Continental carbonate facies of a Neoproterozoic panglaciation, NE Svalbard

IAN J. FAIRCHILD*, EDWARD J. FLEMING*†‡, HUIMING BAO§, DOUGLAS I. BENN†¶, IAN BOOMER*, YURI V. DUBLYANSKY□, GALEN P. HALVERSON◊, MICHAEL J. HAMBREY⌂, CHRIS HENDY☺, EMILY A. MCMILLAN*, CHRISTOPH SPÖTL□, CARL T.E. STEVENSON* AND PETER M. WYNN●

*School of Geography, Earth and Environmental Sciences, University of Birmingham B15 2TT, UK. (E-mail: i.j.fairchild@bham.ac.uk)

†Department of Geology, The University Centre in Svalbard (UNIS), N-9171 Longyearbyen, Norway

‡Present address: CASP, West Building, 181A Huntingdon Road, Cambridge, CB3 0DH, UK

§Department of Geology and Geophysics, E235 Howe-Russell Complex, Louisiana State University, Baton Rouge, LA 70803, USA

¶School of Geography and Geosciences, University of St Andrews, St Andrews KY16 8YA, Scotland, UK

◊Department of Earth & Planetary Sciences/Geotop,3450 University St. Montreal, Quebec H3A 0E8, Canada

⌂Department of Geography and Earth Sciences, Aberystwyth University, Aberystwyth, Ceredigion SY23 3DB, UK

☺Department of Chemistry, University of Waikato, Private Bag 3105, Hamilton 3240, New Zealand.

□Institut für Geologie, Leopold-Franzens-Universität Innsbruck, Innrain 52, A-6020 Innsbruck, Austria

●Lancaster Environment Centre, University of Lancaster, Lancaster LA1 4YQ, UK

Revision submitted to *Sedimentology August 31st* 2015

ABSTRACT

The Marinoan panglaciation (?650-635 Ma) is represented in NE Svalbard by the Wilsonbreen Formation which contains syn-glacial carbonates in the upper 100 m of the 130-175 m-thick formation. These sediments are now known to have been deposited under a CO_2-rich atmosphere, late in the glaciation, and global climate models facilitate testing of proposed analogues. Precipitated carbonates occur in four of seven facies associations: Fluvial Channel (including stromatolitic and intraclastic limestones in ephemeral stream deposits); Dolomitic Floodplain (dolomite-cemented sand and siltstones and microbial dolomites); Calcareous Lake Margin (intraclastic dolomite and wave-rippled or aeolian siliciclastic facies) and Calcareous Lake (slump-folded and locally re-sedimented rhythmic/stromatolitic limestones and dolomites associated with ice-rafted sediment). There is no strong cyclicity and modern analogues suggest that sudden changes in lake level may be characteristic.

Both calcite and dolomite in stromatolites and rhythmites display either primary or early diagenetic replacive growth. Oxygen isotope values (-12 to +15 ‰$_{VPDB}$) broadly covary with $\delta^{13}C$. High $\delta^{13}C$ values of +3.5 to +4.5 ‰ correspond to equilibration with an atmosphere dominated by volcanically degassed CO_2 with $\delta^{13}C$ of -6 to -7 ‰. Limestones have consistently negative $\delta^{18}O$ values, whilst, rhythmic and playa dolomites preserve intermediate compositions, and dolocretes possess slightly negative to strongly positive $\delta^{18}O$ signatures, reflecting significant evaporation under hyperarid conditions. Meltwater compositions inferred as -8 to -15.5 ‰ could reflect smaller Rayleigh fractionation related to more limited cooling than in modern polar regions. A common pseudomorph morphology is interpreted as a replacement of ikaite ($CaCO_3 \cdot H_2O$), which may also have been the precursor for widespread replacive calcite mosaics. Local dolomitization of lacustrine facies is interpreted to reflect microenvironments with fluctuating redox conditions. Although differing in (palaeo)latitude, tectonic setting, and

carbonate abundance, the Wilsonbreen carbonates provide a unique pre-Cenozoic analogue for theMcMurdo Dry Valleys of Antarctica.

Keywords: Cryogenian, oxygen isotopes, carbon isotopes, lacustrine, ikaite pseudomorphs, Snowball Earth

INTRODUCTION

The second of two Neoproterozoic panglaciations, in which ice sheets reached sea level in the tropics, terminated 635 My ago at the base of a transgressive cap carbonate defining the Cryogenian-Ediacaran System boundary (Table 1). Deposits of Cryogenian ice ages are preserved on most continents and are often interpreted as glacimarine (Arnaud et al., 2011). However, in NE Svalbard, the Marinoan-aged 130-175 m thick Wilsonbreen Formation (Halverson, 2011) uniquely contains *non-marine carbonates* as well as subglacial tillites (Fig. 1). These evince hyperarid terrestrial environments (Fairchild et al., 1989) and an atmosphere rich in carbon dioxide during glaciation (Bao et al., 2009), the latter conclusion fulfilling a prediction of the *Snowball Earth* hypothesis (Kirschvink, 1992; Hoffman et al., 1998). Because Wilsonbreen Formation outcrops are restricted to remote icefield nunataks (Fig. 2), they have rarely been visited, and hence our knowledge of the sedimentary architecture has been incomplete. The Wilsonbreen Formation carbonates contain probably the highest carbonate and sulphate $\delta^{18}O$ values and lowest sulphate $\Delta^{17}O$ signatures so far discovered in the geological record, features which evoke one of the most extreme climatic events in Earth history (Bao et al., 2009, Benn et al., 2015). Here we characterize a range of non-marine environments in which the carbonates were precipitated using a combination of field, petrographic and stable isotope evidence, and scrutinize a claim (Fairchild et al., 1989) that these deposits are an analogue of the extreme terrestrial environments of the modern McMurdo Dry Valley region of Antarctica, albeit formed at much lower palaeolatitudes.

The study area in the Svalbard mainland of Spitsbergen (Figs. 1, 2) and the basin continuation to the NE have long been recognized as classic areas for late Precambrian glaciation (Kulling 1934). The first detailed description of the Wilsonbreen Formation was by

Wilson & Harland (1964), although carbonates were discussed only in terms of its bounding dolomites as stratigraphic markers. A later sedimentological synthesis (Hambrey, 1982; Fairchild & Hambrey, 1984) showed that evidence of glacial activity was confined to two glacial units in the Polarisbreen Group: the Wilsonbreen Formation, and a newly discovered, thin, older unit in the Elbobreen Formation (Petrovbreen Member also known as E2) (Table 1). Several distinctive forms of carbonate in association with the glacial deposits. The first of these is dolomitic glacial rock flour, demonstrated in ultra-thin sections (Fairchild, 1983). Subsequently, stable isotope studies demonstrated the presence of glacimarine precipitates in the Petrovbreen Member and glacilacustrine deposits in the Wilsonbreen Formation (Fairchild & Hambrey, 1984; Fairchild & Spiro, 1987; Fairchild et al., 1989). It was also shown that these carbonates contrast with a distinctive marine transgressive 'cap carbonate' (following Williams, 1979) over the Wilsonbreen Formation. Halverson et al. (2004) provided a much more detailed chemostratigraphic framework for the Polarisbreen Group and postulated that both diamictite units belonged to the same, Marinoan glaciation. However, new chemostratigraphic data led to a reversion to the previous two-fold glaciation interpretation of older literature (Halverson, 2006; Halverson et al., 2007). No direct geochronological constraints exist on Svalbard, but the low $^{87}Sr/^{86}Sr$ values at the base of the Polarisbreen Group correlate well with those associated with the first evidence of Neoproterozoic glaciation elsewhere (Halverson et al., 2010). Also, the major marine transgression succession above the Wilsonbreen Formation closely resembles the basal Ediacaran facies elsewhere dated at 635 Ma (Rooney et al., 2015). Palaeomagnetic constraints suggest that NE Svalbard lay in the subtropics throughout the Cryogenian (Li et al., 2013) as part of the equatorially centred, fragmenting Rodinia supercontinent.

The closely similar stratigraphy in central E Greenland (now officially redesignated as Northeast Greenland) (Hambrey & Spencer, 1987; Moncrieff & Hambrey, 1990) indicates that it represents a southwestern basin continuation (Hambrey, 1983; Knoll et al., 1986; Fairchild & Hambrey, 1995), subsequently offset to the south by left-lateral strike-slip faulting (Harland, 1997). The Neoproterozoic succession in western Svalbard is quite different and was probably not deposited in the same basin (Harland et al., 1993).

Cryogenian events and panglaciation

The concept of Neoproterozoic low-latitude glaciation was championed by Harland (1964) who argued from the widespread occurrence of tillites that they were globally distributed and hence should be used in correlation. Subsequently it took an enormous effort from the geological community to establish that the diamictite units are predominantly glacigenic or at least glacially influenced, that convincing evidence of low-latitude glaciation exists, and that the correlations are supported by a geochronological framework (Fairchild & Kennedy, 2007). Despite the complexity of individual successions, summarized in Arnaud et al. (2011), it is now widely recognized that two panglacial events occurred during the Cryogenian period (c. 720–635 Ma), referred to as the Sturtian (or early Cryogenian) and Marinoan (or late Cryogenian) glaciations (Halverson et al., 2007; Macdonald et al., 2010; Hoffman et al., 2012; Calver et al., 2013; Rooney et al. 2014, 2015; Lan et al., 2014). The traditional geological approach recognized the role of low atmospheric CO_2 levels in triggering glaciation, but there was a lack of understanding of whole-Earth behaviour until insights from climate modelling of a possible future "nuclear winter" (e.g. Budyko, 1969) led to the realization that a frozen, high albedo planet represents a stable climatic state. Kirschvink (1992), in coining the term *Snowball Earth*, suggested that low palaeolatitudes of continents in Neoproterozoic times could have had a role in triggering panglaciation, and that to escape the Snowball state a build-up of volcanically derived CO_2 was needed (Caldeira & Kasting, 1992) to overcome the dominance of the albedo of Earth's icy surface on its energy balance and trigger deglaciation. Hoffman et al. (1998) and Hoffman & Schrag (2002) subsequently expanded upon the Snowball model, incorporating insights from planetary modelling to enlighten new stratigraphic, sedimentological and geochemical data. *Snowball Earth* is best regarded as a *theory*, not merely a hypothesis, because it is built around a series of propositions which are, to a greater or lesser extent, interacting (Fairchild & Kennedy, 2007). The theory has stimulated a huge acceleration in data-gathering and modelling which has significantly clarified the extent to which a given phenomenon is essential to or characteristic of a Snowball Earth. Initially (Hoffman et al., 1998), there was a focus on the associated negative carbon isotope anomaly, apparently encompassing the glacial period, which was regarded as a characteristic of low organic productivity in an ocean which became isolated from the atmosphere once the Snowball state had been established. Additionally, the post-glacial cap carbonates and the associated transients in the negative carbon isotope anomalies were regarded as a positive test of the hypothesis, reflecting rapid meltdown and sea-level rise under a post-Snowball, ultra-greenhouse environment. However, after further scrutiny all of these propositions have emerged as flawed or ambiguous (Kennedy et al., 2001a. b; Hoffman & Schrag, 2002; Halverson et al., 2002; Schrag et al., 2002; Trindade et al., 2003; Hoffman et al, 2007; Le Hir et al., 2008, 2009; Kennedy & Christie-Blick, 2011; Hoffman et al., 2012).

In the early years of the Snowball hypothesis, missing was an account of how the glacial formations themselves could be used to provide support for the theory. In the "hard" *Snowball*, the concept is of a universally thick ice cover over the oceans, composed of an upper zone of glacier ice and lower zone of frozen sea water (Pierrehumbert 2005; Pierrehumbert et al., 2011). If so, sea level should be greatly lowered and marine ice margins should occur at much greater depths than the shallow-water pre-glacial sediments. Evidence to support this hypothesis was found in Namibia. On platform tops, glacigenic deposits are rare, whereas on platform margins tidewater-glacier grounding-line phenomena can be demonstrated, inferred to be some hundreds of metres topographically lower (Hoffman, 2011; Domack & Hoffman, 2013). On the other hand, most glacial sedimentologists were hostile to Snowball theory, insisting that there is evidence of repeated advances and retreats of ice in marine environments during glaciation, and also that locally wave- and storm-generated structures are present, indicating open water (Xiao et al., 2004; Etienne et al., 2007; Allen & Etienne, 2008; Le Heron et al., 2011, 2013). Apparent support from models showing equatorial open water (e.g. Hyde et al., 2000) faced the problem that the simulated climate solutions were not stable. Hence, the response of Hoffman (2011) was that "counter arguments [to the Snowball model] 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". A further twist however was that a simulation of a long-lived marine ice-free equatorial fringe was achieved by Abbot et al (2011), in what they term the Jormangund state. Hoffman et al. (2012) noted that whereas the *Jormangund* state preserved the pattern of modern low-latitude climate belts, with a moister equatorial region, the *Snowball* climatic pattern (Pierrehumbert et al., 2011) would result in higher precipitation minus evaporation in the subtropics and an extremely arid equatorial zone. This draws attention to the need for study of continental glacial deposits such as the Wilsonbreen Formation where more direct evidence of climatic conditions can be obtained.

The surviving essential predictions of Snowball Earth theory can be summarized as follows: 1) glaciations must occur synchronously globally, 2) they must be long-lasting (>1 Ma) to allow 3) the build up of atmospheric CO_2 to high levels when 4) sedimentation occurs in a brief period prior to termination. Although earlier geochronological compilations had legitimate doubts about 1) and 2) (Allen and Etienne, 2008), they are now more firmly established (Rooney et al., 2015), although the duration of the Marinoan (?5-15 My) is imprecisely known. The Wilsonbreen Formation has now permitted positive tests of 3) and 4).

Compared with Cenozoic strata, there are very few approaches to determination of atmospheric CO_2 concentrations in the Neoproterozoic. A bold new approach was applied by Bao et al. (2008) based on processes occurring during stratospheric ozone formation which results in an enrichment in the isotope ^{17}O in ozone and carbon dioxide and depletion in oxygen (O_2). This is a non mass-dependent effect which does not influence ^{18}O abundances. The ^{17}O signal can be preserved in the oxygen atoms of sulphate in rocks if atmospheric oxygen is used to oxidize sulphides on the land surface. Sulphate has the peculiar property of not exchanging oxygen atoms with other species over 1000 My timescales at surface conditions, provided there is no microbial redox cycling of sulphur-bearing ions. Bao et al. (2008) tested the idea that at time periods when atmospheric P_{CO_2} was enhanced, there would be ^{17}O-depleted sulphate in the geological record, and indeed found the most significant anomaly that had at that time been discovered, recorded in barite crystal fans occurring in the carbonate succession overlying Marinoan glacial diamictites in South China.

Bao et al. (2009) then focused on lacustrine carbonates of the central part (member W2) of the Wilsonbreen Formation and found more profound ^{17}O-deficiencies in carbonate-associated sulphate (CAS) in limestones, consistent with very high atmospheric P_{CO_2} during glaciation. More recent studies in other geographic regions, coupled with process modelling approaches, have supported this approach in Marinoan cap carbonates (Bao et al., 2012; Cao & Bao, 2013; Killingsworth et al., 2013; Bao, 2014), but the Wilsonbreen Formation remains the only unit where P_{CO_2} can be estimated during glaciation.

Subsequently, Benn et al. (2015) have used the same isotope systematics of CAS on a much larger dataset of limestones from members W2 and W3 to argue that similar high P_{CO_2} values (estimated at between 1 and 10% atmospheric CO_2) occurred throughout the deposition of the Wilsonbreen Formation. Since it would have taken a long time to accumulate CO_2, the inference was that the bulk of the Wilsonbreen Formation was deposited in a relatively short period near the end of the glaciation. In turn, this implies an extended hiatus early in the glaciation which was identified with a permafrosted horizon at the base of the Wilsonbreen Formation.

Coupled ice sheet and atmospheric general circulation model results in Benn et al. (2015) using Snowball Earth boundary conditions demonstrate that at 2% atmospheric P_{CO_2} thick glaciers exist on the continents along with extensive areas of bare ground and that hyperaridity is widespread. Precessional forcing generates movements of ice margins by at least hundreds of kilometres and was linked to the presence of distinct ice advances within the Wilsonbreen Formation (Fig. 1). Although conclusions based on the Marinoan glaciation should not necessarily apply to the much longer Sturtian glaciation, the new results provide a possible route to reconcile the opposed positions stated in Allen & Etienne (2008), of temperate glacial conditions during panglaciation, and Hoffman (2011), of sediment deposition occurring rapidly during meltdown. It is the purpose of the current paper to provide a detailed sedimentological analysis to underpin this new synthesis and in particular to demonstrate the plausibility that the Wilsonbreen carbonates were deposited within a coherent geomorphic-climatic system. As a first step, we need to examine the possible modern analogue.

The Antarctic McMurdo Dry Valleys as analogues for the Wilsonbreen Formation carbonates

Fairchild et al. (1989) previously used the Dry Valleys region as an exemplar for the carbonates in member W2 of the Wilsonbreen Formation. Conversely, leading workers on the modern environment (Lyons et al., 2001) commented that this frigid, dry modern environment might be a valuable analogue for Proterozoic Snowball Earth environments. The Dry Valleys have also been used as analogues for the Martian surface (Marchant & Head, 2007; Dickson et al., 2013). The local “alpine” glaciers of the Dry Valleys are cold-based, but outlet glaciers such as the Taylor Glacier are warm-based. Given the criteria for recognition of glacial thermal regime presented by Hambrey & Glasser (2012), warm-based analogues are needed to understand much of the glacigenic facies of the Wilsonbreen Formation (Fleming et al., 2016). Nevertheless, for the carbonate facies, the context of the Dry Valley region, uniquely for present-day environments, provides parallels for the following evidence provided by Fairchild et al. (1989): (1) a record of extreme evaporation with potentially up to 20 ‰ difference in $\delta^{18}O$ between input waters and those responsible for the precipitation of the carbonates with heaviest $\delta^{18}O$ signatures and (b) alternating deposition of glacial sediment and continental deposits indicating ice advance and retreat, specifically including rhythmic and microbial, presumed lacustrine carbonates and evidence of former evaporites.

Fig. 3 illustrates the context of the modern setting which has been extensively studied over several decades by Antarctic groups led from New Zealand, Japan and the USA, with Taylor valley being the focus since 1993 of the McMurdo Dry Valleys Long-term Ecosystem Research (LTER) Program of the National Science Foundation. The Dry Valleys is a 4800 km^2 region occupying part of the Transantarctic Mountains between the East Antarctic Ice Sheet (EAIS) and the Ross Sea. The individual valleys themselves mostly trend E-W and are up to 80 km long and up to 15 km wide and are internally drained, often with several discrete lake basins in each valley. Two outlet glaciers from the EAIS (Wright Upper and Taylor glaciers) only just cross the regional bedrock altitude divide into the eastward-draining catchments, whilst the Ferrar glacier flows all the way to the coast (Fig. 3). Local valley glaciers extend from the mountains into dry valleys, e.g. Canada Glacier in Taylor Valley (inset in Fig. 3) and ephemeral streams develop for short periods in summer. A variety of lakes occur, including those occupied by ice frozen to the bed in the relatively high-altitude Victoria Valley, highly saline lakes with no ice cover (e.g. Don Juan Pond, Upper Wright Valley) and finally lakes with a 3-5 m ice cover (e.g. Lakes Brownsworth and Vanda in Wright Valley and Lakes Bonney, Hoare and Fryxell in Taylor Valley), which melt only in a marginal moat in summer (Green & Lyons, 2009; Dickson et al., 2013). Microbe-dominated biotas flourish wherever and whenever there is liquid water in the region, including mats on stream and lake beds, and photosynthesizing algae are important in lakes during the spring season (Fountain et al., 1999). These mats take up nutrients rapidly from stream water and the biogeochemical patterns are strongly influenced by exchange of hyporheic waters with stream water (McKnight et al., 1999).

There is a spatial microclimatic zonation comprising a coastal zone with just-thawed soils in summer and a stable cold upland zone with particularly low relative humidity (Marchant and Head, 2007; Marchant et al., 2013). The intervening valleys have a mean annual temperature of -16 to -21 °C and the maximum daily temperature is below zero on average throughout the year (Fountain et al., 1999). The valleys receive less than 10 mm water-equivalent of precipitation per year, almost always as snow. Because of the prevailing low relative humidities (e.g. between 50 and 60 % in Taylor Valley, Fountain et al., 1999), ablation (mostly as sublimation) greatly exceeds precipitation. Wilson (1981) described the consequences of this geographic configuration and climatology using physical and chemical principles. Precipitation rises with altitude, but falls inland further from the Ross Sea. The snowline marks the boundary where ablation exceeds precipitation and it rises inland as precipitation declines. Wilson (1981) attempted to explain the distribution of salts based on an understanding of downslope-increasing aridity, but it seems that variable meteorology confounds the predictions in detail. Nevertheless, it is the case that deliquescent salts flow downhill in the sub-soil above a permanently frozen layer. Lakes with a lid of ice display a balance between ablation at the surface and freeze-on of lake water beneath the lid, replenished seasonally by stream inflow. Once snow has been removed by ablation, sunlight can penetrate through vertical ice crystals and significant solar heating of lakes can occur. The present configuration of salts allows deductions of both long- and short-term history. Specifically, spatial variability in salts requires a long-term (>10^5-10^6 years) stability of the ice-free subaerial valley sides. On the other hand, some lakes have a basal brine layer which diffusion modelling shows originated from a period around 1200 years ago when some lakes were ice-free shallow brines, before being re-filled by fresh meltwater.

One aspect not treated by Wilson (1981) is the effect of wind. An important reinforcement mechanism for aridity is provided by the regional development of katabatic winds flowing off the East Antarctic Ice Sheet. As this air warms adiabatically, humidity decreases, particularly in winter (Nylen et al., 2004). It is now clear that the episodically strong summer winds are actually warm foehn winds, which arise from strong pressure gradients during the occurrence of cyclones over the Ross Sea, the incidence of which depends on hemispheric climatic anomalies. These topographically enhanced and channelled winds flow at typically >5 m s^{-1} westerly along the Dry Valleys and cause very large intra-annual and inter-annual increases in meltwater production and streamflow (Doran et al., 2008; Speirs et al., 2013). The high incidence of these winds in summer 2001/2, when positive degree days increased by an order of magnitude, led to rises in lake level of 0.5-1 m, effectively wiping out the previous 14 years of lowering of lake level in a period of three months (Doran et al., 2008).

These details effectively demonstrate the sensitivity of the environment to climatic changes that will strongly influence the facies deposited.

Barrett (2013) recently reviewed the controversial long-term history of the McMurdo region. Evidence for landscape evolution based on the position of Miocene volcanic ash deposits clearly demonstrates that after the establishment of the current large East Antarctic Ice Sheet in the Miocene, only surficial landscape modification has occurred (Sugden et al., 1995; Lewis et al., 2007). Importantly, the cold, dry climates of the Dry Valleys remained stable, even during significant warming events recorded in the Ross Sea during the Pliocene. This long-term climatic stasis can be compared with the predicted long-term hydrological inactivity anticipated on a Snowball Earth as carbon dioxide levels slowly rose.

