Chapter 55

Glacial sediments and associated strata of the Polarisbreen Group, northeastern Svalbard

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Abstract: Northeastern Svalbard hosts exceptionally well-preserved Neoproterozoic sediments. The glaciogenic Petrovbreen Member and Wilsonbreen Formation (Fm.) of the Polarisbreen Group crop out in a narrow, Caledonian-aged fold-and-thrust belt spanning from Olav V Land on Spitsbergen in the south to western Nordaustlandet in the north. The older Petrovbreen Member is thin (0–52 m) and patchily preserved, comprising mainly poorly stratified, dolomite-matrix diamictite likely deposited in a marine setting. The basal contact of the Petrovbreen Member erosionally truncates the upper Russøya Member, which preserves a large (12‰) negative C-isotope anomaly. The Petrovbreen Member is overlain by 200 m of dark, monotonous shales of the MacDonaldryggen Member, followed by cherty dolomite grainstone and microbiolamintes of the Slangen Member. The upper Slangen Member is an exposure surface in the southern part of the belt, but in northern Spitsbergen and on Nordaustlandet is transitional into sands of the northward-thickening Bråvika Member. The Wilsonbreen Fm. is typically 100–150 m thick and consists mostly of massively to poorly stratified diamictite, with subordinate sand beds, conglomerate lenses, and carbonates deposited in a terrestrial environment. It is overlain by the colourful Dracoisen Fm., which records, at its base, a typical post-glacial negative $\delta^{13}C$ anomaly. There are no direct radiometric age constraints or reliable palaeomagnetic data from the Polarisbreen Group, but it is widely accepted that northeastern Svalbard was contiguous with East Greenland during the Neoproterozoic.

Late Neoproterozoic glaciogenic rocks occur in both western and northeastern Svalbard (Harland *et al.* 1993). The much better studied and preserved glaciogenic sediments and bracketing strata of the upper Hecla Hoek succession in northeastern Svalbard are the subject of this chapter. These rocks occur in Olav V Land and Ny Friesland in northeastern Spitsbergen and in Gustav V Land in northwestern Nordaustlandet (Fig. 55.1). These rocks are best exposed on nunataks in Olav V Land and southern Ny Friesland, but occur also over a large area in Nordaustlandet along the eastern side of Hinlopenstretet, and to a lesser extent, in coastal exposures in northern Ny Friesland. In Spitsbergen, two distinct glacial units, named the Petrovbreen Member (older) and Wilsonbreen Fm. (younger) occur within the middle of the mixed clastic-carbonate Polarisbreen Group (Fig. 55.2). Both the Petrovbreen Member and the Wilsonbreen Fm., together with bracketing strata, are readily identified in Nordaustlandet (Halverson *et al.* 2004), where a separate nomenclature has historically been used (Kulling 1934). Because correlation across Hinlopenstretet (Fig. 55.1) is unambiguous (Kulling 1934; Harland *et al.* 1993; Fairchild & Hambrey 1995), only the better known nomenclature from Spitsbergen (Harland 1997) is used here (Fig. 55.2).

The history of geological investigation of Precambrian glacial deposits was recently summarized in Harland (2007). In short, the Neoproterozoic sedimentary succession in northeastern Svalbard was first studied by Nordenskiöld (1863). A glacial origin for the Polarisbreen diamictites (specifically the Wilsonbreen Fm.) was originally proposed by Kulling (1934), who mapped and named these rocks in Nordaustlandet and northern Ny Friesland during the Swedish–Norwegian Arctic Expedition of 1931. Fleming & Edmonds (1941) further investigated coastal exposures of the Hecla Hoek, including the Wilsonbreen Fm., but it was not until W. B. Harland and C. B. Wilson worked systematically on the upper Hecla Hoek succession on both coastal and inland sections on Spitsbergen (Harland & Wilson 1956; Wilson 1961; Wilson & Harland 1964) that specific attention was focused on the Neoproterozoic glacial rocks of Svalbard and their significance in Earth history (Harland 2007). Since this time, and with their global importance established (Harland 1964), several detailed sedimentological studies on the Polarisbreen glaciogenic sediments have been carried out (e.g. Chumakov 1968; Edwards 1976; Hambrey 1982; Fairchild 1983; Dowdeswell *et al.* 1985), with the results culminating in a monograph by Harland *et al.* (1993). Carbonates within and bounding the Polarisbreen diamictites have also been the focus of coupled sedimentological–geochemical investigations (Fairchild & Hambrey 1984; Fairchild & Spiro 1987; Halverson *et al.* 2004). More broadly, Neoproterozoic sediments in Svalbard and East Greenland were the subject of the seminal chemostratigraphic study by Knoll *et al.* (1986) that first firmly established the intimate association between negative C-isotope anomalies and Precambrian glaciation. The upper Hecla Hoek succession also hosts an unusually rich microfossil record, which has been studied in detail by A. H. Knoll, N. J. Butterfield and colleagues (e.g. Knoll 1982, 1984; Butterfield *et al.* 1988, 1994). The excellent preservation of these sediments and their role in the ongoing debate over the cause and severity of Neoproterozoic glaciations overcompensate for the logistical difficulty involved in studying them. This contribution is intended to update the review by Hambrey *et al.* (1981) in the context of more recent research on the Polarisbreen Group and significant advances in our understanding of the Neoproterozoic Earth system since that time.

