Chapter 2 A history of Neoproterozoic Glacial Geology, 1871–1997

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Chapter 2

A history of Neoproterozoic glacial geology, 1871–1997

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Abstract: Neoproterozoic glacial records have been discovered on 23 palaeocontinents, their rate of discovery changing little since 1871. Yet, half of all the resulting publications appeared since 2000. The history of research before 1998 is described in five stages defined by publication spikes; subsequent work is not covered because historical perspective is lacking. In stage 1 (1871–1907), 'Cambrian' (now Neoproterozoic) glaciation was recognized successively in Scotland, Australia, India, Norway, Svalbard and China. Criteria for recognition included faceted and striated pebbles in matrix-supported conglomerates resting on ice-worn bedrock pavements. In stage 2 (1908–1940), Neoproterozoic glaciation was shown to have been widespread in Africa, Asia and the Americas. Major textbooks summarized these findings, but the rejection of continental drift (to account for late Palaeozoic glacial dynamics) put a chill on research. In stage 3 (1942–1964), the occurrence of glacial deposits within carbonate successions, as well as nascent palaeomagnetic observations, suggested that Neoproterozoic glaciers reached sea-level at low palaeolatitudes, but the belated recognition of sediment gravity flowage caused glacial interpretations to be prematurely abandoned in key areas. In stage 4 (1965–1981), the extent of Neoproterozoic glaciation was rethought in light of plate tectonics. Distinctive chemical sediments (iron $\pm$ manganese formations and cap carbonates) were identified. In basic climate models, modest lowering of solar luminosity resulted in global glaciation due to ice-albedo feedback, and deglaciation due to greenhouse forcing resulted from silicate-weathering feedback in the carbon cycle. Neoproterozoic glacial geologists were blind to these ideas. In stage 5 (1982–1997), reliable palaeomagnetic data combined with glacial marine sedimentation models confirmed that Neoproterozoic ice sheets reached sea level close to the palaeoequator.

The first Neoproterozoic glacial deposits were found in the SW of Scotland in 1871 and the rate of their discovery on 23 palaeocontinents and microcontinents was essentially linear until 1992 (Fig. 2.1). However, half of all the papers ($n = 811$) written on them before the end of 2008 were published in the previous 10 years (Fig. 2.2). This surge in activity is the justification for the present volume, but this history ends at 1997. We lack historical perspective on the last decade and the endpoint eliminates the need to weigh any of the author's own contributions on this topic.

This chapter begins by placing the recognition of the Port Askaig boulder beds of Scotland as glaciogenic in its historical context, coming in 1871 on the heels of the discoveries of late Palaeozoic glacial deposits in India, Australia and South Africa. Their inspiration was the Pleistocene glacial controversy of 1837–1865, which spawned both the orbital and 'greenhouse' theories of bidirectional climate change, and established the principal criteria for the recognition of past glaciation.

The history is divided into five periods punctuated by spikes in publications (Fig. 2.2) associated with the 10th and 17th International Geological Congresses in Mexico City (1906) and Moscow (1937), respectively, the twin palaeoclimate conferences in Newcastle upon Tyne (1963) and Köln (1964), the IGCP Project 38 volume on *Earth's Pre-Pleistocene Glacial Record* (Hambrey & Harland 1981), and the onset of the present surge triggered in 1998.

The first period (1871–1908) saw the recognition of 'Lower Cambrian' glaciogenic formations in SW Scotland (1871), South Australia (1884), NE Norway (1891), western Svalbard (1898), South China (1904) and NW India (1908). It ended with overviews of Late Palaeozoic and Lower Cambrian ('possibly Pre-Cambrian') glaciations by Edgeworth David (1907*a*, *b*) at the 10th IGC, and the winning-over of a skeptical Geological Society in London by a fellow Australian (Howchin 1908). Faceted and striated clasts in diamictites were the decisive criteria for glaciogenesis, with a striated subglacial pavement beautifully exposed beneath 'Reusch's Moraine' (Fig. 2.3) in Finnmark, Norway, for good measure.

The second period (1909–1941) saw Neoproterozoic glacial deposits recognized in southern, central and western Africa, central and northern Asia, eastern and western North America and eastern South America (Fig. 2.1, Table 2.1). It was evident that glaciation 'at or just before the beginning of the Cambrian' had been more extreme than in the Pleistocene, possibly equal to the composite Late Palaeozoic glaciations of Gondwanaland. These findings were highlighted in a *Symposium on Palaeozoic and Pre-Cambrian Climates* at the 17th IGC (1937) in Moscow. Major textbooks made note of these developments (Köppen & Wegener 1924; Brooks 1922, 1926; Coleman 1926).

In the third period (1942–1964), nascent evidence from palaeomagnetism and from the occurrence of glacial deposits atop thick carbonate-dominated successions in the North Atlantic region and within such successions in central and southern Africa implied that Neoproterozoic ice sheets had reached sea level in low palaeolatitudes. This inference met resistance and, after the discovery of turbidity currents as a cause of graded bedding (1950), some diamictites were reinterpreted as mass-flow deposits of non-glacial origin. These issues were discussed at major conferences on *Palaeoclimates* in Newcastle-upon-Tyne, England (1963), and Köln, Germany (1964). The radiation of macrofauna following soon after the 'infra-Cambrian' ice age was often remarked upon.

The fourth period (1965–1981) was notable for a detailed restudy of the classic Port Askaig Tillite, debate over the roles of true polar wander and plate tectonics in the distribution of glaciogenic deposits, and the recognition of widespread post-glacial 'cap dolomites' and syn-glacial iron and manganese formations, features not known from Phanerozoic glaciations. Climate modelling received a kick-start when simple energy-balance calculations suggested that the Earth could freeze over from pole to pole due to ice-albedo feedback in response to lowering of the solar constant by a few percent. Such a 'white Earth' disaster was shown to be self-reversing because of negative climate feedback associated with the geochemical cycle of carbon, but geologists were unaware of these developments.