Large lakes formed during the Last Glacial Maximum in all the major Dry Valleys. This required two conditions: (1) expansion of the Ross Sea Ice sheet to block the marine margins of the Dry Valleys (Hendy, 1980; Hall et al., 2013) and (2) increased meltwater production in the valley (by wind-induced melting, Doran et al., 2008), despite significantly colder conditions. Conversely, dating of aragonitic lacustrine deposits shows that, in interglacial periods, there was an expansion of Taylor Glacier, noted for characteristically light oxygen isotope values, in addition to expansion of local valley-side glaciers, with heavier isotope values (Hendy 1979, 1980). These phenomena presumably reflect higher snow accumulation on the EAIS, enhanced meltwater production, and partial blockage of the valley by ice. Importantly, these inferences draw attention to the potential anti-phasing of global temperature and local glacier advance and the greater importance of regional humidity controls on Milankovitch timescales.

In the current paper, a wealth of new data on the Wilsonbreen Formation carbonates are presented which allows the analogies previously made to be tested and evaluated in much more depth. The interpretation of these data is assisted by modern analogues and new computer simulation studies of Neoproterozoic climates (Pierrehumbert et al., 2011, Benn et al., 2015).

METHODS

Fieldwork (2010-11) was supported by helicopter and by skidoo. Sections, including six entirely new locations, were measured by 30 m tape, orientated by compass and Abney level, and linked to bedding dips, with total thickness checked by GPS with uncertainty of around 5%. The best available sections in each region were logged (Figs. 1, 2), although they vary in quality from almost perfect, to intermittent good outcrop separated by ground-level frost-shattered regolith. The carbonate rocks are chemically fresh, but laboratory study of cut or sectioned samples was necessary to identify facies in many cases. Over 350 samples were sawn, 210 of which were thin-sectioned of which 60 were stained with Alizarin Red-S and potassium ferricyanide and over 30 polished sections studied by cold-cathode cathodoluminescence (CL) at 15 kV.

Carbon and oxygen stable isotope data are presented here as $\delta^{13}C$ and $\delta^{18}O$ in parts per thousand with respect to the VPDB standard. Differences between laboratories are insignificant in relation to the wide range of isotope values (29 ‰ in $\delta^{18}O$ and 8 ‰ in $\delta^{13}C$) in this study. Supplementary data in Bao et al. (2009) included methods and all data collected to that date. New data was obtained at the University of Birmingham using a continuous-flow Isoprime IRMS, with a multiflow preparation system. Samples of between 80-250 μg powdered carbonate were reacted with phosphoric acid at 90°C for 90 minutes. Results were calibrated using IAEA standards NBS-18 and NBS-19. A fluid inclusion study is reported in the supplementary information. Sulphate oxygen and sulphur isotopes are presented in Benn et al. (2015). We draw on previously presented trace element data from the carbonate fraction soluble in dilute nitric acid (Fairchild & Spiro, 1987; Fairchild et al., 1989; Bao et al., 2009), whilst new trace element and other isotope analyses will be presented elsewhere.

In figure captions, the location and stratigraphic position are given in a standard format. For example *W2, Dracoisen, 70 m*, refers to a sample from member W2 from the Dracoisen section, 70 m above the base of the formation (or the base of the section where the base is not seen). The context and oxygen isotope composition of the carbonate from that horizon can be found in the stratigraphic section diagrams: Fig. 4 for Dracoisen and supporting Figs. 1-6 for the other sections. Carbonates in members W1 and W3 are summarized only in Table 2, but full sample logs are given in Benn et al. (2015).

WILSONBREEN FORMATION ARCHITECTURE, COMPOSITION AND POST-DEPOSITIONAL HISTORY

The Wilsonbreen Formation, dominated lithologically by sandy diamictites, contains a wealth of evidence indicating that it is largely glacigenic, and deposited in both aqueous and subglacial settings (Hambrey 1982; Fairchild & Hambrey, 1984; Dowdeswell et al., 1985; Harland et al., 1993). New work (Benn et al., 2015; Fleming et al., 2016), including five previously undescribed sections, has resulted in a coherent stratigraphic-facies reconstruction (Fig. 1). This paper focuses on beds (sandstones, rhythmites, mudrocks) containing precipitated carbonate; the presence of such strata was originally used to define member W2 (Hambrey, 1982). In the northernmost section (Dracoisen), W2 is readily distinguished by three such carbonate-bearing beds separated by diamictites (Fig. 2A), and overall a similar pattern applies in other sections (Fig. 1). However thin rhythmite units also occur in member W1, one containing precipitated carbonate, and most of the thin non-diamictite (sandstone and rhythmites) beds in member W3 contain such carbonate facies (Table 2). Although glacigenic rocks continue NNE to the coast and to Nordaustlandet, these were not included in our study since diamictites are thinner and less well exposed, and precipitated carbonates are absent (Halverson et al., 2004; Hoffman et al., 2012). Neither do such carbonates occur in the equivalent Storeelv Formation of NE Greenland, representing the SW continuation of the basin (Hambrey & Spencer, 1987; Moncrieff & Hambrey, 1990; Hoffman et al. 2012, Fleming, 2014).

Terrigenous detritus

Harland et al. (1993) summarized information from five sites, determining that carbonate clasts make up 40-85%, igneous and metamorphic clasts 5-33% and sandstones and quartzites 10-20% of *stones* (i.e. gravel-sized debris); the basement lithologies are primarily granitoids and gneisses, with some basalt in which ferromagnesian minerals are commonly chloritized. In the wider context of terranes affected by the Caledonian orogen, detrital zircon studies indicate provenance from Archaean and Palaeoproterozoic rocks of the cratonic interior and Mesoproterozoic detritus derived from the eroded remnants of the Grenville-Sveconorwegian orogen (Cawood et al., 2007). A systematic study of carbonate clast compositions will be presented elsewhere, but in summary 80% of pebbles are dolomite rock and 20% are limestone; many of these carbonates resemble units from the underlying succession (Table 1) in lithology and isotope composition; no stratigraphic trends in clast type or isotope composition were found. The sand fraction is dominated by quartz and feldspar with subordinate dolomite and limestone and the mud fraction likewise by quartzo-feldspathic debris and dolomite (Fig. 5). The dominance of silt in the mud fraction is shown in graded rhythmites (Fig. 5B, C); notably detrital calcite occurs in the mud fraction only locally and the proportion of clay minerals is likewise very low. The dolomite matrix of Polarisbreen Group diamictites is predominantly the result of glacial transport and comminution as shown by the presence of sub-micron relic rock flour (Fairchild, 1983) in which clay minerals are virtually absent (Fig. 5D). This detrital matrix contains a complex mixture of both bright and dully luminescent red dolomite when viewed in CL (Fig. 5C). The mean carbon and oxygen isotope compositions of dolomitic matrix in diamictites and wackes ($n = 43$) are +2.4±1.3 and -3.5 ±1.8 ‰ respectively, slightly lower values than for dolomite pebbles. This matrix composition forms a useful reference point for comparison with precipitated carbonates in the Wilsonbreen Formation.

Burial diagenetic modification

The Wilsonbreen Formation is overlain by a marine transgressive succession, but most Ediacaran strata (Table 1) are non-marine playa lake facies, in turn overlain by 950 m of Cambro-Ordovician platform carbonates (Harland, 1997) indicating a minimum burial of 1.5 km. Late Silurian Caledonian folding and local thrusting followed (Fig. 2), but small-scale folding is generally absent, as is penetrative deformation. The Wilsonbreen Formation lies outside the thermal aureoles of Devonian granite plutons in the fold belt (Harland, 1997). Pores in the cap carbonate are filled with bitumen, indicating that the sediments passed through the oil window, but good preservation is indicated by the ability to remove individual striated clasts from diamictite matrix (Hambrey, 1982) and some unusually high oxygen isotope compositions of dolomites (Fairchild et al., 1989; Bao et al., 2009). A fluid inclusion study (see supporting information) indicates that initial pore fluids were advectively replaced by a typical meteoric fluid with δ^2H in the range -60 to -100 ‰ and that ^{18}O is slightly enriched by exchange with the solid phase, but fluid inclusion volumes are orders of magnitude too small to have affected the bulk solid composition. Physical compaction effects are not noticeable because of the low clay content of the sediments and early cementation of the carbonates, but very locally lamina boundaries are styolitic. Uncemented sandstones exhibit straight to concavo-convex boundaries between quartz grains.

Many Wilsonbreen Formation outcrops are reddened, and even in unreddened facies, organic carbon contents are low (below detection levels of 0.2% on three W2 carbonates). Fairchild & Hambrey (1984) regarded the haematite pigment as post-depositional based on the occurrence of Liesegang band structures and ongoing palaeomagnetic studies may throw further light on its timing. Reddening is less pervasive in clean sandstones, implying that it requires a source of iron in the fine fraction or in carbonates. The low preservation of organic carbon may reflect low abundance of clay (Kennedy et al., 2006) as well as a low-productivity continental setting, contrasting with the dark-coloured glacimarine deposits of the early Cryogenian glacial deposits in the study area (Member E2, Table 1).

Limited quartz overgrowths are common in sandstones and on sand grains floating in dolomites, whilst some sandstones display partial poikilotopic calcite cement. Ferroan saddle dolomite and ferroan calcite occlude larger primary pores, in dolomites and limestones respectively. These carbonate phases also occur, locally together with quartz and white mica in crystal pseudomorphs. Saddle dolomite averages -12.2±1.5 ‰ in $\delta^{18}O$ and -1.8±1.5 ‰ in $\delta^{13}C$ and calcite spar likewise -10.5±1.6 ‰ and +2.0±1.5 ‰ (Bao et al., 2009).

FACIES ANALYSIS

Facies Associations

Following Benn et al. (2015) seven facies associations (FAs) are recognized; Table 3 presents summary descriptions and interpretations, whilst Fig. 7A illustrates inferred environments diagrammatically. The glacial and periglacial facies associations (FA1, 6, 7) are discussed in Hambrey (1982), Fairchild & Hambrey (1984), Benn et al. (2015) and Fleming et al. (2016), and only a few salient points are mentioned here. Two distinct glacier advances from the south are recognized, just below and above member W2, by occurrences of FA1: sheared diamictite, sandstones and gravels, locally resting on highly deformed rhythmites (glacitectonites) or striated clast pavements (Fig. 1, labelled FA1 subglacial). The bulk of the Wilsonbreen Formation is however composed of weakly stratified diamictites (FA6) with decimetre to metre-scale lenses of rhythmites, commonly with precipitated carbonate in members W2 and W3 in particular. Local presence of dropstones or isolated gravel clasts (lonestones), till pellets, stratification, and absence of subglacial shearing phenomena, point to a subaqueous ice-rafted origin. Local lensing conglomerates are interpreted as sediment-gravity flows which, at Dracoisen, comprise a conspicuous low-angle cross-stratified unit (Fig. 1, labelled FA6, proximal), interpreted as a grounding-line fan (Benn et al., 2015; Fleming et al., 2016).

Gravel with ventifacts overlying shattered dolomite represents the Periglacial Facies Association (FA7) at the base of the Wilsonbreen Formation and Benn et al. (2015) interpret this as a multi-million year continental hiatus. FA7 also occurs at the base of W2 where a periglacial exposure horizon with sandstone wedges penetrating subglacial till is found (Fig. 1) at South Ormen (ORM), South Klofjellet (KLO) and Ditlovtoppen (DIT) and also at the Golitsynfjellet (GOL) section (Fairchild et al., 1989), which is between McDonaldryggen (McD) and the Backlundtoppen-Kvitfjella Ridge (BAC). The periglacial horizon is overlain by sandstones correlated with those at the base of W2 at Dracoisen (Dracoisen).

The other four Facies Associations contain precipitated carbonate and are described in detail below, complementing the tabular and diagrammatic summaries (Table 3; Fig. 6). Each of the Facies Associations occurs as units 0.1 to 4 m in thickness as presented in the stratigraphic logs (Fig. 4, supporting information Figs. S1-S6).

FA2 (Facies 2S and 2T)

Description

This facies association is represented by erosionally-based, fining-upwards tabular sandstone beds, 0.5-4 m in thickness. Facies 2T typically comprises moderately sorted very fine to medium-grained sandstone with a basal erosional surface capped by a gravel lag. Internal cross-stratification is tabular, typically with low-angle accretion surfaces and set thickness of 0.5-1 m. Such facies are seen at the base of the W2 section at Ditlovtoppen (Supp. Fig. 1), South Ormen (Supp. Fig. 2), East Andromedafjellet (Supp. Fig.3), and Reinsryggen (Supp. Fig. 4), and are also prominent near the top of the member at Dracoisen (Fig. 8H) and at Ditlovtoppen, where thin siltstone beds define low-angle accretion surfaces (Fig. 8G).

Facies 2S is distinguished from Facies 2T by the presence of limestone laminae and intraclasts, and reduced silt content. The best-exposed example is a 4 m-thick unit that forms the basal part of member W2 at Dracoisen (Figs. 2A, 4), resting on massive diamictite with decimetre-scale erosional relief. A thin basal pebble conglomerate passes up into 0.7 m of pebbly sandstone, whilst the main, central part of the bed is very fine- to medium-grained sandstone with local tabular cross-stratification, with set thickness up to 15 cm and current ripples (Fig. 8D). At both levels, there are abundant (ca. 25-50%) layers of limestone, with individual laminae typically 1-3 mm thick (Fig. 8A). Universal features are a domed growth morphology (Fig. 8A) and differentiated microstructures (Fig. 8B) with well developed slightly irregular laminae of micrite, microspar, and detritus-rich carbonate with regular mm-scale fenestrae (Fairchild, 1991; Riding, 2000). Very locally a through-going vertical structure reminiscent of cyanobacterial or algal filament moulds (Fairchild et al., 1991) is observed (Fig. 9B). All these limestones are laterally discontinuous, breaking down to trains of intraclasts that are locally stacked at high angles. In places, extensive crusts, up to 0.5-6 mm thick, of radiaxial calcite cement develop, also broken to form intraclasts (Fig. 8C).

Under CL, the calcite fabrics that are broken into intraclasts fluoresce uniformly brightly, whereas later vein cements show more variable CL (Fig. 9A). Chemically, this is reflected in high Mn contents of several thousand ppm and exceptionally high Mn/Fe of 1 (Bao et al., 2009). Strontium contents are 150-300 ppm, whilst Mg is 4000-6000 ppm (Bao et al., 2009), equivalent to around 2 mole % $MgCO_3$. The radiaxial fabrics appear pristine and microdolomite inclusions are absent. Sulphate content is high (2000-5000 ppm, Bao et al., 2009), whilst the preservation of a negative $\Delta^{17}O$ anomaly demonstrates that the sulphate has not undergone reduction (Bao et al., 2009; Benn et al., 2015).The stable isotope composition of stromatolitic limestones defines a coherent field (Fig. 7A) with $\delta^{18}O$ ranging from -10.5 to -3.4 ‰ and $\delta^{13}C$ from +0.9 to +4.6 ‰, weighted towards higher values, with overall isotopic covariation (Fig. 7B). A micromill traverse through syndepositional calcite reveals lamina to lamina variations in $\delta^{18}O$ of 2 ‰ and in $\delta^{13}C$ of 0.5 ‰ without strong covariation (Fig. 8E, F). The clear petrographic distinction between syn-depositional and later calcite spar cement is reflected also in the low $\delta^{18}O$ signature of the spar of -10 to -12 ‰ (Fig. 8F).

Interpretation

FA2 is notable for a dominance of tractional sediment transport. The consistent presence of laterally extensive basal erosion surfaces imply a channel context for this facies association, whilst for Facies 2T the low-angle accretion surfaces with thin silt beds is reminiscent of point-bar deposits which, although not unexpected in Precambrian sediments, would require more extensive exposures to confirm (Davies & Gibling, 2010). The presence of high-angle cross-stratification, good sorting, and disrupted intraclasts in Facies 2S are characteristics found in either tidal sand flats or in low sinuosity fluvial channels. The limestones contain two features typical of Neoproterozoic microbial deposits: macroscopically domed laminae and differentiated microstructures formed by periodic variations in phenomena such a sediment trapping, gas generation and carbonate precipitation (Fairchild, 1991; Riding, 2000) and so can be referred to as stromatolitic. Limestones were lithified before erosion since cement crusts are broken and derived intraclasts are dispersed and stacked, for example along foresets, as expected for significant tractional flows in shallow water. The fact that all of the thin limestones disappear laterally to be replaced by trains of intraclasts, or else are completely eroded indicates highly variable flow conditions. Further, the locally highly regular microbial lamination, including clastic layers, points to a periodic control on flow rates.

When taken in isolation, a marine origin for facies 2S could be postulated on a general similarity with tidal sandflat deposits and the occurrence of well-preserved radiaxial cements with relatively high $\delta^{18}O$ compositions (Fairchild & Spiro, 1987). However, tidal sandflat deposits in Neoproterozoic successions typically show well-developed herringbone cross-stratification (Fairchild, 1980; Fairchild and Herrington, 1989). There is a clear contrast with the more regular macrostructures of marine stromatolites elsewhere in the basin (Fairchild & Herrington, 1989; Knoll & Swett, 1990; Halverson et al., 2004) and Neoproterozoic deposits more generally (Grotzinger & Knoll, 1999). This may be attributable to a more hostile environment, with highly variable rates of sedimentation. The most significant point may be that Neoproterozoic marine peritidal deposits are invariably dolomitized (Knoll and Swett, 1990), probably a feature of high Mg/Ca in seawater (Hood and Wallace, 2012).

The alternative is a freshwater, fluvial context. We now know that radiaxial fabrics are not diagnostic of marine waters, but also occur in speleothems (Neuser & Richter, 2007) and also that such fabrics are primary and consistent with the relative low Sr content of the calcite (Fairchild & Baker, 2012). Hence, we interpret the radiaxial calcite as the pristine original low-Mg calcite phase. Micrite and microspar fabrics have similar chemistries and are also considered to reflect depositional conditions. The implication is that the Mn content of FA2 calcites is also primary and reflective of low contemporary atmospheric PO_2 (cf. Hood & Wallace, 2014), but not anoxic conditions else sulphate reduction would have occurred and the distinct negative $\Delta^{17}O$ anomaly would have been erased (Bao et al., 2009). Reassuringly, the Mg composition of facies 2S limestone is similar to modern speleothem deposits in a cool Scottish cave depositing from waters with Mg/Ca controlled by dolomite dissolution (Fairchild et al., 2001). The lamina thickness of the stromatolites is very similar to those of modern fluvial microbial tufas which are complex deposits containing both biologically mediated and inorganic precipitates just like Facies 2S (Andrews and Brasier, 2005; Andrews, 2006). Interestingly the lamination is annual and reflects a strong annual variation in discharge and Facies 2S possesses physical sedimentological characteristics consistent with those of ephemeral streams. In this interpretation, the stable isotope compositions, which are similar for radiaxial and microsparry calcites, can also be interpreted as primary, in which case the isotopic covariation (Fig. 7B) would have to be interpreted as an evolutionary trend towards more evaporated equilibrated solutions (Talbot, 1990), rather than the lighter values reflecting higher-temperature recrystallization. Modern Antarctic streams lack the calcite mineralization, but microbial mats are well-developed and are adapted to ephemeral flow conditions, readily reactivating even after being dry for many years (McKnight et al., 2007). The fluvial interpretation will be developed later in the light of the relationship of FA2 to other facies associations.

Facies Assocation 3 (Facies 3D and 3S)

Description

Facies 3D is marked by discrete zones of pronounced dolomite cementation within a sandstone or siltstone; in some cases the sediment is pervasively cemented. Where dolomite is most abundant, detritus floats in a displacive mass of dolomite crystals (Fig. 10B, D, E, F), but other dolomite-cemented silty sands are still clast-supported (Fig. 10C). Locally, highly distinct dolomite-cemented nodules are visible (Fig. 10A) or a structureless dolomite bed can be encountered with a low content of floating silt and sand. The most characteristic structures are mm-scale nodular dolomicrite structures within massive dolomite-cemented layers and associated with calcite-filled fractures. These phenomena are found at one horizon in member W3 (Fig. 10D), as well as in several locations in W2 (e.g. Fig. 10F). A rarer phenomenon is the presence of equant centimetre-scale cauliflower-like pseudomorphs, filled by ferroan saddle dolomite cement (Fig. 11A) occurring at the top of a conglomerate-based fining-upwards cycle (Fig. 8G). Since saddle dolomite is a burial phase (Radke and Mathis, 1980), the implication is that the pseudomorphs were occupied with soluble crystals that dissolved during burial prior to cementation.

Facies 3S refers to dolomitic laminites, with broad cm-scale domed macrostructure, with an aspect ratio of typically 10:1. They are found uniquely in a complex bed, in association with Facies 3D, and overlying facies 2S (at 70 m, Dracoisen, Fig. 4). It has been studied on the

“Multikolorfjellet” cliffs and the “Tophatten” nunatak 1 km to the north. The bed is around a metre in thickness, but with variability in its internal structure. Most commonly there is minor erosion of underlying diamictite at the base of the bed, overlain by crudely laminated very fine- to medium-grained green sandstone, locally with mm-scale limestone layers that are partly disrupted into intraclasts (Facies 2S). In places this can be seen to pass upwards into intensively dolomite-cemented sand in which the rock fabric appears to have expanded, associated with corrosion of quartz detritus by dolomite and the formation of cavities, lined with dolomicrospar, and occluded by calcite. At the “Tophatten” locality, a chaotic breccia unit a few dm thick is locally found at the base of the bed instead of sandstone. Everywhere, the top of the bed is marked by 10-20 cm of dololaminites with a complex microstructure, which alternate on a cm-scale with displacively cemented sands (Fig. 10E). The laminae can be composed of dolomicrite or dolomicrospar and contain common fenestrae (Fig. 9E, H), whilst surface exposure reveal a finely textured microtopography (Fig. 10G). Locally, slightly lower in the bed, limestone laminites (FA5) form a 10 cm horizon overlying a 20-30 cm chaotic carbonate breccia (Fig. 10I) and gradually become disrupted downwards.

Dolomite from FA3 is characteristically bright under CL (Fig. 11B-D). In facies 3S, dolomicrite clots are uniformly bright, whilst adjacent dolomicrospar displays duller growth filling small fenestrae whilst larger fenestrae are filled by dolospar with more variable properties (Fig. 11B). Manganese (3000-4000 ppm), Fe (10000-15000), Na (2000) and Sr (250-350) ppm values are all unusually high (Bao et al., 1989) and our unpublished electron microsope images and microanalyses show enrichments also in many transition metals and rare earths and a consistent chemical zonation within crystals of dolomicrite mosaics. In facies 3D, a difference in mean CL brightness, and hence timing of growth, can sometimes be observed between nodules and surrounding matrix (Fig. 11D), whilst locally zonation within individual crystals growing between siliciclastic sand grains can be observed (Fig. 11C). Sulphate concentrations are high (4000 ppm); there is no $\Delta^{17}O$ anomaly but high $\delta^{18}O$ in sulphate (Bao et al., 2009; Benn et al., 2015).