Structural framework

The Svalbard archipelago, situated on the northwestern corner of the Barents Shelf, comprises three tectonic terranes that were amalgamated in the Silurian–Devonian Ny Friesland (Caledonian) orogeny (Harland & Gayer 1972; Harland *et al.* 1992; Gee & Page 1994; Lyberis & Manby 1999). The thick Hecla Hoek succession occurs in the northern part of the Eastern Terrane, which is the only exposed part of the Barentsia microcontinent (Breivik *et al.* 2002). The Polarisbreen Group is part of the relatively undeformed upper Hecla Hoek succession, which includes the Lomfjorden and Hinlopenstretet supergroups (Fig. 55.2). In Ny Friesland, the basal Lomfjorden Supergroup is juxtaposed against high-grade metasediments of the Stubendorfbreen Supergroup along the Eolussletta Shear Zone (Fig. 55.1; Lyberis & Manby 1999). The nature of this contact and the relationship between the lower and

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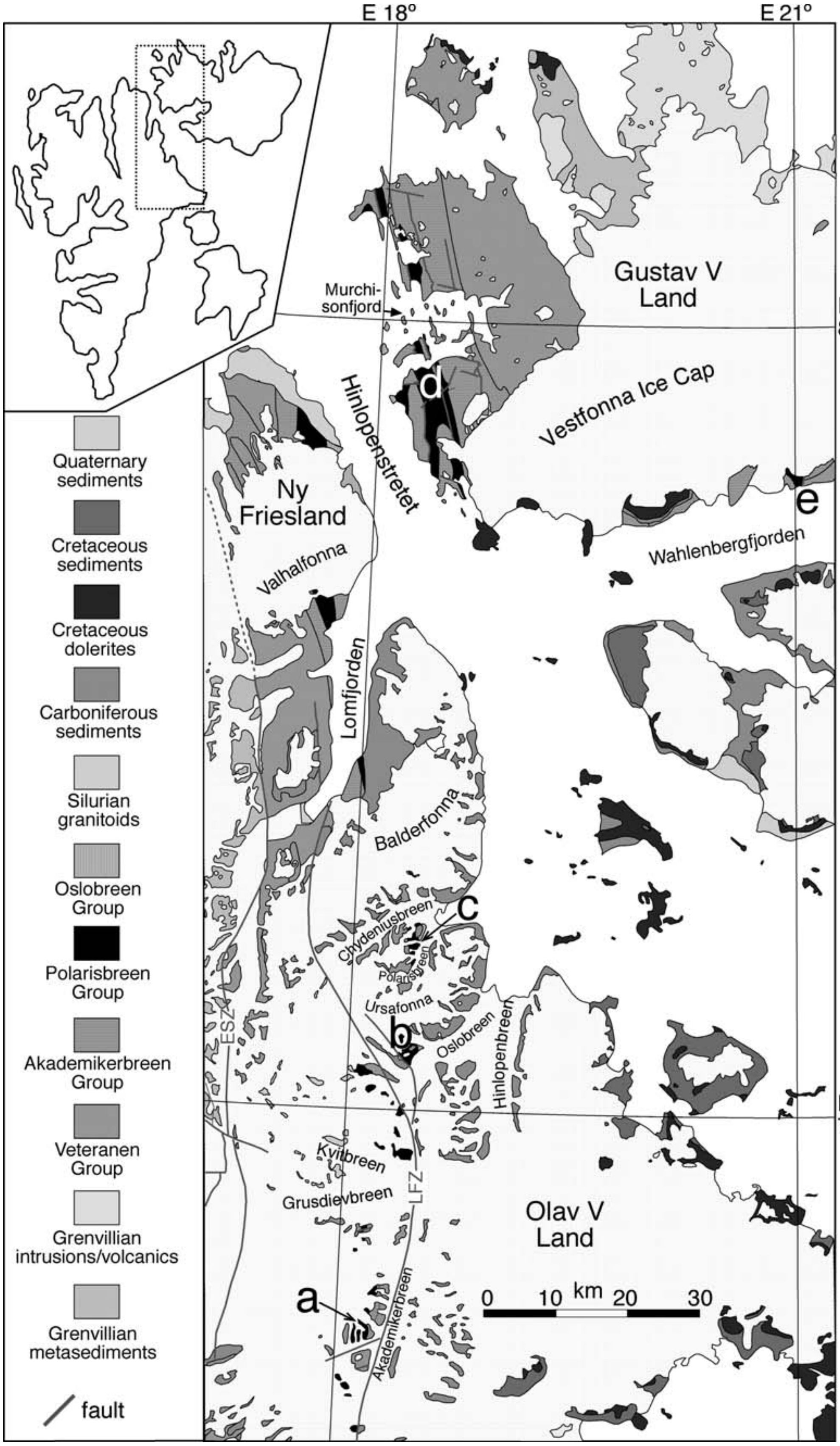

Fig. 55.1. Geological sketch map of the region in northeastern Svalbard where the Polarisbreen Group and equivalent rocks crop out. The mapping is from Dallmann *et al.* (2002), with modifications to the Neoproterozoic strata based on mapping by the author. Letters a to e refer to principal regions in which the Polarisbreen Group has been studied: a, Akademikerbreen area; b, Ditlovtoppen; c, Dracoisen/Oslobreen; d, Sveanor (south coast of Murchisonfjord), e, Aldousbreen (north coast of Wahlenbergfjord). ESZ, Eolusletta Shear Zone (cf. Lyberis & Manby 1999); LFZ, Lomfjorden Fault Zone.

upper Hecla Hoek has been the subject of a long-standing debate, with one side arguing that the two packages form a single, conformable stratigraphic sequence (e.g. Harland 1959; Harland *et al.* 1992). Most recent workers, however, agree that this contact represents a significant hiatus associated with Grenvillian orogenesis (e.g. Gee *et al.* 1995; Manby & Lyberis 1995; Witt-Nilsson *et al.* 1998), an argument that is supported by mapping and geochronology in Nordaustlandet where the Lomfjord Supergroup rests unconformably on early Neoproterozoic volcanics and subvolcanic intrusives (Fig. 55.2), which themselves intrude and overlie high-grade, late Mesoproterozoic metasediments (Gee *et al.* 1995; Johansson *et al.* 2005). The broad,

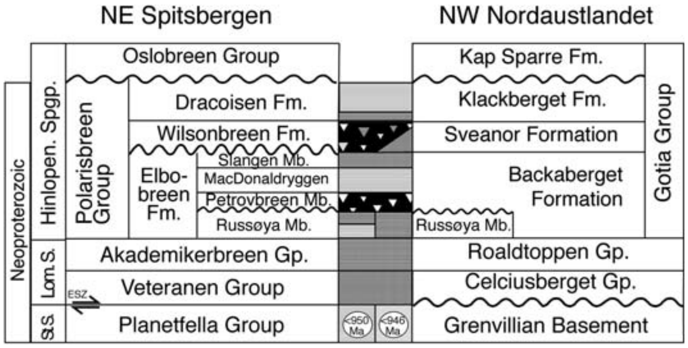

Fig. 55.2. Stratigraphic nomenclature and correlation of the Neoproterozoic successions in northeastern Spitsbergen and northwestern Nordaustlandet. Note that only the nomenclature from Spitsbergen is used in this chapter. Hinlopen. Spgp., Hinlopenstretet Supergroup; Lom. S., Lomfjorden Supergroup; St. S., Stubendorfbreen Supergroup. The 'upper Hecla Hoek succession', referred to in the text and many publications, is the Lomfjorden plus Hinlopenstretet supergroups.

conformable and aggradational nature of the upper Hecla Hoek and its continuity with the equivalent units in East Greenland (Fairchild & Hambrey 1995) imply that the glacial sediments of the Polarisbreen Group were deposited in a thermally subsiding basin (Maloof *et al.* 2006).