In the fifth period (1982–1997), Neoproterozoic diamictites were increasingly interpreted in light of glacial marine sedimentation models and, combined with reliable palaeomagnetic data, proved that ice sheets had reached sea level close to the palaeoequator. The 'Snowball Earth' hypothesis, adapted from climate and planetary science, was advanced as a

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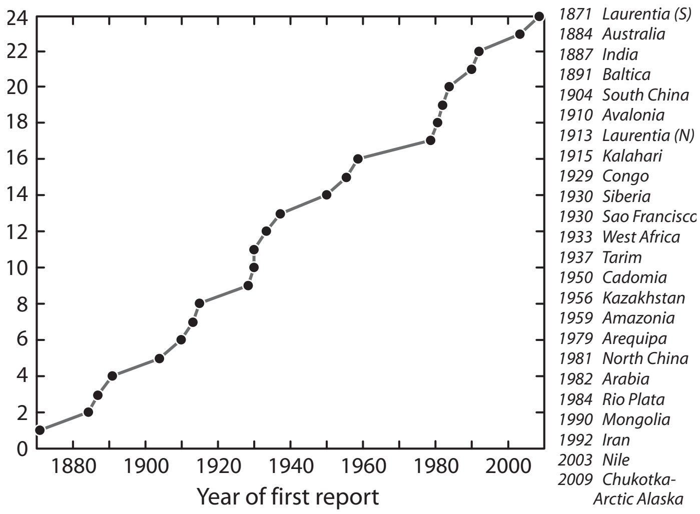

Fig. 2.1. Cumulative discovery of Neoproterozoic glaciogenic deposits by palaeocontinent. See Table 2.1 for locations and references. Note that Laurentia is counted twice because of the large geographical separation between southern (S) and northern (N) palaeohemisphere deposits of present eastern and western Laurentia, respectively.

parsimonious explanation for low-latitude glaciation, associated iron ($\pm$manganese) deposits and (later) cap carbonates. Initially ignored, the strong negative reaction to this hypothesis on the part of Neoproterozoic glacial sedimentologists will in future be seen as the most striking feature of the post-1998 period.

Prologue: discovery of the Port Askaig Tillite and its historical scientific context

> [On the stratified rocks of Islay.] 'If . . . we compare the embedded boulders of granite [in the schist] with the granites found in situ throughout the Highlands, we feel the necessity of tracing them to another source, and hope we do not overstep the bounds of prudent speculation in suggesting that those erratics are the reassorted materials of some great Northern Continent that has yielded to the ceaseless gnawing tooth of time, leaving scattered fragments as wreckage of its former greatness, and that the material of which the mass is composed have in time, deeper than we have hitherto suspected, been transported by the agency of ice.'
>
> James Thomson, F.G.S. 1871. On the stratified rocks of Islay. Report of the 41st Meeting of the British Association for the Advancement of Science, Edinburgh, John Murray, London, 110–111.

The Port Askaig boulder beds

The first Neoproterozoic (then 'Cambrian') formation to be interpreted as glaciogenic (see quote above) was the Port Askaig, exposed on the east side of Islay (Thomson 1871, 1877), an island in the west of Scotland famous for its single malt whisky. The formation is a 750-m-thick succession of 47 mappable heterolithic-boulder diamictites with interbeds of crossbedded sandstone (Kilburn *et al.* 1965; Spencer 1971; Arnaud 2004). In his richly detailed sedimentological study, Spencer (1971) describes the Port Askaig as lying 'at the same horizon, between two carbonate-rich formations, for 700 km from NE Scotland to western Ireland'. It is now thought to be the oldest of three distinct glaciogenic horizons within the Dalradian Supergroup (McCay *et al.* 2006) – older and younger Cryogenian and middle Ediacaran in age. James Thomson's speculation (see above) about a 'great Northern Continent' as the source of the glacial debris turned out to be true (Cawood *et al.* 2003), although that continent did survive the 'gnawing tooth of time' after all. It is North America, or more precisely, its pre-Pangaean antecedent Laurentia. It was merely displaced from the Dalradian in the early Eocene, when Greenland separated from NW Europe. The Port Askaig was deposited at palaeolatitudes near 25°S according to reliable palaeomagnetic data from mafic dykes, lavas and sills of the marginally older (723–718 Ma) Franklin Igneous Suite of Arctic Laurentia (Evans 2000). It accumulated on a marine shelf, subject to repeated subaerial exposure (Spencer 1971; Johnston 1993), situated at the then southeastern margin