Carbon and oxygen isotope values are correlated (Fig. 7), but Fig. 12 illustrates that the two analyses from member W3 lie about 1 ‰ higher in $\delta^{13}C$ than expected from this trend. Facies 3D has a range of $\delta^{18}O$ from -1.9 to +11.4 ‰, but the mean value is biased upwards by multiple analyses from a sample which passes up into facies 3S which tends to have higher values (Fig. 12). The latter facies is notable for possessing possibly the heaviest oxygen isotope values of carbonate rocks so far recorded in the geological record (Bao et al., 2009), with values up to +14.7‰ (VPDB) being found (Fig. 12). A micromill traverse (Fig. 10H, J) demonstrates that these extreme high values are maintained on the mm-scale, but that over petrographic boundaries, $\delta^{18}O$ values can vary by as much as 6 ‰.

Interpretation

For Facies 3D, the dolomicritic, syndepositional, passive to displacive growth with nodular structure and cracks is characteristic of calcretes in which precipitation occurs as a response to evaporative losses at or above a water table. Although rare, the spar-filled pseudomorphs (Fig. 11A), interpreted as after anhydrite (Fairchild et al., 1989), provide further evidence of evaporative conditions. Specifically the evidence for displacive growth, nodules and cracks served to identify alpha calcretes (Wright, 1990) which in Phanerozoic examples tend to occur on non-carbonate substrates and in more arid conditions than the more common beta calcretes containing structures resulting from higher plants (Wright & Tucker, 2009). The high $\delta^{18}O$ values and covariation with carbon isotopes require evaporation (Fairchild et al., 1989) which at the high end of the spectrum reaches extreme proportions and hence requires an extremely arid environment. Dolocretes are rather less common than calcareous calcretes and tend to be better developed when originating from groundwater than when pedogenic, as in Triassic strata of the Paris Basin (Spötl & Wright, 1992). In this example, pedogenic and groundwater dolocretes had a similar range of stable isotope compositions to each other, but their covariance slope (1:1) was steeper than in FA3. Overall the absence of any signal of light carbon from oxidation of organic matter in FA3 is notable, but consistent with the undetectably low organic carbon contents of the rocks. In Phanerozoic rocks, the presence of some biological features (e.g. root structures) can serve to identify pedogenic calcrete, but this is not possible in the Proterozoic.

The Wilsonbreen Formation dolocretes are interpreted as pedogenic primarily because the extremely high $\delta^{18}O$ values would require ground surface conditions for such extremely effective evaporation to occur. As will be discussed later, this interpretation is also consistent with the vertical facies relationships.

Regarding Facies 3S, the differentiated microstructures are again typical of microbial deposits. Such laminites are found in association with soils and intermittently flooded subaerial surfaces (Alonso-Zarza, 2003), although younger examples include root mats from higher plants that are clearly inapplicable here. Klappa (1979) ascribed cm-scale laminated deposits “hard pan” on calcretized limestone substrates as originating from the activities of lichen which colonize, bore into and form accretionary deposits on surfaces. The lichen-formed deposits do exhibit features such as fenestrae, sediment incorporation and variable crystal size which are consistent with the Wilsonbreen Formation example. However, no evidence of alteration of underlying cemented material has been found and the Wilsonbreen Formation microbial laminae are much more distinct and are noticeably domed, contrasting with laminar calcretes. In fact, the microbially influenced layering is indicative of active upward accretion, rather than slow pedogenetic alteration. The accretion took the form both of growth of carbonate-mineralized microbial mats, but also sand laminae. A shallow depression on a floodplain/playa margin seems apposite. A combination of a high water table from which evaporation could occur, or very shallow water inundation followed by drying out and sediment addition, perhaps by aeolian action is indicated. At Dracoisen (Fig. 9I), the gradational relationship between laminated carbonate and underlying chaotic breccia is a classic characteristic of evaporite dissolution breccias. The calcite-cemented nature of the breccia is consistent with removal of one or more horizons of calcium sulphate evaporites either during deposition of the bed, or soon afterwards following resumption of glacial conditions. A possibly similar Mesoproterozoic example is provided by Brasier (2011) from Ontario in which stromatolites are associated with collapse breccias and calcretes, and inferred to form at a playa lake margin. Likewise, the modern McMurdo Dry Valleys contain a record of many shallow saline lakes and salt pans (Wilson, 1981).

Dolomite is known to be capable of precipitating as a primary phase or by replacement of a $CaCO_3$ precursor in a range of surface environments (Warren, 2000), although the initial crystals (protodolomite) may lack well-developed ordering reflections and these can increase over time (Gregg et al., 1992). The petrographic characteristics of FA3 dolomicrite are consistent with very early diagenetic replacement of a precursor carbonate or of primary growth of (proto)dolomite and the latter is clearly the case for zoned dolomicrospar cavity-linings (cf. Hood and Wallace, 2012). The presence of euhedral crystals within displacive fabrics (Fig. 11C) is distinctive. Although Tandon & Friend (1979) interpreted euhedral growth zones in displacive calcite in calcretes as evidence of recrystallization it is more logical to see it as a primary growth fabric, as argued for dolocretes by Spötl & Wright (1992).

The extremely high $\delta^{18}O$ values rule out post-depositional modification and an interpretation of the values as reflective of the depositional environment is also consistent with the trace element chemistry and preserved crystal growth zones. The high Mn content and absence of pyrite implies low pO_2, but not anoxia, although the sulphate oxygen-isotope systematics are indicative of more redox variability than in Facies 2S. Specifically bacterially mediated electron shuttling by Mn-species can catalyze repeated transitions between sulphate and sulphite can permit the erasure of a $\Delta^{17}O$ signature and creation of a high $\delta^{18}O$ in sulphate (Bao et al., 2009). Such processes could catalyze dolomite nucleation given the evidence from other field and experimental studies on the catalytic role of sulphate reduction (Vasconcelos et al., 2005; Zhang et al., 2012).The inferred redox variations may be related to a supply of brine primarily from within the sediment, contrasting with the surface waters from which Facies 2S precipitated. The occurrence of the highest $\delta^{18}O$ values in laminated dolomites of Facies 3S is consistent with their formation by very near-surface evaporation, whilst abrupt variations in $\delta^{18}O$ (Fig. 10J) are suggestive of occasional inundations by less evolved waters. In summary, FA3 provides examples of facies that stretch the boundaries of earth surface phenomena and indicate deposition in unusually arid terrestrial environments.

Facies Association FA4 (Facies 4I, 4R and 4S)

Description

Facies 4R is the most common facies in this association and consists of rhythmic alternations of carbonate and sorted terrigenous sediment, which occur in association with structures such as wave ripple lamination or desiccation structures. The fine carbonate layers are usually dolomitic, or mixed dolomitic-calcitic, but include some limestone (Figs. 7A, 12). Universally, the coarser sediment layers, composed of sediment in the size range coarse silt to fine sand, show signs of tractional sorting, which is a key discriminant from FA5. Wherever laminae are sufficiently thick, undulatory cross-lamination is displayed (Fig. 13B) which can be confidently identified as wave-generated. Locally, symmetrical ripples are preserved in cross-section (Fig. 13A) or on bedding planes (Fig. 13C). Drying out is commonly indicated by desiccation structures with associated small intraclasts (Fig. 13B, H) or salt pseudomorphs (Fig. 13D), although such structures are not present in the majority of samples. Four examples of apparently non-evaporitic crystal pseudomorphs have been found, but these are much better developed in FA5 and are described in that section. Carbonate laminae are micritic in texture and typically uniform, although differentiated clotted microstructures also occur, similar to those described below in FA5, consistent with precipitation beneath benthic microbial mats (Riding, 2000). This facies was locally highly affected by subsequent glacitectonic deformation at the top of W2 at Ditlovtoppen, as described by Fleming et al. (2016).

The isotope traverse of Fig. 13J reveals a shift in isotopes from the sandy layers into dolomicrite consistent with an authigenic origin for fine dolomite, which is confirmed by CL observations (Fig. 14A, B). The limestone laminae in this facies association display a range of $\delta^{18}O$ values from -11.9 to -3.2 with a mean of -8.1 ‰, whereas the dolomite ranges from -5.3 to +1.4 with a mean of -1.9 ‰ (Fig. 7B). The difference of 6 ‰ in mean value, compared with inferred and observed differences of 2.6-3 ‰ for calcite and (proto)dolomite precipitating from equivalent fluids (Land, 1980; Vasconcelos et al., 2005) implies that the dolomites precipitated on average from waters with higher oxygen isotopevalues and the dolomites display isotope covariance (Fig. 7B). Trace element data will be presented elsewhere, but FA4 and FA5 dolorhythmites also show a covariation of Sr (from 100 to 200 ppm) with $\delta^{18}O$ and somewhat higher Sr values in calcites, and like other Wilsonbreen precipitated carbonates, they are Mn-rich (>1000 ppm).

Facies 4I occurs typically as discrete beds, normally 10-20 cm thick, of sharp- to erosionally based intraclastic sandstone with wave-generated lamination. The sand matrix is moderately sorted coarse silt to medium-grained sand and intraclasts are sometimes confined to the lower half of the bed. Several successive beds are shown in Fig. 13H in a section transitional upwards from Facies Association 5, and an example of this facies in thin section is illustrated in Fig. 13G. Two occurrences of ooids (Fig. 13E) have been found. At North Klofjellet, high in a generally poorly exposed W2 section, is a 1 m bed of indistinctly cross-laminated sands alternating with cm-scale desiccated limestone beds containing scattered sand grains. This is the lateral equivalent of fluvial facies (FA2) 1 km away at South Klofjellet. Ooids are found near the top of the unit, but make up less than 5% of the sand fraction. A range of cortices from superficial coatings through to fully developed ooids with no visible core are developed. Fabrics are micritic and microsparitic with a crude concentric structure. The mean oxygen isotope value of the limestone (>90 wt. % $CaCO_3$) is -7.6 ‰. The second example is the occurrence of a small proportion of ooids within thin (0.1 m) cross-laminated green sandstone underlying FA 5 sediments in member W3 at Ormen (Table 3).

Facies 4S is represented by distinct sandstone beds in the Ditlovtoppen and Dracoisen sections. These sandstones are highly uniform, consisting of well-sorted fine- to medium-grained sandstone, with very well-rounded grains. Bedding structures are confined to an indistinct, discontinuous parallel stratification. The Ditlovtoppen example presents as a tabular 3 m-thick bed over the 200 m width of the outcrop. Its lower few decimetres are locally thinly laminated sand and dolomite, changing laterally to uniform sandstone. Two thin beds of dolomite with floating grains are also found at the top of the unit. Grains are very well-rounded and range from very fine to coarse-grained, but most of the rock volume is composed of medium to coarse sand grains (Fig. 13F). The only sedimentary structure displayed is an indistinct cm-scale horizontal lamination with slight variations in grain size, or locally with mm-scale laminae with silty dolomitic matrix. Locally the lamination displays sedimentary deformation, suggestive of upward fluid escape. Oxygen isotopes in several samples are slightly heavier than expected for detrital matrix, consistent with addition of precipitated dolomite. There is a transition downwards (Fig. 10H) to Facies 4R in which mm-cm scale sand laminae between rhythmic carbonates develop cross-lamination and wave-ripple morphology, and within 20-30 cm of the boundary occurs a thin representative of Facies 4I (intraclastic flake breccias) and desiccation structures indicative of emergence.

Interpretation

Facies 4R and 4I display evidence of sorting and reworking of the sediments by wave action. This indicates deposition in a water body that was unfrozen at the time of deposition of the coarser layers which represent distinct time periods with more pronounced wave action. The sharp-based intraclastic beds (Facies 4I) appear to represent distinct storm events in which considerable disruption and transportation of cemented carbonate layers occurred, although at least in some cases these layers were already disrupted by desiccation.

The grains and structures in facies 4R and 4I are consistent with either a marine or a lacustrine origin, although it is noted that there is an absence of demonstrable tide-related phenomenon (cf. Fairchild & Herrington, 1989) and although there is insufficient information available to provide a quantitative description of wave climate (Allen, 1984), no wave phenomena were seen requiring oceanic conditions. Although ooids are best-known from marine environments and thick oolitic units were used as a criterion for warm climates in the

Neoproterozoic context by Fairchild (1993), ooids have been described from Quaternary sediments reworked into Antarctic moraines (Rao et al., 1998) and in various modern alkaline or hypersaline lakes. Lacustrine ooids form in water depths of 1-5 m, with the best development in shallowest water. The Wilsonbreen Formation examples do not resemble hypersaline aragonite ooids with radial structure (e.g. Halley, 1977), consistent with the oxygen isotope composition which does not indicate any evidence for hypersalinity. Ooids in smaller modern lakes tend to be superficial with relatively irregular outlines, whereas fully developed ooids are found on the large Lake Tanganyika in Burundi (Cohen & Thouin 1987) correlating with stronger wave action.

Chemical arguments favour a lacustrine origin for the carbonates. They probably arose as some combination of water column precipitates or within the sediment, e.g. as microbially influenced precipitates. Dolomite could be primary or, given the CL evidence, be a very early diagenetic replacement of $CaCO_3$, although not one with high Sr content. Given the consistent chemical characteristic of Mn-enrichment, it seems highly improbable that burial diagenetic recrystallization took place and so it is most straightforward to interpret the stable isotope values as primary, in which case the wide range of $\delta^{18}O$ compositions is notable because it is much greater than expected from marine waters. In the case of the dolomites, this is much greater than the relative small changes (1-2 ‰) that might be expected to accompany increased ordering from an initial protodolomite to an ordered dolomite (Gregg et al., 1992; Kaczmarek & Sibley, 2014). The formation of dolomite from more ^{18}O-rich, evaporated waters, is consistent with the standard paragenetic model in playa lakes (Dutkiewicz et al., 2000), although changes in source water composition as well as evaporation are likely to have occurred.

The well-rounded sand grains found in Facies 4S are consistent with aeolian transport as in the McMurdo Dry Valleys of Antarctica (Fig. 3C; Calkin & Rutford, 1974; Hambrey & Fitzsimons, 2010). However grains with such a transport history can finally be deposited in aeolian, fluvial or lacustrine settings. The consistent grain-size characteristics of individual laminae and good sorting of the coarse laminae instead point to tractional flows, but lack of cross-stratification rules out aeolian or subaqueous dunes. Hendy et al. (2000) developed the ice-conveyor model to account for sandy deposits on the floors of certain ice-covered Antarctic lakes with floating glacier-ice margins. Wind-blown sand melts its way through the ice in contrast to gravel which remains on the surface where it is transported by ice flow to the distal lake margins. Such an environment can develop the indistinct parallel stratification observed in this facies, but two features of the modern systems not observed are mounded bedding and upward gradation into coarse gravel (Hall et al., 2006). The Ditlovtoppen bed is tabular, whereas in the modern lakes sand transmission to the ice is focused, leading to mounds and ridges on the lake floor.

A plausible alternative for Facies 4I is an inter-dune environment, a setting where wind ripple migration would be expected (Lancaster & Teller, 1988). Such phenomena could give rise to translatent sub-horizontal laminae, representing the set boundaries, without internal ripple cross-lamination being preserved (Mountney & Thompson, 2002). A water table that was at least seasonally high is required to account for dewatering structures and the precipitation of dolomite (incipient dolocrete). In such modern environments seasonal flooding by surface water or emergent groundwater might occur, accounting for occasional dolomicrite laminae. The style of stratification is inconsistent with dune deposition, but is that expected on a sandflat or playa with a high water table and fits with the transition to Facies 4R and 4I observed in Fig. 10H.

Overall FA4 represents shallow-water and exposed sediments associated with a wave-dominated shoreline and susceptible to wind reworking. Although distinction of marine from lacustrine coastal settings is never easy, the wide range of oxygen isotope compositions of micritic Mn-rich carbonates favours a lacustrine origin.

Facies Association 5 (Facies 5D and 5R)

Description

This facies association is dominated by rhythmic alternations of carbonates and poorly sorted clastic sediment (Facies 5R). Locally, gradations are seen to brecciated rhythmites which are distinguished as Facies 5D. Facies Association 5 is only a minor constituent of most of the W2 sections, being more prominent in the S Klofjellet and Backlundtoppen-Kvitfjella Ridge (BAC) sections (Figs. S5 and S6), but it is the dominant carbonate-bearing facies association in member W3 where it alternates with ice-rafted diamictites of FA6.

Fig. 11 illustrates variants of Facies 5R found in member W2. Fig. 11A exhibits highly regular millimetre-scale alternations of limestone and wacke. The clastic sediment includes many microscopic diamictite pellets (arrowed) which are a normal feature found in this facies, whereas the small pseudomorphs crossing lamina boundaries are found more locally. Pebble-sized fragments in clastic layers are seen in Fig. 11C and 11D, the latter displaying a dropstone texture associated with disruption of limestone laminae. An indicator of instability is shown by the minor fault in Fig. 11C across which the number of limestone laminae changes, indicating erosion on the upthrown side and hence that this is a sedimentary growth fault. Larger-scale disturbance is shown by the folds in Fig. 11F in which carbonate laminae display some plasticity, but are also fractured, indicating partial cementation and a sedimentary origin for the folds. Above this, the sediments are visibly disrupted and transitional to Facies 5D. The thickest example of a facies 5D observed was a 0.4 m bed at Ditlovtoppen, containing both intraclastic and terrigenous sediment, and which disappeared laterally within 100 m. It clearly was derived by localized resedimentation of Facies 5R.

In member W3, Facies 5R is present as isolated beds up to 1 m thick exhibiting similar alternations of carbonate laminae and wacke/diamictite as in member W2 (Fig. 16). Fig. 16A illustrates the base of one such bed with clear alternations of thick diamictite laminae and carbonate passing upwards into more carbonate-dominated facies with only thin clastic laminae. The dominance of precipitated carbonate in this facies is shown in Figs. 16C, D, E, the latter being the sole example of precipitated carbonate in member W1. Disturbance by soft-sediment folding is nearly universal and the same combination of ductile and brittle behaviour of carbonate layers is shown (Fig. 16D, E) as in member W2. Fig. 16B illustrates the development of a resedimented bed (Facies 5D) dominated by intraclasts, but with some poorly sorted sediment material, over a horizon with soft-sediment folds.

It is common for carbonate layers in Facies 5R to show irregular lamination or domal structures. There can be upward doming of layer tops (Figs. 15C, 16C) up to centimetre-scale (Fig. 15F). Topography can be inherited from underlying layers (Figs. 15F, 16C), whereas sometimes the base as well as the top of the layer is domed upwards (Fig. 15C). Fig. 15B displays both these features in layers underlain by complex cement crusts with some remaining porosity. The crusts show neither a botryoidal nor euhedral morphology and are composed of polycrystalline calcite mosaics in which each calcite shows the same zonation in CL. There are transitions through beds with only minor clastic debris (Fig. 15F) to more massive limestones with complex microstructures.

Petrographically, carbonate laminae can be regularly (rhythmically) developed, millimetre-scale in thickness. Laminae are often heterogeneous, and may be either dolomitic, calcitic (Fig. 17A, B, D, F) or mixed mineralogy in composition (Fig. 17C). Where heterogeneous, laminae may show an increase in crystal size upwards (Fig. 17A) or display more or less evident clotted textures of micrite within microspar. Under CL, both calcite and dolomite present coherent replacive fabrics (e.g. Fig. 17F), in which crystals with identical zones grow throughout the fabric and enlarge into fenestrae. Rare examples of micritic rods around 10-20 μm in diameter, with minor associated pyrite, are reminiscent of calcified sheaths such as found in the pre-Cryogenian Draken Formation in the study area (cf. Knoll et al., 1993), but are not as distinct. Millimetre-scale convexities on the upper bed surface appear as thrombolitic in texture (Riding, 2000) with irregular fenestrae (Fig. 17B). Clastic laminae tend to level the microtopography (Fig. 17A), whilst individual sand grains, dropstones or diamictite pellets can occur anywhere within the carbonate fabrics (Fig. 17D).

Trace element compositions of carbonate are similar to FA4. The limestone laminae in this facies association display a range of $\delta^{18}O$ values from -5.6 to -12.8 ‰ with a mean of -9.2 ‰ and the $\delta^{13}C$ values also display a large range from -2.1 to +3.5 ‰. Although the full range in $\delta^{13}C$ is shown by member W2, values in member W3 tend to be lower (Fig. 12). FA5 dolomite ranges from -10.3 to +3.7 ‰, with a mean of -3.2 ‰ (Fig. 7B) and as for FA4, the dolomites display isotopic covariance.

Common examples of crystal pseudomorphs occur in FA5, usually as subhedral to euhedral crystals, variably joined into confluent masses embedded within or apparently cutting across lamination (Fig. 18E, F, G). In some cases, trains of crystals are aligned along or concentrated within particular laminae. Size of individual crystals is similar within samples and ranges from 0.1-0.2 mm (Fig. 18F) to 1-3 mm (Fig. 18B, 19A), the most common size being 0.5-1 mm (Fig. 186E, G). The range of cross-sections is dominated by four-sided or six-sided figures of crystals with equant to columnar habit. Pseudomorphs are equally likely to be developed in rhythmites with complex microstructures as in rhythmites with uniform micrite. Limestone hosts for pseudomorphs (n = 15) had a mean $\delta^{18}O$ composition of -8.8 (range -6.2 to -12.7) ‰

and dolomites likewise (n = 4) mean = -1.92, (range -6.4 to +1.7) ‰, that is similar to FA5 as a whole.

Whilst the within-sediment mode of occurrence is found in both members W2 and W3, the most spectacular, upward-growing crystals have only been seen at two levels in the S. Klofjellet section of W3. A polished hand specimen (Fig. 19A) displays three distinct crystal horizons of which the upper two are shown in the photomicrograph of Fig. 18B. The crystals evidently grew freely upwards and crystal terminations are strongly draped by overlying sediment layers indicating that the crystals formed at the sediment-water interface. In different cases, they are draped either by carbonate (e.g. forming rounded masses on the lower horizon of Fig. 18B), or poorly sorted sediment (wacke or diamictite). In the latter case, crystal faces are variably corroded at the contact (Fig. 18B, D). Measurement of the internal crystal angles (=180° minus the apparent interfacial angle) in cut sections of this sample yielded 75 measurements assigned to 10° bins. The results display two modes centred around 40-50° and 90-110° (Fig. 18C). Inspection of the crystal pseudomorphs which grew *within* sediment (Fig. 18E, G) is consistent with these results.

The preservation of the pseudomorphs is typically in the same mineral as the host carbonate, dolomite or calcite as appropriate. In some cases, the infilling phase is wholly cementing (Fig. 18E), whilst in others the variable internal fabrics point to a dominantly replacive origin (Fig. 18G). Such an origin is very clear for the upward-growing crystals where each is pseudomorphed in a mosaic of 20-100 μm calcite crystals, which in stained thin section show a non-ferroan core and a ferroan periphery (Figs. 18D, 20A).