The upper Hecla Hoek crops out in a 120-km-long, north–south belt (Fig. 55.1) that was deformed during the Caledonian orogeny. The Polarisbreen Group sits in the core of the north-plunging Hinlopenstrettet Synclinorium, which spans the length of the belt (Fig. 55.1). The eastern limb of the synclinorium, exposed in Nordaustlandet, is a fold-and-thrust belt, dissected by a network of conjugate NW–SE and SW–NE strike–slip faults (Sokolov *et al.* 1968). The western limb of the synclinorium is less deformed, but is similarly cut by Caledonian-aged reverse and strike–slip faults. (Lyberis & Manby 1999). In the central part of the belt, the upper Hecla Hoek is intruded by Silurian granitoids (Teben'kov *et al.* 1996). The upper Hecla Hoek was eroded and draped by Carboniferous sediments during post-orogenic, west–east extension (Harland 1959), but normal faulting was largely concentrated along the prominent Lomfjorden Fault Zone (LFZ; Fig. 55.1). Dolerites related to the Late Jurassic–Early Cretaceous High Arctic Large Igneous Province (Maher 2001) intrude the Hecla Hoek within the northern part of the belt in the Hinlopenstretet region (Fig. 55.1). The Late Cretaceous–Early Tertiary West Spitsbergen Orogen, which heavily deformed the Western Terrane (Harland 1959), only lightly affected the Eastern Terrane, but did result in reactivation of major lineaments, such as the LFZ (Larsen 1987; Lyberis & Manby 1993).

Stratigraphy

The Polarisbreen Group (formerly Series) was originally defined by Wilson & Harland (1964) in the Dracoisen area in southern Ny Friesland, near the confluence of the Polarisbreen and Chydeniusbreen glaciers (Fig. 55.2). It conformably overlies the thick stack (2 km) of Akademikerbreen Group carbonates and broadly marks a transition from predominantly carbonate to mixed carbonate-clastic sedimentation. The contact between these two groups is easily identified throughout the outcrop belt, with the lithologically distinctive, pale yellow 'Dartboard Dolomite' (Wilson 1961; Wilson & Harland 1964) of the uppermost Akademikerbreen Group contrasting with the laminated black shale and bluish-grey to black limestone of the lower Polarisbreen Group (Halverson *et al.* 2004). The nature of this contact is variable. In some places, it is unambiguously transitional. Elsewhere, the upper Dartboard Dolomite is brecciated and silicified, and contains tepee-like structures, and the contact with the overlying limestone is sharp (Halverson *et al.* 2004). The top of the Polarisbreen Group is bound by the Cambrian Tokammane Fm. (base of the Oslobreen Group; Fig. 55.2). Although no erosional break at this contact is evident from mapping or outcrop exposures, a large depositional hiatus is implied by the biostratigraphy (Knoll & Swett 1987).

The Polarisbreen Group (Figs 55.2 & 55.3) is subdivided into three formations (Wilson & Harland 1964): Elbobreen, Wilsonbreen and Dracoisen. The Elbobreen Fm. is further subdivided into four regionally mappable members, each of which could be accorded formation status in its own right: (from bottom to top) the E1 or Russøya Member, the E2 or Petrovbreen Member, the E3 or MacDonaldryggen Member, and the E4 or Slangen Member (Fig. 55.2). A fifth unit that only occurs in the north of the outcrop belt, the Bråvika member, was tentatively assigned to the Wilsonbreen Fm. (Halverson *et al.* 2004), but is clearly transitional below with Slangen Member, and thus may more appropriately belong to the Elbobreen Fm. The Petrovbreen Member and Wilsonbreen Fm. comprise the two glaciogenic units, both of which are distinct from one another sedimentologically and within the context of their bounding strata (Figs 55.3 & 55.4).

Glaciogenic deposits and associated strata

Russøya Member (E1)

The Russøya Member varies from 75 to 170 m in thickness, thinning from south to north (Halverson *et al.* 2004). It is a lithologically variable unit, being predominantly carbonate in Nordaustlandet and containing an increasing fraction of sandstone and siltstone to the south. Despite its heterogeneity, the stratigraphic geometry of the Russøya Member is consistent, comprising a relatively thick (40–160 m), basal, shoaling-upward sequence, overlain by up to four complete parasequences, 5–20 m thick (Fig. 55.3). The lowest parasequence consists of a basal dark shale that grades upward into increasingly carbonate-rich sediments, including limestone ribbonites and grainstones. Molar tooth structures are prominent in this sequence, but occur exclusively within the ribbonite facies (Fig. 55.3).

The upper parasequences typically consist of shale at the base and sandstone or dolomitic grainstone, stromatolites or microbial laminites at the top. The uppermost parasequence is capped by a 2–6-m-thick biostrome of a spiralling, columnar stromatolite (*Kussiella*) and brecciated and variably truncated on an outcrop scale (Halverson *et al.* 2004). Although erosional truncation below the Petrovbreen Members is <2 m on the outcrop scale, regional correlations, assisted by chemostratigraphy (see below), suggest as much as 50 m of erosion beneath the Petrovbreen Member (Halverson *et al.* 2004).