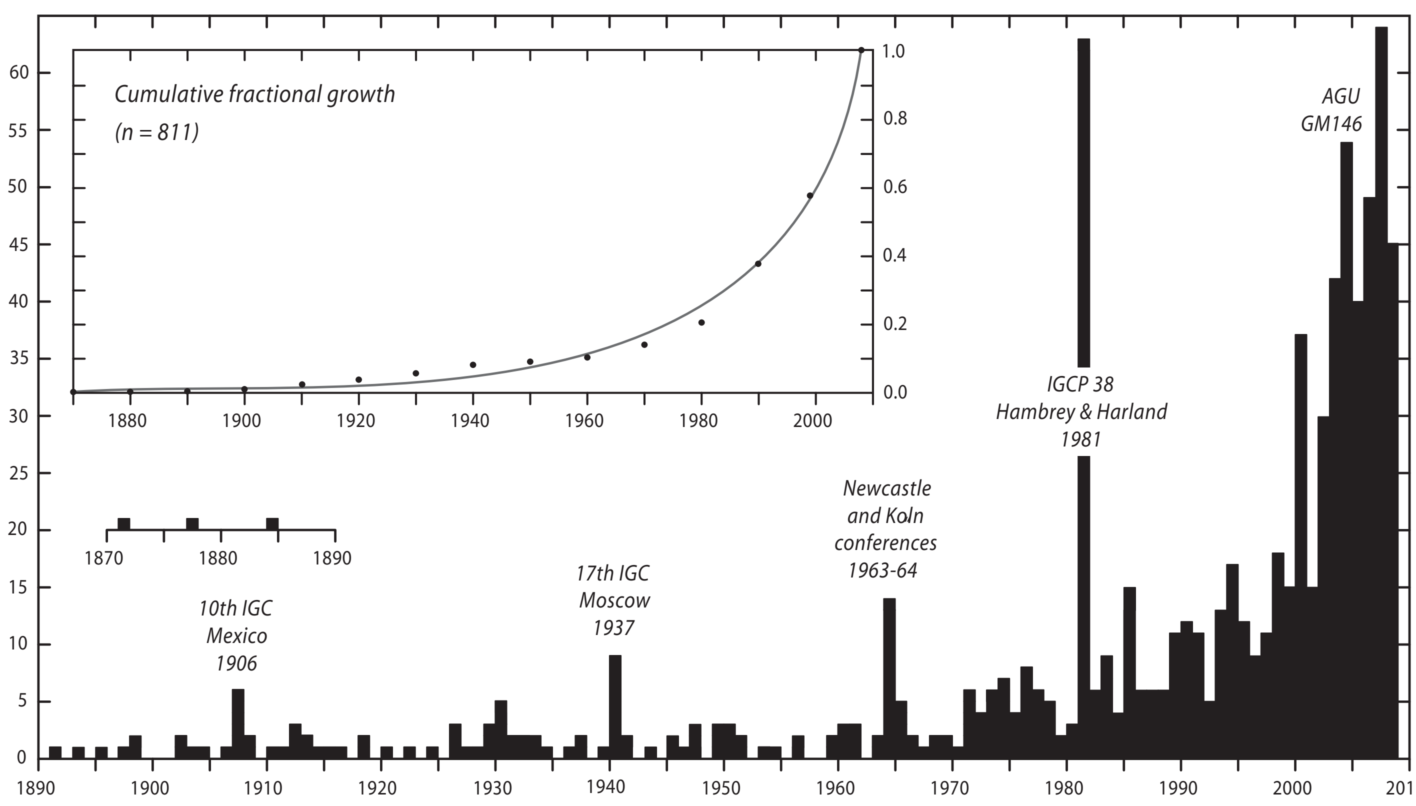

Fig. 2.2. Growth in the annual number of papers concerning Neoproterozoic glaciation, 1871–2008.

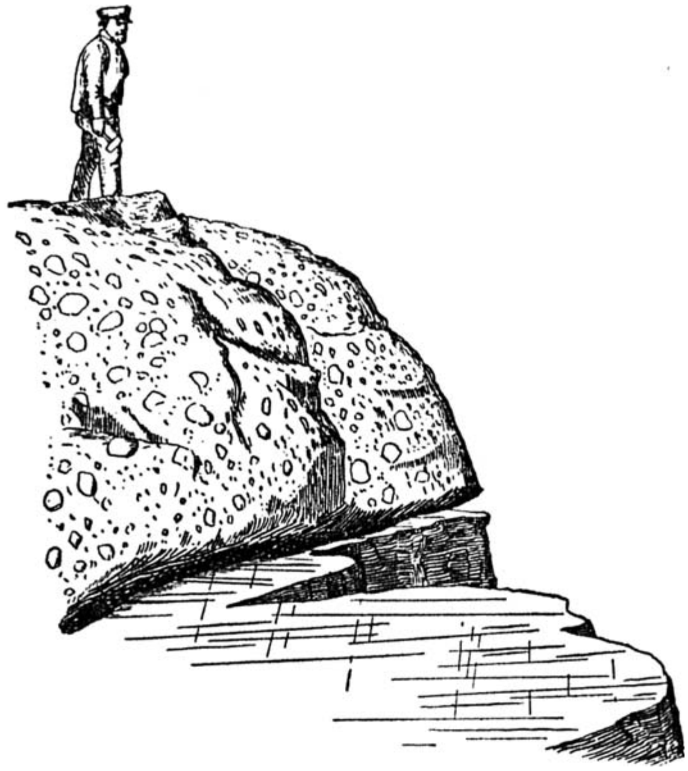

Fig. 2.3. Iconic sketch of the glacially striated pavement beneath the end-Cryogenian Smalfjord diamictite ('Reusch's moraine') at Bigganjargga, Varangerfjord, East Finnmark, northern Norway (from Reusch 1891).

of the palaeocontinent. The Port Askaig nicely illustrates the basic information needed to interpret Neoproterozoic glaciogenic deposits: their distribution, derivation, depositional environment, palaeogeographic setting and age.

Late Palaeozoic glaciation and the theory of continental drift

Thomson's (1871) interpretation of the Port Askaig came at a time of growing interest in pre-Pleistocene glaciation. A subglacial pavement overlain by diamictite in South Australia found in 1859 by A.C. Selwyn proved to be of Permo-Carboniferous age (Howchin 1912). A Permo-Carboniferous diamictite in the Talchir coal basin of NE India (Blanford *et al.* 1859) was shown to have been deposited by grounded, northward(!)-flowing glaciers (Fedden 1875; Koken 1907). The late Carboniferous Dwyka Tillite in Natal, South Africa, rests on a pavement scratched and grooved by onshore(!)-flowing glaciers (Sutherland 1870). Ironically, these findings had been inspired by the glacial interpretation of a Permian-age diamictite in England (Ramsay 1855) that proved to be non-glacial in origin.

