into the Keilberg Member ('cap carbonates') of the overlying Maieberg Formation. We prefer the former attribution to the Auros Formation because no distinct, easily mappable boundary exists in our study area to separate the interval with columnar stromatolites from the one with tubestones, whereas the overlying 'classical' cap carbonates consist of a distinctly different, laterally traceable facies (see below).

The tubular facies occurs only where the Ghaub diamictites are absent or extremely thin. In the Kaokoland on the Hoanib shelf near Ombaatjie (Hoffman, 2002), tubestones occur in the vicinity of only centimetre-thin Ghaub Formation. Profile 35 of the Fransfontein sections of Domack and Hoffman (2011) shows 'geoplumb tubestone stromatolite', which overlies a very thin Ghaub Formation that pinches out to the shelf facies of the Huab Ridge farther east, where thick tubestones occur (Hoffman et al., 2014). In the OML, the tubular structures, commonly more than ten centimetres in length, are well exposed in the Southern area, but also on different farms farther to the east. The tubes are mostly approximately orthogonal to bedding, as indicated by the almost planar microbial laminations between the tubes, which may have concave upward contacts with the outer walls of the tubes. Hoffman et al. (2009) and Domack and Hoffman (2011) described laminites cut by 'geoplumb' tubes, which are not at right angles with one another (see 5.3); this feature was originally taken as an indicator of the slope angle, but Hoffman (2013) later attributed it to tectonic stress. In outcrop, the tubes are centimetre-scale in width, exhibit irregular diameters along their lengths with thickening and thinning (Fig. 10C), and deflect laterally. The tubes frequently start and terminate at sheet cracks (Fig. 10A, B). At their upper ends, some of the tubes show funnel-like features (Fig. 10B). It is unclear if the latter fabrics are typically present; their absence may be related to the two-dimensionality of the outcrops, which is likely also responsible for the above-mentioned thickening and thinning. Other tubes are straight and possess uniform thicknesses (Fig 11A).

When viewed on the bedding planes, the sub-rounded tubes are densely packed with typical distances of about 10 cm, but also occur much more closely spaced. Generally, the silicified tubes protrude from a few millimetres to about one centimetre above bedding planes (Fig. 11B) because of their greater resistance to weathering. In cross-section, the tubes are often approximately ring-shaped, with a small millimetre-sized hole in the middle. The tubes do not exhibit apparent radial and circumferential symmetry, in contrast to the internal fabrics of the columnar stromatolites underneath (Fig. 9).

The tubes are evenly distributed all over the dolomitic rocks, but occur in three intervals, separated by two layers nearly without tubes (Fig. 7). In western Gauss Farm (northeastern part of the Southern area) and around the border fence between Auros and Gauss farms, these tube structures are very common, but are on average shorter than they are elsewhere (between three and five centimetres in length). Rock surfaces parallel to the laminations show abundant small knobs of quartz with interspaces of approximately 1.5 cm and diameters between 0.7 and 1 cm.

Transition zone to Cap carbonates

The tubes diminish upward in number and finally completely disappear within a microbially laminated transition zone between the tubestone facies underneath and the 'cap carbonates' above. This interval has strongly variable thicknesses, from almost non-existent (approximately 1 km WNW of the Auros farmhouse) to about 20 m (about 2.5 km south of the Jakkals Omuramba farmhouse, just south of Pt. 1775; Rittersbacher, 2008). The transition zone consists of more evenly laminated, slightly wavy, much finer crystalline carbonates than the tubestones underneath. The lamination is often barely visible, bed thickness is about 10 cm, and the rock is mostly light grey, lighter in colour than the interval below. The rocks of the transition zone contain almost no vugs or open pore spaces. Silicification is mostly absent in this fine- to medium-grained, recrystallised microbial bindstone (internally a packstone). On the northwestern Gauss farm, a few solitary, silicified stromatolitic cups with diameters of more than 10 cm occur underneath the cap carbonates, with the low relief infilled by massive to slightly bedded packstone of this unit. The single stromatolites are rare on bed surfaces, about one per five m^2 or less. Again, no unconformity has been found, neither at the base, nor within this transitional interval, nor at its top. The presence of this transition zone, which is distinctly different from the 'classical' cap carbonates above, is one of the reasons why we place the tube interval into the Auros Formation.

4.2 Ghaub Formation

The maximum thickness of the Ghaub Formation is 200 m in the central OML (Beukes, 1986) and 1,000 m in the Otavi area to the west (Söhnge, 1957). This formation is present in the Northern area, where we measured a maximum thickness of approx. 170 m, and absent in the Southern area. The diamictites thicken westward and thin out eastward to only few metres, and disappear under Quaternary cover in a valley east of the Jakkals Omuramba farmhouse, near the eastern boundary of Keilberg Farm. South of the OML, the Ghaub Formation is present in the subsurface, at least locally. In the OKS 1 well (Fig. 7), located in the Otjikoto gold mine area, Outjo Basin (about 50 km southeast of Otavi), two distinct diamictite intervals within the coeval Swakop Group occur: a 10-m-thick lower unit and a 40-m-thick upper unit, separated by about 17 m of shale. The petrology of the clasts is highly variable.

4.2.1 Diamictite matrix and clasts

The diamictites are grain-supported and consist of unsorted larger clasts of centimetre–decimetre, rarely up to metre in size (Martin et al., 1985; Bechstädt et al., 2009), mostly firmly encased in the diamictite ‘matrix’, which is formed to a large degree by very small, millimetre- and sub-millimetre-sized clasts. Much of this ‘matrix’ consists of finely reworked, small grains of dolomitic material, with mica and crystalline debris as well in the west. The rounding of clasts shows a wide vertical and lateral spread, from angular to well rounded. Most of the carbonate clasts are not sorted and are angular–subangular (Fig. 12A), in contrast to the mostly well-rounded sandstone, quartzite, and crystalline clasts (Fig. 12B). This difference in rounding was mentioned by Martin et al. (1985), and can be explained by the reworking of already well-rounded components from the Nosib Group, which does not contain carbonate components but sandstones, quartzites, and clasts from metamorphic or plutonic basement. The crystalline basement itself was also most likely a source of clasts.

In the OML, much of the Ghaub Formation weathers to a rusty-brown colour because of its high to very high pyrite content. It is therefore easily distinguishable in the field (Fig. 12C), because of contrast with the under- and overlying lithologies, as well as the sparse plant cover. Ghaub diamictites without or with only little pyrite occur in stratigraphically lower parts of the succession in the east and contain mostly reworked fragments of the Auros Formation. These parts are difficult to differentiate from the Auros carbonate rocks, and contain many breccia intervals within grey dolomites. The definition of the boundary between the Auros and Ghaub formations in these successions is ambiguous.

The amount and variety of the components (e.g. carbonate, quartz, detrital mica, and crystalline components) are characteristic of different intervals and show distinct trends along strike (see 4.2.2 and Bechstädt et al., 2009). The overall ratio of matrix to clasts is roughly 2:3. We cannot corroborate

Martin et al. (1985), who specifically reported, in contrast to Le Roex (1941) and Söhnge (1957), a lack of facetted and striated clasts. Sandstone and quartzite clasts rarely show facetted surfaces, which were not observed in the carbonate clasts. At least some of the facets may be explained by breaking apart along cleavage planes and fissures, with some rounding thereafter. In the OML, we were able to detect small amounts of striated clasts (see example in Fig. 12D), as was also reported by Hoffman (2007) from the same outcrops. Neither he nor we observed grooved or striated pavement underneath the Ghaub Formation, although it is so commonly present at the base of tillites in the Cenozoic, or, for instance, the Palaeozoic Dwyka. A well-known example of the latter is the sub-Dwyka Noitgedacht striated pavement near Kimberley (e.g. Master, 2012).

4.2.2 Lateral and vertical changes of clast zones

The Ghaub Formation shows a wide vertical and lateral variety in terms of composition and fabric (Figs. 12–13 and Bechstädt et al., 2009). Clasts size and lithology vary distinctly in relation to the clast contents and the amounts of intercalated fine, well-sorted, and partly graded clastics. Three zones with different clast contents were described from northern Keilberg Farm, which exhibit distinct lateral and vertical trends (Bechstädt et al., 2009, and Fig. 13). Clast Zone 1 contains clasts from the Abenab Subgroup only, inclusive of Auros oolites and rare stromatolites. The recrystallised neomorphic matrix is dolomitic. Clast sizes are up to 30 cm. Pyrite content is low. This clast zone can be further subdivided into three subzones (Fig. 13; for details see Bechstädt et al., 2009). In these subzones the percentage of clasts to matrix, the matrix colours, and the clast types differ greatly. In Zone 1a, about half of the clasts consist of silicified dolomite; in Zone 1c, this component comprises about one third. Zone 1b is distinctly more clast-rich than the other zones (which have more matrix) and also contains much less silicified dolomite than the other two zones. Because only carbonate clasts occur (see 4.1.1), the boundary between Clast Zone 1 and the underlying Auros Formation is debatable. The thin Clast Zone 2 has some Nosib sandstone/quartzite clasts incorporated, in addition to clasts from the Abenab Subgroup. Clast Zone 3 contains some (recycled?) clasts from the crystalline basement in addition to the still predominant Abenab and Nosib clasts (Fig. 13). In addition to the predominant dolomitic pseudosparite, detritic quartz, feldspar, mica, and some neomorphic sericite occur in the matrix, which are missing in the other clast zones. Zone 1 is restricted to eastern and central parts of the Northern area (sections 1 to 3), and comprises almost all of the Ghaub Formation. A metre-thick interval of Clast zone 3 occurs at the top of Clast zone 1 (Fig. 13), not described by Bechstädt et al. (2009). In section 3 in the central part, Clast zone 1 crops out in the basal third of the Ghaub succession, Clast zone 2 on its top forms a relatively thin interval. Clast zone 3, which also contains larger components, is found in the upper parts of section 3, and builds up the entire Ghaub Formation in section 4 and outcrops farther to the northwest.

The maximum clast size of the Abenab clasts is up to 1 m, but most of the large clasts are up to a few tens of centimetres. Most Nosib clasts consist of well-rounded quartzitic sandstones, size and rounding inherited from the Nosib source. Some clasts are facetted, however. The basement clasts are usually less well rounded and consist mainly of granites or diorites from the local Grootfontein basement complex. The matrix of Clast zone 3 is pyrite-rich (no banded iron has been found), which gives a characteristic rusty weathered colour to this sparsely vegetated diamictite interval.

4.2.3 Well-sorted clastic facies and potential dropstone intervals

Condon et al. (2002) reported glaciomarine rainout intervals from the Ghaub Formation of the Fransfontein area, and therefore questioned a (hard) Snowball setting. More recently, water-laid, well-sorted conglomerate and sandstone facies were described from below and within diamictites (Eyles and Januszczak, 2007; Domack and Hoffman, 2011). The latter authors reported often well-stratified detrital carbonates from different intervals at the base, within, and at the top of the Ghaub Formation, interpreted as undermelt deposits and suspension products from meltwater, as well as iceberg-rafted debris. In the OML, comparable conglomerates and sandstone lenses are rare, but they do occur locally between diamictite intervals (Bechstädt et al., 2009). More common are well-sorted, silt-sized interbeds between unbedded and unsorted diamictite successions (Fig. 12C), which can often be traced laterally for tens of metres. These elongated lenses are mostly around 3–50 cm thick, commonly evenly laminated, and partially ripple-laminated. The lenses occur within Clast zone 3, near the top of the Ghaub succession in the central area, and in correlative basal parts farther to the west (Bechstädt et al., 2009).

4.3 Maieberg Formation

Söhnge (1957) subdivided the Maieberg Formation into two ‘lithozones’: T2, and T3. T1 represents the Ghaub Formation. Hoffmann and Prave (1996) coined the term Keilberg Member for the cap carbonates, which represent the T2 lithozone of Söhnge (1957). The Maieberg Formation is not discussed in any detail in this paper. The cap carbonates several metres in thickness (which in our area consist of dolomites and are therefore often called ‘cap dolomites’) are pale cream or buff in colour, and in fresh cuts typically lightly pinkish in the lower parts.

In the Northern area, the cap carbonates of the Maieberg Formation conformably overlie the Ghaub diamictites. In the metre-thick transitional interval, clasts are oriented along the bedding planes within a grain- to packstone matrix. In most cases, the components are dark dolomites, but quartzites, likely derived from the Nosib Group, are also common. The overlying, often pinkish-coloured cap carbonates are less than 10 m thick; outcrops are mostly discontinuous vertically and laterally, but facies seems to be relatively uniform (Fig. 14A). A few isolated centimetre-sized clasts have been found in the lower parts of the cap carbonates of the Northern area, and show undulations of the encasing fine laminations. Laminations directly around and underneath a few of these clasts are downwarped (Fig. 14B). Most of these single clasts are dark dolomites. The clasts are much less common and distinct than the clasts in the Fransfontein area (Fig. 14C) directly below cap carbonates, interpreted by Condon et al. (2002) and Domack and Hoffman (2011) as dropstones.

Large ripples characterise much of the succession. Only in outcrops near the eastern boundary of the Keilberg Farm are highly slumped and brecciated cap carbonates observed (Fig. 15A, B). In the same area, several megabreccia intervals with metre-scale carbonate clasts crop out in the overlying T3 parts of the Maieberg Formation. The Maieberg T3 intervals bear slump features and multiple clast sheddings; the often coarse-grained breccia intervals reach farther westward with time. No crystalline clasts have been found, in contrast to the underlying Ghaub Formation. The overlying Elandshoek Formation also frequently contains breccias; spectacular outcrops of crackle breccias occur along the unpaved road within the 'Tigerschlucht', Block 648, just east of Keilberg Farm.

Cap carbonates of the Southern area are less than 20 m thick (distinctly thicker than in the north), and show some differences. No diamictites occur underneath, no erosional unconformity was observed, and no slumped beds or breccias were detected. The above-mentioned difference in colour between the often pinkish cap carbonates and the underlying grey 'transitional zone' on top of the tube horizon (see 4.1.2) marks the separation of both intervals. The sediments in the basal part of the cap carbonates are micritic to peloidal dolomites with abundant microlaminations, because of the variable but very low clay and silt contents between the beds. Some of the laminations are undulating or slightly folded. Pyrite and haematite crystals are abundant, and the non-carbonate content is around 10%. Different, somewhat enigmatic sedimentary features such as large ripples (although smaller than those in the Northern area), or crystal rays to fans (Fig. 15C), which are also observed in northwestern Namibia, occur commonly.

In southern Auros Farm, above the small vanadium deposit of Wolkenhaube, thin tubestones crop out directly underneath cap carbonates, which are similar to those of the Southern area at the northern Auros Farm, about 5 km to the northeast. In the T3 unit of the Maieberg succession in the same Wolkenhaube area, spectacular megabreccias and slumps occur (Fig. 16), which are not present in the northern parts of the Southern area, but are similar to many breccia intervals in the Maieberg Formation of the Northern area.

5 Discussion

5.1 Facies evolution

The main trends in facies evolution, with emphasis on sea-level changes and differential subsidence, are shown in palaeogeographic sections from the Northern, more basinal and more subsiding area, to the Southern, less subsiding area (Fig. 17). The realm between the two areas is not outcropping, but because of the strongly differing subsidence faults have to be assumed. Input of sediments was most likely from the east.

5.1.1 Deepening and shallowing-upward trends

The succession of interest begins with the uppermost Gauss Formation. In the Northern area it consists mostly of grey dolomitic wackestones with several interbeds containing millimetre-sized large ooids, which were redeposited. Auros shales follow on top. In the Southern area, the uppermost

Gauss sediments contain microbial mats, overlain by stromatolitic domes (some of which are partly toppled), and some oolites, and are followed by the Auros shales. These changes in facies indicate a distinct deepening trend followed by a shallowing trend, starting with the Auros shales at the base, overlain by columnar stromatolites, and followed by wavy microbial mats with tubes at the top. Both trends can be observed in the Southern area only, the Northern Area was deeper and affected by higher subsidence (Figs. 6, 7, 17).

5.1.2 Facies in the Northern area

Auros Formation

The partly varve-like facies of the shaly lower Auros Formation with carbonate intercalations indicates low-energy, partly stagnant, oxygen-poor to anoxic depositional conditions. Oolitic interbeds within the overlying upper Auros carbonate laminites and micrites, as well as the very good sorting and often grading of these oolites (Fig. 8), imply that the ooids are not in place but were shed from an unknown shallow-water source, which likely existed farther to the east. The facies is comparable to the Gauss oolites below, but the amount of oolite interbeds is much greater. In the oolitic/pisolitic successions of the Mackenzie Mountains, northwestern Canada, stratigraphically below the Marinoan/Elatina glacial equivalents, James et al. (2005) described glendonite pseudomorphs after ikaite. These pseudomorphs indicate that shallow seawater in the area of present-day northwestern Canada was near freezing, despite the presence of ooids. This finding is evidence of a cold-water origin of some Proterozoic oolite deposits, in disagreement with the well-established warm-water genesis of Phanerozoic ooids. The glendonite occurrences indicate that seawater was already cold before diamictite deposition, and had distinctly higher calcium contents than seawater today. We were not able to find glendonite in pre-glacial oolite samples from the OML. Despite deposition in deeper waters, cementation of the oolite pore space must have occurred early, and thus allowed the ooids to resist compaction. Most are perfectly round and show no ‘pressure solution’. The high Ca^{2+} and Mg^{2+} saturation of marine waters due to submarine weathering of large amounts of glass-rich volcanic ridges, assumed by Gernon et al. (2016), may be an important factor for changes in ocean chemistry already in pre-glacial times. In contrast, ooids within local calcitic (dedolomitised?) parts are extremely elongated (Laukamp, 2006) because of the later stress of the Damara Orogen.