Petrovbreen Member (E2)

The Petrovbreen Member is on average only about 10 m thick (Fig. 55.4), but is highly variable (<52 m), its thickness broadly reciprocating the depth of erosion on the sub-glacial sequence boundary (Halverson *et al.* 2004). In Nordaustlandet, the Petrovbreen Member is typically very thin and had been recognized only in northernmost sections (Flood *et al.* 1969; Hambrey 1982; Harland *et al.* 1993), but recent mapping has shown that it can be identified in most sections (Halverson *et al.* 2004). The yellow- to orange-weathering Petrovbreen Member is predominantly composed of poorly stratified and massive diamictite and wackestone with a yellow- to orange-weathering dolomite matrix. Other common facies include dolomitic rhythmites, tabular clast conglomerates and sandy dolostone (Fairchild & Hambrey 1984). Climbing ripples occur near the top of the member in some sections. The clasts in the diamictite range up to 90 cm in diameter, are subangular to angular, and consist mainly of light grey dolomite

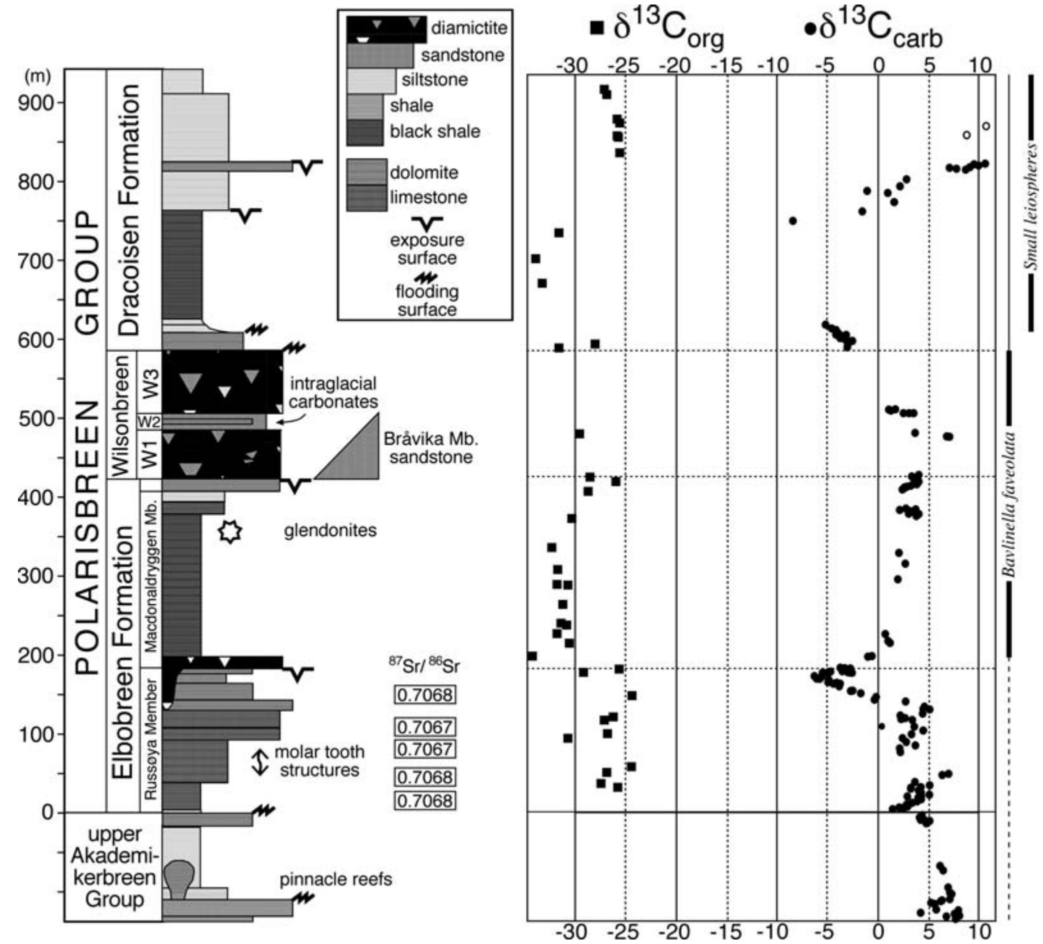

Fig. 55.3. Composite stratigraphic column through the Polarisbreen Group (after Halverson *et al.* 2004) and associated chemostratigraphic and biostratigraphic data. Sr-isotope data are from Halverson *et al.* (2007), C-isotope data (v. VPDB) on organic matter ($\delta^{13}C_{org}$) from Knoll *et al.* (1986) and Kaufman *et al.* (1997) (as compiled in the latter), and C-isotope data on carbonates ($\delta^{13}C_{carb}$) from Fairchild & Spiro (1987), Halverson *et al.* (2004) and Kaufman *et al.* (1997). The two principal acritarch assemblages in the Polarisbreen Group (right side of figure) are summarized from Knoll (1982) and Knoll & Swett (1987).

(grainstone and stromatolite) and black chert. Minor sandstone, siltstone and rare volcanic clasts are also found (Hambrey 1982; Harland *et al.* 1993).

Evidence for direct glacial influence in the Petrovbreen Member comes from rare striated clasts and the abundance of dropstones in finely laminated sediments (Hambrey 1982; Harland *et al.* 1993), as well as evidence of glacial rock flour in the form of submicrometre-sized dolomite in the matrix of the diamictite (Fairchild 1983). Fe and ^{18}O-enrichment of this dolomitic matrix relative to clasts suggests early marine alteration (Fairchild *et al.* 1989). The absence of features indicative of a terrestrial environment, coupled with rapid facies changes and the occurrence of dropstones, suggest glaciomarine (or glaciolacustrine) deposition near an ice-grounding line (Hambrey 1982; Fairchild & Hambrey 1984; Harland *et al.* 1993).

MacDonaldryggen Member (E3)

The contact between the upper Petrovbreen Member and the base of the MacDonaldryggen Member is everywhere sharp but conformable. In Olav V Land, where the MacDonaldryggen Member was named (Harland *et al.* 1993), the base consists of a thin (10–40 cm), finely laminated carbonate (variably limestone and dolomite) with wispy organic-rich laminae (Halverson *et al.* 2004). Elsewhere, the upper Petrovbreen Member is directly overlain by the same finely laminated, olive green to dark grey mudstone that comprises the bulk of the *c.* 200-m-thick MacDonaldryggen Member. The fraction of silt increases up-section, with thin, fine sandstone beds present in some sections. In Nordaustlandet, authigenic calcite nodules up to 2 cm in diameter, and in places densely clustered within distinct layers, speckle the upper MacDonaldryggen Member some 20–40 m beneath the contact with the overlying Slangen Member (Fig. 55.3). These nodules are commonly rosette-shaped and have canted and square prismatic habits (Halverson *et al.* 2004).