By 1907 it was evident that with continents in fixed relative positions, no amount of polar wander could prevent early Permian glaciation in the southern hemisphere from extending to within 10° latitude of the palaeoequator, while in the northern hemisphere tropical warmth stretched to the pole (Koken 1907; Irving 1956). 'The Permian ice age poses an unsolvable problem to all models that do not dare to assume horizontal displacements of the continents', wrote a 32-year-old German meteorologist (Wegener 1912). With the Atlantic and Indian oceans closed up, however, all the glacial action could be contained poleward of 45°S palaeolatitude, comparable to the extent of northern hemisphere glaciation in the Pleistocene. Moreover, a radial pattern of flow could then be inferred, centred on south-central Africa (Martin 1981). 'This would take everything mysterious away from the phenomenon' (Wegener 1912). The distribution of Late Palaeozoic glacial deposits was prime motivation for Wegener's theory of continental displacement, or drift (Martin 1981). Had he lived, vindication in the form of plate tectonics would have come to Wegener at the age of 87.

Table 2.1. *First reported occurrences of Neoproterozoic glaciogenic deposits by palaeocontinent*

No.	Year	Palaeocontinent	Continent	Ref.
1a	1871	Laurentia (south)	Europe	Thomson (1871)
2	1884	Australia	Australasia	Woodward (1884)
3	1891	Baltica		Reusch (1891)
4	1904	South China	Asia	Willis (1904)
5	1908	India		Holland (1908)
6	1910	Avalonia	North America*	Sayles & Laforge (1910)
1b	1913	Laurentia (north)	North America	Hintze (1913)
7	1915	Kalahari	Africa	Rogers (1915)
8	1929	Congo		Beetz (1929)
9	1930	Siberia		Nicolaev (1930)
10	1930	São Francisco†	South America	Moraes Rego (1930)
11	1933	West Africa		Baud (1933), Furon (1933)
12	1937	Tarim		Norin (1937)
13	1950	Cadomia		Wegmann *et al.* (1950)
14	1956	Kazakhstan		Nalivkin (1956)
15	1959	Amazonia		Maciel (1959)
16	1979	Arequipa		Caldas (1979)
17	1981	North China		Mu (1981), Lu *et al.* (1985)
18	1982	Arabia		Gorin *et al.* (1982)
19	1984	Rio Plata		Spalletti & Del Valle (1984)
20	1990	Mongolia		Gibsher & Khomentovsky (1990)
21	1992	Iran		Hamdi (1992)
22	2003	Nile‡		Miller *et al.* (2003)
23	2009	Chukotka-Arctic Alaska		Macdonald *et al.* (2009*b*)

*Before 1980, the Squantum Tillite was thought to be Late Palaeozoic in age.
†Often considered part of the Congo palaeocontinent.
‡Western margin of East African orogen.

The Pleistocene glacial controversy

Interest in pre-Quaternary glaciation arose in the wake of the most acrimonious and far-reaching controversy in 19th-century geology, the brouhaha over the glacial theory for Pleistocene tills, erratic boulders and associated landforms (moraines, drumlins, eskers, kame-and-kettle, fluted ground, grooved bedrock, crag-and-tail, roches moutonnées, whaleback rocks, cirques and hanging valleys, horns and arêtes, U-shaped valleys and fjords). Glacial theory – that most of northern Europe and North America had been sculpted by enormous and dynamic ice sheets in the geologically recent past – was independently proposed by Esmark (1824) in Norway, Dobson (1925) in the USA, Venetz (1830) and Charpentier (1837) in Switzerland and Bernhardi (1832) in Germany. None of these papers – published in major journals by reputable professors (Esmark and Bernhardi), engineers (Venetz and Charpentier) and an industrialist (Dobson) – raised a ripple. No geologist ever overturned conventional wisdom with a single paper.

Conventional theory held that these features were products of floods – meltwater floods resulting from dam bursts in the Alps (cf. de Saussure, von Humboldt, von Buch, de Beaumont) or iceberg-laden floods of Arctic origin in northern Europe and North America (cf. Sefström, Murchison, Hitchcock, Lyell). The controversy erupted when glacial theory was taken up by a young, ambitious, energetic and multilingual Swiss palaeoichthyologist (Agassiz 1837, 1840). A recent convert to glacial theory, Agassiz linked it to the extinction of the boreal megafauna (e.g. mammoth, mastodon, woolly rhinoceros, giant deer), a linkage easily disproved (Forbes 1846). He tied the end of the glacial period to the uplift of the Alps, exactly opposite to the view of Charpentier, his primary glaciological tutor, and laughable to those who knew that Alpine orogeny began in the Eocene. Of the geological establishment, only the former arch-Diluvialist William Buckland (Oxford University) was quickly won over to glacial theory.

The climax of Agassiz's campaign in support of glacial theory came in 1840 (Davies 1968), when he and Buckland toured the British Isles, finding varied and widespread evidence of glacial action (Agassiz 1842). The evidence included tills (boulder clays) with polished, faceted and striated clasts, moraines marking the limits of former glaciers, streamlined bedrock and perched terraces resulting from ice-dammed lakes. At first, Agassiz's findings aroused great interest in the English-speaking world, but within months of his tour the glacial hypothesis was rejected by most geologists as unworkable (contrast Lyell 1840, 1841; Hitchcock 1841). Thereafter, glacial theory would languish until a new generation of Scottish geologists undertook systematic surveys that provided (to their own surprise) overwhelming evidence that glacial theory was correct after all (Ramsay 1860; Jamieson 1862, 1863, 1865; A. Geikie 1863, 1865; J. Geikie 1874). Agassiz had to wait 25 years for vindication. Had he lived, Jens Esmark would have been 99 years old.