In the Fransfontein Ridge, ooid sheddings occur together with coarse components in the Franni Aus Member of the Ombaatjie Formation, a correlative unit of the Auros Formation of the OML. This member has been interpreted to represent a glacioeustatic ‘falling stand wedge’ (Domack and Hoffman, 2011). If a platform with a platform break would have existed at that time, as assumed by the mentioned authors, a strongly falling sea-level should expose the platform and upper slope. The ‘ooid factory’ would be strongly restricted in size and would hardly be able to provide the thick oolites present in the Fransfontein and Otavi areas underneath the Ghaub Formation. If ramp topography was present instead of the assumed platform, this problem would not exist.

In the Northern area of the central OML, the Auros ooid-sheddings interfingering with laminated carbonates can be traced eastward into coarse clastic debris-flow intervals. These units are intercalated within recrystallised micritic, thick-bedded units. The largest clast sizes (a few decimetres) are reached near the eastern boundary of Keilberg Farm. No clear dividing horizon between the Auros clastics and the overlying Ghaub diamictites exists in this area, where slump features and breccias were also found in the overlying Keilberg cap carbonates (T2), and megabreccias in the Maieberg T3 interval. These features indicate the proximity to a major synsedimentary fault, which played a role from at least Auros to Maieberg time, and possibly also thereafter.

Ghaub Formation: Glaciogenic or glacier-related deposits?

One group of authors interpreted the Ghaub Formation as formed solely by mass flows and other types of sediment shedding, without or with only indistinct glacial influence (Schermerhorn, 1974; Martin et al., 1985; Henry et al., 1986; Eyles and Januszczak, 2004, 2007). The latter authors also interpreted intervals beneath the Ghaub Formation as formed by debris flows and water-laid sedimentation in rift basins, and argued against a glacial setting. The lower part of the described succession, several tens of metres below the marker interval of the cap carbonates in the Fransfontein area, clearly indicates higher water energy. Hoffman and Halverson (2008), as well as Domack and Hoffman (2011), excluded this part of the succession from the Ghaub diamictites and placed these ‘often highly-silicified oolite debris flows’ (Hoffman, 2007) in the uppermost member (Franni-Aus Member, slope facies) of the underlying Ombaatjie Formation. Comparable rocks are to be found underneath the Ghaub diamictites in the oolitic and laminated Auros Formation of the OML, as mentioned above.

According to authors such as Hoffman and Schrag (1999, 2002) and Domack and Hoffman (2011), the Ghaub Formation in Kaokoveld and the Fransfontein Ridge is a product of grounded ice. The latter authors mapped the Fransfontein Ridge in great detail (see also Hoffman et al., 2014), measuring many different sections. The observed textures are interpreted as a product of an ice grounding-line wedge on the distal foreslope, where mainly tillites and associated facies (undermelt deposits, suspension from meltwater, iceberg-rafted debris, and sediment gravity flows) were deposited. Domack and Hoffman (2011) assume that the main glacial phase (the time of a totally frozen Snowball Earth) is not represented by the Ghaub diamictites but by the unconformity at its base. In the fence diagram shown by Hoffman et al. (2014), an unconformity occurs only at the base of the Ghaub diamictites in the Kranspoort trough (see below), and not in the area in the west (although mentioned in the text), or in the east, where the Ghaub Formation is missing and exposure of the shelf should have occurred. The Ghaub sediments themselves are interpreted to indicate improvement of the climate and to mirror several advances and retreats of ice. In the Fransfontein area, the difference in depth between the distal foreslope (the assumed depositional site of the Ghaub Formation) and the platform margin was said to amount to about 1,500 m (Domack and Hoffman, 2011). A concomitant sea-level lowstand was suggested, which would have enabled ice cover of the platform and large parts of the slope, and was followed by extreme sea-level rise. In a later erratum to the earlier paper, Hoffman (2013) and Hoffman et al. (2014) reduced this value to ≥ 500 m (see 5.3).

Domack and Hoffman (2011, and data repository 2011063) assumed for the Fransfontein area a uniform regional subsidence during Ghaub time, and a mostly flat-lying depositional interface on the platform with Marinoan glacial erosion but few depositional features. On the middle to distal slope, an 18-km-wide erosional valley several hundred metres deep is present, the Duurwater trough (called the Kranspoort trough by Hoffman et al., 2014). This incised valley was interpreted as Chuos glacial topography, partly filled by a Ghaub ice stream, which deposited a moraine inside the trough upon retreat (Hoffman, 2005; 2011b) (see also 5.3).

The postglacial cap carbonates of the Fransfontein area were said to conformably overlie the Ghaub Formation on the distal slope, and disconformably overlie the Ombaatjie Formation on the upper foreslope and platform, where the Ghaub Formation is missing and erosion is thought to have occurred during glaciation (e.g. Domack and Hoffman, 2011; Hoffman et al., 2014). The authors considered the Otavi platform as above sea-level but beneath an ice sheet for most of the Ghaub glaciation. The assumed sub-glacial unconformity was said to have more than 80 m of relief between a raised outer platform and a mostly dish-shaped inner platform, outcropping much farther to the west (Hoffman, 2007). The raised outer rim (locally with assumed aeolianites: Halverson et al., 2002) and the upper foreslope of the platform are untypically devoid of glacial deposits. Fairly thin and very discontinuous carbonate diamictites (possibly lodgement tillites according to Domack and Hoffman, 2011, data repository item 2011063) were described from the inner platform of the Hoanib river gorge in the Kaokoveld, about 200 km to the west. According to Hoffman (2011a), the thick Ghaub diamictites of the OML with granitic and quartzitic clasts are underlain and overlain by shallow-water carbonates. This interpretation assumed an inner platform depositional setting of the central OML, positioned about 20 km to the north of the projected platform break in the Otavi valley (e.g. Domack and Hoffman, 2011, and Fig. 1). As discussed by Bechstädt et al. (2009) and more thoroughly in this paper (see 5.2), the Ghaub outcrops of the central OML do not represent inner platform deposits at all, and the under- and overlying facies are not of shallow-water origin.

Indicators of glaciation such as striated clasts are absent from the Fransfontein Ridge according to Eyles and Januszczak (2004, 2007), Hoffman (2007), and Domack and Hoffman (2011). The almost exclusively present carbonate clasts weather strongly on their exposed surfaces, and the non-exposed parts are firmly encased in the rock matrix; it is therefore almost impossible to observe if glacial striations of clasts exist (Hoffman, 2007). In a Snowball setting, other typical glacial features such as *roches moutonnées*, grooved and striated pavement, and widespread pebble plasters are expected across large areas of the platform and upper slope, deposited at the latest during the melting phase, characterised by short glacial advances and retreats (Domack and Hoffman, 2011). The Dwyka glaciation bears numerous examples of such features (e.g. Masters, 2013). From the Chuos Formation of the Kaokoland and of the Otavi Mountainland impressive glacitectonic deformation features were reported (Le Heron et al., 2012; Busfield and Le Heron, 2013), including impact-related structures, but striated and faceted clasts and striated pavements are absent. The soft-sediment deformation structures indicate dynamic oscillations of the ice grounding-line and were critical in determining the presence and influence of grounded ice at Chuos time. Soft-sediment structures indicative of mass movements were not found. In the Ghaub succession of the OML (Northern area) such synglacial fabrics were not observed. During and after the melting of ice, strong chemical erosion would have followed on the platforms, soon succeeded by deposition of marine lag sediments or reworking of tillites when the platform was flooded. Assuming Snowball Earth, it is surprising that all these fabrics and sediment types do not occur in the Fransfontein Ridge, especially in the areas where the Ghaub Formation is absent. Halverson et al. (2002) interpreted fine-grained strata on the assumed emerged outer platform as loess-type aeolianite, deposited during late Ghaub time. One would expect the above-mentioned marine lag sediments to occur as well in association with this aeolianite.

As described above, three different prograding clast intervals occur in the Ghaub Formation in the OML (Fig. 13). Clast Zone 1 is quite variegated and restricted to the eastern and central part of the Northern area. The lithostratigraphic boundary with the Auros Formation is indistinct in the easternmost successions. Clast zone 1 does not show any clear indication of glacial origin. Only Clast zones 2 and especially 3 may have a relation with a glacial setting, but no clear indication for glacial till has been found. In the western part, the brownish weathered diamictites of Clast zone 3, which include Nosib clasts and crystalline components, are thickest, and also contain some striated clasts (Fig. 12D). All clast zones unconformably cover the Auros Formation oolites, breccias or laminites described above.

Several levels of diamictite beds of Clast zone 3 are separated by well-sorted and partially graded water-laid breccias and microconglomerates, but especially by sand/silt intercalations (Fig. 12C). These interbeds are positioned in the uppermost parts of the sections in the central area, and in basal parts farther to the west. These observations indicate stacked, imbricated layers (Bechstädt et al., 2009). The diamictite units are clearly prograding and thickening westward. The latter can also be observed in under- and overlying formations. This demonstrates that for a long time subsidence increased from the east toward a local basin in the west, and was much higher than in the Southern area (Fig. 17). No exotic clasts have been found, and the diamictite source appears to be local. Over time, erosion cut deeper and deeper in the strongly uplifted and unroofed hinterland (Bechstädt et al., 2009), reaching the Nosib Group and, at last, probably the crystalline basement.

Discrete diamictite sheddings were followed by time intervals during which currents washed out, winnowed, and sorted the sediment, including the formation of current ripples. These fabrics point to an active hydrological cycle during at least parts of Ghaub time. Comparable intervals also occur in successions of the Fransfontein Ridge, and were interpreted by Domack and Hoffman (2011) as inter- or sub-ice deposits produced by glacial meltwaters. The assumed repeated presence of meltwater indicates, according to these authors, climatic oscillations. The changes in the CIA weathering index (Bahlburg and Dobrzinski, 2011) provide additional evidence for climatic variations. In the OKS 1 well (Fig. 7), about 50 km to the south of the OML, two distinct matrix-rich diamictite intervals are separated by about 17 m of shale. At least two separate shedding events occurred in this slope to basinal area of the Damara Ocean.

All of these observations in the OML and its environs indicate that the diamictites correspond to debris or (in the case of oolites) grain flows into deeper marine settings. No robust evidence for the presence of tillites was found, but much of the sediment may have represented recycled periglacial material. Thus, the diamictites may represent originally glacial to glaciofluvial sediments shed into a basinal area in front of a fan delta. The clasts may have been fed by a glacial environment located at some distance, possibly in mountainous areas to the east (see also the discussion by Eyles and Januszczak, 2004, 2007). Unroofing of the source area must have stripped off a voluminous sedimentary pile. This inference is further corroborated by published data, indicating that the Ghaub Formation is about 1000 m thick in the Otavi area at the western margin of the OML (Söhnge, 1957).

Dropstones

Evidence for a glaciation at some distance and not in the depositional area of the Ghaub Formation of the OML is also derived from the presence or absence of glacial rainouts and from the facies found in the Southern area, as discussed below. In the Fransfontein area, Condon et al. (2002) reported glacial rainout facies from the Ghaub Formation. The several-metres-thick Bethanis Member of the upper Ghaub Formation (Hoffman and Halversson, 2008; Hoffman, 2011b; Domack and Hoffman, 2011) contains an allodapic unit from which the above-mentioned authors reported dropstones, which in our opinion also show typical features associated with such an origin (Fig. 14C). The member occurs on top of diamictites and directly underneath the cap carbonates, and there is a close relationship between glacial rainouts and diamictites. Eyles and Januszczak (2007) denoted uncertainty with respect to the dropstone interpretation. Despite intense search, we were not able to find exotic components in the Fransfontein area in this interval, or underneath or in the cap carbonates above.

In the uppermost parts of the Ghaub Formation/lowermost cap carbonate in the OML, only questionable dropstones were identified. Some outsized clasts in the upper, unstratified Ghaub Formation may represent dropstones, but because of the lack of laminations, this interpretation is debatable; they may also represent lonestones. In a transitional interval < 1 m in thickness between the Ghaub Formation and cap carbonates, the clasts are oriented along the bedding planes of fine clastic sediments without clear impact on the lamination below; their origin is therefore again debatable. Within the cap carbonates, a few isolated clasts were found, consisting mostly of dark dolomite fragments. A few show impact structures (Fig. 14B), downward depressed rims where slight liquefaction may have occurred. These fabrics and the occurrence inside the laminated cap carbonates make an origin as dropstones the only viable interpretation for these clasts. These and other, more questionable dropstones are several orders of magnitude less common than the reported dropstones (e.g. Domack and Hoffman, 2011) in the Bethanis Member in Fransfontein Ridge (Fig. 14C). In the Southern area of the OML (see below), not only are the diamictites missing, but neither the tubestones nor the cap carbonates contain any clasts that might be interpreted as dropstones.

According to the Snowball Earth (e.g. Hoffman et al., 1998b) or Panglacial Earth (Hoffman, 2009) scenario, the distribution of dropstones should be fairly uniform, and dropstones would be expected to occur on top of flooded platforms or ramps, related to deglaciation. However, this is not the case; in the OML and Fransfontein Ridge, varying amounts of dropstones have been reported from areas where the Ghaub diamictites are present (e.g. Condon et al., 2002), but they are not found when the Ghaub Formation is very thin or missing. In a Snowball and/or Panglacial scenario, dropstones should occur not only at the top of glacial successions, but might also be expected in the interval directly below the basal Ghaub unconformity, assuming that the Ghaub itself represents a meltback (Hoffman and Schrag, 1999), whereas the main glaciation is represented by the unconformity underneath the Ghaub Formation (Domack and Hoffman, 2011). Glacial rainout should have formed seasonally at the time when thick ice was building up *before* the main glaciation occurred. In the Fransfontein Ridge, this stage is represented by the Franni-Aus Member, interpreted as a ‘falling stand wedge’ by Domack and Hoffman (2011) and Hoffman (2011b). No dropstones were described from this member or found in the corresponding Auros Formation in the OML.

A post-Snowball strong and sudden melting phase would have provided subglacial melting and masses of icebergs, which would have released their clast contents in a short interval of time. If climatic improvement took longer, strong melting and glacial calving of the ice-cover would have proceeded poleward more slowly, and the variety of dropstone types may be more locally restricted. According to both scenarios, however, during the main transgression, one would expect the release of large amounts of glacial rainout material. This phenomenon would form an areally widespread and laterally correlatable dropstone interval, present not only in basinal areas but also on top of flooded platforms or ramps, which is not the case. The difference between the dropstone-rich Fransfontein area, the dropstone-poor Northern area of the OML, and the lack of dropstones in the Southern area is an argument for only local presence of glacial ice reaching the sea. Only in such areas with greatly varying amounts of undermelting and glacial calving, corresponding local iceberg alleys may have been present. The glacial rainout consisted of local material, similar to the situation in parts of Greenland and Newfoundland today. Elsewhere, dropstones formed much more rarely or were absent. However, the occurrence of a few dropstones in the lower part of the cap carbonates of the OML indicates that no super-greenhouse setting was present at cap-carbonate-time, the climate may still have been cool. These observations indicate non-uniform ice occurrences, not a worldwide Snowball or Panglacial Earth. A Waterbelt Earth (Pierrehumbert et al., 2011; Rose, 2015) or High-Obliquity Earth (e.g. Williams et al., 2016) with more regional patches of ice on the continents are much better suited to explain the described facies.

5.1.3 Facies in the Southern area

Columnar stromatolites

The large stromatolitic columns in the interval overlying the Auros shales (Figs. 7, 9, 17 stage 2) indicate deposition in a neritic, photic environment with low energy. The absence of stromatolites in the underlying Auros shales may have been caused by the distinct clay input and/or depositional settings below the euphotic zone, too deep for stromatolites to grow. Subsequently, clay input ceased, and undisturbed growth of columnar stromatolites began. Accommodation was high and water energy low enough to allow the build-up of these columns to heights in metres. Coarse debris of reworked columns is very rare, and no decapitated or toppled columns were found. Because of the fragility of the columns, which were a few decimetres thick and often more than one metre in height, the setting was likely protected, for instance by an ooid barrier. Therefore, deposition most likely took place below the local storm-wave base. The stromatolites were quite healthy, without any indication that light was a limiting factor for their growth. How deep the photic zone reached in this area in the late Proterozoic can only be speculated upon. Penetration of light was most likely not influenced by plankton in the water (if plankton were present at all), but only by some suspended clay. Depositional depths were likely less than the deeper limit of the euphotic zone today, which is around 200 m, but certainly not several hundreds of metres.