Interpretation

The lack of size-sorting in FA5 sediments indicates they were laid down in a water body lacking significant current activity, but on unstable slopes as suggested by the soft-sediment folds, interpreted as slump structures. All transitions occur from disturbed and slump-folded Facies 5R to resedimented diamictites (Facies 5D, interpreted as debris flows) composed largely of Facies 5F blocks with some exotic clasts, are seen. The clastic sediment in Facies 5R is clearly glacially derived because it is invariably very poorly sorted, contains diamictite pellets likely to be derived by ice rafting (till pellets) and local ice-rafted clasts (dropstones), and gravel is present in thicker laminae. Also FA5 transitions upwards and downwards into stratified diamictites interpreted by Fleming et al. (2016) as representing more continous deposition from floating ice. In contrast, no distinct fine-grained sediment-gravity flows were observed, implying lack of proximity to fluvial input to the water body.The carbonate laminae commonly display evidence (thrombolitic domal growth morphology and complex microstructures) of a microbial origin (Fairchild, 1991; Riding, 2000), including much evidence for in-situ carbonate precipitation, coupled with some siliciclastic sediment incorporation. A combination of photosynthesis and favourable nucleation of carabonate within extracellular polymeric substance and dead cellular material present in microbial mats can be envisaged as promoting carbonate precipitation (Riding, 2000; Bosak & Newman, 2003). However, in some cases, there is no clear microbial structure within carbonate layers and we cannot rule out settling of water-column precipitates in these cases.

The highly regular nature of the mm-scale carbonate-siliciclastic couplets, particularly in the thickness of the carbonate layers, is striking (Figs. 15, 16) and there are many modern and late Quaternary lacustrine analogues in which such couplets are annual, i.e. varves. Although there is no barrier to such sediments forming under marine conditions, it is true that photosynthesis raises carbonate saturation faster in low ionic strength waters than in seawater

(Fairchild, 1991) and no modern marine analogues have been described. The dominant process creating carbonates in Alpine lakes is water-column photosynthesis from algal blooms (Kelts & Hsü, 1978), and this has guided many interpretations of late Quaternary varves, allowing inference of a succession of events through the year (e.g. Neugebauer et al., 2012). However lakes vary greatly in their hydrology, internal structure, salinity and state of carbonate saturation and there are many different patterns, e.g. carbonate mineral production may continue through to the autumn (Shanahan et al., 2008) or may partly depend on periods when there is input of ions from riverine input (Stockhecke et al., 2012), or may in part relate to winter freezing conditions (Kalugin et al., 2013). Variations in redox conditions over time influence the preservation of varves in modern lakes and in the case of the Gotland deep of the brackish Baltic Sea episodic oxygenation episodes may trigger a chain of events leading to characteristic Mn-carbonate layers superimposed on the annual pattern (Virtasolo et al., 2011). The absence of burrowing organisms in Cryogenian times led to continuous preservation of varve structure, but the Mn-chemistry of these carbonates is quite consistent (Fairchild et al., 1989; Bao et al., 2009) implying less strong redox fluctuations in the water body.

The crystal pseudomorphs also present important environmental evidence. The facies occurrence of these pseudomorphs argues against an evaporative origin. Although dolomite samples with high isotope values indicates some evaporation, the oxygen isotope composition of the dominant limestone and dolomite occurrences is typical for facies association 5 and lacks evidence for increased salinity. This, together with the carbonate composition, implies a carbonate precursor. The very regular mode of replacement by calcite with internal crystal-growth zonation (Fig. 17E) indicates that the precursor was more soluble than calcite, but still capable of forming euhedra, and hence crystalline rather than amorphous. Vaterite and monohydrocalcite can be ruled out because they invariably form spherulites or microcrystalline precipitates (Dahl & Buchardt, 2006; Pollock et al., 2006; Rodriguez-Blanco et al., 2014). Although aragonite typically forms mosaics of microcrystalline orthorhombic fibres, it is capable of forming radiating pseudo-hexagonal twinned "ray" crystals of similar size to those seen in the present study (Fairchild et al., 1990, Riccioni et al., 1996). However, where six-sided cross-sections are seen in the Wilsonbreen Formation (Fig. 16E), they are often elongated rather than equant, and the most elongated crystals in cross-section show a pair of terminating faces, not a basal pinacoid characteristic of aragonite. Strontium content of Wilsonbreen rhythmites is also low, whereas it is typically high in formerly aragonitic limestones (e.g Fairchild et al., 1990).

Ikaite is a high-pressure phase, metastable at Earth surface conditions, but becomes relatively more stable at cold temperatures (Kawano et al., 2009), and nearly always forms naturally at cool temperatures (-1.9 to +7° C, Huggett et al., 2005), on the sediment surface in spring-fed alkaline lakes and fjords and within marine sediment (Buchardt et al., 2001). It readily disintegrates to form calcite unless the solution contains an inhibitor for calcite precipitation. Bischoff et al. (1993) found that phosphate was most effective in this respect and indeed significant phosphate levels are typical of modern ikaite occurrences (Huggett et al., 2005; Selleck et al., 2007). It is acknowledged that ikaite may have been much more widely present as a primary phase than had been realized (Shearman & Smith, 1985; Bischoff et al., 1993; Buchardt et al., 2001). This is being borne out by new discoveries such as in sea ice (Fig. 19C, Nomura et al., 2013) and as millimetre-scale crystals in cold lakes (Fig. 19D, E; from the Patagonian Argentinian Laguna Potrok Alke, Oehlerich et al., 2013).

In the discussion above, ikaite as the precursor phase for the pseudomorphs in the FA5 was deduced by a process of elimination, but it is important to demonstrate that the observed properties are consistent with this identification. Ikaite is a monoclinic mineral that varies considerably in habit, from equant (Sekkal & Zaoui, 2013) to elongate prismatic (Buchardt et al.

2001; Last et al., 2013) and the dominant crystallographic forms vary greatly, and may be stepped or curved (Shearman & Smith, 1985), which make its positive identification difficult. Most of our knowledge of its likely morphology comes from pseudomorphs, including the "bipyramidal" aggregates of crystals known as glendonites (David et al., 1905; Fig. 18A), found in shales associated with Permian glacial deposits and other cool-water environments and the "thinolites" of Quaternary lakes of the western Great Basin, USA, figured by Dana (1884)[Fig. 19B] and interpreted as ikaite pseudomorphs by Shearman & Smith (1985) and Shearman et al. (1989). Swainson & Hammond (2001) reinforced this identification following determination of refined lattice cell parameters of a= 8.8, b = 8.3 and c = 11.0 Å with angle β between a and c of 110°.

The pseudomorphs of the Wilsonbreen Formation are interpreted to represent a combination of forms. Prism and pinacoid forms meet at internal angles of 90° and 110° when cut at a high angle to the faces. This would account for the higher mode in the crystal angle distribution (Fig. 18C) and the common "rhomb" shapes in section, similar to the crystals in Fig. 19C. Secondly the 30-60° mode is interpreted as representing the junctions between prismatic faces such as those seen to terminate crystals in Figs. 18A, 19B, E. It is notable that ikaite is only one-third as dense as calcite and so pseudomorphs in calcite would be expected to be initially highly porous even if the $CaCO_3$ was precipitated locally. This is consistent with the styles of preservation observed (Fig. 16), the example of Figs. 17E and 18D comparing well with examples in Larsen (1994), Huggett et al. (2005), Selleck et al. (2007) and Frank et al. (2008). An important corollary is that the replacive calcite mosaics observed in FA5 (and FA4) limestones more generally (Figs. 15B, 17F) are likewise also likely to be after ikaite.

Identification of a diagenetic process, ikaite replacement, that is expected to be syngenetic, and preservation of the crystal growth zones in CL, lends weight to the preservation of primary isotopic chemistry in FA5 (and FA4) carbonates. Furthermore, for FA5 the alternation with ice-rafted sediment taken to indicate temperatures consistently close to freezing and hence given the low $\delta^{18}O$ signatures, the water body must have been fresh. Application of Kim & O'Neil's (1997) experimentally determined fractionation factors extrapolated to 0° C, indicates that water compositions on the VSMOW scale are approximately 2.7 ‰ higher than calcite compositions on the VPDB scale, i.e. -8.3 to -15.5 ‰, a range which is likely to reflect mixing of different water sources. Dolomite facies have values on average 6 ‰ higher, implying formation from waters of on average higher salinity and higher mean Mg/Ca (Müller et al., 1972).

There are plenty of modern analogues for calcareous microbial laminae in cold lakes, even in such extreme environments today as the ice-covered lakes of the McMurdo Dry Valleys of Antarctica (Parker et al., 1981), or reducing solution hollows beneath the Great Lakes (Voorhuis et al., 2012). One curious phenomenon are thick cement crusts found in both FA5 and FA2. A possible origin for these is the phenomenon observed in certain Antarctic lakes of the localized lifting of mats by gas generation (Parker et al., 1981). Early cementation of the mat would allow a more gradual fill of the resulting fenestrae as found in modern Antarctic lakes (Wharton et al., 1982).

In respect of seasonality, many well-studied modern lakes show a turnover associated with cooling in winter and may have a frozen surface in that season. In that case, winter sedimentation is typically dominated by clay and organic matter (e.g. Lauterbach et al., 2011, Kalugin et al., 2013), but both these components are scarce in the Wilsonbreen Formation examples. In Antarctica, carbonate precipitation is linked to peak water column or microbial mat photosynthesis in late Spring to early summer (e.g. Wharton et al., 1982; Lawrence & Hendy, 1985), limited also by nutrient availability. Higher sediment input would be expected in the late summer to autumn when ice cover was at a minimum.

A clue to the overall similarity of the microbial fabrics in the different facies associations may be discerned in the work on Antarctic stromatolites. The first major mat-former to be identified (*Phormidium frigidum* Fritsch) was known to be pre-adapted to cold environments and low light conditions (Parker et al., 1981) and tolerates conditions from fresh to saline and anoxic to oxygen-saturated. Simmons et al. (1993) amplified that this species is found not only in lakes, but also in glacial meltstreams, soils and cryoconite holes on ablated glacier surfaces, that is easily spanning the range of environments encountered for microbial deposits in the Wilsonbreen Formation. Voorhuis et al. (2012) found that a species of *Phormidium* also dominated the mat community at low oxygen levels in a 23 m-deep Great Lakes sinkhole and demonstrated that the genus has genetic properties that facilitate toleration of sulphide or utilization of it for anoxygenic photosynthesis. It therefore seems that it is likely that the Wilsonbreen Formation microbial communities were of low diversity, as expected for extremophiles, with *Phormidium* or a similar ancestor, dominating the biota.

Vertical and lateral facies relationships

Vertical facies transitions can be used to establish whether sedimentary successions have predictably cyclic, Markov properties (Powers & Easterling, 1982). A transition matrix derived from all the logged W2 sections (Fig. S7) demonstrates that there are vertical transitions between all of the facies associations, except FA1/FA7 which are only found bounding W2, not within it. This indicates a stochastic element to the facies accumulation, but there are also clearly preferred transitions. Although data are insufficient for formal statistical analysis, since total occurrences of each Facies Association are similar, the relative number of transitions is a useful guide (Fig. 22). Two sets of the most common transitions concur with the interpretations made earlier. Firstly, fluvial channels (FA2) most commonly pass up into floodplain sediments (FA3) and vice-versa. Such a close relationship could not be anticipated if FA2 were marine. Secondly, calcareous lake sediments (FA5) pass into FA6 (glacilacustrine) and vice-versa (and this is also the dominant pattern in member W3). This relationship emphasizes the transitional nature of change between major amounts of floating ice and reduction in ice sufficient to permit carbonate accumulation in microbial mats. Other transitions, such as those between ice-rafted sedimentation (FA6) and floodplain and lake-marginal sediments (FA3 and 4), would require more sudden changes in lake level.

Overall, the Northeast Greenland-NE Svalbard Neoproterozoic sedimentary basin is envisaged as elongate, with basement exposed in the far SW in Greenland (Fairchild & Hambrey, 1995) and axial glacier flow to the NNE in NE Svalbard (Fleming et al., 2016). In the Formation as a whole (Fig. 1) the inferred palaeogeography is reflected in sourcing the subglacial advances from the SSW correlated with grounding-line facies in W3 only in the northern sections. Within member W2 however, the facies mosaics do not illustrate this overall pattern so clearly. Fig. 22 illustrates the cumulative thickness of each of the facies associations and demonstrates that there is no simple spatial trend in facies within W2, except perhaps for the high incidence of fluvial facies (FA2) in the southernmost section. Such a complex facies mosaic implies that water and sediment are derived from multiple sources.

Integration of sedimentological, geochemical and modelling evidence

In this section we extend the arguments as to why the Wilsonbreen Formation environments can be considered as a coherent whole and hence be considered by a modern analogue in a single climatic setting. Considering first the sedimentological evidence, FA5 contains clear evidence of syn-glacial deposition (ice-rafted sediment and ikaite formation) and Facies 4R and 4I show similar types of carbonate accumulation, and localized dropstone deposition, but more evidence of reworking by waves in shallow water. FA3 evinces a hyperarid terrestrial environment that nevertheless borders accumulation of microbial tufas in streams with regularly fluctuating discharge (Facies 2S). Facies 4S evokes a landscape with significant aeolian sediment transport. All of these facies are consistent with a cold, arid terrestrial environment, but few resemble standard Neoproterozoic facies between glaciations (Fairchild, 1993).

Secondly, the geochemical evidence from sulphate isotope systematics (Bao et al., 2009; Benn et al., 2015) demonstrates that throughout the deposition of all the carbonate facies in W2 and W3, atmospheric P_{CO_2} was very high (probably of the order of 0.1 bar) and yet these are icehouse, not greenhouse sediments.

Thirdly, a variety of modelling constraints indicate that glaciation and high P_{CO_2} can only coincide in the case of a Snowball Earth, that is a planet that is undergoing a hysteresis of CO_2 and climate in which initial low P_{CO_2} triggers glaciation which at a critical stage of development becomes globally distributed because of ice-albedo feedbacks. A necessary condition to escape from such an extreme ice age is build-up of high P_{CO_2}. Modelling of continental environments with the Wilsonbreen case in mind demonstrates that at sufficiently high P_{CO_2}, glacial advances and retreats can be triggered by precessional forcing (Benn et al., 2015), but that the climate remains uniformly cold, changing primarily in terms of the accumulation of snow and ablation of snow and ice. Continental ice volumes change little and so significant sea level change would not be expected. Deglaciation to warm, ice-free conditions is not reversible without a necessarily slow (multi-million year) fall in P_{CO_2} (Le Hir et al., 2009). In this perspective, once one has demonstrated non-marine conditions for part of the Wilsonbreen Formation, then the expectation would be that it would all be continental in character. Marine transgression would not be expected until glaciation is over, and indeed such a distinct transgression is found truncating the Wilsonbreen Formation (Fairchild & Hambrey, 1984; Halverson et al., 2004). Overall this implies that in Svalbard, isostatic depression by neighbouring ice sheets was much smaller than eustatic sea level fall due to build-up of ice (Benn et al., 2015). In summary, the Wilsonbreen facies are likely to reflect a mosaic of arid terrestrial environments subjected to glaciation, which brings us back to the modern analogue.

Comparison with the McMurdo Dry Valleys and synthesis

The aim of this section is both to assess the degree of similarity of the modern and Neoproterozoic settings and to gain further insights from modern and Quaternary phenomena into the likely controls on Wilsonbreen deposition. We start by examining three topics where differences might be expected: tectonic setting, (palaeo)latitude, and the composition of bedrocks undergoing weathering with implications for carbonate mineralogy, and then use insights from the Dry Valleys to aid synthesizing the interpretation of stable isotope fields of the carbonates.

The Wilsonbreen Formation represents a small part of a largely marine depositional basin that persisted for more than 300 My and deposited a pile of sediments (Hecla Hoek Supergroup) many kilometres thick. The outcrop belt has a NNE-SSW trend, and the original basin may have been an intracratonic rift with a similar trend (Cawood et al., 2007) leading to the tabular nature of the Formation from north to south. The Dry Valleys region occupies a hinge between the uplifting Transantarctic Mountains and the subsiding McMurdo Sound (Etienne et al., 2007). Whilst upland areas not covered by ice have neither eroded nor accumulated sediment, significant Quaternary sediment accumulations have been drilled in Lower Taylor Valley (McKelvey, 1981). Meltwater is supplied both from outlet glaciers and local valley glaciers, and ice sheet advance from the adjacent Ross Embayment has occurred repeatedly. In contrast, since the Svalbard area was already at sea level prior to glaciation, and a panglaciation then ensued with no thick ice cover locally, a pronounced eustatic sea level fall should have isolated the basin from the sea. The proximity of upland areas to source glaciers is also unknown, but this basin configuration offers the possibility of glacier advance both along and across-strike, leading to creation and destruction of lakes, as has been the case in the late Quaternary in the Dry Valleys. Such externally imposed changes in lake level or glacial advances and retreats would be superimposed on any intrinsic tendency for facies migration and are the most likely reason for the myriad vertical facies transitions and complex lateral geometries in the Wilsonbreen Formation.

The high palaeolatitudes of the Dry Valleys ensures a high seasonality of temperature such that meltwater production is limited to a couple of months in summer. The Wilsonbreen Formation is thought to have lain in the sub-tropics (Li et al., 2013) where seasonality today primarily relates to moisture balance. This also applies in the late stages of a Snowball Earth, since general circulation modelling indicates that alternating seasons in which precipitation exceeds ablation in the summer and vice-versa in the winter would have been a strong feature of the climate in latitudes 20-30° from the equator (Benn et al., 2015). The Snowball climate also has greatly enhanced diurnal and annual temperature variability at a given latitude compared with today (Pierrehumbert, 2005) since temperature responds directly to insolation, with less effective smoothing than today by lateral latent heat transport by moisture. The implications remain to be fully explored, but Benn et al. (2015) gives an indicative plot (their Fig. S15) demonstrating that, in the month of April, although mean temperatures remain below zero, lowland areas in the tropics and sub-tropics should experience a significant number of positive degree days when melting can occur. The overall effect is that, open water in Wilsonbreen lakes was clearly extensive, at least seasonally, contrasting with the perennial ice cover in the Dry Valleys, apart from marginal moats.

Another latitude-related factor is the relative sensitivity to climatic change related to orbital forcing. Whilst the Dry Valley region is susceptible to facies changes related to subtle environmental changes on inter-annual to multimillennial scales, the most significant environmental changes related to glacial advances and retreats are on orbital timescales (Hendy 1980). Fig. 1 illustrates that for the Wilsonbreen Formation as a whole, there are about ten periods of cessation of glacial deposition, with uncertainties because of the thin and lensing (possibly eroded) intervals in W3 in particular. Accumulation rates of annually laminated carbonate facies, both fluvially (FA2) and in lakes (FA4/FA5) is of the order of 1 mm/year: hence such retreat phases represent *minimum* periods of 10^3-10^4 years since periods of non-deposition due to lack of local accommodation space or sediment supply are likely. Such timescales are approaching the present-day Milankovitch scale of climatic fluctuation, an observation which stimulated the modelling of precessional cycles in Benn et al. (2015). Given that precession (on the ca. 20 ky timescale) is greatly enhanced during higher eccentricity (on the 100 ka timescale), a plausible interpretation of members W2 and W3 is that they represent one or two eccentricity cycles, within each of which several precession-related fluctuations are recorded.

Modern Antarctica has very little carbonate bedrock. Nevertheless, alkalinity is generated both by carbonate and silicate weathering, permitting carbonate precipitation, particularly in lakes, controlled by seasonal photosynthesis. By contrast, the Wilsonbreen Formation reflects the erosion of carbonate-rich catchments and calcite precipitation is recognized also in streams. Although limestone makes up 20% of the gravel fraction, dolomite is the only carbonate in the fine fraction of detritus because of preferential calcite dissolution, implying that unmodified molar Mg/Ca ratios will on average be <1 , although will increase as calcite is reprecipitated within the basin. Data from streams and most lakes in the Dry Valleys also have Mg/Ca < 1, although extensive $CaCO_3$ precipitation in Lake Fryxell leads to enhanced Mg/Ca (Green et al., 1988) and presumably such waters were responsible for precipitation of aragonite in many late Quaternary lakes (Hendy, 1979). Rather than aragonite, ikaite is recognized as a precursor to calcite in the Wilsonbreen case. Modern examples of ikaite formation are associated with high aqueous phosphate which acts as an inhibitor to calcite precipitation (Bischoff et al., 1993) and yet modern Antarctic lakes are oligotrophic. However, Burton (1993) stressed that the role of inhibitors is complex because there are many potential inhibiting species (e.g. magnesium, phosphate, sulphate, organic radicals and many other species depending on concentration) and they can interact. One possible factor is that a circum-neutral pH would be expected in Wilsonbreen waters because of high atmospheric P_{CO_2} which has the effect of increasing the balance of HPO_4^{2-} to PO_4^{3-} and strengthening the phosphate-inhibition effect on calcite; this effect is also accentuated by high sulphate (Burton, 1993). Modern dolomite has not yet been recognized in Antarctica, yet it is common in the Wilsonbreen facies, apparently aided by factors such as evaporation and fluctuating redox, and possibly also the widespread occurrence of dolomite rock flour as nuclei.

In the previous text, extensive evidence has been presented for the development of stable mineralogy in the environment by primary precipitation or early diagenesis, a stability favoured by low environmental temperatures. In turn this argues for treating the stable isotope data as direct indicators of depositional conditions. Given the very wide range of $\delta^{18}O$ values, the effects of variation in depositional temperature, or minor post-depositional exchange, are minimized. Carbon isotopes are in any case resilient to secondary alteration (Banner & Hanson, 1990). A synthesis of the Wilsonbreen carbonates in relation to the stable isotope fields is given in Fig. 23. The $\delta^{13}C$ signature is linked to acquisition of carbon from bedrock and oxidation of organic sources and variable equilibration with atmospheric carbon dioxide, whilst variability in $\delta^{18}O$ is interpreted as due to mixing of meltwaters or different compositions of primary meltwaters coupled with evaporation. We now discuss these as factors in the Dry Valley context.