The upper MacDonaldryggen shales and siltstones are everywhere transitional into flaggy dolomites that mark the bottom of the Slangen Member. In southern sections, this transition includes an increase in carbonate, including limestone ribbons. Mud-cracks have been reported from this level (Fairchild & Hambrey 1984), but appear to be rare.

Slangen Member (E4)

The Slangen Member is typically a 20–30 m regressive parasequence of cherty dolostone. The dominant facies in the Slangen Member is a vuggy, trough and tabular cross-bedded grainstone; however, fenaestral microbial laminates also occur as thin interbeds within the grainstone and as distinct beds up to 5 m thick in the upper Slangen Member. Ribbons occur mainly in the transition with the underlying MacDonaldryggen Member, but are also important in the uppermost Slangen Member in Nordauslandet. The occurrence of length-slow chalcedony, high Na concentrations, and rare anhydrite in the Slangen Member suggest deposition in a restricted environment such as a coastal sabkha (Fairchild & Hambrey 1984).

The upper contact of the Slangen Member is variable. In most Spitsbergen sections, it is sharp, silicified and brecciated, with clasts showing no evidence for reworking (Fairchild & Hambrey 1984). At Ditlovtoppen (Fig. 55.1) the grainstone passes transitionally upward into 2.5 m of mud-cracked siltstone beneath the basal diamictites of the Wilsonbreen Fm. In Nordaustlandet, the grainstones and microbial laminites are commonly overlain by silty dolomite ribbons, which grade upwards through ripple cross-laminated dolomitic siltstone and fine sandstone and into dolomitic sandstone of the basal Bråvika Member (Fig. 55.4).

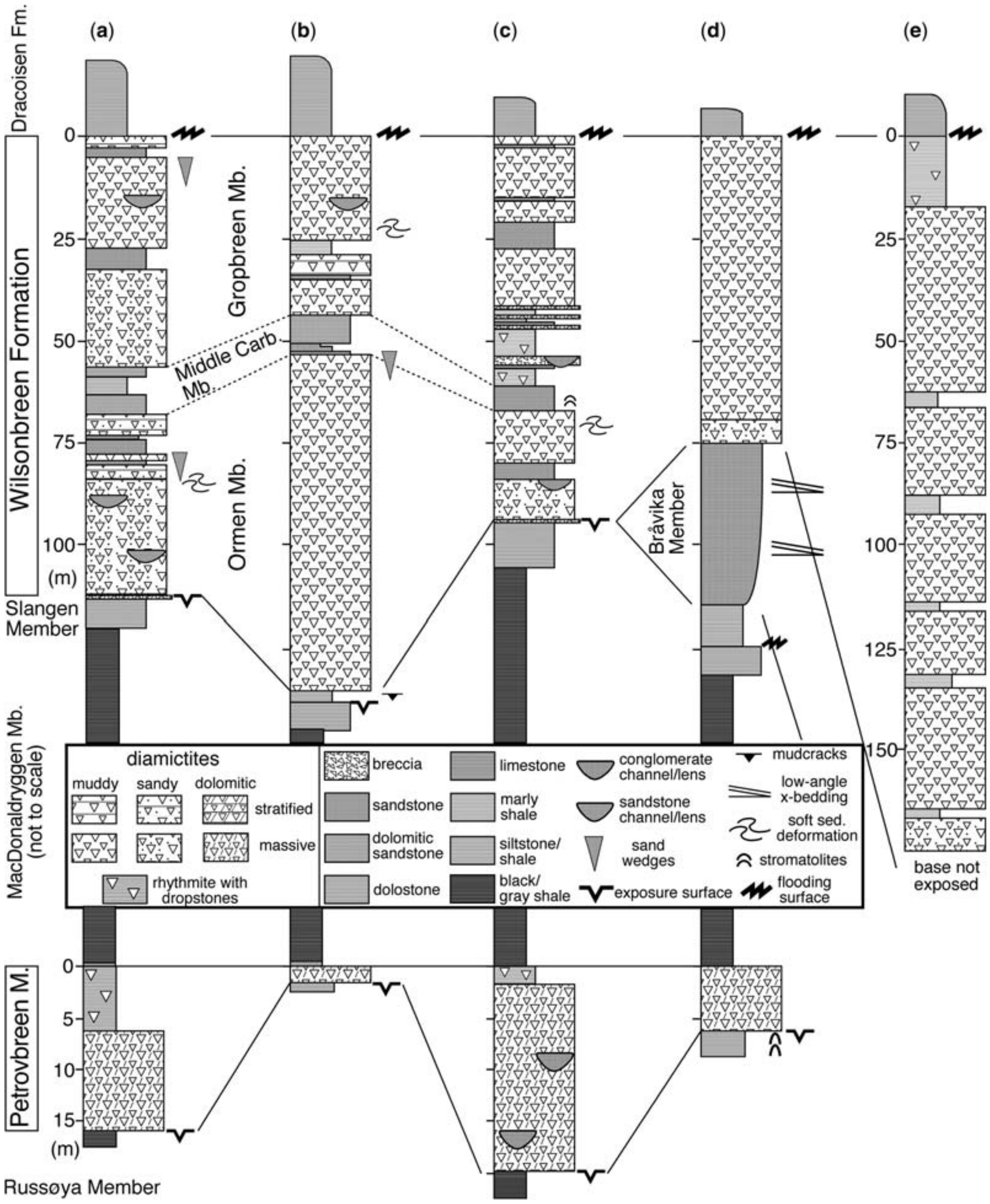

Fig. 55.4. Characteristic stratigraphic sections of the Wilsonbreen Fm. and Petrovbreen Member glacials from each of principal regions in which the glacial intervals have been studied (locations a to e are linked to Fig. 55.1; the Petrovbreen Member does not occur at location e). Stratigraphic depth is given relative to a 0 m datum at the top of each glacial unit. Note the difference in scale between the Wilsonbreen Fm. and Petrovbreen Member. The measured sections are compiled from published columns in Edwards (1976), Harland *et al.* (1993), Halverson *et al.* (2004) and unpublished data of the author. The breakdown and naming of members here follows that of Harland *et al.* (1993), but is slightly different from that used in previous papers, where the names of the Ormen and Gropbreen Members are inverted and a fourth Wilsonbreen member is included (e.g. Hambrey 1982).