Toward the top of the Auros Formation, the stromatolitic columns become smaller and the facies changes to microbial laminites, which show some reworking. These findings indicate an increase in energy and almost certainly a decrease in accommodation and upward shallowing, possibly because of sea-level fall, which may have begun already in the earlier Auros time (see 5.3 and Fig. 17).

Tubestone facies

The tubular structures (Figs. 7, 10, 11, 17 stage 2) within microbial laminites on top of the columnar stromatolites are a feature not only of many Namibian successions, but also of several other ‘Marinoan-aged’ carbonates on various continents. For instance, such tubestones were described from Brazil (Rodrigues-Nogueira et al., 2003), the United States (Cloud et al., 1974; Marenco and Corsetti, 2002; Corsetti and Grotzinger, 2005), Canada (James et al., 2001), and elsewhere in Namibia (e.g. Hegenberger, 1987, 1993; Kennedy et al., 2001a, b; Hoffman and Schrag, 2002), including on the Kalahari side of the Damara Ocean. However, the microbial laminites with tube features show some variability, and may not be of exactly the same type with the same genesis everywhere. Differing interpretations of such fabrics include: (i) gas and/or fluid escape structures; (ii) an origin related to microbial stromatolite morphology, or (iii) metazoan burrows and root casts, which are not applicable for the Proterozoic and therefore not discussed here.

(i) The tubular structures and sheet cracks represent an interconnected network of fluid or gas escape structures fed from shallow tidal (?) sediments below (e.g. Cloud et al., 1974; Hegenberger, 1987, 1993). In Hegenberger’s model, gas was produced because of the decay of microbial mats, mainly during low tides. Gases including CO_2, CH_4, NH_3, and H_2S would have built up along sheet cracks and escaped vertically when the pressure was sufficient to break through the overlying sediment. Fluid or gas pressure opening horizontal fissures along surfaces of weakness between the microbial laminations (Fig. 10) would feed into tubes where pressure escaped subvertically (Fig. 11). However, the problem of this model is why in geologic history such structures are not much more common in rocks containing comparable microbial facies. Kennedy et al. (2001b) suggested that tubes and sheet cracks could be methane escape structures resulting from sudden destabilisation of clathrates during climate warming in the glacial aftermath. Methane release was assumed to have also provided a mechanism for the formation of the overlying cap carbonates with their distinctly negative ^{13}C signature of around −3‰ PDB-V (Kennedy et al., 2001b; Jiang et al. 2003). Hoffman et al. (2002) disputed this interpretation. In a thick interval of the Maieberg Formation (T3), above the cap carbonates, even more negative carbon isotope values (around −5‰ PDB-V) occur (Kennedy et al., 2001b; Halverson et al., 2005). These are approximately 2‰ more negative than the values in the cap carbonates above the interval with tubes. This finding makes an origin of the cap carbonates due to destabilisation of gas hydrate less likely, but a certain influence of methane may have persisted.

(ii) James et al. (2001), Hoffman and Schrag (2002), and Corsetti and Grotzinger (2005) proposed that the synoptic depressions on the surfaces of stromatolitic beds were preserved as the stromatolites aggraded. According to their model, the tubes represent sediment-filled isolated depressions preserved by the post-glacial, highly carbonate-saturated seawater (Corsetti and Grotzinger, 2005).

However, the tube fabrics of the OML differ from the last mentioned fabrics: they are clearly cavities infilled with quartz or chert, carbonate cement, and, on rare occasions, also sulfides. Therefore, we favour the gas-escape model. The tubular facies was placed lithostratigraphically into the uppermost part of the Auros Formation (Söhnge, 1957; Beukes, 1986; King, 1994). In a major revision of the stratigraphic correlation of the Otavi Group, Hoffmann and Prave (1996) placed the tubestones into the lower cap carbonate interval (Keilberg Member), overlain by 'classic' cap carbonates. In this correlation, the facies containing tubes represent the base of the post-glacial Maieberg Formation, and therefore the time of warming. We originally agreed with this modified stratigraphic setting (Bechstädt et al., 2009), but now suggest to retain the earlier lithostratigraphic assignment to the Auros Formation (see Fig. 2) because no distinct boundary with the columnar stromatolites below can be mapped, whereas the one with the cap carbonates is more distinct, inclusive of a mostly present interlayer of microbial mats without tubes.

The facies of stromatolitic columns changed to laminites, likely because of a relative sea-level fall during Auros time (Fig. 17). Accommodation diminished, and the sedimentation rate might have been low, but the assumed sea-level drop was neither strong enough nor fast enough to expose parts of these sediments. No unconformity was found in the whole succession. In contrast to a few reworked stromatolitic columns from the Auros Formation, no reworked clasts of this tubular facies were ever observed in the Ghaub diamictite. This finding may indicate that the microbial laminites hosting the tubes have been deposited in syn- or late-glacial times, but the formation of the tubes themselves followed slightly later, at the initiation of warming, but before distinct lithification.

Three events in time must be differentiated: (i) sedimentation of microbial laminites, which lithostratigraphically belong to the Auros Formation, but perhaps to the Ghaub time; (ii) Early diagenetic formation of tubes (methane release?), possibly during interstadial or early postglacial warming, eventually in a repeated manner (different intervals of more or fewer tubes in the laminites); (iii) partial cementation of the interior cavity of the tubes by blocky carbonates, later partly silicified. A sub-recent setting has shown a possible explanation for the assumed succession of events (Förstner et al., 1968, their figs. 27, 28). These authors investigated the recent carbonates of Lake Constance, southern Germany, where numerous vugs form in the uppermost metres below the sedimentary surface due to degassing. For the preservation of cavities in semi-lithified sediments, the timing of compaction versus lithification is crucial. About 2 m below the sedimentary interface, the vugs change to half-closed, laterally elongated fenestral fabrics because of fast compaction and slow lithification. In deposits underneath, these vugs are completely closed. If lithification were to occur faster in this sub-recent setting, the vugs would remain, with their size and geometry depending on the timing of compaction versus lithification.

The gas release may have been related to the decay of microbial material (model of Hegenberger, 1987, 1993), or melting of clathrates (Kennedy et al., 2001b). These were originally stored during the cold glacial climate and released during warming, glacial retreat, and the start of melting of ice at the end of deposition of the Ghaub Formation and the initiation of cap carbonates. The different episodes of glacial waxing and waning (Rieu et al., 2007; Bahlburg and Dobrzinski, 2011) may even have caused episodic formation of clathrates and punctuated methane release, as indicated by the three observed horizons with different amounts of clustered tubes.

The tubes formed when the rock was in a semi-plastic state after the growth of microbial laminites, before the sediment was fully lithified. Possibly a much faster cementation occurred than in other times, because of strongly increased carbonate contents in seawater. These unusually high Ca and Mg contents (see also James et al., 2005; Gernon et al., 2016) may have caused early stabilisation and lithification of the walls of the tubes. Compaction was low immediately after tube formation, and the gas escape structures remained intact and open. The infill of the tubes is mainly blocky carbonate cement or SiO_2. A fibrous, wall-lining early cement, which is common in the Abenab and Gauss formations, is absent, which corroborates the interpretation that cementation occurred some time after tube formation and not directly below the sedimentary interface. The carbon isotope values of these tube-occluding carbonate cements are comparable to the surrounding, recrystallised carbonates, and not distinctly more negative. Therefore, these cements formed in a later diagenetic stage and do not give any clue to the process of tube formation (presence or absence of methane).

5.2 Synsedimentary tectonics and shelf evolution

According to Hoffman (e.g. 2005) and Domack and Hoffman (2011), a north–south crustal stretching in the Fransfontein and Hoanib areas during Nosib time created the Huab and Makalani horst blocks (see Fig. 3). This rifting episode affected the successions until the base of the Ombaatjie Formation. Subsequent draping of the relief filled an assumed small depression between the horst blocks and the Congo Craton in the north. Broad regional subsidence with structural conformity is assumed after the upper Ombaatjie Formation (Abenab Subgroup) (Domack and Hoffman, 2011). Dürr and Dingeldey (1997) and Prave (1996, 1997) proposed strongly different geodynamic scenarios (see below).

In the interpretation of Hoffman and Halverson (2008), no structural movements occurred during Ghaub time on the platform or its margin, although Halverson et al. (2002) mention a tensional phase. It affected the platform margin represented by the Makalani Ridge during Ghaub time, and even caused a block rotation in the Outjo Basin to the south. According to Hoffman (2005, and thereafter), there is no distinct progradation of the Otavi carbonate platform toward the basinal area to the south, and the margin was supposedly stable for a very long time. Despite this assumed lack of progradation, slope gradients like that at the distinctly prograding western margin of the Great Bahama Bank were tentatively employed for early calculations of water depths of 500–800 m (Hoffman, 2006; Hoffman et al., 2007), which were subsequently revised more than once (see 5.3).

The outer platform is described as raised and the inner platform as mostly dish-shaped, with some relief between the two. Outer platform deposits are reported from the upper Huab area in western Namibia, where a purported dolomitic aeolianite on the platform margin may replace the Ghaub Formation (Halverson et al., 2002). The inner platform crops out only at a long distance farther to the west, along the Hoanib River gorge in Kaokoveld, which shows some erosion at the base of strongly discontinuous and quite thin (if present) diamictites, interpreted as lodgement till (e.g. Domack and Hoffman, 2011, and data repository). In the Fransfontein Ridge, a thick succession of the Ghaub Formation occurs on the distal foreslope (Domack and Hoffman, 2011). More recently, Hoffman et al. (2014) showed sections of the upper slope and platform successions from the Huab Ridge in a poster abstract without any detailed description. The authors stated that the location and development of the shelf break in their large transect (consisting of 64 measured sections) are not structurally controlled. No significant syndepositional fault was found. The dominant control on sedimentation, according to the authors, was < 1.8 km of local palaeotopography on the Chuos glacial surface, which is underlain by crystalline basement. The exact location of the shelf break during Ghaub time is said to have been dictated by > 0.5 km of relief on the glacial surface, remaining from Chuos time. However, in our opinion, distinct differences in thickness of partly neighbouring sections indicate strong local differences in subsidence, as is the case in the OML (Fig. 17). In addition, the relief shown is partly caused by the arrangement of the datum level, which is different in Hoffman et al. (2014) and Domack and Hoffman (2011).

The shelf break inferred by Domack and Hoffman (2011, and elsewhere) is said to have continued from the Fransfontein area eastward to the Otavi and Kombat area of the southernmost OML (see Fig. 1). The Ghaub Formation in the latter region occurs about 20 km northward of this proposed platform edge and is described as unstratified and 'abnormally thick for a platform section', underlain and overlain by shallow-water carbonate facies (Hoffman, 2007, 2011a; Domack and Hoffman, 2011). However, in the area of Ghaub outcrops, the deposits over- and underlying the diamictites are mostly of deeper water origin (Fig. 17; see also chapter 4). The shelf does not have a single break in the southernmost OML, but has much more complex palaeogeography.

The Auros Formation bears large breccia bodies in sections of the northeastern Keilberg Farm, and oolites were shed downslope farther to the west. The Ghaub succession of the Northern area represents mostly deeper-water facies, and consists of almost entirely Abenab carbonate clasts in the east (with the exception of a thin interval with Nosib and crystalline clasts on the top). It incorporates Nosib and crystalline components in the sections farther to the west (see 4.2). This predictable change of the clast content (Bechstädt et al., 2009) is a clear indication of a westward prograding succession, as well as a continuous, deeper dismantling of the clastic source. The Nosib and crystalline basement clasts belong to an interval about 2,000 m below the Ghaub Formation in the area of research. Even if the eroded formations may have been thinning laterally, the source area must have been strongly uplifted.

At the end of diamictite deposition, the area had shallowed (Fig. 17); the amount of sea-level fall is discussed in chapter 5.3. The former relief was filled by sediments, and despite the inset of sea-level rise, the overlying cap carbonates are in the photic environment (Jäger et al., 2010). A platform marginal fault most likely played its role from Auros to Maieberg time and thereafter; the cap carbonates of the Maieberg Formation (T2) in the easternmost outcrops on the Keilberg Farm (and nowhere else) are slumped and brecciated (Fig. 15A, B). Megabreccia beds occur during T3 a few metres above the slumped T2 (cap carbonates). The Maieberg T3 interval bears multiple carbonate-clastic sheddings, from microturbidites to chaotic breccias. Breccia bodies prograded westward, where initially fine-grained turbiditic beds are separated from the cap carbonates by a shaly interval a few tens of metres thick. No reworked Ghaub diamictites occur in the Maieberg Formation. No Nosib and/or crystalline clasts were shed in the Northern area after the Ghaub Formation. The source area of the Ghaub clasts was therefore no longer effective, it may have become flooded, and/or sediment deposition was now possible closer to the source. The frequent clast sheddings in the Maieberg Formation are related to manifold slope failures.

Synsedimentary tectonics occurred repeatedly throughout the depositional interval of the whole Otavi Group up to the Mulden Formation (Laukamp, 2006; Bechstädt et al., 2009). In the OML, strongly differing thicknesses and lateral pinchout of strata occur in the Ghaub succession, as well as in underlying and overlying intervals (Bechstädt et al., 2009). There is a distinct westward thickening of the Gauss, Auros, Ghaub and Maieberg formations. These data demonstrate changes in accommodation and differential subsidence throughout the mentioned interval, and possibly thereafter. Parts of the OML clearly did not represent a stable continental margin; large-scale growth faults, slump features (Figs. 15A, B, 16B) and external as well as internal breccias (Laukamp, 2006;

Bechstädt et al., 2009, with further references) were formed during formation of the Otavi Group deposits. This inference contrasts strongly with the assumption of a tectonically quiet Fransfontein depositional area by e.g. Domack and Hoffman (2011) and Hoffman et al. (2014). In the Northern area (Fig. 17), westward of an assumed synsedimentary fault, a local, more strongly subsiding restricted basinal area, possibly an intra-platform furrow, existed from Auros to Maieberg time and thereafter, filled by a Ghaub fluvioglacial fan delta that shed and prograded from the east into the furrow farther to the west. No clear indication for tillites was found, whereas clues for synsedimentary tectonics are much more robust. One cannot envisage a glacier, even a wet-based one, able to (i) erode consistently down to the basement, and (ii) deliver the sediment to a marine area without tectonic uplift of the source area. The main argument for the presence of glacial ice does not come from the diamictite fabrics themselves but from the rare dropstones in the cap carbonates (Fig. 14B) and a few striated clasts, mainly quartzites (Fig. 12D).

During Ghaub time, a carbonate platform or ramp existed in the Southern area (Fig. 17). The latter was generally not affected by synsedimentary tectonics in Auros, Ghaub and early Maieberg time, and Ghaub diamictites are not present. Only in T3, tectonics revived on the southern Auros Farm (Wolkenhaube area), where slumps and large synsedimentary breccias crop out. This area is close to the margin of the Outjo Basin in the south, for which the tectonic processes are described in detail by several authors (e.g. Nascimento et al., 2016).

Synsedimentary tectonics affected most of the Otavi Group. Some authors overlooked parts of these tectonic movements, which is likely the reason for the differing interpretations of the plate-tectonic development and configuration published to date: (i) A rifted margin passing from Nosib rifting to Otavi passive margin, to Mulden molasse (Martin and Porada, 1977). Other authors assumed the tectonically stable passive margin was initiated later, at the end of the Abenab Subgroup (Hoffman et al., 1998a, b; Halverson et al., 2002; Hoffman et al., 2017). (ii) A rifted margin developing to a "passive" one, which experienced, however, tectonic activity and subsidence through extension in different phases of its development. One such extension phase is documented by upper Maieberg strata (Dürr and Dingeldey, 1997); (iii) A continental margin passing already in middle Otavi time to an active foreland basin setting (Prave, 1996, 1997). The shelfal facies of the Congo craton prograded and retreated during deposition of the Tsumeb Subgroup; highs and local intra-platform basins came into existence (Prave, 1996, 1997), an intra-Otavi angular unconformity was reported. Carbonate lithoclast breccias in the upper Maieberg Formation were interpreted to indicate exposure and erosion of older carbonate rocks. These clasts are, however, resedimented slope deposits and indicate in our opinion large slope failures, and not exposure. The uplift and erosion of the Nosib Group and crystalline basement during the Ghaub Formation might be of more importance in this respect. A major break in sedimentation does not occur before Mulden time.

Whatever the plate-tectonic setting, there was not a uniform shelf break from a continent in the north to the Damara Ocean in the south, no stable passive margin existed. The continental shelf of the

Congo craton represented an unstable continental margin, not passive at all, reacting to the processes in the Damara Ocean, especially the Outjo Basin, and possibly also to the closing of the Adamastor Ocean in the west (Prave, 1996).