Regarding carbon isotopes, in the Dry Valleys the $\delta^{13}C$ composition of a carbonate source is not well defined, but a potential organic source is found in the form of dark-coloured cryoconite holes, which periodically are flushed, providing nutrients for ephemeral streams and lake basins (Bagshaw et al., 2013). Addition of respired carbon from this source can be seen in the chemistry of some Antarctic streams with $\delta^{13}C$ values as light as -9.4 ‰ (Lyons et al., 2013). Most values are -3 to +2 ‰, possibly reflecting carbon contribution from a carbonate source rock (Leng & Marshall, 2004) and ranging up to 5 ‰ which Lyons et al. (2013) attribute qualitatively to progressive equilibration with atmospheric CO_2. Chemical equilibration between atmospheric CO_2 and an Alpine meltstream was shown to be attained within a few hundred metres of flow (Fairchild et al., 1999), although this situation falls short of *isotopic* equilibration, requiring more extensive exchange until the large gaseous source of carbon dominates (Fairchild & Baker, 2012, chapter 5). More efficient processes for modifying $\delta^{13}C$ are found in Dry Valley lakes, where vertical trends caused by water-column photosynthesis at shallow depths (leading to positive values of $\delta^{13}C$), organic matter oxidation (causing a decrease in $\delta^{13}C$ with depth), and local methanogenesis (releasing CO_2 with complementary high $\delta^{13}C$ values at depth) have been described (Lawrence & Hendy, 1985; Neumann et al., 2004). Cold lakes in other settings can sometimes develop very low $\delta^{13}C$ values through release of methane from solid hydrates (Propenko and Williams, 2005).

For oxygen isotopes in streams, Gooseff et al. (2006) gathered data on $\delta^{18}O$ variation for glacier ice, snow, streams and lakes, although interpretation was complicated by evident strong inter-annual variations. In Taylor Valley the mean composition of glacier ice in each sub-basin feeding a specific lake varied from -21 to -33 ‰ and some samples were as light as -45 ‰. Stream samples varied from -42 to -22 ‰ and lay close to the meteoric-water line up to -32 ‰, but diverged from it at higher values, reflecting the consequences of evaporation. This corresponded also to stream lengths of greater than 2 km. In addition to simple surface evaporation, mixing with isotopically heavy shallow subsurface (hyporheic) water was also a factor, indicative of the effectiveness of evaporation of subsurface water at a shallow water table. Further isotope fractionation occurs where water resides longer in saline lakes. Matsubaya et al. (1979) measured and modelled the ^{18}O-enrichments in the most saline ponds in the Dry Valley area and found up to 20 ‰ higher values in the unfrozen Don Juan Pond and the east lobe of Lake Bonney compared with the source. Likewise, Nakai et al. (1975) documented a 23 ‰ variation in $\delta^{18}O$ composition of calcites with the highest values representing evaporative deposits on land surfaces. Such pronounced ^{18}O enrichments along streamcourses and the ca. 20 ‰ total variation are only permitted because of the meteorological factors leading to persistently low relative humidities of 50-60%.

For the ancient carbonates, it is useful to consider the presence or absence of covariance of $\delta^{18}O$ and $\delta^{13}C$ (Figs. 7, 23). Talbot (1990) compiled data from a variety of modern and ancient lakes and found that a lack of isotopic covariation is typical of open lakes in which controlling factors for the variation in the two isotopes are decoupled. This can be compared with the limestones of FA4 and FA5 (Figs. 7b, 23). Conversely, covarying trends in lacustrine carbonates

(FA4 and FA5 dolomites) were identified as a characteristic feature of closed lakes and reflected a combination of evaporation to cause increase in $\delta^{18}O$, and residence time, permitting equilibration with atmospheric CO_2 as a first-order control on ultimate $\delta^{13}C$ values. Although $\delta^{13}C$ and $\delta^{18}O$ data on Antarctic streams (Gooseff et al., 1006; Lyons et al., 2013) was obtained separately and so has not been cross-plotted, these environments too can be anticipated to display covariation. The slopes of covariation in Talbot's (1990) data were found to be quite varied, with lower slopes interpreted as reflecting broad, shallow lakes in which evaporation would play a more prominent role. The data were primarily from limestone and water-column precipitates, although examples were shown where benthic carbonate and dolomites fitted the trends. Apart from attaining high absolute values for $\delta^{18}O$ in FA3, the Wilsonbreen Formation carbonates show covarying slopes and ranges within those presented by Talbot (1990), providing a confirmation that the evaporation-equilibration explanation for data trends is reasonable.

First we consider $\delta^{13}C$ data: where covariations are absent and at the starting points for covaration. The initial carbon isotope composition of a glacial meltstream may not reflect any atmospheric influence because many meltwaters derive from ice with low air content and have P_{CO_2} values well below atmospheric (Fairchild et al., 1994). The mean $\delta^{13}C$ of detrital dolomite in the Wilsonbreen Formation is +2.4 ‰ (Fig. 6B), but with significant local variations (±1.3 ‰) and additional uncertainty because limestone appears to have preferentially dissolved from the matrix. A combination of moderately positive $\delta^{13}C$ from detritus and organic carbon could have fixed the starting $\delta^{13}C$ value of around +1 ‰ for the FA2 covarying trend. The higher starting point for FA3 reflects the more evolved nature of these fluids which are consistently forming dolomite. The trend for FA4 and FA5 dolomites starts at lower values of 0 to +1 ‰, within the range of calcites in these lacustrine facies. Carbon isotope signatures < +1 ‰, that is lower than those in fluvial facies, presumably reflects addition of carbon from an organic source, but overall variations are not large enough to suggest a role for methanogenesis. In the open lakes implied by the lack of isotope covariation in FA4 and FA5 calcites, the high $\delta^{13}C$ values could reflect either the effects of photosynthesis or greater CO_2-equilibration without evaporation. The most extreme high $\delta^{13}C$-low $\delta^{18}O$ limestones in FA4 are from member W3 (Fig. 12); these are intraclastic breccias implying possible exposure which would have aided atmospheric equilibration. No difference is noted between obviously microbial and other limestones, a common pattern in the Neoproterozoic (Fairchild, 1991). Low $\delta^{13}C$ values occur in each section studied location and are almost always stratigraphically close to values that are much higher, possibly indicative of changing lake levels. The FA5 sediments from W3 all have relatively low values which might reflect a relatively deep water setting, consistent with the dominance of ice-rafted sedimentation in this member in the southern sections and the universal disruption of FA5 sediments by slumping.

Now we consider the theoretical $\delta^{13}C$ end-point resulting from equilibration. Carbonate precipitated from a solution in isotopic equilibrium with atmospheric CO_2 is expected to display $\delta^{13}C$ values heavier than the atmosphere by 10.4 ‰ at 0 °C (falling to 9.1 ‰ at 20 °C), as calibrated by the experimental work of Mook et al. (1974). The long-term (>10^8 year) $\delta^{13}C$ composition of the atmosphere should show variation largely in parallel with ocean water with which it tends to equilibrate, and ocean water in turn has a composition constrained by the proportional burial of isotopically light organic carbon. As a result, short-term variations can be expected because of flux variabilities, as demonstrated by direct measurement of past (pre-industrial Holocene) atmospheres from ice cores showing a range from -6.3 to -6.6 ‰ (Elsig et al., 2009). A further mass-balance constraint is the bulk Earth mean composition of carbon (Berner, 2004). The latter can be estimated from mantle and meteorite samples as around -7 ‰, but volcanic gases are typically somewhat heavier (Javoy et al., 1986). An atmosphere with a composition around -6 ‰ would be in equilibrium at zero degrees with carbonates around +4.4 ‰.

The carbon dioxide level in the Snowball Earth atmosphere should have progressively risen because of sustained input from volcanic sources and limited removal, mainly by dissolution in the ocean wherever gaps in the ice cover occurred (Le Hir et al., 2008). The limited opportunities for back-exchange from the oceans imply that the atmosphere should have provided a good sample of the carbon isotope composition of volcanic emissions. For the W2 data, each of the facies association fields (Figs. 12, 23) tops out at around 3.5 to 4.5 ‰ which is close to the expected values for equilibration with the atmosphere dominated by volcanic emissions as discussed above. Member W3 dolocretes have $\delta^{13}C$ values 1 ‰ higher than considered so far (Fig. 12), but this difference is difficult to interpret without an overall data trend. One possibility is that there is a local contribution by freezing, which Lyons et al. (2013) invoke to explain $\delta^{13}C$ values in the range +5 to +12 ‰ on carbonate-encrusted rocks on the Dry Valleys' land surface.

We now focus attention on $\delta^{18}O$, starting with the range of values in covarying trends. The range of $\delta^{18}O$ in FA2 limestones is rather less than observed in the Dry Valley streams, demonstrating the feasibility of evaporation as a driver for variability, although some variation in source water composition is also possible. Similar remarks apply to the dolomites of FA4 and FA5, although largely open lacustrine environments such as these are not found in the Dry Valleys. The slope of covariation is much lower for the FA3 than for the other facies associations, which is consistent with a surficial origin leading to more efficient evaporation (Talbot, 1990). Allowing for a 3 ‰ offset between dolomite and calcite (Land, 1980), the evaporative trend (Figs. 7B, 12) extends beyond the composition of fluvial limestones by a further 11 ‰ for dolocretes (Facies 3D) and 15 ‰ for stromatolitic laminites (Facies 3S). Evaporation of coastal seawater, under typical high humidity conditions would cause an increase in δ^{18}O of at most 6 ‰, whereas 17 ‰ increase from a -10 ‰ starting point was observed in a freshwater Texan pond under conditions of less than 50% humidity (Lloyd, 1966). The extreme enrichments in FA3 stromatolitic crusts, require similarly low humidities to this example and the Dry Valleys, and could only be possible for facies developed at the land surface.

The FA4 and FA5 lacustrine calcites have lower δ^{18}O compositions than those of FA2 fluvial limestones. By analogy with modern environments, this is likely to reflect a local, low altitude source of water for FA2, whereas the variability in δ^{18}O within FA4 and FA5 could reflect varying meltwater sources, including large glaciers with low δ^{18}O. The range of δ^{18}O is actually rather less in FA2 than in the Dry Valley streams and given that the values are typically heavier than the lakes, a relatively local, low-altitude source for meltwater is implied. For the Carbonate Lake Margin FA4, the δ^{18}O values are similar to Carbonate Lake FA5 and there is a lack of isotopic covariation with both sets of data. This would imply an open lake (Talbot, 1990), but care is needed with such an interpretation because the data represent several different lakes that formed successively. As derived earlier, the range of $\delta^{18}O_{water}$ values implied from calcite precipitation at 0° C is around -8.5 to -15.5 ‰ on the V-SMOW scale. This range might be explained by variable mixing of local snow, local glacier ice and melt from larger or higher glaciers, by analogy with the Dry Valley region.

The ultimate driver for Rayleigh fractionation in the atmosphere, which leads to ^{18}O-depleted values in atmospheric precipitation, is partial condensation and removal of vapour as a result of a fall in temperature of the air mass. Inferred Wilsonbreen meltwater δ^{18}O values higher than those in the modern Dry Valleys implies a lesser degree of fractionation, as originally noted by Fairchild et al. (1989). A comparable modern glacial area for the

Wilsbonbreen in terms of $\delta^{18}O$ is the Vatnajökull ice cap of Iceland below which Robinson et al. (2009) found a representative ice- and snowmelt composition of -12 ‰. Note that indirect evidence for more isotopically light Neoproterozoic meltwater has been found in two records from South China (Zhao & Zheng, 2010; Peng et al., 2013), although these do not necessarily relate to a panglacial and in the former case is a younger, Ediacaran glaciation. A more pertinent record is that of Kennedy et al. (2008) in calcite-cemented Marinoan tidal siltstones of South Australia where calcite as light as -25 ‰ was analyzed, and interpreted to reflect input of meltwater from highly fractionated low-latitude ice sheets, although there may be other possible interpretations of these data bearing in mind the active decomposition of clathrates at this site. In summary, the distinctively heavy isotopic signature of Wilsonbreen meltwater may be a characteristic feature of low-latitude glaciation and clearly requires study by an isotope-enabled general circulation model.

CONCLUSIONS

1. Carbonate in the Wilsonbreen Formation confirms the non-marine environment of deposition in a glaciated basin supplied with debris from erosion of platform carbonates. The consistently dolomitic nature of the detrital matrix of Wilsonbreen Formation sediments contrasts with the common presence of limestone in coarser debris and demonstrates the preferential dissolution of calcite over dolomite. Carbonate-rich meltwaters were thus able to precipitate $CaCO_3$ once supersaturation was achieved by processes such as photosynthesis or evaporation.

2. Evidence from physical sedimentary structures allows four facies associations to be distinguished in which carbonate was precipitated, distinct from three facies associations

dominated by glacial and periglacial processes. The Fluvial Channel Facies Association (FA2) commonly contains microbial limestone laminae with mm-scale lamination and notable syn-depositional radiaxial calcite cements. These compare physically with modern microbial mats in ephemeral Antarctic streams. The Dolomitic Floodplain Facies Association (FA3) consists of soil zone/playa surficial dolocretes and dolomitic stromatolites in which dolomite was probably a primary precipitate. The Carbonate Lake Margin Facies Association (FA4) typically displays micritic and locally microbial laminae interlaminated with wave-sorted silts and sands with local development of intraclastic breccias. Finally, the Carbonate Lake Facies Association (FA5) displays mm-scale (varved) alternations of micritic carbonate or laminated microbial carbonate (including microsparitic and fenestral laminae) with poorly sorted sediment containing recognizable ice-rafted debris. Locally in member W2, and pervasively in member W3, partly lithified sediments were disturbed by slump folding and locally transformed to carbonate-rich debris flows.

3. In both FA4 and FA5, there are textural indicators of mineralogical replacements. Dolomite can be seen to replace a $CaCO_3$ precursor. Although some calcite is likely to be primary, calcite pseudomorphs after ikaite ($CaCO_3.6H_2O$) are common. The ikaite formed individual crystals within the sediment, formed crusts which grew centripetally into pores, and locally grew upwards at the sediment-water interface. This paragenesis is now becoming better known in modern cold lakes.

4. Stable isotope data demonstrate that carbonates in the different facies associations form distinct fields which are all interpreted as consistent with primary depositional conditions. Limestones in FA4 and FA5 lack $\delta^{13}C$- $\delta^{18}O$ covariation and were primarily influenced by mixing of meltwater sources, the variable addition of light carbon from organic decomposition, and some re-equilibration with the atmosphere. Other sub-sets demonstrate covariation,

interpreted as a combination of evaporation and equilibration with the atmosphere. This allows the $\delta^{13}C$ composition of CO_2 released from volcanism during the glaciation to be constrained to -6 to -7 ‰. Direct evidence of primary fluid compositions is unavailable because of secondary fluid migration into inclusions, despite the presence of primary trace element growth zones. Nevertheless, the very wide range of $\delta^{18}O$ values must be primarily related to changes in water composition, given the consistently cool depositional conditions. The exceptionally high $\delta^{18}O$ signatures of FA3 dolomites, up to +14.7 ‰$_{VPDB}$, attest to the hyperaridity of the environments. Conversely, the inferred $\delta^{18}O$ compositions of the input meltwaters (-8 to -15 ‰$_{VSMOW}$) are more comparable to modern Iceland than to present-day polar regions. This is likely to reflect relatively limited Rayleigh fractionation in the atmosphere because of its relative warmth linked to enhanced absorption of infra-red radiation from high CO_2 levels.

5. Although preferred facies transitions occur, there is no development overall of multi-facies cyclicity. The strong isotopic covariations associated with closed lakes and streams, and rhythmic carbonate laminae are strong motifs of non-marine facies. In these environments, the landscape was repeatedly transformed by damming and draining of lakes as the glaciers advanced and retreated. In turn these are likely to have represented amplified geomorphic responses to subtle climatic shifts in a persistently hyper-arid setting for which the McMurdo Dry Valleys provide a rich modern analogue. Although Wilsonbreen lakes were not perennially ice-covered as in the modern environment, many points of similarity between the modern Antarctic and the ancient environments have been drawn in this study, specifically including the rates and styles of sediment deposition, biogeochemistry, and extreme ^{18}O-enrichment related to the hyperarid climate.

6. The carbonates of the Wilsonbreen Formation are distinctive and include unique facies and record-breaking isotope compositions. They represent, along with interbedded

diamictites, the complex environmental response to changing rates of accumulation and ablation forced by a series of precessional cycles late in the evolution of a Snowball Earth.

ACKNOWLEDGEMENTS

The fieldwork and subsequent analyses were funded by Natural Environment Research Council grant GR3/ NE/H004963/1 within project GAINS (Glacial Activity In Neoproterozoic Svalbard). We are grateful to the Norwegian authorities in Svalbard and the logistical department of the University in Svalbard in facilitating our fieldwork, and other members of the GAINS team for their collegiality and scientific zest. The senior author records his personal appreciation to Paul Hoffman for his inspiration, and support for this project at a critical stage, to Martin Kennedy for his friendship, and rigorously critical feedback on Neoproterozoic data and ideas, and to Tony Spencer for his insight, and careful editorial comments on a draft. Journal referees, including Alex Brasier, gave invaluable advice that led to a restructuring of the paper and sharpening of its arguments.

REFERENCES

Abbott, D.S., Voigt, A. and **Koll, D.** (2011) The Jormungand global climate state and implications for Neoproterozoic glaciations. *J.Geophys. Res.*, **116**, D18103, doi:10.1029/2011JD015927.

Allen, P.A. (1984) Reconstruction of ancient sea conditions with an example from the Swiss Molasse. *Mar. Geol*., 60, 455-473.

Allen, P.A. and **Etienne, J.L.** (2008) Sedimentary challenge to Snowball Earth. *Nat. Geosci.*, **1,** 817-825.

Alonso-Zarza, A.M. (2003) Palaeoenvironmental significance of palustrine carbonates and calcretes in the geological record. *Earth-Sci. Rev.,* **60**, 261-298.

Andrews, J.E. (2006) Palaeoclimatic records from stable isotopes in riverine tufas: Synthesis and review. *Earth-Sci. Rev.*, **75**, 85-104.

Andrews, J.E. and **Brasier, A.T. (**2005) Seasonal records of climate change in annually laminated tufas: short review and future prospects. Journal of Quaternary Science 20, 411 – 421.

Arnaud, E., **Halverson, G. P.** and **Shields-Zhou, G.** (eds) (2011) The Geological Record of

Neoproterozoic glaciations. *Geol. Soc. London Mem.,* **36**.

Bagshaw, E.A., Tranter, M., Fountain, A.G., Welch, K., Basagic, H.J. and **Lyons, W.B.** (2013) Do cryoconite holes have the potential to be significant sources of C, N, and P to downstream depauperate ecosystems of Taylor Valley, Antarctica? *Arctic Alpine Antarctic Res.*, **45,** 440-454.

Banner, J.L. and **Hanson, G.N.** (1990) Calculation of simultaneous isotopic and trace-element variations during water-rock interaction with applications to carbonate diagenesis. *Geochim. Cosmochim. Acta*, **54**, 3123-3137.

Bao, H. (2014) Sulfate: A time capsule for Earth's O_2, O_3 and H_2O. *Chemical Geology*, 396, 108-118.

Bao, H., Lyons, J.R., and **Zhou, C.** (2008) Triple oxygen isotope evidence for elevated CO_2 levels after a Neoproterozoic glaciation. *Nature*, **453**, 504-506.

Bao, H., Fairchild, I.J., Wynn, P.M. and **Spötl, C.** (2009) Stretching the envelope of past surface environments: Neoproterozoic glacial lakes from Svalbard. *Science*, **323**, 119-122.

Bao, H., Chen, Z.Q. and **Zhou, C.M.** (2012) An O-17 record of late Neoproterozoic glaciation in the Kimberley region, Western Australia. *Precambrian Research*, **216-219**, 152-161.

Barrett, P.J. (2013) Resolving views on Antarctic Neogene glacial history – the Sirius debate. *Earth Env. Sci Trans.Roy. Soc. Edin.* **104**, 31-53.

Benn, D.I, Bao, H., Dumas, C., Fleming E.J., Hambrey, M.J., Le Hir, G., McMillan, E.A., Petronis, M.J., Ramstein, G., Stevenson, C.T.E., Wynn, P.M. and **Fairchild, I.J.** Orbitally forced climatic fluctuation in Snowball Earth: Geological evidence and model results. *For submission to Science.*

Berner, R.A. (2004) *The Phanerozoic Carbon Cycle*. OUP.

Bischoff, J L., Fitzpatrick, J.A. and **Rosenbauer, R.J.** (1993) The solubility and stabilization of ikaite ($CaCO_3 \cdot 6H_2O$) from 0° to 25°C: Environmental and palaeoclimatic implications for thinolitic tufa. *J. Geol.*, **101**, 21-33.

Bosak, T. and **Newman, D.K.** (2003) Microbial nucleation of calcium carbonate in the Precambrian. *Geology*, **31**, 577-580.

Brasier, A.T. (2011) Searching for travertines, calcretes and speleothems in deep time: Processes, appearances, predictions and the impact of plants. *Earth-Sci. Rev.* **104**, 213-239.

Buchardt, B., Israelson, C., Seaman, P. and **Stockmann, G.** (2001) Ikaite tufa towers in Ikka Fjord, southwest Greenland: their formation by mixing of seawater and alkaline spring water. *J. Sedim. Res.*, **71**, 176-189.

Budyko, M.I. (1969) The effects of solar radiation variations on the climate of the Earth. *Tellus*, **XXI**, 611–619.

Burton, E.A. (1993) Controls on marine carbonate mineralogy: review and assessment. *Chem. Geol.*, **105**, 163-179.

Caldeira, K. and **Kasting, J.F**. (1992) Susceptibility of the early Earth to irreversible glaciation caused by carbon dioxide clouds. *Science,* **359,** 226–228.

Calkin, P.E. and **Rutford, R.H.** (1974) The Sand Dunes of Victoria Valley, Antarctica. *Geog. Rev*., 64,189-216.

Calver, C. R., Crowley, J.L., Wingate, M.T.D., Evans, D.A.D., Raub, T.D. and **Schmitz, M.D.** (2013) Globally synchronous Marinoan deglaciation indicated by U-Pb geochronology of the Cottons Breccia, Tasmania, Australia. *Geology*, **41**, 1127-1130.

Cao, X. and **Bao, H.** (2013) Dynamic model constraints on oxygen-17 depletion in atmospheric O_2 after a snowball Earth. *Proc. Natl. Acad Sci. USA*, **110**, 14546-14550.

Cawood, P.A., Nemchin, A., Strachan, R., Prave, T. and **Krebbendam, M.** (2007) Sedimentary basin and detrital zircon records along East Laurentia and Baltica during assembly and breakup of Rodinia. *J. Geol. Soc. Lond.*, **164**, 257-275.

Cohen, A.S. and **Thouin, C.** (1987) Nearshore carbonate deposits in Lake Tanganyika. *Geology*, **15**, 414-418.

Dahl, K. and **Buchardt, B.** (2006) Monohydrocalcite in the Arctic Ikka Fjord, SW Greenland: First reported marine occurrence. *J. Sedim. Res.*, **76,** 460-471.

Dana, E.S. (1884). A crystallographic study of the thinolite of Lake Lahontan. U.S. Geol. Surv. Bull. **12,** 429–450.

David, T.W.E., Taylor, J.G., Woolnough, W.G. and **Foxall, H.G**. (1905) Occurrence of the pseudomorph glendonite in New South Wales. *Rec. Geol. Surv. N.S.W.,* **3**, 161-179.