Lyell, who had 'an undeniable penchant for the skilful defence of lost causes' (Cunningham 1990, p. 247), never completely gave up on his decidedly non-uniformitarian iceberg-drift hypothesis. Conversely, it is insufficiently appreciated today that many of the criteria still used to distinguish ancient glaciogenic deposits were clearly spelled out over 170 years ago by Esmark, Dobson, Venetz, Bernhardi, Charpentier and Agassiz.

The problem of bidirectional climate change and the discovery of the 'greenhouse' effect

The idea of a glacial period, or 'Ice Age' (Schimper 1837), was not controversial to geologists per se: characteristic fauna and flora now live *c.* 1000 km farther north of their occurrences in the 'Drift' (Smith 1836, 1839; Forbes 1846). However, to physicists interested in climate, it came as a shock. They had assumed the the Earth's climate was slowly cooling over time due to the dissipation of primordial heat in the Solar System according to the nebular hypothesis. This was consistent with undisputed palaeobotanical evidence (e.g. palm fronds in Switzerland) that most regions were warmer in the Cretaceous and early Cenozoic than they are today. If the climate could change dramatically in either direction, previously unknown factors must be at work in the climate system.

It fell to the Irish physicist, mountaineer and orator John Tyndall (1861, 1863) to demonstrate experimentally that certain gaseous molecules in the atmosphere, notably water vapour and carbon dioxide (CO_2), absorb infrared radiation ('obscure rays') emitted by the Earth but are transparent to sunlight. Fourier (1824) had previously mentioned the possibility of such a 'greenhouse' effect, but had rejected it as quantitatively unimportant. For Tyndall (1863, p. 204), water vapour provides 'a blanket more necessary to the vegetable life of England than clothing is to man'. Radiative energy balance is maintained because the atmosphere 'constitutes a local dam, by which the temperature of the Earth's surface is deepened: the dam, however, finally overflows, and we give to space all that we receive from the Sun' (Tyndall 1863, p. 205).

Tyndall understood that water vapour cannot be the ultimate cause of climate change, because its concentration is itself a strong function of temperature. It amplifies any given climate change but cannot be its ultimate cause. On the other hand, the content of CO_2 in the atmosphere, or any of the hydrocarbon gases, can vary independently. 'It is not, therefore, necessary to assume alterations in the density and height of the atmosphere to account for different amounts of heat being preserved to the Earth at different times', wrote Tyndall (1861, p. 277), 'a slight change in its variable constituents would suffice for this. Such changes may in fact have produced *all the mutations of climate which the researches of geologists have revealed*' [italics added].

Unknown to Tyndall, a prematurely deceased French ceramicist had already established the basis for the changes in CO_2 that he required, through the geochemical cycle of carbon (Ebelmen 1845, 1847; see also Hunt 1880; Berner & Maasch 1996).

1871–1908: pioneering discoveries

Following the first identification of glaciogenic deposits now known to be Neoproterozoic in age in Scotland (Thomson 1871), similar discoveries were made in Australia (Woodward 1884), Norway (Reusch 1891), Svalbard (Garwood & Gregory 1898), South China (Willis 1904) and India (Holland 1908). The deposits in the extreme NE of Norway made the biggest initial impact. Hans Reusch (1852–1922), then Director of the Norwegian Geological Survey, found them around the head of Varangerfjorden, not far from the Russian border. Reusch (1891) describes hills underlain by up to 50 m of non-stratified, matrix-supported conglomerate in which cobbles of Archaean gneiss and granite predominate, but which also carry smaller clasts of dolomite with facets and non-parallel striations identical to those found in Quaternary tills throughout Norway. On the coast of the fjord near the lighthouse at Bigganjargga, about 40 minutes' walk east from Karlebotn, Reusch (1891) found an isolated ridge of diamictite *c.* 70 m long × 8 m wide × 3 m high (Bjørlykke 1967; Edwards 1975). The ridge belongs to the older (Smalfjord Formation) of two glaciogenic intervals in the region, and is onlapped and draped by shallow-marine sandstone. Just above the waterline of the fjord, the diamictite has eroded away to reveal a glacial pavement on crossbedded quartzite carrying glacial striations in two main orientations (Fig. 2.3). The less distinct (east–west) set of striations preserves lateral ridges and streaks of cataclasite, which suggest that the overlying diamictite is the melt-out tillite of a stagnant ice-cored moraine (Edwards 1975). 'Reusch's Moraine' has been restudied by generations of geologists (Strahan 1897; Schiøtz 1898; Dal 1900; Holtedahl 1918; Rosendahl 1931, 1945; Føyn 1937; von Gaertner 1943; Crowell 1964; Reading & Walker 1966; Bjørlykke 1967; Edwards 1975, 1997; Jensen & Wulff-Pedersen 1996; Rice & Hofmann 2000; Bestmann *et al.* 2006; Arnaud 2008) and, despite differences in interpretation, it remains an icon of Neoproterozoic glacial geology.