5.3 Amount of sea-level change

In the Snowball Earth/Panglacial Earth model, important sea-level changes were proposed (Hoffman et al., 1998a, b, 2009; Hoffman and Schrag, 2000, 2002; Hoffman, 2009; Domack and Hoffman, 2011): the sea-level dropped down on the slope, and the platform was emergent. After early estimates of 500–800 m (Hoffman, 2006; Hoffman et al., 2007), about 1,500 m of sea-level fall in pre-Ghaub time was calculated from outcrop data in the Fransfontein Ridge (e.g. Hoffman, 2011b; Domack and Hoffman, 2011), corresponding to a Marinoan ‘global average ice-sheet thickness of ~3.0 km on all continents and oceanic plateaus’, ‘consistent with geologic evidence’ (Hoffman et al., 2009). This calculation was later retracted (Hoffman, 2013), but we will discuss it here because this extreme sea-level fall is cited frequently. The calculation was based on various assumptions:

(i) The Ghaub sediments were deposited on the distal slope, to a large degree by grounding ice. The authors discuss (and reject) alternative depositional models.

(ii) A rimmed platform with a distinct break existed; the possibility of a ramp was not discussed.

(iii) A clinoform with a regular dip of 15 degrees existed. This gradient was quite ingeniously deduced from palaeovertical ‘geoplumb’ tubes, which normally are positioned orthogonal to the microbial laminites, but show instead a 15 degrees difference. Hoffman et al. (2009) and Domack and Hoffman (2011) applied this value to the whole slope, from proximal to distal.

(iv) The distance between the depositional area of the Ghaub diamictites and the position of the platform margin break can be determined. There are no outcrops of the interpreted shelf break paralleling the main Ghaub depositional area; these exist only a few tens of kilometres to the east. The shelf break was therefore projected from the easternmost sections containing microbial laminites with tubes to the west (Domack and Hoffman, 2011: their figs. 1 and 3). However, the resulting position after this large-distance projection is not robust. The eastward projection of the platform margin from the Fransfontein Ridge to the OML has been shown to be incorrect (see 5.2), and the westward prolongation into an area without outcrops may be as well.

Several authors (e.g. Kenter, 1990; Blendinger, 2001; Keim and Schlager, 2001; Seeling et al., 2005; Kenter and Schlager, 2009) discussed the slopes of the Triassic Alpine carbonate platforms and compared these slopes with sub-recent ones of the Bahamas. The authors showed the dependency of the inclination of clinoforms from the specific sedimentary fabric, with varying degrees of early cementation, including automicrite formation, as well as sea-level fluctuations. The Triassic platforms had very steep and often uniform clinoforms, which at the toe bent sharply into the flat basin floor. Their inclination of often 30–40 degrees was caused by carbonate sand and rubble, produced mainly on the platform, and frequently high amounts of microbial automicrite 'cement', formed in situ, stabilising the slope. The recent Bahaman clinoforms generally have steep angles only in the upper approx. 150 m, the range of Quaternary sea-level fluctuations. The remaining slope dips at differing degrees, depending on the windward or leeward position, and the sedimentary fabric and underlying tectonic features. On the windward side, e.g. the Tongue of the Ocean, the slope can be almost vertical. The clinoforms on the leeward side only possess a few degrees dip and are prone to creep and slumping because of the high mud content of the sediment transported from the platform interior by the trade winds.

As discussed above, in the OML and the Fransfontein Ridge, pre-diamictite oolites were cemented early, which indicates high carbonate content in the seawater. Whether precipitation of automicrite occurred on Proterozoic slopes or not is unknown, but it is not unlikely. In addition, synsedimentary tectonics would make the depth calculations of Domack and Hoffman (2011), Hoffman (2011b) and Hoffman et al. (2014) obsolete. As mentioned above, Halverson et al. (2002) assumed a tensional phase during the Ghaub Formation affecting the platform margin (Makalani Ridge), when block rotation also occurred in the Outjo Basin to the south, and olistolithic blocks broke off from the shelf slope (Nascimento et al., 2016). In the OML, complex palaeogeographical patterns existed, which, if present in the Fransfontein Ridge as well, would strongly affect the calculations of water depth.

Domack and Hoffman (2011) and Hoffman (2011b) postulated a 'falling stand wedge' directly below the Ghaub diamictites, represented in the Fransfontein area by the Franni-Aus Member, which consists to a large degree of shed ooids and debrites. Deeper-water conditions are assumed in the succession below. In addition, in the OML, the Auros succession underneath the Ghaub diamictites is often largely oolitic. Because the sea-level fall is assumed to have started already in pre-diamictite time, the proposed platform margin would have been subaerially exposed, and the 'ooid factory' must have migrated down on the slope, restricting the possible production area. We doubt that in such restricted sectors, the observed high amounts of ooids could be produced. However, it is not clear if a platform with a platform break, as suggested by Domack and Hoffman (2011) and Hoffman (2011b), existed at all. A ramp setting may be more appropriate, and would pose less problems for ooid production during the 'falling stand wedge' of the Franni-Aus Member or (in the OML) the upper Auros Formation.

Hoffman (2013) stated in a corrigendum to his 2011b paper that the 15-degree angles between the geoplumb tubes and laminations are not synsedimentary but caused by tectonic strain. This difference can therefore not be used to decipher the gradient of the clinoforms, and thus the palaeodepth of the Ghaub Formation. In the corrigendum, Hoffman assumed a glacial base-level fall of $\geq$ 500 m. A comparable 'most probable" value of 620 m was modelled by Liu and Peltier (2013), which was found to be 'broadly consistent with expectations', based on inferences from Otavi Group deposits. This result clearly depends on the input estimates of the ice volumes present. Hoffman (2013) calculated a slope angle of about 6 degrees, which according to the data of Kenter (1990) would be typical for mud-supported carbonate clinoforms, not present in the Cryogenian of Namibia. Angles of up to 40 degrees are easily possible for grain-supported platform flanks, especially if cemented early, as seems to have been the case.

Domack and Hoffman (2011), Hoffman (2005, 2011b), and Hoffman et al. (2014) described one distinct erosive channel, a 'Chuos palaeovalley', not occurring on the platform but on the slope. The Ghaub infill of this Duurwater trough (later called the Kranspoort trough) was interpreted as the product of the confluence of two tributary Ghaub ice streams, which deposited the medial Duurwater moraine upon retreat. Most of the infill of this trough feature is older (Gruis Formation) than the Ghaub diamictites. Hoffman et al. (2014) postulated, in their poster abstract showing 64 sections, that the dominant control of the Ghaub shelf break is < 1.8 km of glacial topography formed by the Chuos glaciation. On the sides of the Kraansport trough, the Chuos Formation crops out, but is not present in the trough centre, and seems to be eroded there. On the laterally present Chuos rocks, or on thin post-Chuos carbonates in the trough, a very thick and laterally restricted 'grainstone prism' of the pre-Ghaub Gruis Formation follows. The trough feature, later filled by the Ghaub Formation, must therefore have formed after the Chuos glacial period, not by this glaciation. The Chuos and overlying successions represent a time of strong rifting; thicknesses of the sections shown are strongly variable, including the Gruis Formation in the trough. This trend indicates strong synsedimentary variations in accommodation caused by changes in rift-related subsidence, and not solely a buried palaeorelief of the Chuos glaciation. This very persistent presence of important palaeogeographical constituents in the Fransfontein Ridge is comparable to own observations in the OML.

The location of the shelf break in Ghaub time is determined by assumed > 0.5 km of relief (Hoffman et al., 2014: conservative offset) on the sub-Ghaub glacial surface, which ramps across the Ombaatjie Formation on top of the Huab Ridge toward the west. The relief is shown by the way the datum level for the sections is arranged by Hoffman et al. (2014), and in addition by a map ("Kransport salient"). The authors deduced the erosion of the marginal platform from the postulated cut-off of the Ombaatjie Formation east of the Ghaub outcrops. This assumed truncation should be better substantiated in the future to rule out interfingering of facies; on none of the sections depicting the truncated Ombaatjie Formation is erosion marked at the top, below, or above often thick tubestones on the shelf, overlain in some sections by different facies of cap carbonates (Hoffman et al., 2014). To us, the setting of the Huab Ridge shelf and the Kranspoort Trough looks comparable to the setting of Northern and Southern areas in the OML (see Fig. 17).

According to e.g. Domack and Hoffman (2011) and Hoffman (2013), the shelf was highly emergent during glacial time, eroded by flowing ice, and subsequently strongly affected by sea-level rise. A laterally correlative unconformity is mentioned, but not adequately described, by Domack and Hoffman (2011) from areas west of Fransfontein. Ghaub diamictites occur on the platform in only a few localities, and are almost entirely restricted to slope occurrences. On the platform (or inner ramp), one would expect moraine ridges, *roches moutonnées*, glacial striations of the rock surfaces underneath, or at least widespread, laterally consistent plastering by glacial pebbles, formed especially upon ice retreat and short glacial advances, as described from the Cenozoic or from the Dwyka glaciation (e.g. Master, 2013). One would expect strong reworking of the older sediments, especially at the platform margin, caused at first by the continental ice flowing down the slope, and later by the rising sea-level flooding the platform or ramp. All these features are missing, the base of cap carbonates in the areas devoid of Ghaub diamictites shows hardly any brecciation at all. If ice covered the platforms, tillites with soft-sediment deformation structures as reported by Busfield and Le Heron (2013) from the Chuos Formation could be expected in the Ghaub Formation as well.

This hypothetical strong sea-level fall, lowstand, and subsequent extreme rise assumed for the Snowball Earth scenario contradicts our observations in the OML (compare the palaeogegraphic sketches of Fig. 17). Modest relative sea-level changes were initiated in the Southern area already during deposition of the Gauss Formation underneath the Auros shales. A distinct gain of accommodation is clear, highlighted by the change from laminated to domal stromatolites in the uppermost Gauss Formation, followed by toppled stromatolitic mounds and locally restricted, partly anoxic conditions during Auros shale sedimentation, representing a depositional setting below storm wave base. This interval is interpreted to represent maximum flooding (sea-level 1 of Fig. 17). After the Auros shales, the depositional area shallowed in the Southern area, whereas the Northern area remained in a deeper-water setting, without clues on eventual changes of water depth. In the former area, the shallowing can be identified by passage into the photic zone, which enabled the growth of columnar stromatolites, and finally to a shallow subtidal zone, but not subaerial exposure conditions for the formation of the tubestones (sea-level fall to level 2 and 3-4 of Fig. 17). This shallowing without unconformity was stopped by the capping dolomites of the Keilberg Member.

The mutually exclusive distribution of tubular facies and Ghaub diamictites mentioned above also corresponds to the descriptions by e.g. Domack and Hoffman (2011) and Hoffman et al. (2014). The origins of both facies are related to differences of water depth. The Northern area, with partly varve-like laminites, intercalated oolite sheddings, and Auros breccia deposits underneath the Ghaub diamictites, was clearly deeper than the Southern one, characterised by different types of stromatolites. The thicknesses of deposits in both areas are dissimilar as well, and mirror accommodation differences; in the Northern area, the thickness from the base of the Auros shales to the base of the cap carbonates is about 520 m (80 m of it Auros shale), and in the Southern area, only about 90 m, with 30 m of Auros shale (Fig. 17).

Is this shallowing trend observed in the southern Auros Formation caused by a (possibly eustatic) sea-level fall, by diminished subsidence, or by a combination of sea-level and subsidence changes? In contrast to the Northern area, sediment deposition was clearly higher than the gain of accommodation space (combined sea-level change and subsidence) and caused the observed shallowing trend. Can this trend be explained by a setting without any sea-level drop at all? The stromatolitic columns seem to have formed in water depths below fairweather wave base, but still in the photic zone. If we tentatively assume a water depth of 80 m for the growing stromatolitic columns (Fig. 17, stage 2), and a depth of twenty metres for the tubestones, not reaching tidal influence, and also consider the about 60-m-thick succession of both units above the Auros shales, the observed shallowing trend can be explained by sedimentation filling up the available accommodation. However, this inference depends of course on the water depths assigned to the different facies, and would imply a stop in subsidence. The question arises of how long was the time interval involved. If we take into account the more than 500 m in thickness of the correlatable succession in the Northern area, compared with the few tens of metres of the succession in the south (Fig. 17), the time represented should not be very short. In this time span, there was probably some subsidence also in the south, but there is no indication of a tectonically enhanced subsidence compensating for eventual high rates of sea-level fall; the Southern area was tectonically stable (see 5.2). The eventual eustatic sea-level drop (assumed in Fig. 17, sea-level drop from stage 1 to stage 3-4) was therefore not sudden, had a value of slightly more than the amount of subsidence, but was not able to expose the platform sediments. Thus, the amount of sea-level fall was distinctly smaller than the revised value of 500 m or more suggested by Hoffman (2013), possibly around 100 – 200 metres, or even less. The value depends on the amount of subsidence; in Fig. 17, for graphic reasons, a sea-level drop of less than 100 m was assumed. The observed long-lasting shallowing trend may also imply that eventual glaciations started long before sedimentation of the Ghaub diamictites, as was also assumed by Domack and Hoffman (2011) and Hoffman (2011b).

During the supposed postglacial transgression, starting with the cap carbonates (Fig. 17, sea-level changed from level 3-4 to 5), the platform or ramp in the Southern area was slowly flooded, and the area deepened without any indications of reworking. The situation in the Northern area seems to have been comparable at the beginning; transgressive lag deposits have not been observed in either of the two areas, and no spectacular, sudden melting of ice is indicated. Distinct differences between the Northern and the Southern areas remain. The Northern area was already affected by local synsedimentary tectonics during the cap dolomite sedimentation (Fig. 15A, B), and more severely soon after, during Maieberg T3 time. Turbiditic sheddings, mass flows and slump features are indicative of strong tectonic activity, strong subsidence, and slope failure. There is no way to determine the amount of eustatic sea-level rise in the north, in addition to subsidence. In the Southern area, the transgression started with the cap carbonates as well, but more slowly, because of low subsidence. In the upper parts of the cap rocks, bundles of millimetre-thick but decimetre-long crystal rays after aragonite (Fig. 15C) occur frequently, which are indicative of highly supersaturated, subtidal, very low-energy conditions. The change of accommodation was in this case mainly caused by the assumed eustatic sea-level rise, starting in a very shallow setting, in addition to some subsidence. The Maieberg succession above the cap carbonates consists mostly of micritic carbonates, with thinner shale intercalations. In the reconstruction of Fig. 17, the rising sea-level of stage 5 reached approximately the value of stage 1.

If we employ a different scenario for the Southern area, that the Auros platform/ramp and stratigraphic equivalents would have strongly subsided during Auros time (in contrast to Fig. 17), counterbalanced by a very distinct sea-level fall, the succession would have stayed for some time in shallow water. But thereafter, if flooded by ≥ 500 m of sea-level during the Maieberg Formation, as suggested by Hoffman (2013), carbonate production should soon have ceased. Not much peri-platform ooze would have been available, even less if other less subsiding platforms were emergent and only slowly became flooded, and calcareous plankton did not exist. Therefore, hardly any carbonates would have been deposited in the Southern area during the Maieberg T3 interval on top of the cap carbonates. This is not the case; carbonate sedimentation continued, which supports the assumption of a very moderate, postglacial sea-level rise.

The amplitude of the reconstructed sea-level change is therefore not very different from the changes observed in the Pleistocene and Holocene. Christie-Blick (1997) and Christie-Blick et al. (1999) assumed a similar magnitude based on the depth of incised valleys in Utah of less than 200 m. Other candidates for a much larger sea-level change during Snowball Earth were found in South Australia: the Wonoka canyons, which are in places more than 1 km deep (Christie-Blick et al., 1990). The origin of these canyons was under debate (Christie-Blick et al., 2001), but more recently, they have been interpreted as submarine channels (Giddings et al., 2010).

The observations and reconstructions mentioned above make the presence of a thick ice cover on all continents, including at tropical latitudes, unlikely to be ‘consistent with geologic evidence’ (Hoffman et al., 2009). Far less ice would have existed on Earth, on the continents and in the oceans, possibly less than during the Quaternary. In the OML, carbonate platforms were not eroded or incised. If these processes occurred elsewhere, it may have been dependent on the amount of local subsidence, or even uplift, distance to eventual glaciations, and much less on eustatic fluctuations during Ghaub time.

6 Conclusions

The Ghaub diamictites and underlying or correlative shallow-water carbonates are of equal importance for the interpretation of the Cryogenian glaciations. In the Fransfontein area, Hoffman (2011b), Domack and Hoffman (2011) and Hoffman (2013) assumed a platform with a raised rim and a deeper, dish-like interior as the substrate for continental ice flowing from the shelf down to the distal slope. It is difficult to envisage that the platform rim and interior were not distinctly eroded during this process. In the Ghaub Formation of the OML and in correlative carbonates no indications of ice flowing over a platform down the slope were found. In both areas, the OML and the Fransfontein Ridge, the paleotopography is unclear, and the assumed platform was more likely a ramp. A comparable facies of thick ooid sheddings was found in both areas in the successions underneath the Ghaub Formation. A considerable decline in sea-level during this period, as suggested by the above-mentioned authors, should have fully exposed the platform. In such a setting, the ‘ooid factory’ restricted to shallow-water settings had to move distinctly downslope and should have been positioned in only a small area on the slope, hardly able to deliver the observed large volumes of ooids. If a ramp existed, the production area could move up and down the clinoform more easily and would maintain its extent and productive capacity.