Daves, N.S. and **Gibling, M.R.** (2010) Paleozoic vegetation and the Siluor-Devonian rise of fluvial lateral accretion sets. *Geology*, **38**, 51-54.

Dickson, J.L., Head, J.W., Levy, J.S. and **Marchant, D.R.** (2013) Don Juan Pond, Antarctica: Near-surface $CaCl_2$-brine feeding Earth's most saline lake and implications for Mars. *Sci. Rep.* **3,** 1166. DOI: 10.1038/srep01166.

Domack, E.W. and **Hoffman, P.** (2013) An ice grounding-line wedge from the Ghaub glaciations (635 Ma) on the distal foreslope of the Otavi carbonate platform, Namibia, and its bearing on the snowball Earth hypothesis. *Geol. Soc. Am. Bull.*, **123,** 1448-1477.

Doran, P.T., McKay, C.P., Fountain, A.G., Nylen, T., McKnight, D.M., Jaros, C. and **Barrett, J.E.** (2008) Hydrologic response to extreme warm and cold summers in the McMurdo Dry Valleys, East Antarctica. *Antarct. Sci.*, **20**, 499-509.

Dowdeswell, J.A., Hambrey, M.J. and **Wu, R.T**. (1985) Clast shape and fabric in Precambrian and modern glacial sediments. *J. Sedim. Petrol.* **55**, 691-704.

Dutkiewicz, A., Herczeg, A.L. and **Dighton, J.C.** (2000) Past changes to isotopic and solute balances in a continental playa: clues from stable isotopes of lacustrine carbonates. *Chem. Geol.*, **165**, 309-329.

Elsig, J., Schmitt, J., Leuenberger, D., Schneider, R., Eyer, M., Leuenberger, M., Joos, F., Fischer, H. and **Stocker, T.F.** (2009) Stable isotope constraints on Holocene carbon cycle changes from an Antarctic ice core. *Nature*, **461**, 507-510.

Etienne, J.L., Allen, P.A., Rieu, R. and **Le Guerroué, E.** (2007) Neoproterozoic glaciated basins: a critical review of the Snowball Earth hypothesis by comparison with Phanerozoic glaciations. In: Hambrey, M.J., Christoffersen, P., Glasser, N.F. & Hubbard, B. (eds.) Glacial Processes and Products. *Special Publication of the International Association of Sedimentologist*s, doi: 10.1002/9781444304435.ch19.

Eyles, N. and **Januszczak, N. (**2004). 'Zipper-rift': a tectonic model for Neoproterozoic glaciations during the breakup of Rodinia after 750 Ma. *Earth-Sci. Rev.,* **65**, 1-73.

Fairchild, I.J. (1980) Sedimentation and origin of a late Precambrian 'dolomite' from Scotland. *Journal of Sedimentary Petrology*, 40, 423-446.

Fairchild, I.J. (1983) Effects of glacial transport and neomorphism on Precambrian dolomite crystal sizes. *Nature*, **304**, 714-716.

Fairchild, I. J. (1991) Origins of carbonate in Neoproterozoic stromatolites and the identification of modern analogues. *Precambrian Research*, **53,** 281-299.

Fairchild, I.J. (1993) Balmy shores and icy wastes: the paradox of carbonates associated with glacial deposits in Neoproterozoic times. In Wright, V.P. (ed.) *Sedimentology Rev.*, **1**, 1-16.

Fairchild, I.J. and **Hambrey, M.J.** 1984. The Vendian succession of northeastern Spitsbergen – petrogenesis of a dolomite-tillite association. *Precamb. Res.*, **26**, 111-167.

Fairchild, I.J. and **Spiro, B.** (1987) Petrological and isotopic implications of some contrasting Precambrian carbonates, NE Spitsbergen. *Sedimentology*, **34**, 973-989.

Fairchild, I.J. and Herrington, P.M. (1989) A tempestite-evaporite-stromatolite association (late Vendian, East Greenland): a shoreface-lagoon model. *Precambrian Research*, **43,** 101-127

Fairchild, I.J. and **Hambrey, M.J.** (1995) Vendian basin evolution in East Greenland and NE Svalbard. *Precambrian Res.,* **73**, 217-233.

Fairchild, I.J. and **Kennedy, M.J.** (2007) Neoproterozoic glaciation in the Earth System. *J. Geol. Soc, London,* **164**, 895-921.

Fairchild, I.J. and **Baker, A.** (2012) *Speleothem Science – from Process to Past Environment.* Wiley, Chichester.

Fairchild, I.J., Hambrey, M.J., Spiro, B. & Jefferson, T.H. (1989) Late Proterozoic glacial carbonates in northeast Spitsbergen: new insights into the carbonate-tillite association. *Geol. Mag.,* **126**, 469-490.

Fairchild, I.J., Marshall, J.D. & Bertrand-Sarfati, J. (1990) Stratigraphic shifts in carbon isotopes from Proterozoic stromatolitic carbonates (Mauritania) – influences of primary mineralogy and diagenesis. Am. J. Sci., **290A**, 46-79.

Fairchild, I.J., Bradby, L., Sharp, M. and **Tison, J.-L.** (1994) Hydrochemistry of carbonate terrains in Alpine glacial settings. *Earth Surf. Proc. Land.*, **19**, 33-54.

Fairchild, I.J., Killawee, J.A. Hubbard, B. and **Dreybrodt, W.** (1999) Interactions of calcareous suspended sediment with glacial meltwater: a field test of dissolution behaviour. *Chem. Geol.,* **155**, 243-263.

Fairchild, I.J., Baker, A., Borsato, A., Frisia, S., Hinton, R.W., McDermott, F. and **Tooth, A.F.** (2001) High-resolution, multiple-trace-element variation in speleothems. *Journal of the Geological Society*, *London*, **158**, 831-841.

Fleming, E.J. (2014) *Magnetic, structural and sedimentological analysis of glacial sediments: insights from modern, Quaternary and Neoproterozoic environments.* University of Birmingham, unpubl. PhD. thesis, http://etheses.bham.ac.uk/5136/

Fleming, E.J., Stevenson, C.T.E., Petronis, M.S., Benn, D.I., Hambrey, M.J. and **Fairchild, I.J.** (2016) Criteria for the recognition of subglacial processes during a Neoproterozoic panglaciation, NE Svalbard" *In revision for this issue, Sedimentology*

Fountain, A.G., Lyons, W.B., Burkins, M.B., Dana, G.L., Doran, P.T., Lewis K.J., McKnight, D.M., Moorhead, D.L., Parsons, A.N., Priscu, J.C., Wall, D.H., Wharton, R.A. and **Virginia, R.A.** (1999) Physical controls on the Taylor valley ecosystem. *Bioscience*, **49,** 961-971.

Frank, T.D., Thomas, S.G. and **Fielding, C.R.** (2008) On using carbon and oxygen isotope data from glendonites as paleoenvironmental proxies: A case study from the Permian System of eastern Australia. *J. Sedim Res,* **78**, 713-723.

Gooseff, M.N., Lyons, W.B., McKnight, D.M., Vaughn, B.H., Fountain, A.G. and **Dowling, C.** (2006) A stable isotopic investigation of a polar desert hydrologic system, McMurdo Dry Valleys, Antarctica. *Arctic, Antarctic, Alpine Res.,* **38**, 60-71.

Green, W.J. and **Lyons, W.B.** (2009) The saline lakes of the McMurdo Dry Valleys, Antarctica. *Aquat. Geochem.* **15**, 321-348.

Gregg, J.M., Howard, S.A. and **Mazzullo, S.J.** (1992) Early diagenetic recrystallization of Holocene (<3000 years old) peritidal dolomites, Ambergris Cay, Belize**.** *Sedimentology,* **39**, 143-160.

Grotzinger, J.P. and **Knoll, A.H.** (1999) Stromatolites in Precambrian carbonates: Evolutionary mileposts or environmental dipsticks? *Ann. Rev. Earth Planet. Sci.*, **27**, 313-358.

Hall, B.L., Hendy, C.H. and **Denton, G.H.** (2006) Lake-ice conveyor deposits: Geomorphology, sedimentology, and importance in reconstructing the glacial history of the Dry Valleys. *Geomorphology*, **75**, 143-156.

Hall, B.L., Denton, G.H., Sonte, J.O. and **Conway, H.** (2013) History of grounded ice sheet in the Ross Sea sector of Antarctica during the Last Glacial Maximum and the last termination. *In:* Hambrey, M.J., Barker, P.F., Barrett, P.J., Howman, V., Davies, B., Smellie, J.L. and Tranter, M. (eds.) Antarctic Palaeoenvironments and Earth-Surface Processes. Geol. Soc. London Spec. Publ., 381, 167-181.

Halley, R.B. (1977) Ooid fabric and fracture in the Great Salt Lake and the geologic record. *J. Sedim. Petrol.*, **47**, 1099-1120.

Halverson, G.P. (2006) A Neoproterozoic Chronology IN: Xiao, S. & Kaufman, A.J. (Eds.) Neoproterozoic Geobiology and Paleobiology, Springer, New York, 231-271.

Halverson, G.P. (2011) Glacial sediments and associated strata of the Polarisbreen Group, northeastern Svalbard. In: Arnaud, E., Halverson, G. P. and Shields-Zhou, G. (eds) (2011) The Geological Record of Neoproterozoic glaciations. *Geol. Soc. London Mem.,* **36**, 571-579.

Halverson, G.P., Hoffman, P.F., Schrag, D.P. and **Kaufman, A.J.** (2002) A major perturbation of the carbon cycle before the Ghuab glaciation (Neoproterozoic) in Namibia: Prelude to snowball Earth? *Geochem. Geophys. Geosyst.*, **3**, 1035. doi: 10.1029/2001GC000244.

Halverson, G.P., Maloof, A.C. and **Hoffman, P.F.** (2004) The Marinoan glaciation (Neoproterozoic) in northeast Svalbard. *Basin Res.,* **16**, 297-324.

Halverson, G.P., Hoffman, P.F., Schrag, D.P., Maloof, A.C. and **Rice, A.H.N.** (2005) Towards a Neoproterozoic composite carbon-isotope record. *Geol. Soc. Am. Bull.*, **117**, 1181-1207.

Halverson, G.P., Dudás, F.Ö., Maloof, A.C. and **Bowring, S.A.** (2007) Evolution of the $^{87}Sr/^{86}Sr$ composition of Neoproterozoic seawater. *Palaeogeog. Palaeoclimatol. Palaeoecol.*, **256**, 103-129.

Halverson, G.P., Wade, B.P., Hurtgen, M.T. and **Barovich, K.M.** (2010) Neoproterozoic chemostratigraphy. *Precamb. Res.*, **182,** 337-350

Hambrey, M.J. (1982) Late Precambrian diamictites of northeastern Svalbard. *Geol. Mag.*, **119**, 527-551.

Hambrey, M. J. (1983) Correlation of Late Proterozoic tillites in the North Atlantic region and Europe. *Geol. Mag.* **120**, 209-232.

Hambrey, M.J. and **Spencer, A.M.** (1987) Late Precambrian glaciation of central East Greenland. *Medd. Grønland Geosci.* **19.**

Hambrey, M.J. and **Fitzsimons, S.J. (**2010) Development of sediment-landform associations at cold glacier margins, Dry Valleys, Antarctica. *Sedimentology*, 57, 857-882**.**

Hambrey, M.J. and **Glasser, N.** (2012) Discriminating glacier thermal and dynamic regimes in the sedimentary record. *Sedim. Geol.*, **251**, 1-33.

Harland, W.B. (1964) Critical evidence for a great Infra-Cambrian glaciation. Geol. Rundsch., **54**, 45-61.

Harland, W.B. (1997) *The Geology of Svalbard*. *Geol. Soc. London Mem.,* **17**.

Harland, W. B., Hambrey, M. J. and **Waddams, P.** (1993) Vendian Geology of Svalbard*. Norsk. Polarinst. Skr.* **193**, 150 pp.

Hendy, C.H. (1979) Late Pleistocene glacial chronology of the Taylor Valley, Antarctica, and the global climate. *Quat. Res.*, **11**, 72-84.

Hendy, C.H. (1980) Late Quaternary lakes in the McMurdo Sound region of Antarctica. *Geografiska Annal.*, **82A**, 411-432.

Hendy, C.H. Sadler, A.J., Denton, G.H. and **Hall, B.L.** (2000) Proglacial lake-ice conveyors: a new mechanism for the deposition of drift in polar environments. *Geogr. Annal.* **82A**, 249-274.

Hoffman, P.F. (2011) Strange bedfellows: glacial diamictite and cap carbonate from the Marinoan (635 Ma) glaciations in Namibia. *Sedimentology*, **58**, 57-119.

Hoffman, P.F. and **Schrag,** D.P. (2002) The snowball Earth hypothesis: testing the limits of global change. *Terra Nova,* **14**, 129-155.

Hoffman, P.F., Kaufman, A.J., Halverson, G.P. and **Schrag, D.P.** (1998) A Neoproterozoic Snowball Earth. *Science*, **281**, 1342-1346.

Hoffman, P.F., Halverson, G.P., Domack, E.W., Husson, J.M., Higgins, J.A. and **Schrag, D.P.** (2007) Are basl Ediacarn (635 Ma) post-glacial "cap dolostones" diachronous? *Earth Planet. Sci. Lett.*, **258**, 114-131.

Hoffman, P.F., Halverson, G.P., Domack, E.W., Maloof, A.C., Swanson-Hysell, N.L. and **Cox, G.M.** (2012) Cryogenian glaciations on the southern tropical paleomargin of Laurentia (NE Svalbard and East Greenland), and a primary origin for the upper Russøya (Islay) carbon isotope excursion. *Precambrian Res.*, **206-207**, 137-158.

Hood, A.V.S. and **Wallace, M.W.** (2012) Synsedimentary diagenesis in a Cryogenian reef complex: Ubiquitous marine dolomite precipitation. *Sedimentary Geology*, **255-256,** 56-71.

Hood, A.V.S. and **Wallace, M.W.** (2014) Marine cements reveal the structure of an anoxic, ferruginous Neoproterozoic ocean. *J. Geol. Soc. London*, **171**, 741-744.

Huggett, J.M., Schultz, B.P., Shearman, D.J. and **Smith, A.J.** (2005) The petrology of ikaite pseudomorphs and their diagenesis. *Proc. Geol. Assoc.*, **116,** 207-220.

Hyde, W.T., Crowley, T.J., Baum, S.K. and **Peltier, W.R.** (2000) Neoproterozoic 'snowball Earth' simulations with a coupled climate/ice-sheet model. *Nature*, **405**, 425-429.

Javoy, M., Pineau, F. and **Delorme, H.** (1986) Carbon and nitrogen isotopes in the mantle. *Chem. Geol.,* **57**, 41-62.

Kaczmarek, S.E. and **Sibley, D.F.** (2014) Direct physical evidence of dolomite recrystallization. *Sedimentology*, **61**, 1862-1882.

Kalugin, I., Darin, A., Rogozin, D. and **Tretyakov, G.** (2013) Seasonal and centennial cycles of carbonate mineralisation during the past 2500 years from varved sediment in Lake Shira, South Siberia. *Quat. Int.*, **290-291**, 245-252.

Kawano, J., Shimobayashi, N., Miyake, A. and **Kitamura, M.** (2009) Precipitation diagram of calcium carbonate polymorphs: its construction and significance. *J. Phys. Condens. Matter*, **21**, doi:10.1088/0953-8984/21/42/425102.

Kelts, K. and **Hsü, K.J.** (1978) Freshwater carbonate sedimentation. In: Lerman, A. (Ed.), *Lakes. Chemistry, Geology, Physics.* Springer Verlag, New York, pp. 295-323.

Kennedy, M.J., Christie-Blick, N. and **Sohl, L.E.** (2001a) Are Proterozoic cap carbonates and isotopic excursions a record of gas hydrate destabilization following Earth's coldest intervals? *Geology*, **29**, 443-446.

Kennedy, M.J., Christie-Blick, N. and **Prave, A.R.** (2001b) Carbon isotopic composition of Neoproterozoic glacial carbonates as a test of paleoceanographic models for snowball Earth phenomena. *Geology*, **29**, 1135-1138.

Kennedy, M., Droser, M., Mayer, L.M., Pevear, D. and **Mrofka, D.** (2006) Late Precambrian oxygenation: Inception of the clay mineral factory. *Science*, **311**, 1446-1449.

Kennedy, M.J., Mrofka, D. and **von der Borch, C.C.** (2008) Snowball earth termination by destabilization of equatorial permafrost methane clathrate. *Nature*, **453**, 642-645.

Kennedy, M.J. and **Christie-Blick, N.** (2011) Condensation origin for Neoproterozoic cap carbonates during deglaciation. *Geology*, **39**, 319-322.

Killingsworth, B.A., Hayles, J.A., Zhou, C. and **Bao, H**. (2013) Sedimentary constraints on the duration of the Marinoan Oxygen-17 depletion (MOSD) event. *Proc. Natl. Acad. Sci.,* **110**, 17686-17690.

Kim, S.T. and **O'Neil, J.R. (**1997) Equilibrium and nonequilibrium oxygen isotope effects in synthetic carbonates. *Geochim. Comochim. Acta*, **61**, 3461-3475.

Kirschvink, J.L. (1992) Late Proterozoic Low-Latitude Global Glaciation: the Snowball Earth. In: Schopf, J.W., Klein, C. (Eds.), *The Proterozoic Biosphere: A Multidisciplinary Study.* Cambridge University Press, pp. 51-52.

Klappa, C.F. (1979) Lichen stromatolites: Criteria for subaerial exposure and a mechanism for the formation of laminar calcretes (caliche). *J. Sedim. Petrol.*, **49**, 387-400.

Knoll, A.H. and **Swett, K.** (1990) Carbonate deposition during the Late Proterozoic Era: an example from Spitsbergen. *Am. J. Sci.*, **290-A**, 104-132.

Knoll, A.H., Hayes, J.M., Kaufman, A.J., Swett, K. and **Lambert, I.B.** (1986) Secular variation in carbon isotope ratios from Upper Proterozoic successions of Svalbard and East Greenland. *Nature*, **321**, 832-838.

Knoll, A.H., Fairchild, I.J. and **Swett, K.** (1993) Calcified microbes in Neoproterozoic carbonates – implications for our understanding of the Proterozoic Cambrian transition. *Palaios*, **8**, 512-525.

Kulling, O. (1934) Scientific Results of the Swedish--Norwegian Arctic Expedition in the summer of 1931. XI. 'The Hecla Hoek Formation' around Hinlopenstretet. *Geogr. Annlr. Stockh.,* **16,** 161-254.

Lan, Z, Liu, X., Zhu, M., Chen, Z., Zhang, Q., Li, Q., Lu, D., Liu, Y. and **Tang G.** (2014) A rapid and synchronous initiation of the widespread Cryogenianglaciations. *Precamb. Res.,* http://dx.doi.org/10.1016/j.precamres.2014.10.015

Lancaster, N. and **Teller, J.L.** (1988) Interdune deposits of the Namib Sand Sea. *Sediment. Geol.* **55**, 91-107.

Land, L.S. (1980) The isotopic and trace element geochemistry of dolomite: the state of the art. In (D.H. Zenger, J.B. Dunham & R.L. Ethington eds.), Concepts and Models of Dolomitization, *SEPM Spec. Publ.,***28**, 87-110.

Larsen, D. (1994) Origin and paleoenvironmental significance of calcite pseudomorphs after ikaite in the Oligocene Creede Formation, Colorado. *J. Sed. Res.*, **A64**, 593-603.

Last, F.M., Last, W.M., Fayek, M., Halden, N.M. (2013) Occurrence and significance of a cold-water carbonate pseudomorph in microbialites from a saline lake. *J. Paleolimnol.*, **50**, 505-517.

Lauterbach, S., Baruer, A., Andersen, N., Danielpol, D.L., Dulski, P., Hüls, M., Milecka, K., Namiotko, T., Obremska, M., von Grafenstein, U. and **Declakes Participants.** (2011) Environmental responses to Lateglacial climatic fluctuations recorded in the sediments of pre-Alpine Lake Mondsee (northeastern Alps). *J. Quat. Sci.*, **26,** 253-267.

Lawrence, M.J.F. and **Hendy, C.H.** (1985) Water column and sediment characteristics of Lake Fryxell, Taylor Valley, Antarctica. *NZ J. Geol. Geophys.*, **28**, 543-552.

Le Heron, D.P., Cox, G., Trundley, A. and **Collins, A.** (2011) Sea ice-free conditions during the Sturtian glaciations (early Cryogenian), South Australia. *Geology*, **39**, 31-34.

Le Heron, D.P., Busfield, M. And **Kamona, F.** (2013) An interglacial on Snowball Earth? Dynamic ice behaviour revealed in the Chuos Formation, Namibia. *Sedimentology*, **60**, 411-427.

Le Hir, G., **Ramstein, G., Donnadieu, Y.** and **Goddéris, Y.** (2008) Scenario for the evolution of atmospheric pCO_2 during a snowball Earth. *Geology*, **36**, 47-50.

Le Hir, G., Donnadieu, Y., Goddéris, Y., Pierrehumbert, R.T., Halverson, G.P., Macouin, M., Nédélec, A. and **Ramstein, G.** (2009) The Snowball Earth aftermath: Exploring the limits of continental weathering processes. *Eath Planet. Sci. Lett.*, **277**, 453-463.

Leng, M.J. and **Marshall, J.D.** (2004) Palaeoclimate interpretation of stable isotope data from lake sediment archives. *Quat. Sci. Rev.*, **23**, 811-831.

Lewis, A.R., Marchant, D.R., Ashworth, A.C., Hemming, S.R. and **Machlus, M.L.** (2007) Major middle Miocene global climate change: Evidence from East Antarctica and the Transantarctic Mountains. *Geol. Soc. Am. Bull.,* **119**, 1449-1461.

Li, Z.-X., Evans, D.A.D. and **Halverson, G.P.** (2013) Neoproterozoic glaciations in a revised global palaeogeography from the breakup of Rodinia to the assembly of Gondwanaland. *Sed. Geol.,* **294**, 219-232.

Lloyd, R.M. (1966) Oxygen isotope enrichment of sea water by evaporation. *Geochimica et Cosmochimica Acta,* 30, 801-814.

Lyons, W.B., Priscu, J.C., Laybourn-Parry, J., Moorhead, D., McKnight, D.M., Doran, P.T. and **Tranter, M.** (2001) The McMurdo Dry Valleys long-term ecological research program: new understanding of the biogeochemistry of the Dry Valley lakes: a review. *Polar Geog.*, **25**, 202-217.

Lyons, W.B., Leslie, D.L., Harmon, R.S., Neumann, K., Welch, K.A., Bisson, K.M. and **McKnight, D.M.** (2013) The carbon stable isotope biogeochemistry of streams, Taylor Valley, Antarctica. *Appl. Geochem.*, **32**, 26-36.

Macdonald, F.A., Schmitz, M.D., Crowley, J.L., Roots, C.F., Jones, D.S., Maloof, A.C., Strauss, J.V., Cohen, P.A., Johnston, D.T. and **Schrag, D.P.** (2010) Calibrating the Cryogenian. *Science*, **327**, 1241-1243.