The Wilsonbreen Fm.

In most sections, the Wilsonbreen Fm. consists of *c.* 100–150 m of green- to red-weathering, massively to poorly bedded diamictite; however, the diamictite thins dramatically northwards in Nordaustlandet, being reciprocated by a northward-thickening wedge of quartz sands, informally named the Bråvika member (Halverson *et al.* 2004; Fig. 55.4). In the northernmost exposure of the Wilsonbreen Fm. (at Langrunesset), the Bråvika member is 250 m thick and diamictite is absent.

In Spitsbergen, the Wilsonbreen Fm. is subdivided into the Ormen (W1), Middle Carbonate (W2) and Gropbreen (W3) members (Harland *et al.* 1993; Fig. 55.4). The Ormen Member varies from 20 to 75 m in thickness and consists mainly of poorly stratified, shaley diamictite, but also includes lenses and interbeds of sandstone, conglomerate and breccia (Fig. 55.4; Fairchild & Hambrey 1984). Clast lithology is variable, but dominated by carbonates derived from the underlying Elbobreen Fm. and Akademikerbreen Group. Chert, silt, sandstone and basement clasts, including exotic fragments of gneiss and undeformed pink granite, are also found. Striated clasts occur throughout the unit and dropstones, although not common in most sections, are abundant at Aldousbreen (Fig. 55.4e; Edwards 1976). Sandstone wedges occur near or at the top of the Ormen Member in the Akademikerbreen area (Fig. 55.4a) and at Ditlovtoppen (Fig. 55.4b) (Hambrey 1982; Fairchild & Hambrey 1984).

The Middle Carbonate Member is thin (<25 m) and comprises mainly dolomitic, medium-grained quartz sandstone. It is, however, recognized and distinguished by the presence of thin (<1.5 m) beds, lenses and fragments of primary carbonate. The nature of the carbonate is variable, being corrugated silty dolomitic microbial laminites at Ormen (in the Akademikerbreen area), mixed limestone and dolomite rhythmites and ribbons at Ditlovtoppen (Fig. 55.4b), and isolated limestone stromatolites within sandstone at Dracoisen (Fig. 55.4c). Based on their geochemistry, these carbonates are interpreted to have formed under the influence of evaporation in periglacial lakes (Fairchild *et al.* 1989).

The Gropbreen Member is lithologically similar to the Ormen Member, with dominantly weakly stratified to massive maroon-weathering diamictite, and subordinate sandstone lenses and beds. Clast composition is also similar to the Ormen Member, but with a higher proportion of basement clasts. In many sections, a medium brown sandstone up to a few metres thick occurs at the very top of the Gropbreen Member and is continuous with well developed sandstone wedges that penetrate up to 2 m into the underlying diamictite. Whether the sand is present or not, the contact between the Wilsonbreen Fm. and the overlying Dracoisen Fm. is always sharp.

In Nordaustlandet, the Wilsonbreen is subdivided into two members. The Bråvika member overall is a coarsening-upwards package of quartz sandstone, with irregular dolomite lenses and intraclasts, and in the northernmost sections, minor limestone rhythmites and ribbons. The northward-thickening sandstone contains both planar and trough cross-bedding throughout, indicating a palaeoflow direction to the north (Halverson *et al.* 2004). The prevalence of low-angle cross-sets, coupled with pinstripe laminations and a bimodal grain population, suggests an aeolian origin, although the trough cross-bedding and dolomite precipitation indicate that the sand was in places reworked subaqueously. In the southern exposures in Nordaustlandet, the Bråvika sandstones are transitional with overlying diamictites, but in the Sveanor (Fig. 55.4d) are also transitional below with the Slangen Member (Halverson *et al.* 2004).

The diamictite-dominated unit overlying the Bråvika member in Nordaustlandet resembles the Ormen and Gropbreen members on Spitsbergen, only containing a higher percentage of undeformed igneous clasts. Rhythmites with dropstones and fine interbedded turbidites are well preserved in the southernmost Nordaustlandet section at Aldousbreen (Fig. 55.4e), in Wahlenbergfjorden (Edwards 1976). Unstratified siltstones, interpreted as loessites (Edwards 1976), also occur in the same section (Fig. 55.4).

The Dracoisen Fm.

The colourful lower part of the Draocoisen Fm. is a superb marker interval (Wilson & Harland 1964). A yellow-weathering dolomite bed everywhere overlies the Wilsonbreen Fm., with no evidence for reworking or depositional hiatus. This dolostone varies in thickness from as much as 18 m at Ditlovtoppen to less than 3 m in parts of Nordaustlandet (Halverson *et al.* 2004) and always comprises the transgressive component of a thick (*c.* 170 m) sequence (Fig. 55.3). The dolostone is mostly finely laminated dololutite, with minor low-angle erosional truncations and cross-laminations. Inversely graded laminae and pockets of peloids are also common. Large-scale oscillation ripples with amplitudes of up to 40 cm and wavelengths up to 6 m (Allen & Hoffman 2005) occur in the upper part of the dolostone in most northern sections.

The top of the dolostone is transitional into marly red siltstone, which then passes upwards into green then black shales. The change to black shale, at *c.* 30–40 m above the base of the Dracoisen Fm., marks the maximum flooding surface of the sequence. Black shales with common carbonate concretions persist for over 100 m and are transitional into mud-cracked, variegated, finely bedded siltstone and shale that constitutes much of the upper part of the formation. The interval of mud-cracked siltstone is interrupted by 9 m of microbial-laminated dolomite, internally deformed by cauliflower chert.

Chemostratigraphy

In one of the earliest systematic chemostratigraphic studies of a Neoproterozoic succession, Knoll *et al.* (1986) produced a coupled $\delta^{13}C_{carb}$–$\delta^{13}C_{org}$ record through the Veteranen, Akademikerbreen and Polarisbreen groups in both Spitsbergen and Nordaustlandet (as well as East Greenland). Although by modern standards the resolution of this record is low, with only a handful of data points from the Polarisbreen Group, this paper established the concurrence of glaciations and negative C-isotope anomalies, as well as the prevalence of high $\delta^{13}C$ values during the much of the Neoproterozoic. Subsequent C-isotope stratigraphy on the Polarisbreen Group was published by Fairchild & Spiro (1987), Kaufman *et al.* (1997) and Halverson *et al.* (2004, 2005). Separate studies have focused on the isotopic composition of dolomites and limestones within the Wilsonbreen Fm. as a means of reconstructing the palaeoenvironment of syn-glacial carbonate precipitation (Fairchid & Spiro 1987, 1990; Fairchild *et al.* 1989).