In the central OML, the Ghaub Formation has been found only in the north, where it shows a distinct westward thickening and changes in clast content from base to top and from southeast to northwest. Progradation of different clast sheddings was observed, and is interpreted as debris flows, interrupted by fine-grained waterlaid deposits. The succession bears no fabrics typical of tillites. Some scratched and facetted clasts are present and represent in our opinion redeposited glacial material. The changes of clast content in time and space, including crystalline clasts in the youngest and uppermost unit, indicate deeper and deeper erosion of the hinterland in the east because of uplift of the source area on the order of a few thousand metres. In the Northern area, there is considerable evidence of long-lasting synsedimentary tectonics, which also affected the successions underneath and above the Ghaub Formation. In especially the eastern parts of this area, thick breccia bodies crop out under and over the Ghaub Formation.

Iceberg-rafted debris is extremely rare in the Ghaub Formation of the OML, in contrast to the Fransfontein Ridge. In both areas, dropstones are only found in the presence of Ghaub diamictites. From the Fransfontein area, various authors have described classic glacial rainout deposits, mainly from the uppermost parts of the Ghaub Formation, directly underneath the cap carbonates. In the central OML, dropstones occur in the Northern area only rarely, and even in the cap carbonates. These sparse dropstones indicate that the cap rocks did not form during a super-greenhouse setting, in contrast to the Snowball Earth hypothesis. According to the hard Snowball or Panglacial Earth scenarios, after severe flooding of shelves following massive deglaciation, the distribution of glacial rainout facies should be much more uniform worldwide than is actually observed, and dropstones would be expected to occur almost everywhere, even when diamictites are absent.

A local ice stream may have existed in the Fransfontein area, possibly reaching the ocean. The diamictites of the Ghaub Formation in the OML appear to have been deposited at the forefront of a glacier at some distance. Here, periglacial debris was shed downslope into a deeper, marine realm, possibly by a fan delta. The glaciation itself may have been restricted to mountainous areas farther apart in the east, which must have been strongly uplifted because of synsedimentary tectonics during the late Auros and Ghaub time, and became erosionally downstripped. A wet glacier alone, without tectonics involved, would not have been able to erode thousands of metres of sediment down to the Nosib Group and the crystalline basement without uplift of the source area. The continuous erosion during Ghaub time caused a consistent change in the clast content of this formation, with the stratigraphically deepest rocks on top of the clastic column, and respectively farthest northwestward in the prograding sediment succession. The reason for the strong uplift of the clastic source area(s) of the Ghaub Formation is debatable; a foreland basin (Prave 1996, 1997) or a tectonically unstable shelf (Dürr and Dingeldey, 1997) may be invoked.

Slumped and brecciated cap carbonates were found only in the northeastern part of the Northern area. In the Maieberg Formation overlying the cap carbonates, numerous breccia sheddings and slump features were observed, as well as turbiditic sheddings. These indicate continued basin-wide tectonic impacts, which persisted into the Elandshoek Formation. No quartzitic or crystalline clasts were detected. The source area of the Ghaub clasts was no longer available; only carbonate clasts in slope facies occur in the Maieberg Formation, which indicates widespread slope failures. The coarsest breccia bodies were found in the Maieberg T3 unit close to the mentioned slumped cap carbonates. To the west, breccia bodies only crop out higher in the same Maieberg stratigraphic succession. This finding further confirms westward progradation of sheddings, as in the underlying Ghaub Formation. This scenario of long-lasting synsedimentary tectonics in the north is in sharp contrast with the Southern area, where almost no breccia beds occur. Only in the southernmost part of the study area, at the margin of the Outjo Basin, some intra- and extraformational breccias and slump features were found higher in the Maieberg succession. This Southern area therefore was mostly tectonically stable. The described differences between Northern and Southern area from Auros to Maieberg time mirror a complex palaeogeographic pattern of platforms/ramps and intrashelf basins or basinal inlets. The shelfal area in the south was too high and/or too isolated to be reached by diamictite or other sheddings. In contrast, the Northern area was affected again and again by fine- and coarse-grained sediment input as well as breccia sheddings.

In our opinion, the enormous sea-level change postulated by, for example, Hoffman (2011b) and Domack and Hoffman (2011), as well as its distinctly reduced version (Hoffman, 2013), is based on several not robust assumptions and interpretations, as discussed in detail in the preceding chapters. The shallowing-upward trend observed in our working area in the south, with change from the Auros shales to columnar stromatolites to wavy microbial mats with tubes (without any diamictites) overlain by cap carbonates of shallow-water origin might be explained by sediment infill, but only, if subsidence came to a complete halt, which is implausible. The shallowing-upward succession can be interpreted as product of a modest sea-level fall largely counterbalanced by subsidence. A fall of 100 – 200 m or less can be assumed, which brings the sedimentary interface from below storm wave base to the photic zone below fairweather wave base, and finally to shallow subtidal depths. Of particular importance is the missing unconformity within this shallowing-upward succession. The sea-level fall was not pronounced enough and its rate too slow to subaerially expose the sediments in the Southern area. This scenario implies a distinctly smaller sea-level drop than that proposed for the Fransfontein Ridge and upscaled globally from there. The assumed sea-level fall corresponds with published data from Utah, where evidence for ≥ 160 m drawdown of sea-level was found, based on the depths of incised valleys (Christie-Blick, 1997; Christie-Blick et al., 1999). All estimates on the amount of (eustatic) sea-level fall strongly depend on the subsidence in the respective area. In the scenario of Fig. 17, a very low subsidence of the Southern area is adopted; sea-level fall is therefore reduced. If a stronger subsidence would be assumed, the sea-level drop would be accordingly larger. In the Northern area, subsidence and sedimentation rates were much stronger, it is almost impossible to apply water depths to the successions in this area. More robust conclusions on sea-level fluctuations were gathered from the comparison of the data about the Northern and Southern area.

This assumed moderate sea-level change is therefore not very different from Quaternary ones, which, however, exposed carbonate platforms and shelves. The amount of sea-level drop, not exposing a carbonate platform or upper ramp, provides strong evidence against a glaciation with a thick ice cover on all continents and thickly frozen oceans even at equatorial latitudes. Far less ice would have been present than is assumed in the Snowball/Panglacial hypotheses. Glaciations on equatorial continents may have been regional, concentrated in patches of glacial ice, and/or restricted to higher mountains, providing ice rafted debris only locally at the forefront of some icestreams that reached the ocean. A temporal and facies relationship with tectonism, especially the fragmentation of Rodinia, is evident (Eyles and Januszczak, 2004, 2007; Eyles, 2008). Much of the continents may have been covered by thin ice (model of Pollard and Kasting, 2005), if at all, whereas the equatorial oceans were at least partially ice-free.

How could evolution continue and pro- and eukaryotic biota survive during a Snowball Earth? The discussion clearly shows that for the survival of these organisms, 'refugia' on or under the ice would not be sufficient. The palaeontological data do not allow a (hard) Snowball scenario, but require at least some open oceanic water in the tropics, which is confirmed by many other lines of evidence discussed. One of the main arguments against thick ice is the relatively moderate observed fluctuation of sea-level. An extreme and fast sea-level rise because of a sudden "greenhouse stage", causing melting of thousands of metres of ice is not possible. A Waterbelt Earth is therefore far more likely than a (hard) Snowball or Panglacial Earth. Modest sea-level change would also be compatible with a glacial climate of a High-Obliquity Earth (e.g. Williams et al., 2016), which would imply an only very partial glaciation of Earth.

The few studies that have applied the CIA weathering index for the 'Marinoan' of Namibia and Oman yielded evidence of variable climate conditions, oscillating between dry-cool/glacial and warm-humid/interglacial. The cold climate modes of the Cryogenian were cyclical, punctuated with well-defined warm-humid interglacial periods. Functioning hydrological cycles and operative sediment routing systems were also active throughout the glacial intervals, which requires substantial open ocean water (Rieu et al., 2007; Bahlburg and Dobrzinski, 2011).

According to Hoffman (2011b), the arguments against Snowball Earth that rely on wet-based glacial sedimentology do not consider 'that the preserved glacial record reflects the end of the Snowball Earth, when melting was bound to emerge triumphant'. The unconformity beneath the diamictites represents the period of the hard Snowball according to this author. If all Ghaub deposits are associated with late- or postglacial melting, and the main glacial phase is represented by an unconformity, where is the record of the encroaching glaciers below? Where is the laterally correlatable unconformity on the shelves and platforms/ramps? In Namibia, features diagnostic for advance and for melting of Snowball-like large, widespread glaciations, such as pebble plasters, moraines, correlatable tillite occurrences, and scratched surfaces below tillites, are missing in the shelfal areas. No glacial soft-sediment deformation structures, as described by Busfield and Le Heron

(2013) from the Chuos Formation of the Otavi Mountainland, have been observed in the Ghaub Formation. In addition, in a Snowball world, shelves should have been heavily flooded after the Snowball deglaciation; clear reworking features and lag deposits would occur again and again, especially in shallow-water settings, but evidence of these phenomena is not observed. The lack of these features indicates much smaller glaciations in a Waterbelt or High-Obliquity Earth.

Williams et al. (2016) presented an appealing interpretation of the effects of a High-Obliquity Earth in the Proterozoic that explains many controversial observations on seasonality in the equatorial Cryogenian successions. Important findings are large ‘trains’ of sand wedges and non-tidal, seasonal (annual) oscillations of sea-level, explained by an eccentric lunar orbit. A very interesting point raised by these authors is the apparent absence of Neoproterozoic glaciations in polar regions, where, in a High-Obliquity Earth, fallen snow would have melted in summer; therefore, glacial records are restricted to equatorial regions, not only in the Neoproterozoic but also before. This implies a much smaller sea-level drop than in a Snowball world. The problem of this scenario, as mentioned by these authors, is how and when Earth could have achieved its present obliquity, which is difficult based on our present understanding. Williams et al. (2016) criticised that in the research on the ‘Neoproterozoic climatic paradox’, conflicting observations or anomalies have often been explained away with additional assumptions, which weaken the original model. The authors cited Kuhn (1970), who noted that scientists may, when faced by anomaly, ‘devise numerous articulations and ad hoc modifications of their theory in order to eliminate any apparent conflict.’ The assumption that the main phase of the Snowball Earth glaciation is hidden in the unconformity beneath the diamictites (Hoffman, 2011b) and can therefore not be properly investigated is one of these counterproductive later modifications of a theory. ‘Geology is about what happened, not what should have happened’ (Grotzinger et al., 2011); the sedimentary record provides the ultimate test for the theory.

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FIGURE CAPTIONS

Fig. 1. Geologic overview of northern Namibia with the main structural units. The Fransfontein Ridge is situated north of Khorixas and Outjo, the Otavi Mountainland between Otavi, Tsumeb and Grootfontein. Modified after Geological Survey of Namibia, 1999; Hoffman, 2004; Laukamp, 2006; http://www.snowballearth.org/slides/Ch1-16.gif.

Fig. 2. Stratigraphic setting of the Damara Supergroup in the Cryogenian and Ediacaran of the Otavi Mountainland (northeastern Namibia). Modified after Geological Survey of Namibia (1999), with ages of some strata: Askevold Formation, Hoffman et al. (1996); Ghaub Formation, Hoffmann et al. (2004); Mulden Group, Melcher et al. (2003).

Fig. 3. Stratigraphic/palaeogeographic setting of the Damara Supergroup in the Fransfontein Ridge (after Hoffman 2002, and thereafter, slightly modified). The Huab and Makalani ridges strongly influence sedimentation of the shelf succession at the margin toward more open marine areas. In the Otavi Mountainland, counterparts of these ridges exist in the Grootfontein area (see Fig. 4).

Fig. 4. Geological sketch map of the Otavi Mountainland (modified after Geological Survey of Namibia, 1999; Laukamp, 2006). Encroachment of the Damara succession onto the crystalline ridges is clearly visible. The main farmhouses in the working area are marked (J = Jakkals Omuramba; A = Auros; G = Gauss).

Fig. 5. Geologic map of the Jakkals Omuramba farm area, including the (older) Keilberg and Maieberg farm plots. The Ghaub Formation occurs only in the Northern area and thins out southeastward. The tubestone facies of the Auros Formation occurs only in the Southern area.

Fig. 6. Idealised stratigraphic section of the Keilberg North area from the uppermost Gauss to the lower Maieberg Formation. The thick oolite intervals in the Auros Formation are characteristic for the northwestern part; southeastward they are partly replaced by laminites and breccia intervals. Only the

approximate thickness of the Ghaub Formation is shown; for details to the Ghaub Formation see Fig. 13. CC = cap carbonates of the lowermost Maieberg Formation (= Keilberg Member).

Fig. 7. Idealised stratigraphic sections of the Keilberg South area (left) and part of the cored OKS1 well (right). Abbreviation: CC = cap carbonates of the lowermost Maieberg Formation (= Keilberg Member).

The Keilberg South section shows the interval from the uppermost Gauss to the lower Maieberg Formation. The tubestone facies in the upper Auros Formation is concentrated in three intervals, separated by two layers nearly without tubes.

The OKS1 well, about 50 km southwest of the Otavi Mountainland, drilled the Swakop Group slope to basinal deposits (coeval to the Otavi Group shelf successions). The measured part of the cores shows a shaly-silty succession at the base (likely an Auros Formation equivalent), overlain by the Ghaub Formation (with two diamictite intervals separated by silty shales comparable to those below the diamictites), followed by the Maieberg Formation. The petrology of the diamictite clasts in the cores is highly variable.

Fig. 8. Uppermost Auros Formation, Northern area. (A) Partly silicified calcitic-dolomitic laminite interval (lower part) overlain by bedded, strongly silicified oolite; contact just above the tip of the hammer. White bands in “zebra-laminites” are dolomite; dark bands are calcite. Small slump structure in the lower part. Hammer for scale. (B) Well-sorted oolite with mm-sized ooids.

Fig. 9. Columnar stromatolites, Southern area. (A) Tightly grown, strongly silicified dolomitic columnar stromatolites, which weather out distinctly. Sedimentary infill between the columns is much less silicified dolomite. Keilberg South. Hammer for scale. (B) Perpendicular and oblique section of some columns, showing circumferential symmetry. Pen for scale. (C) Detail of same outcrop; symmetric growth is especially apparent because of the laterally variable amounts of silicification. Northern Auros Farm, near farmhouse. Hammer for scale.

Fig. 10. Tubestones with different types of cracks, Southern area. (A) Tubes perpendicular to bedding are often associated with bed-parallel sheet cracks (Keilberg South). Lens cap for scale. (B) Detail of association of sheet cracks with one funnel-like feature, and tubes perpendicular to bedding (northeast of Auros farmhouse). Bar in upper left corresponds to a pen. (C) The tubes exhibit irregular diameters along their lengths, are thickening and thinning and show lateral deflections (northeast of Auros farmhouse). 2.3 cm coin for scale.

Fig. 11. Silicified tubestones with cm-sized tubes, Southern area. (A) In contrast to Fig. 10C, the tubes are almost perfectly perpendicular to bedding and have uniform thicknesses. The contrast to Fig. 10 may be related to differences in lithification at the time of tube formation (Keilberg South area). Hammer for scale. (B) Surface of tubestone bed with small, silicified knobs. These knobs protrude from a few mm to about 1 cm above bedding planes because of their greater resistance to weathering, and represent the upper end of tubes (north of Auros farmhouse). Lens cap for scale.

Fig. 12. Ghaub Formation clasts, Northern area. (A) Most of the carbonate clasts are not well sorted and are angular–subangular. Hammer for scale. (B) Well-rounded Ghaub Formation clasts consist mostly of sandstones, quarzites and basement lithologies. The difference to the angular carbonate clasts (12A) is likely caused by reworking of already well-rounded components from the underlying Nosib Group. (C) Silt-stringer, interbedded within the Ghaub Formation. Much of the formation weathers rusty-brown because of its high pyrite content, which might be responsible for the sparse plant cover of the Ghaub interval, in contrast to the under- and overlying lithologies. Hammer for scale. (D) Ghaub clast with striations at high angle to the metamorphic lamination. 2.3 cm coin for scale.

Fig. 13. Four sections of the Ghaub Formation (Northern area), showing the lateral changes of the three main clast zones.

Fig. 14. Cap carbonates (Otavi Mountainland, A, B) and Franni Aus Member (Fransfontein Mountains, C). (A) The several-metre-thick cap carbonates are irregularly laminated and show variable, wavy bedding. Northern area. Hammer for scale. (B) Bedding surface with two differently sized dropstones, showing impact structures. The isolated, mm- to cm-sized clasts show downward depressed rims around the component where liquefaction may have occurred. The radial pattern of many fissures around the larger clast indicates that the sediment was already semi-plastic at the time of impact. Lower parts of the cap carbonates, Northern area. Scale in cm. (C) Classic dropstones, characterised by depressions under many clasts and draping above, which demonstrates that the glacial rainout occurred at a time of finer clastic sedimentation. 2.3 cm coin for scale.

Fig. 15. Cap carbonates. (A) Slumped cap carbonates near the eastern boundary of Keilberg North. Lens cap for scale. (B) Irregularly brecciated cap carbonates, same locality as 15A. 2.3 cm coin for scale. (C) Thin and elongated crystal ray bundles and fans in cap carbonates (Keilberg South). 2.2 cm coin for scale.