Although they were not the first to recognize glacial action in 'Cambrian' (later Adelaidean) strata of South Australia, Walter Howchin in Adelaide and Edgeworth David in Sydney combined to bring its extent and importance to wide attention (Cooper 2009). Howchin (1901, 1903, 1908) traced the older of two glaciogenic intervals from its type section in the Sturt River gorge near Adelaide over an area of nearly $800 \times 400\ km^2$ (Sprigg 1986), suggesting that 'the ice gathered on a plateau of comparatively low relief' (David 1907*b*). The younger (Marinoan) glaciogenic interval was first recognized by Jack (1913; see also Mawson 1949*b*; Preiss 1987). At first, Howchin's glacial interpretation was greeted skeptically in Australia, but his convincing photographs of striated clasts (Howchin 1908) were accepted without dissent by the Geological Society in London. If Howchin was responsible for much of the legwork in South Australia, David was intrumental in getting the word out overseas (Cooper 2009). Reports of the 10th IGC (Mexico City, 1906) include his 46-page synthesis (David 1907*a*), as well as two shorter papers (one in French) on pre-Cenozoic glacial epochs globally, with emphasis on the Australian glacial record. He also visited India, Great Britain and the USA during his 1906 travels and doubtless raised the awareness of 'Cambrian' glaciation.

In the lower gorges of the Yangtze River in western Hubei Province, South China, Bailey Willis and Eliot Blackwelder described a body of 'till' of probable 'Lower Cambrian' age that carries heterogeneous boulders exhibiting polish and striae of 'unquestionably glacial' origin, based on examination of specimens by Quaternary glacial geologist T.C. Chamberlin (Willis 1904). The Nantuo tillite (Lee & Chao 1924) extends for 1200 km from NE to SW across the South China platform (Lee 1936; Lee & Lee 1940) and represents the younger of two discrete Cryogenian glaciations in the region (Wang *et al.* 1981). Three years later, a similar and likely correlative body (Jiang *et al.* 2003) was described in the Blaini Formation of the Lesser Himalaya, northern India (Holland 1908).

1909–1941: period of globalization

The number of palaeocontinents subject to glaciation in what is now known to be Neoproterozoic time grew from five in 1908 to fourteen in 1937 (Fig. 2.1, Table 2.1), with new finds in eastern and western North America, central and northern Asia, eastern South America, and southern, central and western Africa. By the time of the 17th IGC in Moscow in 1937, Neoproterozoic glaciation was known on every continent except Antarctica.

In 1913, a boulder slate at Big Cottonwood Canyon in the Wasatch Mountains of Utah was interpreted as glacial in origin (Hintze 1913; see also Blackwelder 1932). Earlier, the Squantum 'Tillite' member of the Roxbury Conglomerate in the Boston Basin of eastern Massachusetts was interpreted as glaciogenic (Sayles & LaForge 1910; Sayles 1914). This unit has a checquered history. As a result of miscorrelation, it was long considered to represent the only Carboniferous glaciation in the northern hemisphere. Its glacial origin was repeatedly questioned (e.g. Dott 1961) and later its age. Only after 1980 was its mid-Ediacaran age recognized through geological mapping, geochronology and palynology (Lenk *et al.* 1982; Thompson & Bowring 2000). In Ediacaran time, the Boston Basin was part of the ribbon continent Avalonia–Cadomia, located off North Africa. It did not become part of Laurentia until the late Silurian (Wilson 1966).

In southern Africa, glaciogenic diamictites of Neoproterozoic age were first recognized in the Numees Formation of northwesternmost South Africa and adjacent Namibia (Rogers 1915). 'Anyone familiar with the southern Dwyka [Tillite] who was suddenly put down on the Numees beds would feel sure he was on the Dwyka again' comments Rogers (1915, p. 89). In fact, distal facies of the Dwyka (Carboniferous) rest unconformably on the Numees, 'which resembles the Dwyka tillite of the southern Karoo more closely than the Dwyka beds in the same neighbourhood do' (Rogers 1915, p. 90).

There were no new discoveries between 1915 and 1929 – who knows how many careers were cut down in World War I? – but four book-length syntheses of ancient climates appeared in rapid succession (Brooks 1922, 1926; Köppen & Wegener 1924; Coleman 1926). In the more theoretical of his books, the English meteorologist Brooks (1926) explicitly recognized and attempted to rationalize the existence of two distinct climate states, 'glacial' (eo-Cambrian, late Palaeozoic, late Cenozoic) and 'non-glacial' (most of Phanerozoic time). His names are preferable to the later terms 'greenhouse' and 'icehouse' used synonymously. Köppen & Wegener (1924) is justly famous – despite having never been translated into English – for its detailed treatment of Phanerozoic palaeoclimates in terms of continental drift, as well as for inserting the first of Milankovic's graphs of northern-hemisphere summer insolation calculated for different latitudes over the past 650 000 years. It devotes three pages to the Neoproterozoic ('Kambrium') glacial record. Coleman (1926), who had earlier proposed a glacial origin for the 'slate conglomerates' of Huronian age (early Palaeoproterozoic) in Ontario, Canada, when they were thought to be the oldest known sedimentary rocks (Coleman 1907), devotes a full chapter to glaciation 'at or just before the beginning of the Cambrian' (Coleman 1926). He concludes that the eo-Cambrian glaciation was more extreme than the Pleistocene and possibly equal to the composite late Palaeozoic glaciations, which were known to be diachronous, younging from west to east (Du Toit 1922). The same conclusion was reached by Kulling (1934), who updated the global picture (Table 2.1) at the end of a detailed account of Neoproterozoic stratigraphy in the NE of the Svalbard Archipelago of Arctic Europe.

The pace of new discoveries picked up again from 1929 to 1937 with the recognition of glaciogenic formations in the Katangan Series of southern Zaire (Beetz 1929), beneath the Bambuí Group on the São Francisco craton of central Brazil (de Moraes Rego 1930; see also Isotta *et al.* 1969), on the Yenisey Ridge at the western margin of the Siberian craton (Nikolaev 1930), in the Taoudeni Basin of the West African craton (Baud 1933; Furon 1933) and in the central Tien Shan of NW China (Norin 1937).