Marchant, D.R. and Head, J.W. (2007) Antarctic dry valleys: Microclimate zonation, variable geomorphic processes, and implications for assessing climate change on Mars. *Icarus*, 192, 187-222.

Marchant, D.R., Mackay, S.L., Lamp, J.L., Hayden, A.T. and **Head, J.W.** (2013) A review of geomorphic processes and landforms in the Dry Valleys of southern Victoria Land: implications for evaluating climate change and ice-sheet stability. *In:* Hambrey, M.J., Barker, P.F., Barrett, P.J., Howman, V., Davies, B., Smellie, J.L. and Tranter, M. (eds.) Antarctic Palaeoenvironments and Earth-Surface Pro.cesses. *Geol. Soc. London Spec. Publ.***,** **381**, 319-352.

Matsubaya, O., Sakai, H., Torii, T., Burton, H. and **Kerry, K.** (1979) Antarctic saline lakes -stable isotope ratios, chemical compositions and evolution. *Geochim. Cosmochim. Acta,* **43**, 7-25.

McKelvey, B. (1981) DVDP

McKnight, D.M., Niyogi, D.K., Alger, A.S., Bomblies, A., Conovitz, P.A. and **Tate, C.M.** (1999) Dry Valley streams in Antarctica: Ecosystems waiting for water. *Bioscience*, 49, 985-995.

McKnight, D.M., Tate, C.M., Andrews, E.D., Niyogi, D.K., Cozzetto, K., Welch, K., Lyons, W.B. and **Capoine, D.G.** (2007) Reactivation of a cryptobiontic stream ecosystem in the McMurdo Dry Valleys, Antarctica: A long-term geomorphological experiment. *Geomorphology*, **89**, 186-204.

Moncrieff, A.C.M. and **Hambrey, M.J.** (1990) Marginal-marine glacial sedimentation in the late Precambrian succession of east Greenland. In: Dowdeswell, J.A. and Scourse, J.D. (eds.) Glacimarine Environments: Processes and Sediments. Geologicl Society Special Publication, **53**, 387-410.

Mook, W., Bommerson, J. and **Staverman, W.** (1974) Carbon isotope fractionation between dissolved bicarbonate and gaseous carbon dioxide. *Earth Planet. Sci. Lett*. **22,** 169–176.

Mountney, N.P. and **Thompson, D.B.** (2002) Stratigraphic evolution and preservation of Aeolian dune and damp/wet interdune strata: an example from the Triassic Helsby Sandstone Formation, Cheshire Basin, UK. *Sedimentology*, **49**, 805-833.

Müller, G., Irion, G. and **Förstner, U.** (1972) Formation and diagenesis of inorganic Ca-Mg carbonates in the lacustrine environment. *Naturwissenschaften*, **59**, 158-164.

Nakai, N., Wada, H., Kiyosu, Y. and **Takimoto, M.** (1975) Stable isotope studies on the origin and geological history of water and salts in the Lake Vanda area, Antarctica. *Geochem. J.*,, **9**, 7-24.

Neugebauer, I., Brauer, A., Dräger, N., Dulski, P., Wulf, S., Plessen, B., Mingram, J., Herzschuh, U. and **Brande, A.** (2012) A Younger Dryas varve chronology from the Rehweise palaeolake record in NE-Germany. *Quat. Sci. Rev.*, **36,** 91-102.

Neumann, K., Lyons, W.B., Priscu, J.C., Desmarais, D.J. and **Welch, K.A.** (2004) The carbon isotopic composition of dissolved inorganic carbon in perennially ice-covered Antarctic lakes: searching for a biogenic signature. *Annals Glac.*, **39**, 518-524.

Neuser, R.D. and **Richter, D.K.** (2007) Non-marine radiaxial fibrous calcites – examples of speleothems proved by electron backscatter diffraction. *Sediment. Geol.*, **194**, 149-154.

Nomura, D.*,* **Assmy, P., Nehrke, G., Granskog, M.A., Fischer, M., Dieckmann, G.S., Fransson, A., Hu, Y.** and **Schnetger, B.** (2013) Characterization of ikaite ($CaCO_3 \cdot 6H_2O$) crystals in first-year Arctic sea ice north of Svalbard. *Ann. Glaciol.*, **54,** 125-131.

Nylen, T.H., Fountain, A.G. and **Doran, P.T.** (2004) Climatology of katabatic winds in the McMurdo dry valleys, southern Victoria Land, Antarctica. *J. Geophys. Res. Atm.* **109**, D03114. doi: 10.1029/2003JD003937.

Oehlerich, M., Mayr, C., Griesshaber, E., Lücke, A., Oeckler, O.M., Ohlendorf, C., Schmahl, W.W. and **Zolitschka, B.** (2013) Ikaite precipitation in a lacustrine environment – implications for palaeoclimatic studies using carbonates from Laguna Potrok AIke (Patagonia, Argentina). *Quat. Sci. Rev.*, **71**, 46-53.

Parker, B.C., Simmons, G.M., Love, F.G., Wharton, R.A. and **Seaburg, K.G. (**1981) Modern stromatolites in Antarctic Dry Valley lakes. *Bioscience*, **31**, 656-661.

Peng, Y., Bao, H., Zhou, C., Yuan, X. and **Luo, T.** (2013) Oxygen isotope composition of meltwater from a Neoproterozoic glaciation in South China. *Geology*, **41**, 367-370.

Pierrehumbert, R.T. (2005) Climate dynamics of a hard snowball Earth. *J. Geophys. Res.*, **110**, D01111, doi:1029/2004JD005162.

Pierrehumbert, R.T., Abbot, D.S., Voigt, A. and **Koll, D.** (2011) Climate of the Neoproterozoic. *Ann. Rev. Earth Planet. Sci.*, **39**, 417-460.

Pollock, M.D., Kah, L.C. and **Bartley, J.K.** (2006) Morphology of molar-tooth structures in Precambrian carbonates: Influence of substrate rheology and implications for genesis. *J. Sedim. Res.*, **76**, 310-323.

Powers, D.W. and **Easterling, R.G.** (1982) Improved methodology for using embedded Markov chains to describe cyclical sediments. *J. Sedim. Petrol.*, 52, 913-923.

Propenko, A.A. and **Williams, D.F.** (2005) Depleted methane-derived carbon in waters of Lake Baikal, Siberia. *Hydrobiologia*, **544,** 279-288.

Radke, B.M. and **Mathis, R.L.** (1980) On the formation and occurrence of saddle dolomite. *J. Sedim. Petrol.*, **50**, 1149-1168.

Rao, C.P., Goodwin, I.D. and **Gibson, J.A.E.** (1998) Shelf, coastal and subglacial polar carbonates, East Antarctica. *Carbonates and Evaporites*, 13, 174-188.

Riccioni, R.-M., Brock, P.W.G. and **Schreiber, B.C.** (1996) Evidence for early aragonite in paleo-lacustrine sediments. *J. Sedim Res.*, **66**, 1003-1010.

Riding, R. (2000) Microbial carbonates: the geological record of calcified bacterial-algal mats and biofilms. *Sedimentology*, **47** (suppl. 1), 179-214.

Robinson, Z.P., Fairchild, I.J. and **Arrowsmith, C.** (2009) Stable isotope tracers of shallow groundwater recharge dynamics and mixing within an Icelandic sandur, Skeiðarársandur. Hydrology in Mountain Regions: Observations, Processes and Dynamics (Proceedings of Symposium HS1003 at IUGG2007, Perugia, July 2007). I*nt. Assoc. Hydrol. Sci. Publ.* **326**, 119-125.

Rodriguez-Blanco, J.D., Shaw, S., Bots, P., Roncal-Herrero, T. and **Benning, L.G.** (2014) The role of Mg in the crystallization of monohydrocalcite. *Geochim. Cosmochim. Acta*, **127**, 204-220.

Rooney, A.D., Macdonald, F.A., Strauss, J.V., Dudás, F.O., Hallmann, C. and **Selby, D.** (2014) Re-Os geochronology and coupled Os-Sr isotope constraints on the Sturtian snowball Earth. *Proc. Natl. Acad. Sci.*, **111**, 51-56.

Rooney, A.D., Strauss, J.V., Brandon, A.D. and **Macdonald, F.A.** (2015) A Cryogenian chronology: Two long-lasting synchronous Neoproterozoic glaciations. *Geology*, 43, 459-462.

Schrag, D.P., Berner, R.A., Hoffman, P.F. and **Halverson, G.P.** (2002) On the initiation of a snowball Earth. *Geochem. Geophys. Geosyst.* **3**, 1036. doi: 10.1029/2001GC000219.

Sekkal, W. and **Zaoui, A.** (2013) Nanoscale analysis of the morphology and surface stability of calcium carbonate polymorphs. *Sci. Rep.*, **3**, 1587. Doi: 10.1038/srep01587.

Selleck, B.W., Carr, P.F. and **Jones, B.G.** (2007) A review and synthesis of glendonites (pseudomorphs after ikaite) with new data: assessing applicability as recorders of ancient coldwater conditions. *J. Sedim. Res.*, **77**, 980-991.

Shanahan, T.M., Overpeck, J.T., Beck, J.W., Wheeler, C.W., Peck, J.A., King, J.W. and **Scholz, C.A.** (2008) The formation of biogeochemical laminations in Lake Bosumtwi, Ghana, and their usefulness as indicators of past environmental changes. *J. Paleolimnol.*, **40**, 339-355.

Shearman, D.J. and **Smith, A.J.** (1985) Ikaite, the parent mineral of jarrowite-type pseudomorphs. *Proc. Geol. Assoc.*,96, 305-314.

Shearman, D.J., McGugan, A., Stein, C. and **Smith, A.J.** (1989) Ikaite, $CaCO_3 \cdot H_2O$, precursor of the thinolites in the Quaternary tufas, and tufa mounds of the Lahontan and Mono Lake Basins, western United States. *Geol. Soc. Am. Bull.,* **101**, 913-917.

Simmons, G.M., Vestal, J.R. and **Wharton, R.A.** (1993) Environmental regulators of microbial activity in continental Antarctic lakes. In: W. J. Green and E. I. Friedmann (eds.), Physical and Biogeochemical Processes in Antarctic Lakes *Am. Geophys. Union Antarct. Res. Ser.*, **59**, 165-195.

Speirs, J.C., McGowan, H., Steinhoff, D.F. and **Bromwich, D.H.** (2013) Regional climate variability driven by foehn winds in the McMurdo Dry Valleys, Antarctica. *Int. J. Climatol.*, **33**, 945-958.

Spencer, A.M. (1971) Late Pre-Cambrian glaciation in Scotland. *Mem. Geol. Soc. Lond.* **6,** 5-102.

Spötl, C. and **Wright, V.P.** (1992) Groundwater dolocretes from the Upper Triassic of the Paris Basin, France: a case study of an arid, continental diagenetic facies. *Sedimentology*, **39**, 1119-1136.

Stockhecke, M., Anselmetti, F.S., Meydan, A.F., Odermatt, D. and **Sturm, M.** (2012) The annual particle cycle in Lake Van (Turkey). *Palaeogeog. Palaeoclimatol. Palaeocol.*, **333-334**, 148-159.

Sugden, D.E., Marchant, D.R., Potter, N., Souchez, R.A., Donton, G.H., Swisher, C.C. and **Tison, J.-L.** (1995) Preservation of Miocene glacier ice in East Antarctica. *Nature*, **376**, 412-414.

Swainson, I.P. and **Hammond, R.P.** (2001) Ikaite, $CaCO_3 \cdot 6H_2O$: Cold comfort for glendonites as paleothermometers. *Am. Mineral.*, **86**, 1530-1533.

Talbot, M.R. (1990) A review of the palaeohydrological interpretation of carbon and oxygen isotopic ratios in primary lacustrine carbonates. *Che. Geol. (Isot. Geosci.)*, **80,** 261-279.

Tandon, S.K. and **Friend, P.F.** (1989) Near-surface shrinkage and carbonate replacement processes, Arran Cornstone Formation, Scotland.*Sedimentology*, **36,** 1113-1126.

Trindade, R.I.F., Font, E., D'Agrella, M.S., Nogueira, A.C.R. and **Riccomini, C.** (2003) Low-latitude and multiple geomagnetic reversals in the Neoproterozoic Puga cap carbonate, Amazon craton. *Terra Nova*, **15**, 441-446.

Vasconcelos, C., McKenzie, J.A., Warthmann, R. and **Bernasconi, S.M.** (2005) Calibration of the δ^{18}O paleothermometer for dolomite precipitated in microbial cultures and natural environments. *Geology*, **33**, 317-320.

Virtasolo, J.J., Leipe, T., Moros, M. and **Kotilainen, A.T.** (2011) Physicochemical and biological influences on sedimentary-fabric formation in a salinity and oxygen-restricted semi-enclosed sea: Gotland Deep, Baltic Sea. *Sedimentology*, **58**, 352-375.

Voorhuis, A.A., Biddanda, B.A., Kendall, S.T., Jain, S., Marcus, D.N., Nold, S.D., Sheldon, N.D. and **Dick, G.J.** (2012) Cyanobacterial life at low O_2: community genomics and function reveal metabolic versatility and extremely low diversity in a Great Lakes sinkhole mat. *Geobiology*, **10**, 250-267.

Warren, J. (1989) Evaporite Sedimentology. Prentice-Hall, New Jersey.

Warren, J. (2000) Dolomite: occurrence, evolution and economically important association. *Earth-Sci. Rev.*, **52**, 1-81.

Wharton, R.A., Parker, B.C., Simmons, G.M., Seaburg, K.G. and **Love, F.G.** (1982) Biogenic calcite structures forming in Lake Fryxell, Antarctica. *Nature*, **293**, 293-295.

Williams, G.E. (1979) Sedimentology, stable-isotope geochemistry and palaeoenvironment of dolostones capping late Precambrian glacial sequences in Australia. *J. Geol. Soc. Aust.*, **26**, 377-386.

Williams, G.E. (2008) Proterozoic (pre-Ediacaran) glaciation and the high obliquity, low-latitude ice, strong seasonality (HOLIST) hypothesis: Principles and test. *Earth-Sci. Rev.*, **87**, 61-93.

Wilson, A.T. (1981) A review of the geochemistry and lake physics of the Antarctic dry areas. In (ed. L.D. McGinnis). Dry Valley Drilling Project pp. 185-192. Washington: Am. Geophys. Union.

Wilson, C.B. and **Harland, W.B.** (1964) The Polarisbreen Series and other evidences of late Pre-Cambrian ice ages. *Geol. Mag.,* **101**, 198-219.

Wright, V.P. (1990) A micromorphological classification of fossil and recent calcic and petrocalcic microstructures. In: Douglas, L.A. (Ed.) *Soil Micromorphology: A Basic and Applied Science*. Elsevier, Amsterdam, pp. 401– 407.

Wright, V.P. and Tucker, M.E. (2009) Calcretes: an introduction. In Wright, V.P. & Tucker, M.E. (eds.) *Calcretes. Int. Assoc. Sedimentol. Reprint Series,* 1-24.

Xiao, S., Bao, H., Wang, H., Kaufman, A.J., Zhou, C., Li, G., Yuan, X. and **Ling, H.** (2004) The Neoproterozoic Quruqtagh Group in eastern Chinese Tianshan: evidence for a post-Marinoan glaciation. *Precamb. Res.* **130**, 1-26.

Zhang, F., **Xu, H., Konishi, H., Kemp, J.M., Roden, E.E.** and **Shen, Z.** (2012) Dissolved sulphide-catalyzed precipitation of disordered dolomite: Implications for the formation mechanism of sedimentary dolomite. *Geochimi. Cosmochim. Acta*, **97**, 148-165.

Zhao, Y.Y. and **Zheng, Y.F.** (2010) Stable isotope evidence for involvement of deglacial meltwater in Ediacaran carbonates in South China. *Chem. Geol.*, **271**, 86-100.

Table 1: Neoproterozoic chronostratigraphy and NE Svalbard lithostratigraphy (after Halverson et al., 2007, Halverson, 2011, updated by unpublished data). Glacial units highlighted in red. This paper deals with the Wilsonbreen Formation, representing the younger of the two Cryogenian glaciations.

Geological System	Group	Formation	Member	Thick-ness (m)	Lithologies	Interpreted environment
Ediacaran	Polarisbreen	Dracoisen	D4 to D7	265	Sandstones and mudstones (D4 and D6); dolomite D5 and D7	Playas (D4 to D6); Coastal (D7)
			D1 to D3	200	Cap carbonate (D1) transitional (D2) to black shale D3)	Transgressive coastal (D1) to offshore (D2-D3)
Cryogenian		**Wilsonbreen**	W3 (Gropbreen)	65-95	Diamictites and sandstones with minor limestone	Glacilacustrine; subglacial at base
			W2 (Middle Carbonate)	20-30	Three main intervals of carbonate-bearing sandstones and siltstones with intervening diamictites	Carbonate intervals are fluvial and lacustrine with glacial influence; diamictites are glacilacustrine rainout deposits
			W1 (Ormen)	55-85	Brecciated underlying dolomite locally overlain by sandstones and conglomerates passing up into diamictites and sandstones with local rhythmites	Basal periglaciated surface, locally succeeded by fluvial deposits, then glacilacustrine rainout deposits and sediment gravity flows
		Elbobreen	E4 (Slangen)	20-30	Oolitic dolomite	Regressive peritidal
			E3 (Macdonaldryggen)	200	Finely laminated dolomitic silty shale	Offshore marine
			E2 (Petrovbreen)	10-20	Dolomitic diamictites, rhythmites and conglomerates	Glacimarine
(base to be defined)			E1 (Russøya)	75-170	Dolomites overlain by limestone with molar tooth structure, black shale and dolomite	Shallow marine
ian	niker- en	Backlund-toppen		530	Limestone and dolomite	Carbonate platform
		Draken		350	Intraclastic dolomite	Peritidal

Table 2. Summary of characteristics of carbonates in members W1 and W3. Analyses are of calcite except where italicized (*dolomite*). s= standard deviation; n = number of samples.

				$\delta^{18}O$		$\delta^{13}C$		
Section	**Member**, m above member base (from member top)	**Facies association**	**Lithology**	mean	s	mean	s	n
AND	**W3** 30 (-15)	5	Intraclastic stromatolitic rhythmites (carbonate layers 5-10 mm thick) with intervening diamictite. Some calcite-filled pseudomorphs	-9.22	0.20	0.27	0.33	5
AND	**W3** 38 (-7)	5	Stromatolitic limestone (laminae 5-10 mm thick) with diamictite laminae, variably broken up within diamictite. Observed over 100 m laterally.	-8.16	0.55	0.23	0.70	6
AND	**W3** 40.5 (-4.5)	5	Brecciated limestone	-	-	-	-	-
REIN	**W3** 56 (-6.5)	5	Limestone with scattered sand	-8.92	-	-0.44	-	1
KLO	**W3** 25 (-18)	5	Carbonates interlaminated with diamictite at base of bed. EFK15 closely resembles the middle carbonate layer at section AND.	-7.97 *-5.03*	- -	1.83 *1.04*	- -	1 *1*
KLO	**W3** 28 (-15)	5	Laminated carbonates in diamictite. Some crystal pseudomorphs both within sediment and growing upwards.	-8.86 *-4.04*	0.4 -	-0.38 *1.16*	0.1 -	3 *1*
KLO	**W3** 31 (-12)	5	Laminated carbonates in ice-rafted sediments with prominent upward growing crystal pseudomorphs (in dolomite) of ikaite.	-9.44	2.38	0.31	1.09	4
KLO	**W3** 35 (-8)	5	Similar to horizon 4 m lower in section.	-	-	-	-	-
PIN	**W3** n/a (-26)	5	Slump folded and brecciated thickly laminated stromatolitic limestone in red silty sandy diamictites. Traceable laterally for 30 m	-12.53	0.28	-0.04	0.29	3
ORM	**W1** 3.5 (-20.5)	5	Slump-folded rhythmites with distinct mm-scale stromatolitic limestone laminae, some with bulbous tops, separated by diamictite.	-12.5	-	0.32	-	2
ORM	**W3** 28 (-30)	3	Sandstone with convolute bedding with nodular dolocrete (floating quartz, micro-nodules and cracks) along specific laminae.	*-9.78*	-	*-5.60*	-	*2*
ORM	**W3** 33 (-25)	4	Diamictite rests on massive stromatolitic limestone with pseudomorphs and slump folds or intraclasts over greenish cross-laminated sandstone with ooids.	-10.76	1.12	2.20	1.56	6
ORM	**W3** 46.5 (-11.5)	5	Red, thickly laminated and slump-folded stromatolitic limestone in diamictite. Locally massive with lot of early cement.	-9.44	2.38	0.31	1.09	4

Table 3. Facies and facies associations of the Wilsonbreen Formation. The facies ass numbered to represent an environmental continuum as depicted in Fig. 7A.

Facies Association	Constituent Facies	Environmental In
1: Deformed Diamictite Detailed facies analysis in Fleming et al. (2016)	Dominated by massive diamictites, locally with deformed stratification with lenticular sand and gravel bodies (1-5 m thick) with disrupted margins. Locally associated with upwards-increasing shear of underlying sediment and striated boulder pavements.	Subglacial till and chann glacitectonites.
2: Fluvial Channel	2S**:** Beds of 0.5-4 m very fine to medium-grained, locally cross-stratified sandstone with erosional conglomeratic base. Some variants have low-angle accretion surfaces with silt-dominated beds. Stromatolitic limestone is either absent or abundant, forming dm packages with mm-cm-scale laminae separated by sand laminae and laterally eroded into trains of intraclasts. 2T: As above, but lacking carbonate precipitates	Ephemeral stream chann strongly seasonal flows. is a widespread carpet o mats.
3: Dolomitic Floodplain	All facies contain dolomite with a positive δ^{18}O composition that actively cements, displaces and/or upwardly accretes the sediment. 3D: Nodules, cm- to dm-scale of dolomite-cemented silts and sands transitional to structureless dm- to m-scale beds with floating silt and sand in dolomite, internal nodules and calcite-cemented fractures 3S: Stromatolitic dololaminites, dm-scale and mm-laminated	Dolocretes representing typically above fluvial de raised water table 3N: Nodular dolocrete 3D: Bedded dolocrete 3S: Subaerial stromatoli
4: Calcareous Lake Margin	4R: Rhythmites with mm-scale dolomite, limestone or mixed mineralogy laminae, commonly desiccated, alternating with 1-10 mm wavy cross-laminated silty sandstones. Carbonate laminae are often stromatolitic. Almost invariably reddened. 4I: Intraclastic, dm-scale, very fine- to medium-grained sandstones. Very locally contain ooids. 4S: Indistinctly horizontally stratified well-sorted fine- to medium-grained sandstone, with well-rounded grains, locally with cm-scale dolomite laminae.	3R: Shallow lake to playa wave-reworked sedimen microbial mat accretion. 3I: Discrete storm horizo lake/playa reworking sa carbonate 3S: Aeolian sandflat dep
5: Carbonate Lake	5R: Rhythmites with mm-scale limestone, dolomite or mixed mineralogy laminae, usually with stromatolitic microstructure and locally building dm stromatolites with cm-scale relief. Alternate with 0.1-5 mm laminae of diamictite/wacke, commonly with till pellets and occasional lonestones. Slump folding and brecciation common. 5D: Discrete gravel to diamictite units with significant intraclastic debris in addition to terrigenous gravel.	5R: Carbonate microbial lacustrine environment ice-rafting and slope ins 5D: Slope-related resedi
6: Glacilacustrine Detailed facies analysis in Fleming et al. (2016)	Dominated by massive and stratified diamictites with common lonestones and till pellets. Decimetre- to m-scale intervals of silty rhythmites occur locally, especially in member W1. May contain lenses of conglomerate, sometimes channelled, and lenses or thin beds of sandstone forming packages which can be inclined at up to 20°.	Ice-rafted glacilacustrine reworked as sediment-g Inclined packages form (termed proximal glacila and rhythmites likewise retreat.
7: Periglacial See Fairchild & Hambrey (1984) and Benn et al. (2015)	Decimetre-wide sandstone wedges penetrating up to 2 m into underlying sediment from a discrete surface that may have a gravel lag. At base of Wilsonbreen Formation, gravel with ventifacts overlying shattered dolostone.	Exposed periglaciated b wedges from exposed p environment.