Fig. 16. Maieberg Formation, Wolkenhaube area (southern part of Auros farm). (A) Breccia body, overlain by unbrecciated bed. Hammmer for scale. (B) Slumped interval. Hammer (near small, isolated plant) for scale.

Fig 17. Palaeogeographic sketches from the Northern, mostly basinal area, to the Southern, mostly shallow water area. Input of clastic sediment is assumed from the east. The Ghaub Formation and stratigraphic equivalents in between the Northern and the Southern area are not outcropping, but because of the strongly differing subsidence of both areas, synsedimentary faults have to be assumed. The horizontal distance is not to scale; see geologic map of Fig. 5. Sea-level fluctuations 1–5 are partly counterbalanced by subsidence and differences in sedimen input; sea-level changes would be larger, if higher subsidence is assumed.

Stage 1, Auros shale. Already at that time the succession is thicker in the north than in the south. Sea-level 1 was high.

Stage 2, Oolites and columnar stromatolites of the Auros Formation. In the north, thick oolite sheddings interfinger with laminated carbonates, which dominate at the top. In the south, the succession above the Auros shales consists of columnar stromatolites, deposited in a much shallower depositional setting than the shedded oolites in the north. Despite a sea-level fall (level 2), the Northern area remained deep because of strong subsidence, only partly compensated by sediment input. The Southern area subsided much less than the Northern area and shallowed.

Stage 3. Diamictites of the Ghaub Formation, Tube facies of the upper Auros Formation. Subsidence of the Northern area continued creating further accommodation for the emplacement of the Ghaub diamictites. Despite continued subsidence, the sea-level fall (reaching its minimum stand at level 3–4) and the high sedimentation rates of prograding diamictites caused shallowing. The Southern area had little subsidence, laminites containing the tube facies were deposited and the area further shallowed.

Stage 4. Cap carbonates of the Lower Maieberg Formation (T2). The sea-level 4 corresponds either largely to the one of the preceding stage 3, or it already started to rise; during the waning glaciation cap carbonates were deposited in Northern and Southern area. The environmental setting was shallower in the south and possibly more restricted.

Stage 5. Upper Maieberg Formation (T3). Subsidence was re-initiated, much more strongly in the Northern than in the Southern area, and sea-level 5 rose strongly. In the north, after deposition of calcareous shales, turbidites and debris flows follow. The environment in the south was distinctly shallower than in the north and not affected by deposition of coarse clastics.

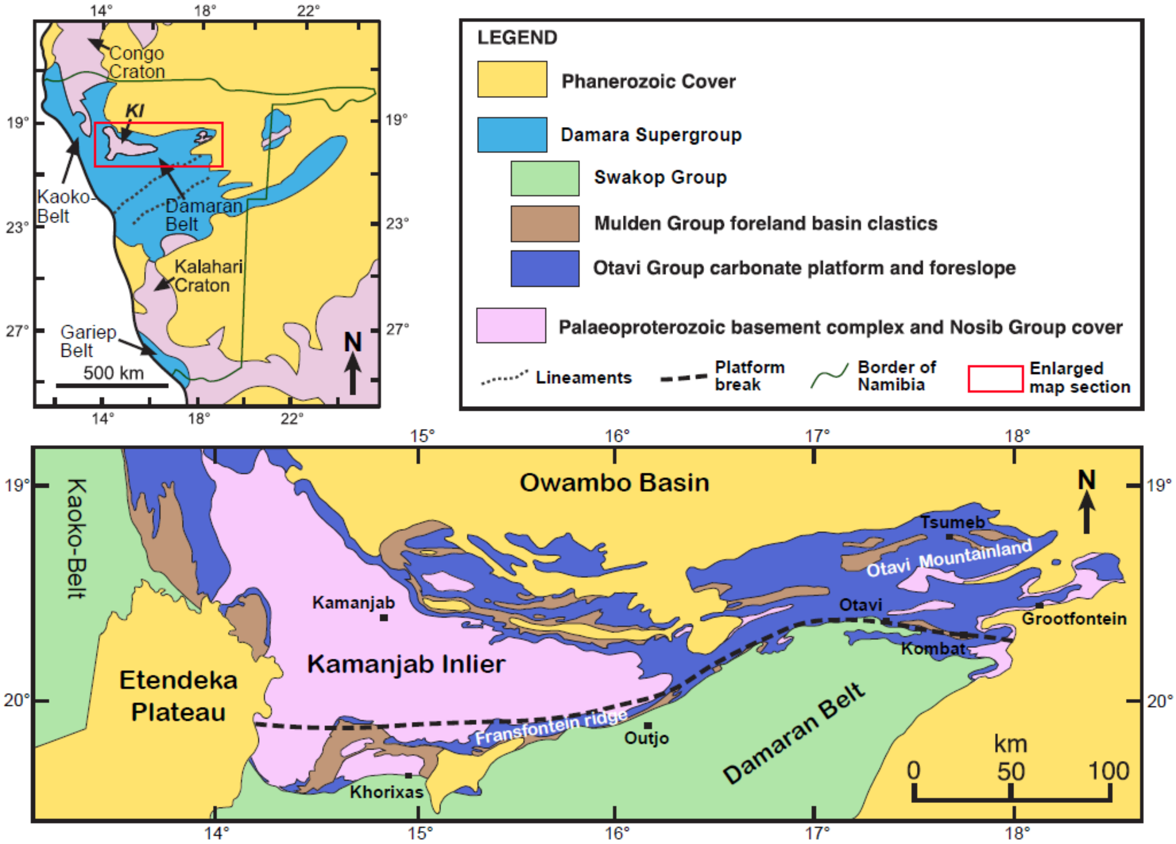
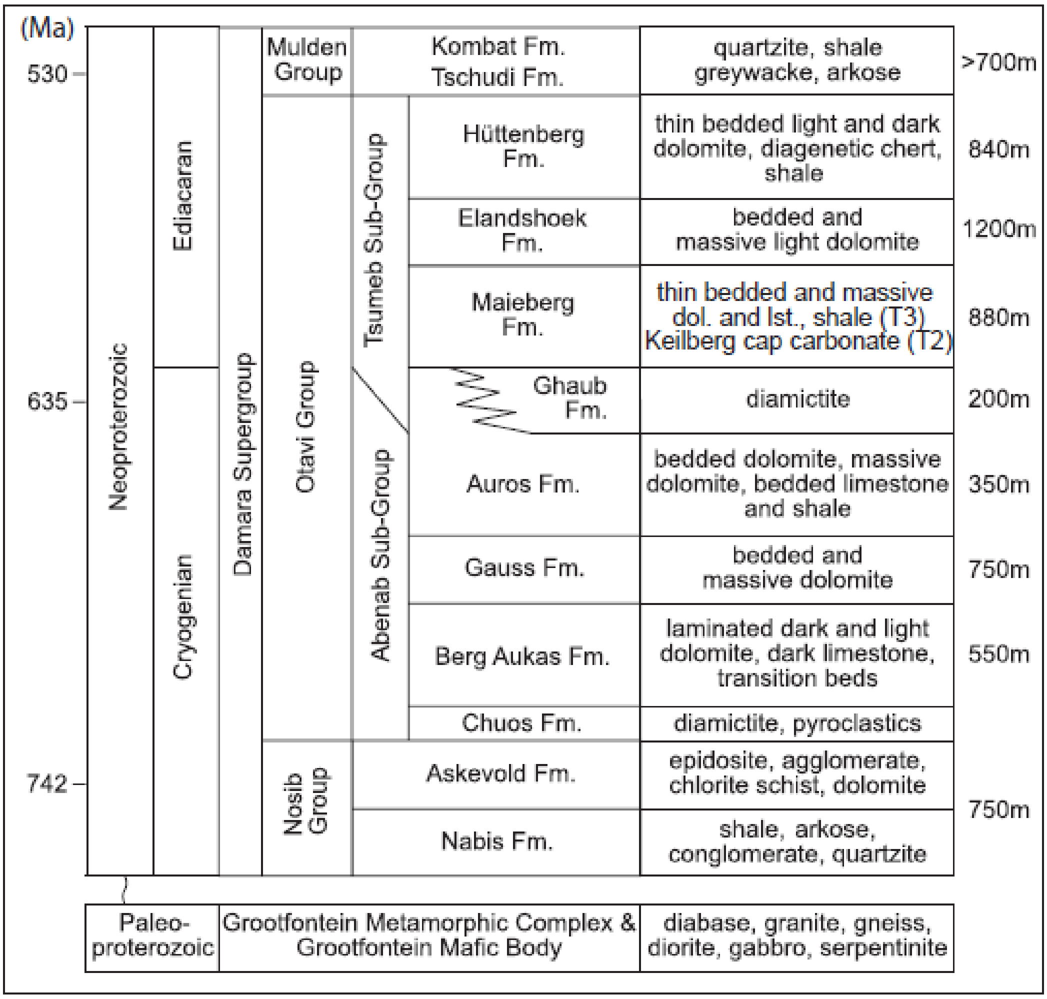
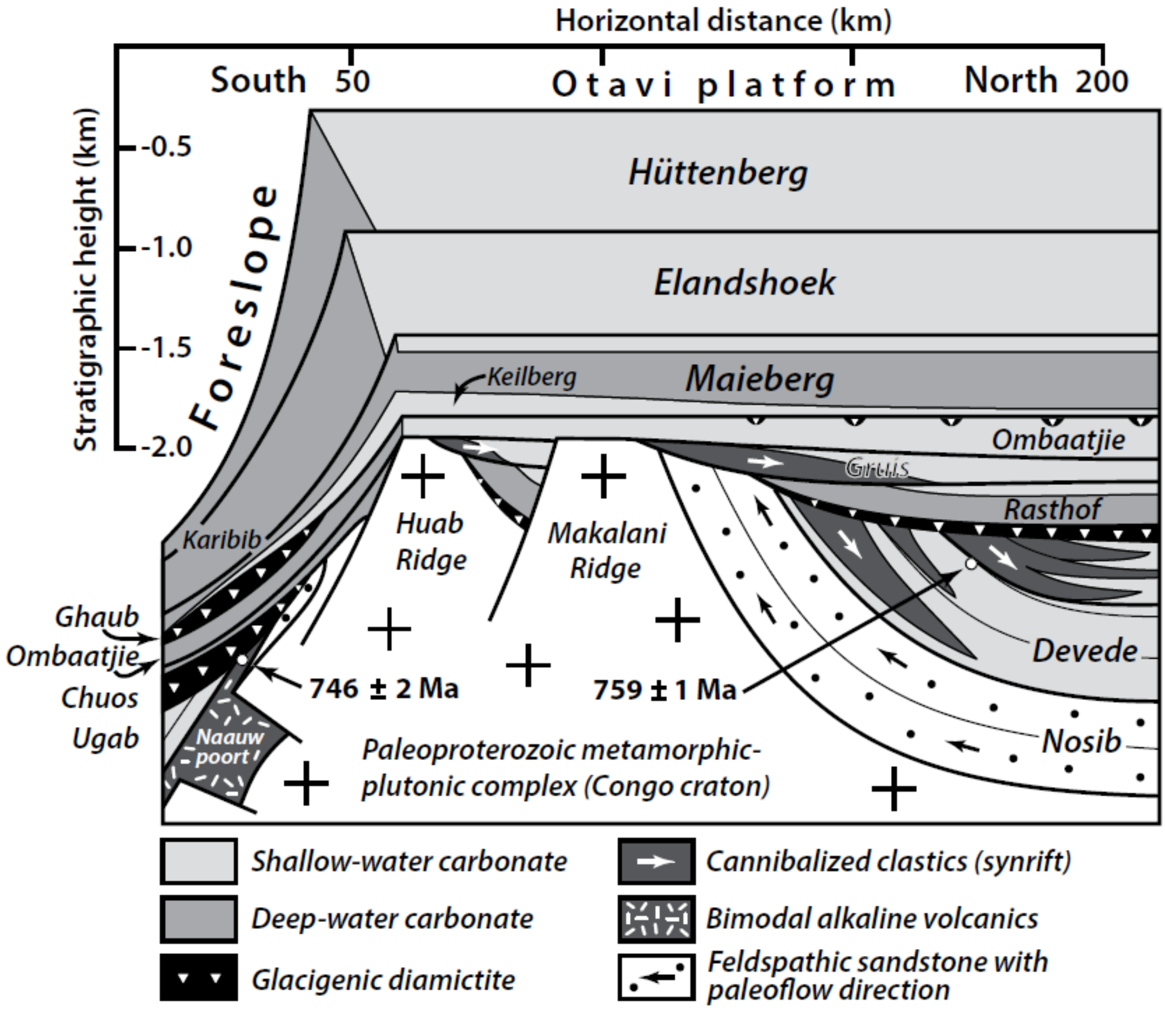

14°
18°
22°
19°
23°
27°
Congo Craton
KI
Kaoko-Belt
Damaran Belt
Kalahari Craton
Gariep Belt
500 km
N
LEGEND
Phanerozoic Cover
Damara Supergroup
Swakop Group
Mulden Group foreland basin clastics
Otavi Group carbonate platform and foreslope
Palaeoproterozoic basement complex and Nosib Group cover
Lineaments
Platform break
Border of Namibia
Enlarged map section
15°
16°
17°
18°
19°
20°
Owambo Basin
Kaoko-Belt
Tsumeb
Otavi Mountainland
Otavi
Grootfontein
Kombat
Kamanjab
Kamanjab Inlier
Etendeka Plateau
Fransfontein ridge
Outjo
Khorixas
Damaran Belt
km
0
50
100
14°

Figure 1



(Ma)
530
635
742
Neoproterozoic
Ediacaran
Cryogenian
Damara Supergroup
Mulden Group
Kombat Fm.
Tschudi Fm.
quartzite, shale
greywacke, arkose
>700m
Otavi Group
Tsumeb Sub-Group
Hüttenberg Fm.
thin bedded light and dark dolomite, diagenetic chert, shale
840m
Elandshoek Fm.
bedded and massive light dolomite
1200m
Maieberg Fm.
thin bedded and massive dol. and lst., shale (T3)
Keilberg cap carbonate (T2)
880m
Ghaub Fm.
diamictite
200m
Abenab Sub-Group
Auros Fm.
bedded dolomite, massive dolomite, bedded limestone and shale
350m
Gauss Fm.
bedded and massive dolomite
750m
Berg Aukas Fm.
laminated dark and light dolomite, dark limestone, transition beds
550m
Chuos Fm.
diamictite, pyroclastics
Nosib Group
Askevold Fm.
epidosite, agglomerate, chlorite schist, dolomite
Nabis Fm.
shale, arkose, conglomerate, quartzite
750m
Paleo-proterozoic
Grootfontein Metamorphic Complex &
Grootfontein Mafic Body
diabase, granite, gneiss, diorite, gabbro, serpentinite

Figure 2



Horizontal distance (km)
South
50
Otavi platform
North
200
Stratigraphic height (km)
-0.5
-1.0
-1.5
-2.0
Foreslope
Hüttenberg
Elandshoek
Keilberg
Maieberg
Ombaatjie
Gruis
Rasthof
Karibib
Huab Ridge
Makalani Ridge
Devede
Ghaub
Ombaatjie
Chuos
Ugab
746 ± 2 Ma
759 ± 1 Ma
Nosib
Naauw poort
Paleoproterozoic metamorphic-plutonic complex (Congo craton)
Shallow-water carbonate
Deep-water carbonate
Glacigenic diamictite
Cannibalized clastics (synrift)
Bimodal alkaline volcanics
Feldspathic sandstone with paleoflow direction