A major symposium on Palaeozoic and Pre-Cambrian climates was held at the 17th IGC in Moscow in 1937 (published in 1940 as Report 6). Ten substantial papers describe Neoproterozoic glacial deposits in different parts of Asia, Africa, Australia, Europe and North America. Highlights include first-hand accounts of various pre-Dwyka glacial deposits throughout southern Africa (Gevers & Beetz 1940), the vast extent of the Nantuo Tillite in South China (Lee & Lee 1940) and the existence of glacial deposits in central Africa close to the equator (Davies 1940; Robert 1940). A consensus emerged that 'eo-Cambrian' glaciation was matched only in the late Palaeozoic, and that invertebrate animals arose following retreat of the ice sheets.

However, the rejection of continental drift in the late 1920s following intense debate (Newman 1995; Oreskes 1999) had a chilling effect on pre-Pleistocene palaeoclimate research. If late Palaeozoic glaciation was allowed to remain 'a hopeless riddle', as Wegener put it, what chance was there to understand glacial periods twice their age?

1942–1964: rebutting challenges

The consensus that emerged from the Moscow Congress of 1937 carried over initially into the postwar period. In a 1948 Royal Society lecture in Sydney, Australia, the normally reserved Sir Douglas Mawson (Sprigg 1986) proclaimed that 'the world must then have experienced its greatest Ice Age' and speculated that prolonged genetic isolation during the period of refrigeration contributed to the subsequent Cambrian radiation of life (Mawson 1949*a*). However, Mawson was a fixist and therefore unduly impressed by the evidence for glaciation in equatorial Africa (Beetz 1929; Davies 1939, 1940).

But what were mobilists to do? Plotting 'infra-Cambrian' glacial deposits on Pangaea reconstructions, assuming them to be closer to reality than the present geography, still left some deposits close to the palaeoequator (cf. Harland 1964*a*, *b*; Harland & Rudwick 1964). On the other hand, if the continents were tightly clustered, they might conceivably have crossed the poles diachronously in infra-Cambrian time due to polar wandering and/or continental drift. In that case, the extent of glacial deposits could far exceed the extent of ice at any time. What was needed were means of determining the palaeolatitudes of individual deposits, so that the ice extent inferred would not depend on correlation.

Palaeomagnetism and the meridional extent of Neoproterozoic ice sheets

In the early 1950s, statistical corroboration of the geocentric–axial–dipole hypothesis by Jan Hospers (Frankel 1987; Irving 2008) had opened the way for palaeomagnetic and palaeoclimatic testing of polar wandering and continental drift (Irving 1956, 1959; Runcorn 1961). Wegener and Du Toit were soon vindicated, drift was unavoidable, and the revolution in the interpretation of ocean basins followed in 1962–1967. Palaeomagnetism offered a means by which the palaeolatitudes of ancient glacial deposits could be determined. Irving (1957*a*) had shown that the average magnetic inclination in Carboniferous glacial varves in Australia is subvertical, implying high-latitude deposition, consistent with Wegener's (1912, 1929) assumption based on their glacial origin. Preservation of primary (depositional) natural remnant magnetization in red clastic sediments of Precambrian age having been demonstrated (Irving 1957*b*; Irving & Runcorn 1957), it fell to Harland & Bidgood (1959) to first attempt this method in infra-Cambrian glacial deposits. Their preliminary results (from a small subset of samples) implied equatorial palaeolatitudes for glaciogenic formations in southern Norway and East Greenland. Harland's palaeomagnetic work lacked the sophistication and reach of Runcorn's group three miles to the west (there was little contact between the two Cambridge University laboratories) and it would be 27 years before a truly reliable palaeomagnetic result was obtained for a Neoproterozoic periglacial marine formation (Embleton & Williams 1986).

Palaeomagnetists have long been interested in the palaeolatitudes of climate-sensitive sedimentary facies (e.g. thick carbonates, evaporites, redbeds, aeolianites, tillites) as a test of the hypothesis that the time-averaged geomagnetic field closely approximated a geocentric axial dipole over Phanerozoic and Proterozoic times (Blackett 1961; Opdyke 1962; Briden & Irving 1964; Briden 1970; Evans 2006). Thick shallow-water carbonates are today found only equatorwards of 35° latitude (Rodgers 1957) and their zonal distribution did not change between glacial and non-glacial periods of the Phanerozoic (Briden 1970; Ziegler *et al.* 1984; Opdyke & Wilkinson 1990; Witzke 1990; Kiessling 2001). There are two reasons for this. First, the saturation state of seawater with respect to calcium carbonate or dolomite varies with the relative, not the absolute, temperature. Second, in glacial periods, higher alkalinity fluxes due to glacial erosion in high latitudes are compensated by reduced weathering rates in the cooler tropics. The occurrence of glacial marine strata directly above or sandwiched between thick carbonate successions in different areas (e.g. Harland & Wilson 1956; Ziegler 1960; Katz 1961; Martin 1965*a*) suggested that, unless the glaciogenic and carbonate strata are everywhere separated by stratigraphic gaps of large magnitude, tidewater glaciers existed in the warmest parts of the surface ocean, consistent with nascent palaeomagnetic evidence for low palaeolatitudes (Girdler 1964). If glaciers occurred in relatively warm areas, including marine platforms where no mountains existed, higher latitudes and elevations must also have been glaciated. One is compelled to infer glaciation at all palaeolatitudes, independent of any assumption with respect to correlation (Harland 1964*a*, *b*). This is the crux of the Neoproterozoic glacial conundrum.