A compilation of C-isotope data from these various publications is plotted alongside a composite stratigraphic column in Figure 55.3. The Polarisbreen Group $\delta^{13}C_{carb}$ and $\delta^{13}C_{org}$ records broadly mirror one another (Fig. 55.3), with an average isotopic difference of *c.* 30‰, which is typical for the Neoproterozoic (Hayes *et al.* 1999). The most striking feature of the C-isotope record is a 12‰ negative anomaly just below the Petrovbreen Member. This anomaly is reproduced in multiple sections and essentially spans the upper two parasequences of the Russøya Member (Halverson *et al.* 2004). $\delta^{13}C$ values are still negative in carbonates within the lowermost MacDonaldyrggen Fm., but quickly rise to positive values, where they remain into the Wilsonbreen Fm. Intraglacial carbonates show a range in $\delta^{18}O$ and $\delta^{13}C$ values between -11 and $+11$‰ and $+1$ and 5‰, respectively (Fairchild & Spiro 1987; Halverson *et al.* 2004). Fairchild & Spiro (1987, 1990) argued that the Wilsonbreen carbonates were precipitated by evaporation in a periglacial lake based, in part, on these unusually high $\delta^{18}O$ compositions. More recently, Bao *et al.* (2009) reported $\Delta^{17}O$ (triple oxygen isotopes) compositions in carbonate-associated sulphate (CAS) within these carbonates of as low as -1.6‰, which are the lowest values ever reported from natural specimens. Bao *et al.* (2009) interpreted these anomalous values to record either extraordinarily high pCO_2 or unusually sluggish O_2 cycling during the glaciation.

A second negative C-isotope anomaly is recorded in the basal Dracoisen dolostone and reproduced in multiple sections, where $\delta^{13}C$ values gradually decline from -3 to -5‰ (Halverson *et al.* 2004). $\delta^{13}C$ values then spike to >10‰ in evaporitic dolomites in the upper Dracoisen Fm. (Fig. 55.3).

Four Sr-isotope data from the Polarisbreen Group were first published by Derry *et al.* (1989), and a few new analyses from the original sample set were added in Jacobsen & Kaufman (1999). Additional $^{87}Sr/^{86}Sr$ data from the Polarisbreen were published in Halverson *et al.* (2005, 2007). The highest-quality data from all these studies consistently show $^{87}Sr/^{86}Sr = 0.7067$–0.7068 in the Russøya Member (Fig. 55.3). The absence of unaltered marine limestone above the Russøya Member precludes application of Sr-isotope chemostratigraphy to the remainder of the Polarisbreen Group.

Palaeolatitude and palaeogeography

Barentsia is widely assumed to have been contiguous with East Greenland during the Neoproterozoic based on the remarkable stratigraphic similarity between the two regions (Harland & Gayer 1972; Knoll *et al.* 1986; Fairchild & Hambrey 1995). However, the precise location of Eastern Svalbard relative to the East Greenland Caledonides (Johansson *et al.* 2005) and the geometry and origin of the East Greenland–eastern Svalbard platform within the broader framework of the connection and eventual rifting between Laurentia, Baltica and Amazonia are controversial. In the conventional configuration, western Baltica is the conjugate margin to eastern Greenland (and Barentsia), virtually identical to the Caledonian fit (e.g. Weil *et al.* 1998). In a different reconstruction, Hartz & Torsvik (2002) placed the East Greenland–eastern Svalbard platform adjacent to the southern peri-Urals of Baltica and proposed that it developed as a sinistral pull-apart basin between Baltica and Amazonia. This model has been challenged by more recent palaeomagnetic data from Baltica (Cawood & Pisarevsky 2006), and in a modification of earlier west-Baltica–eastern Greenland fits, Pisarevsky *et al.* (2008) proposed that the west Norwegian margin was conjugate to southeastern Greenland. This model implies that the eastern Svalbard and the

Neoproteroozic successions in northern Norway were part of a long passive margin isolated from the other coeval sedimentary basins to the south.

Direct palaeomagnetic constraints for Svalbard's Neoproterozoic palaeography are limited to a new suite of palaeomagnetic data from the Akademikerbreen Group that show a stable, pre-Caledonian remanent magnetization indicating that the EGES platform resided in tropical latitudes ($I \approx 15^{\circ}$) during the mid-Neoproterozoic (Maloof *et al.* 2006). No reliable palaeomagnetic results have been obtained from the Polarisbreen Group, and Svalbard's complicated and uncertain tectonic history since this time precludes any extrapolation of palaeolatitude for the glacial deposits from elsewhere on the Laurentia craton (Evans 2000).

Geochronological constraints

No direct radiometric ages have been obtained on the upper Hecla Hoek. U–Pb ages on detrital zircons in the Planetfjella Group in Spitsbergen and presumed primary zircons from volcanics and subvolcanic intrusives beneath the Veteranen Group in Nordaustlandet indicate a maximum age for the base of the Lomfjorden Supergroup of *c.* 946 Ma (Johansson *et al.* 2000, 2005; Fig. 55.2). Similar ages have been obtained from igneous clasts in the Wilsonbreen Fm. in Nordaustlandet (Johansson *et al.* 2000).

The top of the Polarisbreen Group, which is in disconformable contact with the overlying Cambrian Tokamanne Fm. (Knoll & Swett 1987), is inferred to be <542 Ma (the age of the Precambrian–Cambrian boundary; Amthor *et al.* 2003). If the Dracoisen Fm. is correlative with the Maieberg Fm. in northern Namibia and the Doushantuo Fm. in South China (Hoffmann *et al.* 2004; Condon *et al.* 2005), as implied by the chemostratigraphy and biostratigraphy of the Polarisbreen Group, then the Wilsonbreen glaciation ended at 635 Ma (Halverson *et al.* 2005).