Figure 3

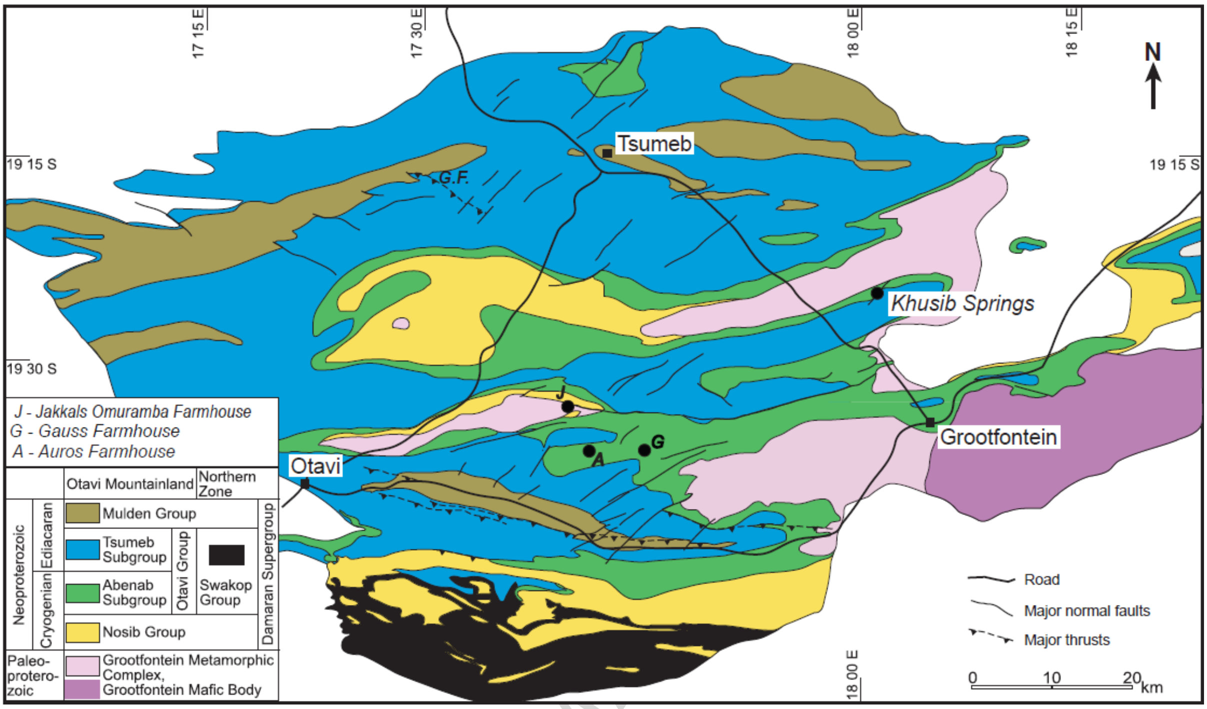
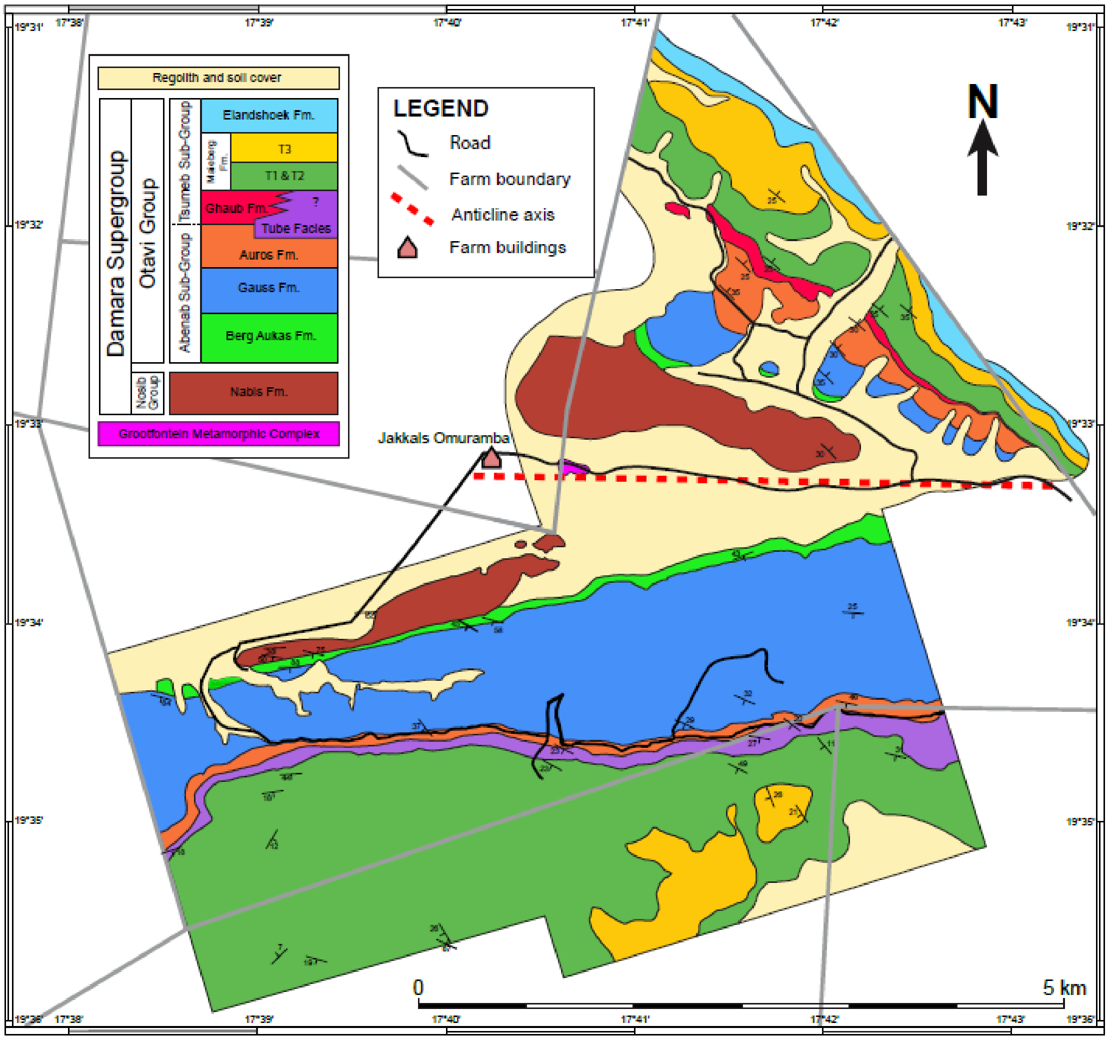
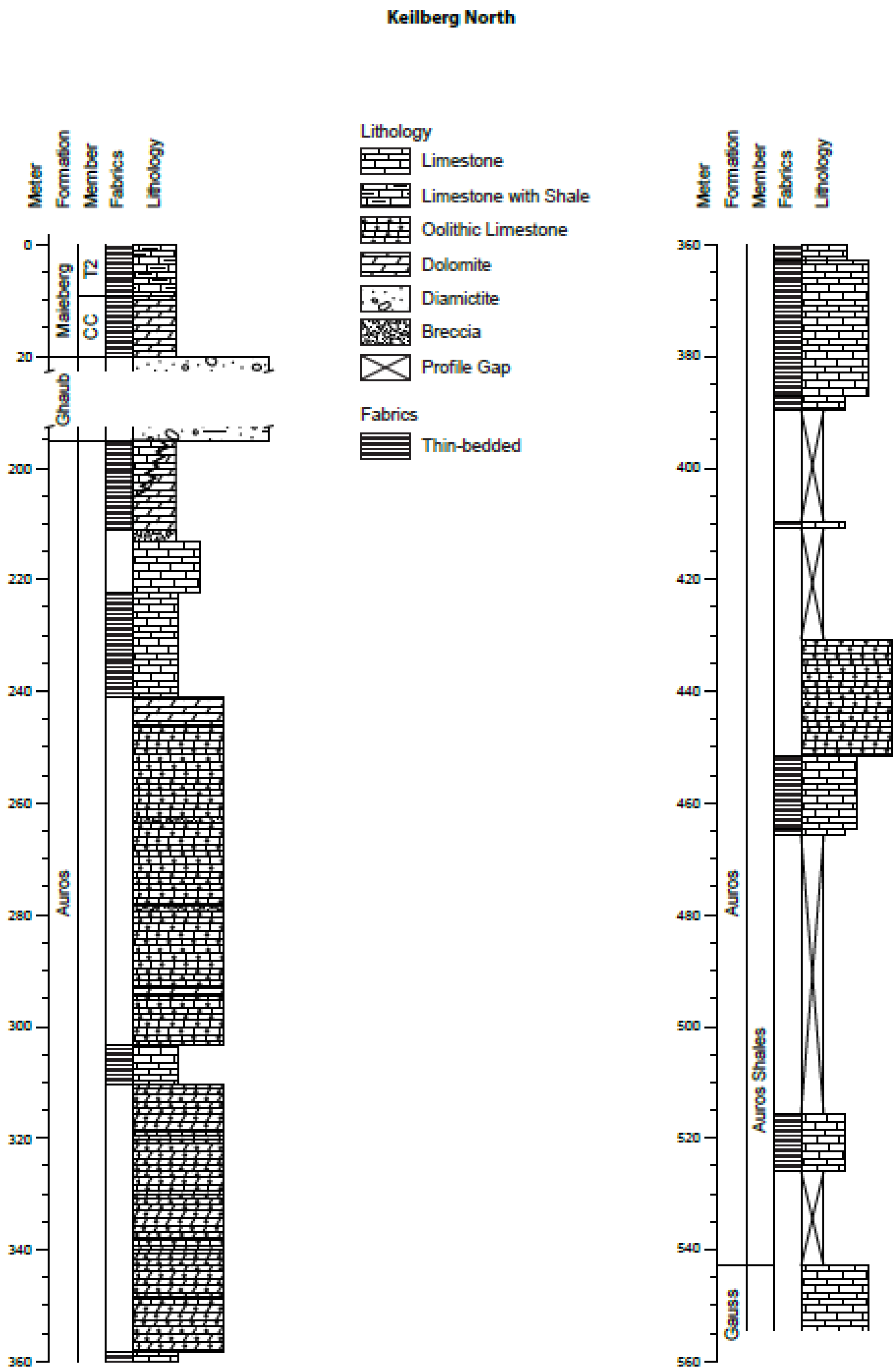

17 15 E
17 30 E
18 00 E
18 15 E
N
19 15 S
19 15 S
19 30 S
Tsumeb
G.F.
Khusib Springs
J
A
G
Otavi
Grootfontein
J - Jakkals Omuramba Farmhouse
G - Gauss Farmhouse
A - Auros Farmhouse
Otavi Mountainland
Northern Zone
Neoproterozoic
Cryogenian
Ediacaran
Mulden Group
Tsumeb Subgroup
Abenab Subgroup
Otavi Group
Swakop Group
Nosib Group
Damaran Supergroup
Paleo-protero-zoic
Grootfontein Metamorphic Complex,
Grootfontein Mafic Body
Road
Major normal faults
Major thrusts
0
10
20
km

Figure 4



Regolith and soil cover
Damara Supergroup
Otavi Group
Tsumeb Sub-Group
Abenab Sub-Group
Nosib Group
Elandshoek Fm.
Maieberg Fm.
T3
T1 & T2
Ghaub Fm.
?
Tube Facies
Auros Fm.
Gauss Fm.
Berg Aukas Fm.
Nabis Fm.
Grootfontein Metamorphic Complex
LEGEND
Road
Farm boundary
Anticline axis
Farm buildings
N
Jakkals Omuramba
0
5 km

Figure 5



Keilberg North
Meter
Formation
Member
Fabrics
Lithology
Maieberg
T2
CC
Ghaub
Auros
Auros Shales
Gauss
0
20
200
220
240
260
280
300
320
340
360
380
400
420
440
460
480
500
520
540
560
Lithology
Limestone
Limestone with Shale
Oolithic Limestone
Dolomite
Diamictite
Breccia
Profile Gap
Fabrics
Thin-bedded

Figure 6

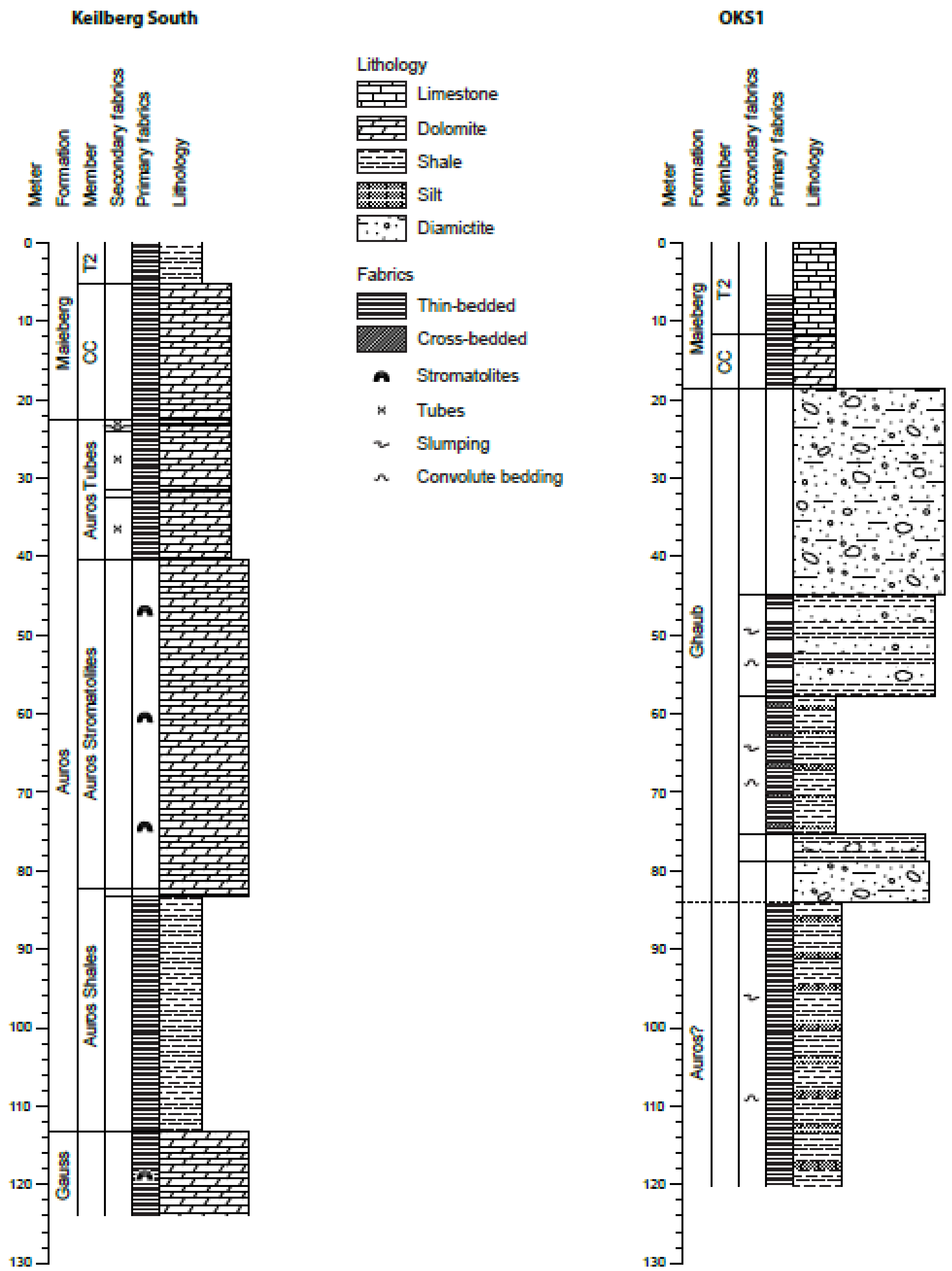

Keilberg South
OKS1
Meter
Formation
Member
Secondary fabrics
Primary fabrics
Lithology
Maieberg
T2
CC
Auros
Auros Tubes
Auros Stromatolites
Auros Shales
Gauss
Ghaub
Auros?
Lithology
Limestone
Dolomite
Shale
Silt
Diamictite
Fabrics
Thin-bedded
Cross-bedded
Stromatolites
Tubes
Slumping
Convolute bedding