Figure Captions

Fig. 1. Reconstruction of the sedimentary architecture and palaeoenvironments of the Wilsonbreen Formation and its constituent members (W1 to W3). Precipitated carbonate is present in the palaeoenvironmental group called "Carbonate Lacustrine and Fluvial" throughout W2 (except locally at the top and base), also in W3 and at one of the locations in W1 as listed in Table 2. The Svalbard archipelago is shown bottom right and Spitsbergen is the main island of the group, whilst Nordaustlandet is the island to the NE. The rectangle on the Svalbard map shows the study area as enlarged upper right with Wilsonbreen Formation outcrops (red) within nunataks (grey) rising from the highland snowfield. From north to south, study locations are: DRA (Dracoisen); here the main section is located on a nunatak informally known as Multikolorfjellet with some additional sampling from W2 at a second nunatak we term Tophatten 1 km to the north; DIT (Ditlovtoppen); AND (East Andromedafjellet); REIN (a ridge on South Andromedafjellet informally known as Reinsryggen); KLO (South Klofjellet); with some additional observations from a partial W2 section 1 km away, at North Klofjellet; McD (MacDonaldryggen); GOL (Golitsynfjellet, intermediate between McD and BAC) – a partial W2 section was illustrated by Fairchild et al. (1989); BAC (Backlundtoppen-Kvitfjellet ridge); PIN (an unnamed nunatak informally termed Pinnsvinryggen); SLA (Southeast Slangen) and ORM (South Ormen).

Fig. 2. Examples of studied section outcrops. **A.** Member W2 at Multikolorfjellet, Dracoisen (DRA) illustrating the three groups of glacial retreat facies beds (numbered 1 to 3) separated by diamictites which also make up members W1 and W3. **B.** Member W2 at ORM (South Ormen) with pale sand-dominated units and dark red finer units. Bedding is inverted and dips steeply away from the photographer. **C.** REIN (Reinsryggen) section on the south flank of Andromedafjellet

illustrating Wilsonbreen Formation members overlain by cap carbonate member D1. **D.** KLO (South Klofjellet) section of the entire Wilsonbreen Formation on steep slopes cut by minor faults, one of which is highlighted. **E.** BAC (Backlundtoppen-Kvitfjellet ridge) photomontage from helicopter hovering above the glacier Wilsonbreen. The visible section is vertical (thrust fault shown) and the accessible part defines a narrow ridge between the cliff and a snowbank on the ridge crest.

Fig. 3. A. Oblique aerial view of the McMurdo Dry Valleys (1/1/1999 imagery from US Geological Survey via Google Earth) with location arrowed in inset of Antarctica (upper right). EAIS = East Antarctic Ice Sheet. LGM = Last Glacial maximum. Abbreviations on upper right inset: EA = East Antarctica, WA =West Antarctica, RIS = Ross Ice Shelf. The lower left inset shows an oblique aerial photograph looking west up Taylor Valley with cold-based valley glaciers on hillside on left, ice-covered Lake Bonney (valley floor, right) with the tip of the Taylor Glacier beyond. **B.** Despite sub-zero temperatures, runoff occurs from the Lower Wright Glacier (an outlet glacier of the coastal Wilson Piedmont Glacier) and feeds the Onyx River. This stream flows inland to the west, eventually to Lake Vanda, visible in **A.** in the central part of the valley. Note the aeolian sands banked agains the glacier. **C.** Oblique aerial view of Victoria Lower Glacier (E end of Victoria Valley) which feeds a stream flowing inland, westwards to Lake Vida, seen in the distance. Aeolian dunes tranverse to the stream are visible on right-side of the the valley.

Fig. 4. Profile of member W2 at Dracoisen. The lithological log emphasizes physical characteristics (see key, upper right) whilst the assignment to facies associations (centre) also draws on petrological and stable isotope information. FA1 to FA6 represent an environmental continuum (cf. Fig. 6) and FA7 is grouped with FA1 as the terrestrial glacial end-member. Each datapoint

represents the stratigraphic position of a studied sample or observed lithological boundary in the field. The oxygen isotope composition of precipitated calcites and dolomites (the latter adjusted by -3 ‰) are shown. Mixed-mineralogy samples (<90% calcite or dolomite in calcite-dolomite mixtures) are not plotted. The same conventions are used for profiles of the other studied sections, which are presented as supporting figures (Figs. S1-S6).

Fig. 5. Detrital textures and minerals. **A.** and **B.** Paired images of a polished thin section under CL and in transmitted light respectively of the matrix of a silty sandstone (W2, 81.2 m, Dracoisen) illustrating the abundance of faint blue-luminescing feldspar and varied fragments of both bright and dull red-luminescing dolomite in the mud fraction. **C.** Thin section, transmitted light. Siltstone graded rhythmites (W3, 60.4 m, East Andromedafjellet) displaying several sand-sized till pellets. **D.** Crossed polars. Matrix at the top of a graded silt layer illustrating quartzo-feldspathic debris and micron-sized dolomite, but no clay minerals (W1, 10 m Slangen).

Fig. 6. Summary cartoon of facies associations. Facies Associations 2 to 5 are colour-coded here and in isotope plots. The margin of an ice sheet is depicted, terminating partly on land and, in the foreground, in the lake. FA1 is shown in a subglacial environment (subglacial sediments are coloured brown); periglacial phenomena are also grouped in this facies association. FA2 and FA3 both occur in a fluvial setting (sediments coloured yellow), whilst FA5 and FA6 were deposited in lakes (sediments coloured light brown). FA4 includes both shallow lacustrine and coastal sediments.

Fig. 7. A. Summary of stable isotope compositions of precipitated carbonate, together with the mean composition of dolomitic detritus. Dolomite and calcite groupings are separated and only

samples with >90% of either dolomite or calcite in the carbonate fraction are plotted. **B.** Stable isotope fields illustrating degree of covariance of the sample groups.

Fig. 8. Facies association 2 (Fluvial Channel). **A**- **F** are Facies 2S (W2, Dracoisen, at around the 60 m level), whereas **G** and **H** are facies 2T. **A.** Sandstone with microbial laminites and low-domed stromatolites variably broken into intraclasts. **B.** Stained thin section in transmitted light of stromatolite microstructure with micrite (M), microspar (S), fenestral (F) and detrital (D) laminae. **C.** Stained thin section in transmitted light showing the broken edges of stromatolitic intraclasts with radiaxial calcite cement crusts in a calcareous sandstone matrix. **D.** Cross-stratified sandstone (set 30 cm high) with stromatolite intraclasts overlain by current ripple forms. **E.** Polished rock slice with arrow denoting micromilled traverse shown in **F**. **F.** Micromill isotope traverse across two micrite/microspar lamina and a central zone of calcite spar which has a much lower isotope signature. **G.** Sandstone body (with pebbly base) showing accretionary surfaces (e.g. dashed line). Palaeohorizontal shown by solid black line. (W2, Ditlovtoppen, 119 m). **H.** Photomontage of tabular sandstone unit of FA2 with a pebble horizon near its top (arrowed). It rests erosively on floodplain (FA3) silts and is overlain by red lake margin (FA4) sediments; ruler is 25 cm long (W2, Dracoisen, 83 m).

Fig. 9. Photomicrographs of stromatolitic limestones from FA2. **A.** Paired transmitted light (left)and CL (right) micrographs. Stromatolitic laminae of orange-luminescing microspar (MS) with subhedral authigenic quartz and clastic layer (M) including quartz and feldspar grains (the latter luminesces dark blue). Large fenestra is filled by radiaxial calcite (R) seen in both transverse and basal sections and displaying brighter earlier growth and duller later growth. These fabrics are cut by a vein (V) filled with bright to dull luminescing calcite. W2, Dracoisen, 58.5 m. **B.** Stained thin

section, transmitted light. Alternating micrite, microspar and clastic laminae with prominent irregular vertical “filamentous” structure of clear calcite. W2, Dracoisen, 58.5 m.

Fig. 10. FA3 (Dolomitic floodplain). Facies 3D is shown in **A-D**, Facies 3S in the others, and both facies in **E**. All of the Facies 3S images come from W2, Dracoisen, 70 m. **A.** Nodular dolocrete with calcite-lined vugs in siltstone with scale in mm (W2, East Andromedafjellet, 35 m). **B.** Stained thin section in plane polarized light showing matrix-supported fabric of dolomite cementation of sandy siltstone. Dolomicrospar lines a fenestra which is occluded by calcite. W2, Dracoisen, 70 m. **C.** Thin section in plane polarized light. Grain-supported dolomicrite cement of silty sandstone. The dolomite has a $\delta^{18}O$ composition of +2.7 ‰ and has a uniform texture in contrast to clastic dolomite of Fig. 4D. W2, South Klofjellet, 57 m. **D.** Displacive dolomite cement supporting silt and sand grains. Well-developed structure of dark nodules which show different CL characteristics from surrounding dolomite from which they are separated by curved cracks. W3, South Ormen, 78 m. **E.** Stained thin section of interlaminated dolomite-cemented sand and microbial laminae with fenestrae, some occluded by ferroan dolomite (turquoise arrow) or ferroan calcite (purple arrow). **F.** Stained thin section illustrating similar fabric to (**D.**), but with calcite cementation of cracks and larger pores (W2, Dracoisen, 83 m) **G.** Field photograph of textured bedding surface of dololaminite identified as microbial mat texture (W2, Dracoisen, 70 m). **H.** Polished rock chip of microbial dololaminites with arrow marking position of 5.2 mm micromill traverse (shown in **J.** below). **I**. Microbial laminites draping downwards into underlying laminate whose brecciation is attributed to evaporite dissolution collapse. Outlined ruler is 20 cm long. **J.** Stable isotope profile (in ‰ with respect to V-SMOW) of microbial dololaminites along line illustrated in **H**. The isotopes covary over a magnitude of 6 ‰ for $\delta^{18}O$ and 1 ‰ for $\delta^{13}C$.

Fig. 11. FA3 (dolomitic floodplain). **A.** Transmitted light, stained thin section. Sandy dolocrete (FA3) containing equant nodule cemented by ferroan saddle dolomite (turquoise), with local late calcite (red), interpreted as a fill of a small anhydrite nodule. W2, Backlundtoppen-Kvitfjellet ridge, 74.7 m. **B.** Facies 3S stromatolite with fenestrae. Paired transmitted light (left) and CL (right) images. Brightly luminescing dolomite may be primary or an early replacement of a precursor. W2, Dracoisen 69.95 m. **C.** Paired transmitted light (left) and CL (right) images. Dolocrete showing very fine-grained quartz sand grains floating in dolo(micro-)spar with crystals displaying a common zonation of bright to dull CL. Displacive primary dolomite growth is the preferred interpretation. W2, Ditlovtoppen, 118.5 m. **D.** Paired transmitted light (left) and CL (right) images. Dolocrete, similar to Fig. 10D, F, with sparse floating quartz and feldspar (black and blue respectively in CL) and calcite-filled cracks(centre) and pores (base). Uniformly luminescing dolomicrite crystals, differ in brightness within nodules presumably forming at different stages. Calcite-filled pores show CL zonation (base) or no CL (cracks, centre). W2, Dracoisen, 82.9 m.

Fig. 12. Stable isotope plot, differentiating facies within FA3 and (in purple and larger symbols) samples from member W3.

Fig. 13. FA4 (Calcareous Lake Margin). Images **A**, **B**, **D** and **H**-**J** are Facies 4R; **C**, **E** and **G** are Facies 4I and **F** is Facies 4S. **A.** Laminated rhythmic limestones and silty sandstones with conspicuous isolated wave ripple structure in centre of view. W2, Dracoisen, 90 m. **B.** Sand-rich example of facies 4R with cross-laminated silty sands and dolomitic rhythmites, in part desiccated or eroded to form intraclasts. White areas near top are mineral (probable salt) pseudomorphs. W2, Ditlovtoppen, 109 m. **C.** Wave ripples with 15 cm wavelength on bed top W2, Dracoisen (Tophatten), approximately equivalent to the 87 m level on Fig. 4. **D.** Carbonate rhythmite surface

cut by desiccation cracks and bearing salt pseudomorphs. W2, Reinsryggen, 82.5 m. **E.** Stained thin section, plane polarized light of oolitic intraclastic sandstone. Ooids are bimineralic (calcite and dolomite) with dominant concentric structure. W3, South Ormen, 84 m. **F.** Part of scanned thin section in transmitted light. Well-sorted quartzose sandstone with very well-rounded grains and low-angle lamination marked by subordinate interstitial dolomite. W2, Ditlovtoppen, 114.2 m. **G.** Stained thin section in transmitted light. Sandstone with conspicuous stromatolitic rhythmite limestone intraclasts. W2, Backlundtoppen-Kvitfjellet ridge, 76.5 m (supporting Fig. 6). **H.** Two-metre high section at of Ditlovtoppen (108.5-110.5 on suppl. Fig. 1) showing poorly stratified sandstone bed (Facies 4S) overlying Facies 4R with several discrete graded intraclastic sandstone beds (Facies 4I). **I.** Polished slab of dolomitic rhythmites with desiccation cracks (C). The 3.2 mm-long micromill isotope traverse of **J** is indicated. W2, South Ormen, 31.3 m. **J.** Isotope results from the micromill traverse with lighter isotope values corresponding to detrital dolomite matrix and the heaviest values (at right) indicative of precipitated dolomite composition.

Fig. 14. FA4 illustrating contrast between detrital and replacive dolomite. **A and B.** Paired transmitted light and CL images respectively. Enlargement of the boundary between a dolomicrite lamina (below) and a detrital lamina (top) of the same sample as in Fig. 13I, J. The detrital layer shows quartz (black in CL), feldspar (blue), a dolomite clast (bright ring) whilst the dolomicrite shows a consistent zonation of crystals with a bright core as well as some fine silt-sized siliciclastic debris. The dolomicrite is interpreted as an early diagenetic replacement of an early carbonate phase. W2, South Ormen, 31.3 m.

Fig. 15. FA 5 (Calcareous Lake) in member W2. All illustrate Facies 5R, but transitions to Facies 5D are shown in **D** and **E**. **A.** Thin section in transmitted light. Calcareous rhythmites with subordinate

clastic sediment including till pellets (e.g. yellow arrows). White areas, 1-2 mm across, are mineral pseudomorphs. Reinsryggen, 81 m. **B.** Calcareous rhythmites with irregular lamina tops indicative of microbial structure and growth domes above vuggy areas with syn-depositional calcite cements. Note scale in mm. Reinsryggen, 80.5 m. **C.** Sawn and polished hand specimen illustrating a growth fault across which the stratigraphy of stromatolitic rhythmite layers changes. Note scale in cm. East Andromedafjellet, 30.3 m – note this is a thin rhythmite occurrence within diamictite (supporting Fig. 1). **D.** Stained thin section in transmitted light. Dolomite dropstone (d) deforms underlying limestone rhythmite and lies at the base of a coarser resedimented layer including limestone intraclasts. Ditlovtoppen, 108 m. **E.** Lower half of a discrete 40 cm diamictite debris flow unit with rhythmites deformed by slumping at the base. Numerals 1 cm apart on tape, lower right corner. Ditlovtoppen, 109 m (wedging out over 100 m to the section shown in supporting Fig. 1). **F.** Stromatolitic rhythmites, alternately pure white and impure sediment-bearing limestone. Backlundtoppen-Kvitfjellet ridge, 77 m. Location of micromill traverse of **G** illustrated. **G.** Micromill traverse as in **F** illustrating a systematic shift in δ^{18}O, but within a relatively narrow range of 1 ‰, similar to range of uncorrelated variation in δ^{13}C.

Fig. 16. FA5 (calcareous lake) in members W3 and W1. All show Facies 5R with various transitions to Facies 5D. **A.** Diamictites of faces association 6 becoming interlaminated (yellow arrows) with limestone rhythmites. Slump folds in upper left. Lens cap for scale. W3, South Klofjellet, 116.5 m. **B.** Limestone rhytmites developing slump folds upwards and transitioning to an intraclastic diamictite. W3, South Klofjellet, 116.5 m. **C.** Polished hand specimen illustrating erosional truncation of stromatolitic rhythmite laminae by diamictite with prominent cm-scale pebbles. W3, South Klofjellet, 120 m. **D.** Thin section in transmitted light of partly brecciated rhythmite with laminae up to cm scale with interstitial ice-rafted sediment. Equant white areas a few mm across

are mineral pseudomorphs. W3, East Andromedafjellet, 63 m. **E.** Slump-folded 1-3 mm limestone rhythmite layers with interstitial green ice-rafted sediment. Scale is in cm. This is the only precipitated carbonate horizon in W1 (South Ormen, 3.5 m).

Fig. 17. Stromatolitic fabrics in FA5. **A.** Transmitted light. Millimetre-scale calcite laminites separated by thinner ice-rafted laminae and locally containing till pellets (P). Calcite laminites display peloidal clots and local fenestrae and have variably bulbous tops. W2, Reinsryggen, 79.8 m. **B.** Transmitted light. Similar horizon to **C.** displaying clastic lamina overlain by clotted and fenestral microbial lamina with bulbous top. W2, East Andromedafjellet, 13.5 m. **C.** Stained thin section, transmitted light. Dolomitic microbial laminate containing dolomicrite and dolomicrospar laminae and floating detritus (white). Conspicuous fenestrae are filled by pink-stained calcite. W2, South Ormen, 31.4 m. **D.** Transmitted light. W2, Reinsryggen, 79.8 m. Faintly clotted (peloidal) micrite (examples arrowed) and microspar laminae with intervening calcite-filled fenestra. Dark patches are micro-till pellets (some are labelled P), now partly silicified.

Fig. 18. Crystal pseudomorphs. **B-G** are all FA5. **A.** Permian glendonite from South Australia: ikaite pseudmorphs that original grew in glacimarine mudrocks. Pencil for scale. Sample provide by Malcolm Wallace. **B.** Transmitted light view of interlaminated diamictites and laminated limestones, microbial in part, and displaying two distinct horizons of upward-growing crystals. Crystals influence subsequent sedimentation pattern and hence grew into the water column. W3, South Klofjellet, 116.5 m. **C.** Histogram of apparent interfacial angles from thin sections of samples shown in **B.** Modes are most consistent with an ikaite precursor (see text) **D.** Transmitted light, stained thin section of same sample as B. Pseudomorphs are composite of mosaics of zoned, mostly ferroan (bluish) calcite crystals. Outer edges of crystals have in part been dissolved and in

part silicified. W3, South Klofjellet, 116.5 m. **E.** Transmitted light. Calcite-cemented crystal pseudomorphs, morphology consistent with ikaite, in stromatolitic limestone . W3, South Ormen, 85 m. **F.** Transmitted light. Indistinct calcite-cemented probably ikaite pseudomorphs, morphology probably consistent with ikaite, in stromatolitic limestone. W2, Reinsryggen, 81.2 m. **G.** Transmitted light. Dolorhythmites hosting dolomite pseudomorphs, inferred to be after ikaite, with varying micrite-microspar-spar replacive textures. W2, North Klofjellet, 67 m.

Fig. 19. Wilsbonbreen crystal pseudomorphs (**A**) compared with ikaite crystals (**C**) and pseudomorphs of different ages and contexts (**B, D, E**). **A.** Polished hand specimen (same sample as Fig. 16B. Note small pink pebble to left in diamictite layer overlying top crystal layer. Crystals grew upwards at three horizons and were draped by overlying sediments before being replaced by calcite as illustrated in Figs. 18D and 20A. W3, South Klofjellet, 116.5 m. **B.** Examples of "thinolite" crystals from Dana (1884), interpreted as ikaite pseudomorphs by Shearman et al. (1989). Scale not given in the original, but crystals are typically cm-dm-scale. **C.** Profiles of ikaite crystals recovered from Arctic sea ice by partial melting (Nomura et al., 2013). **D.** and **E.** ikaite pseudomorphs from a Patagonian lake (Oeherlich et al., 2013). **D,** illustrates crystals with equant habit and stepped faces as seen by scanning electron microscopy. **E.** illustrates pseudomorphs attached to moss filaments.

Fig. 20. Calcite fabrics of FA 5. **A.** Transmitted light, stained thin section. Enlargement of the ikaite pseudomorphs of Fig. 15A showing relic crystal outlines in dark micrite and replacive calcite (micro-)spar mosaic of zoned euhedral non-ferroan calcite overgrowth by ferroan calcite. W3, South Klofjellet, 116.5 m. **B.** Paired transmitted light (left) and CL (right) images. Fenestral microbial fabric similar to Fig. 17D displaying consistent crystal zonation: brighter to duller in

micritic matrix and sharper zones within fenestral cements. Fabric is consistent either with primary calcite growth within extracellular polymeric substance or replacement of ikaite, followed by cementation of fenestrae. W2, Reinsryggen, 80.6 m.

Fig. 21. Vertical (upward) facies transitions in member W2. For this purpose FA1 and FA7 are conflated. The width of the arrows is proportional to the number of transitions minus one. The transition matrix shown in supporting Fig. 8 indicates that at least one vertical transition occurs between each of the facies associations except FA1. See text for discussion.

Fig. 22. Relative thickness of strata belonging to the different facies associations in the five sections with comparably thick carbonate facies preserved in W2, from south to north. See text for discussion and abbreviations in Fig. 1.

Fig. 23. Diagrammatic summary of the processes responsible for variation in isotope composition of the Wilsobreen carbonate and, interpretations of their parageneses with (at base) a cartoon environmental profile. Large numbers refer to Facies Associations 2 to 5.

N.B. Supporting Information is supplied as a free-standing pdf and an Excel document; supporting figure captions are not repeated here.