Biostratigraphy

In contrast to the rich and diverse fossil assemblage from the underlying Akademikerbreen Group (Knoll 1982; Butterfield *et al.* 1994, and references therein), the biostratigraphic record of the Polarisbreen Group is most notable for its depauperate microfossil assemblage, despite the generally mild thermal history of the rocks (Knoll & Swett 1987). In the MacDonaldryggen Member and in shales within the Wilsonbreen Fm., the acritarch *Bavlinella faveolata* dominates the microfossil assemblage (Knoll 1982; Knoll & Swett 1987). *B. faveolata* also occurs in the Dracoisen Fm. and is locally abundant, but overall is subordinate to a low diversity assemblage of small, smooth-walled leiosphere acritarchs (Fig. 55.3). The diverse assemblage of the large, acanthomorphic acritarchs that characterize middle–upper Ediacaran successions in Australia (Grey *et al.* 2003) and elsewhere (Knoll 2000) is absent in the Polarisbreen Group (Knoll & Swett 1987). Coupled with the lack of Ediacaran fauna, this lacuna strongly suggests that the contact between the Polarisbreen Group and the overlying Oslobreen Group represents a significant depositional hiatus (Knoll & Swett 1987) and that at least the last 35 myr of the Ediacaran Period are missing in Svalbard.

Discussion

The Neoproterozoic succession in Svalbard has played a prominent role in questions and controversies regarding Precambrian glaciations and environmental change since Harland (1964) first proposed a pan-global *infra-Cambrian* glaciation. Although a few researchers maintained scepticism over the glacial origin of diamictites in the Polarisbreen Group (Krasil'shchov 1973; Schermerhorn 1974), the detailed sedimentological work of Edwards, Hambrey, Fairchild and others (e.g. Edwards 1976; Hambrey 1982; Fairchild & Hambrey 1984; Dowdeswell *et al.* 1985) effectively put to rest any doubt. Evidence for a glacial environment, specifically in the Wilsonbreen Fm., is found in the abundance of striated and faceted clasts, dropstones, the extrabasinal origin of some stones, and sandstone wedges.

The Petrovbreen Member diamictite and associated lithofacies are typically interpreted to have been deposited in relatively deep water, most likely in a marine environment (Fairchild & Hambrey 1984; Harland *et al.* 1993). This interpretation is consistent with the lack of evidence for any obvious change in relative sea level at the contact between the Petrovbreen and MacDonaldryggen Members (Halverson *et al.* 2004). In contrast, the heterogeneous Wilsonbreen Fm., which includes diamictites interpreted as lodgement tills (Dowdeswell *et al.* 1985), direct evidence for subaerial exposure (sandstone wedges), and lacustrine carbonates is inferred to have been deposited in a predominantly terrestrial setting (Fairchild & Hambrey 1984; Dowdeswell *et al.* 1985; Halverson *et al.* 2004). Dowdeswell *et al.* (1985) argued that the Wilsonbreen diamictites were deposited beneath ice sheets or ice caps (rather than valley glaciers) based on the preponderance of exotic clasts and a lack of angular supraglacial debris.

A great deal of effort has been expended on attempting to correlate the Polarisbreen diamictites with other Neoproterozoic glacial units in the North Atlantic (e.g. Harland & Gayer 1972; Hambrey 1983; Nystuen 1985; Fairchild & Hambrey 1995) and globally (e.g. Kaufman *et al.* 1997; Kennedy *et al.* 1998; Halverson *et al.* 2005). Based on biostratigraphic and chemostratigraphic data, it has generally been argued that the Polarisbreen Group is late 'Vendian' in age, meaning essentially that it post-dates the earliest Neoproterozoic glaciations. Unfortunately, the initial optimism that the chronostratigraphy and correlations would be firmed up by geochronological and palaeomagnetic data (Hambrey 1983) has not been realized, and no undisputed correlation scheme exists for the North Atlantic region, let alone the whole of the globe.

In a twist on the conventional Vendian interpretation of the age of the Polarisbreen Group diamictites, Halverson *et al.* (2004) argued that both the Petrovbreen Member and Wilsonbreen Fm. belonged to a single glacial episode, ending at 635 Ma (the approximate age of the end of the Ghaub glaciation in Namibia and Nantuo glaciation in South China; Hoffmann *et al.* 2004; Condon *et al.* 2005). This model was based on (i) the similar stratigraphic context and magnitude of the upper Russøya negative $\delta^{13}C$ anomaly (Fig. 55.3) and the so-called Trezona anomaly, which precedes the Ghaub glaciation in Namibia (Halverson *et al.* 2002), coupled with lack of evidence for an analogous anomaly preceding any of the older glaciations; (ii) the occurrence of glendonites in the upper MacDonaldryggen Member; and (iii) the virtually identical geochemistry, sedimentology and stratigraphy of the Dracoisen and post-Ghaub Maieberg (and other) cap-carbonate sequences. In light of new Sr-isotope data from the level of the negative anomaly in the Russøya Member ($^{87}Sr/^{86}Sr = 0.7068$ v. 0.7072 in the Trezona anomaly; Halverson *et al.* 2007) and evidence that the lowermost of three distinct glacial units in the Dalradian Supergroup is preceded by a negative C-isotope anomaly of similar magnitude to that in the upper Russøya Member (McCay *et al.* 2006), it now seems more likely that the Petrovbreen Member diamictite must represent a separate, older glaciation. Although it remains likely that the Wilsonbreen Fm. is *c.* 635 Ma, the ongoing ambiguity in Cryogenian–Ediacaran age constraints precludes any unequivocal assignment of age to either of the Polarisbreen glaciations.

The more conventional two-glaciation interpretation also raises the question of the climatic significance of the glendonite nodules in the upper MacDonaldryggen Member. Glendonites have also been reported in possibly equivalent-aged strata in the Windermere Supergroup in the Mackenzie Mountains (James *et al.* 2005). Insofar as the MacDonaldryggen pseudomorphs originated as ikaite, which forms at near-freezing temperatures in alkaline-charged waters (Buchart *et al.* 1997), then their occurrence in continental shelf sediments with no other evidence of glaciation, and just below sabkha deposits, poses yet another Neoproterozoic climatic conundrum. Thus, it seems certain that the Polarisbreen Group will continue to play a central role in the ongoing effort to understand Neoproterozoic climate.

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