Figure 7

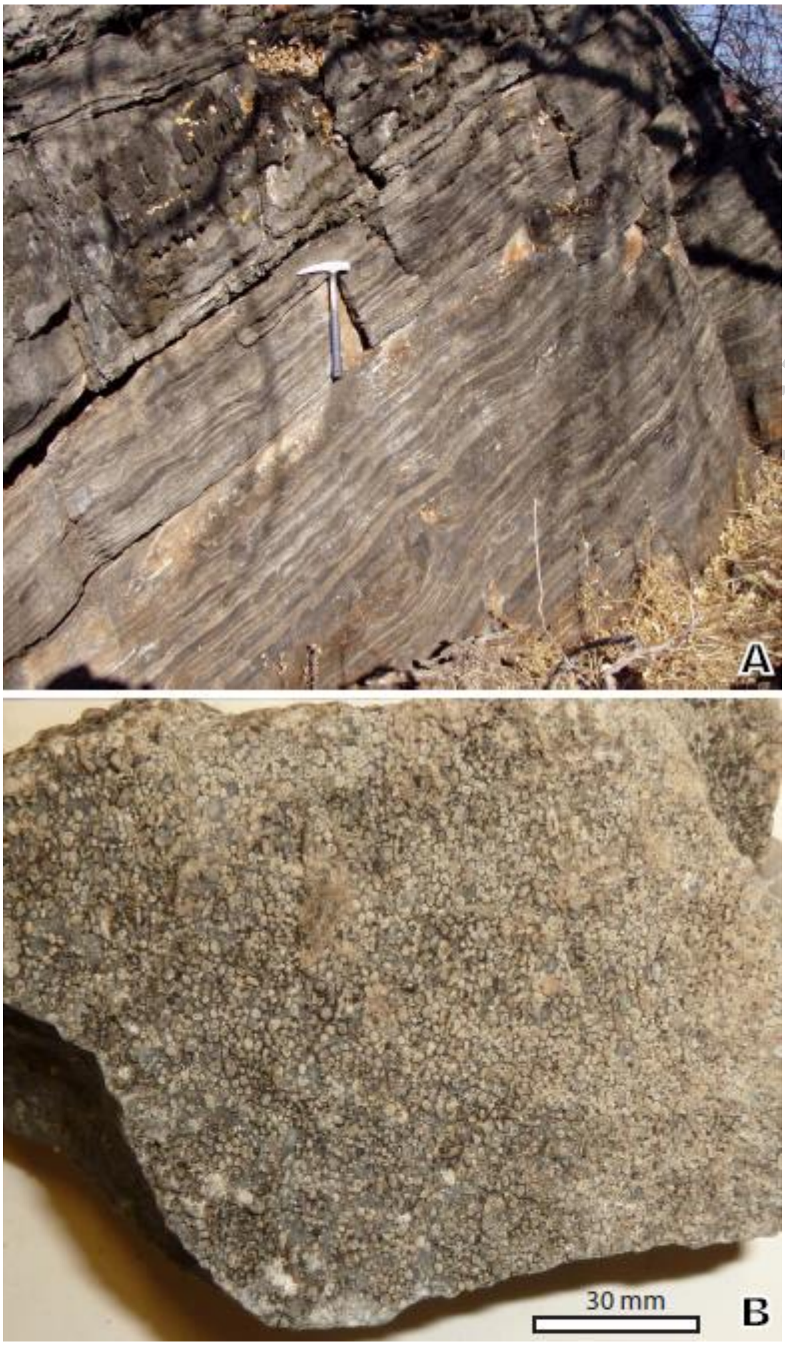


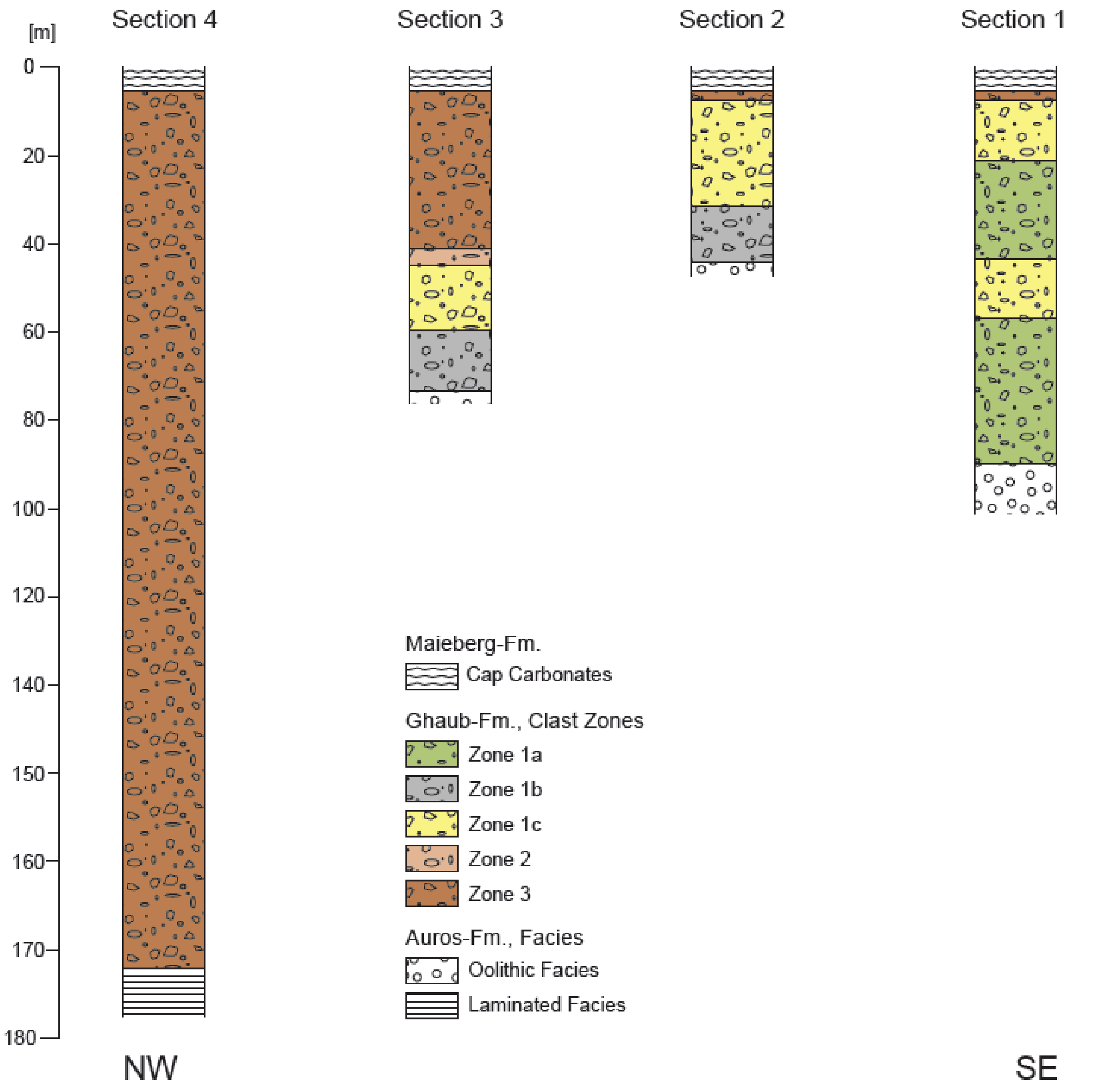
Figure 8

Figure 9

Figure 10

Figure 11

Figure 12



Section 4
Section 3
Section 2
Section 1
[m]
0
20
40
60
80
100
120
140
150
160
170
180
Maieberg-Fm.
Cap Carbonates
Ghaub-Fm., Clast Zones
Zone 1a
Zone 1b
Zone 1c
Zone 2
Zone 3
Auros-Fm., Facies
Oolithic Facies
Laminated Facies
NW
SE

Figure 13

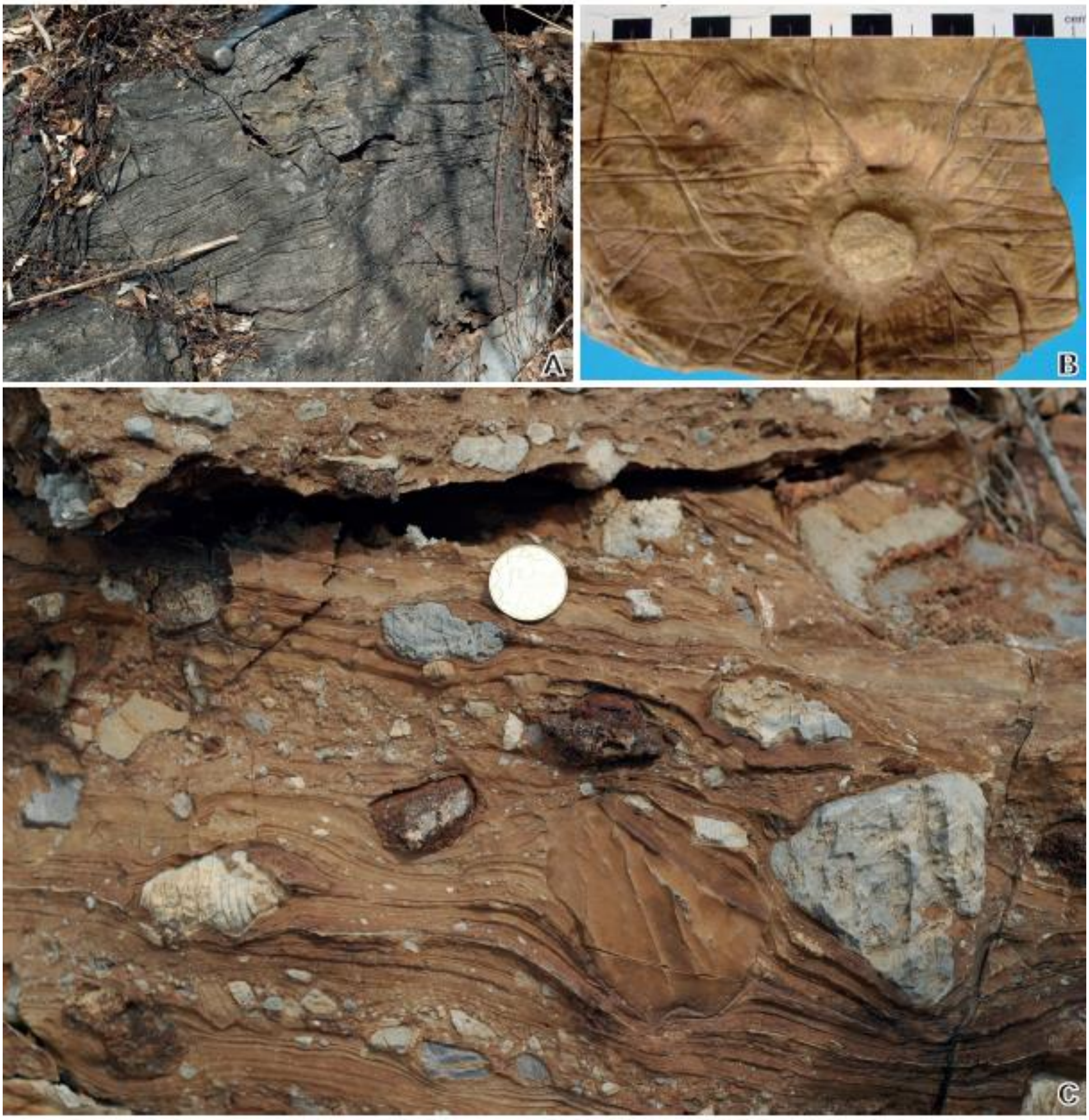


Figure 14

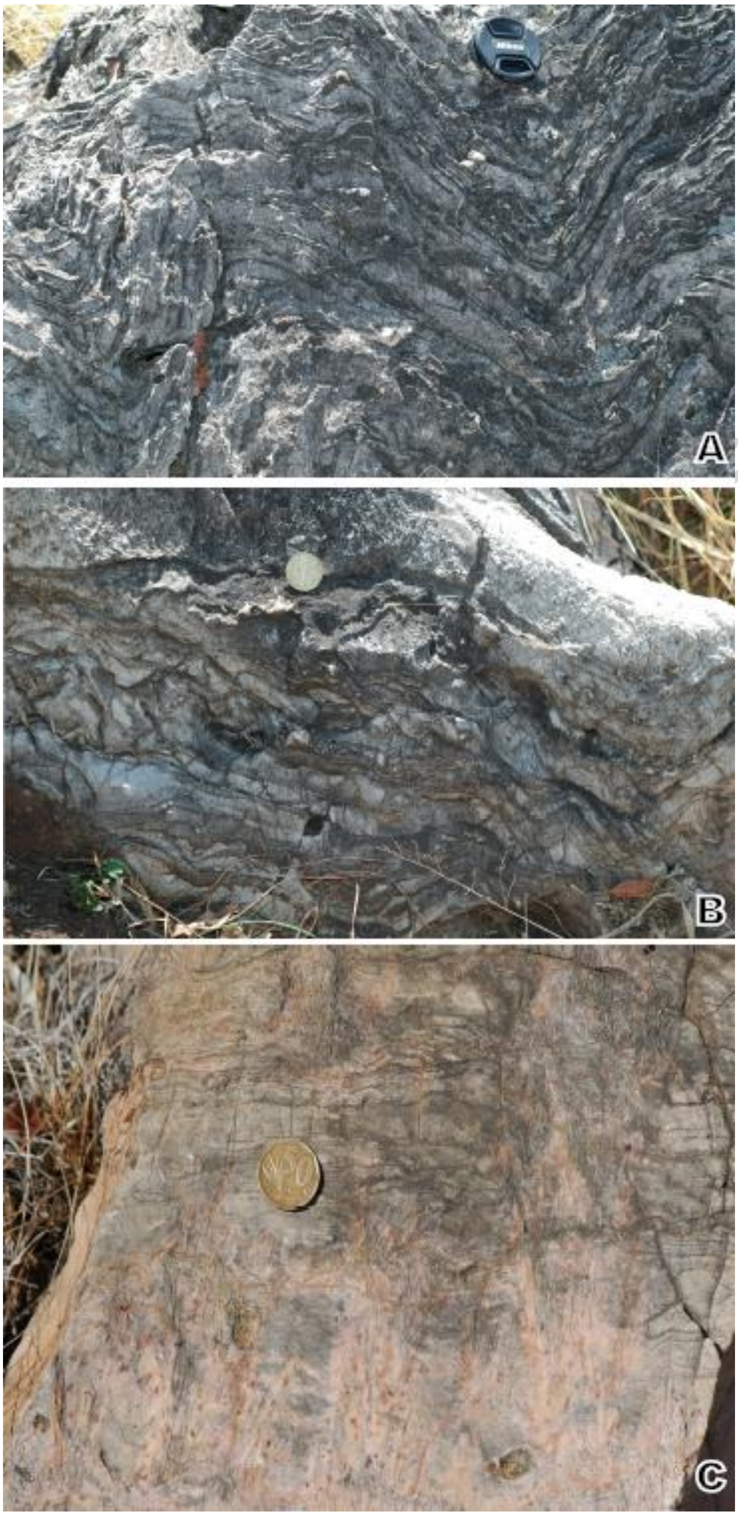


Figure 15

Figure 16

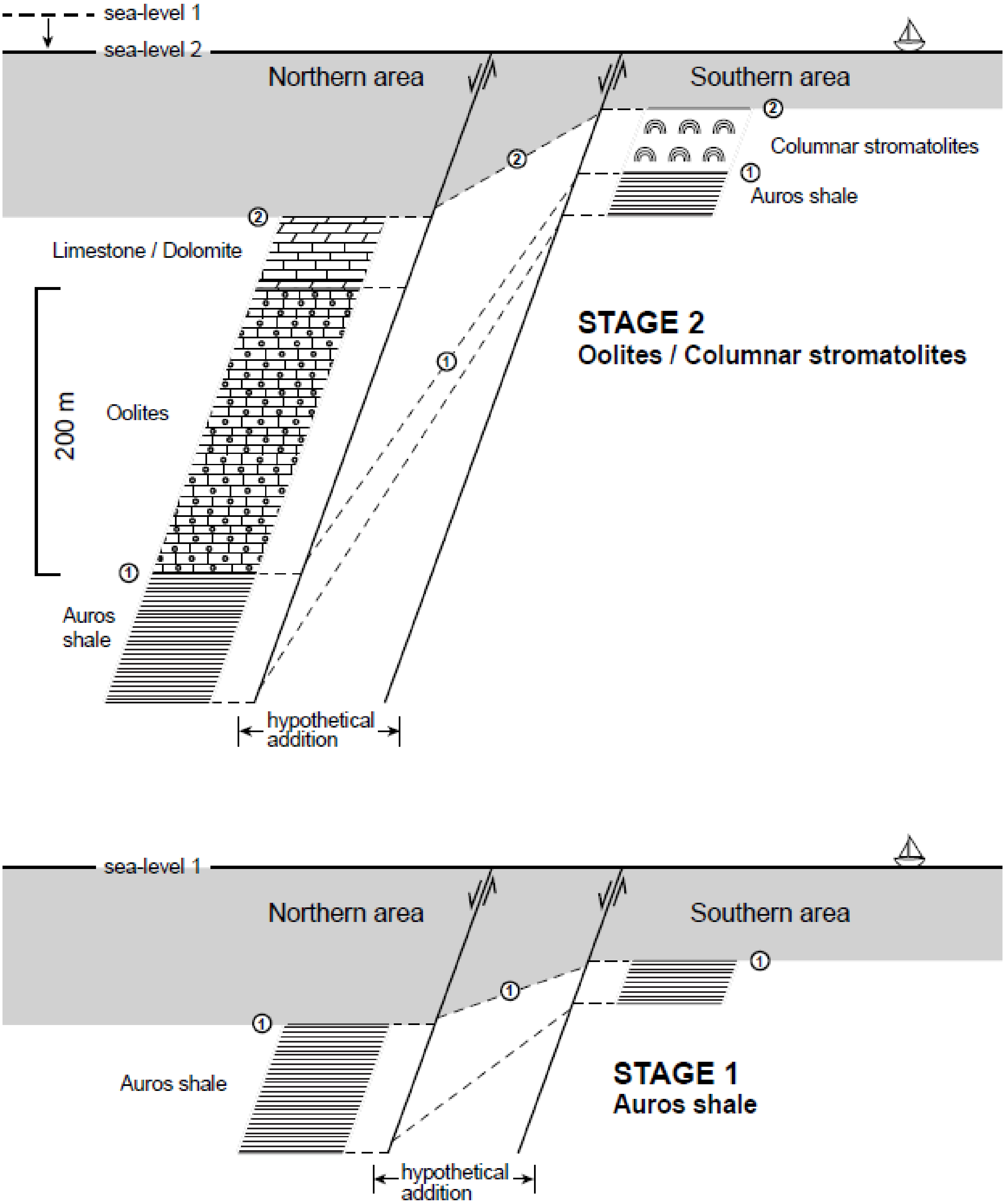

sea-level 1
sea-level 2
Northern area
Southern area
②
Columnar stromatolites
①
Auros shale
②
②
Limestone / Dolomite
STAGE 2
Oolites / Columnar stromatolites
①
200 m
Oolites
①
Auros shale
hypothetical addition
sea-level 1
Northern area
Southern area
①
①
①
Auros shale
STAGE 1
Auros shale
hypothetical addition

Figure 17