The Newcastle and Köln palaeoclimate conferences of 1963–1964

The spectre of glaciation at low palaeolatitudes caused some to question the glacial origin of Neoproterozoic 'pebbly mudstones' (Crowell 1957; Schermerhorn & Stanton 1963; Schermerhorn 1974). Since the recognition of turbidity current as a cause of graded bedding (Kuenen & Migliorini 1950; Natland & Kuenen 1951), mass-flows loomed as alternatives to glaciers in some areas. This would not, however, account for the disproportionate occurrence of diamictites in late Neoproterozoic times. Martin *et al.* (1985) suggested that mass-flows at low palaeolatitudes might have been triggered by glacioeustatic changes driven by high-latitude glaciation, but late Palaeozoic glaciation did not result in low-latitude diamictites comparable in setting and extent to those of the late Neoproterozoic.

These were among the issues discussed at a pair of palaeoclimate conferences that proved to be a watershed for Neoproterozoic glacial geology (Fig. 2.2). The NATO Palaeoclimates Conference held at Newcastle-upon-Tyne in January 1963 was the first to highlight the topic since the Moscow symposium in 1937. Girdler (1964) reviewed the palaeolatitudes of continents during pre-Mesozoic glacial periods according to existing palaeomagnetic data. The Permo-Carboniferous and Huronian (early Palaeoproterozoic) glaciations occurred at high palaeolatitudes, affirming the data, but middle and low palaeolatitudes were registered for the eo-Cambrian glaciations. There were three possible explanations (Girdler 1964): the (eo-Cambrian) palaeomagnetic data were faulty, the rocks are not glacial in origin, or the whole globe was glaciated. An entire session was devoted to the recognition of glacial sediments and of till-like deposits of non-glacial origin (Schwarzbach 1964*a*; Crowell 1964; Heezen & Hollister 1964). In the Discussion, geologist Wally Pitcher and others pushed back against non-glacial interpretations, citing detailed stratigraphic studies on the Port Askaig boulder beds (Kilburn *et al.* 1965). Harland (1964*a*) provided a global synthesis of Precambrian glacial deposits and their stratigraphic settings, concluding that infra-Cambrian glaciation had extended into the marine environment at low palaeolatitudes, unlike any younger ice age. An early tectonic 'mobilist', his views were informed by extensive fieldwork on the diamictite-bearing Hecla Hoek succession in Svalbard (e.g. Harland & Wilson 1956; Wilson & Harland 1964). He urged that such a glaciation provided an exceptional basis for global temporal correlation. Rudwick (1964) discussed the glacial aftermath as an ecological cradle for the Cambrian radiation. Four months after the Newcastle meeting, another global synthesis of Palaeozoic and Precambrian glaciation appeared, graced with fine images of Neoproterozoic glacial debris from central Africa (Cahen 1963).

In March 1964, Manfred Schwarzbach convened an international symposium on palaeoclimates in Köln, simultaneous with the publication in English translation of his book *Climates* *of the Past* (Schwarzbach 1964*b*). The book has a chapter on eo-Cambrian glaciation, the subject of 7 of 37 papers from the Köln symposium published in the *Geologische Rundschau*. Harland (1964*b*) again summarized the global infra-Cambrian glacial record, and Chumakov (1964, 1992) reviewed the growing number of Precambrian tilloids in the (European) USSR. Spjeldnaes (1964, p. 38) affirmed the glacial origin of strata in different parts of Norway and Svalbard, noting that the eo-Cambrian deposits had been studied 'with the same methods, in the same regions and by the same scientists as have the more easily interpreted Quaternary deposits'.

Perhaps the defining summation of the infra-Cambrian glacial problem from this period is Harland & Rudwick (1964) in the popular science magazine *The Scientific American* – which is still worth reading.

1965–1981: in the wake of the plate tectonic revolution

Plate tectonics revolutionized the Earth sciences, and Neoproterozoic glacial geology was no exception. It freed the distribution of ancient glacial deposits from the assumption of fixed relative positions of the continents, as foreseen by Harland (1964*a*) and (earlier) Wegener (1912), just as the concept of polar wandering (Gold 1955) had freed them from fixed position with respect to the poles. Equally important, plate tectonics provided a basis for making sense of geochemical cycles, and thereby to understand the controls on and changes to the chemical compositions of the oceans and atmosphere (Siever 1968). This opened up the possibility of understanding secular changes in atmospheric CO_2 concentration – the dream of Ebelmen (1845, 1847), Tyndall (1861, 1863) and Chamberlin (1898, 1899) – and consequent changes in climate through 'greenhouse' radiative forcing. This led to renewed interest during the period 1965–1981 in certain chemical sediments characteristically associated with Neoproterozoic glaciogenic deposits (banded iron- and manganese-formations, and post-glacial 'cap' carbonates). Finally, the period saw the publication of an influential textbook (Sugden & John 1976) that did much to invigorate the study of processes of glacial erosion, sedimentation and landscape development.

Centenary of Thomson's (1891) study of the Port Askaig Tillite

There could have been no finer centennial celebration than the publication of Anthony (Tony) Spencer's Geological Society Memoir on *Late Pre-Cambrian Glaciation in Scotland* (Spencer 1971). It remains the gold standard in Neoproterozoic glacial sedimentology. The centenary was also marked by papers symbolizing the disunity over the hot-button issue of diachronous or synchronous glaciation (Crawford & Daily 1971; Dunn *et al.* 1971). If the continents were tightly clustered, it was conceivable that continental drift or polar wandering could move them all at different times through the polar region, as was the case for Gondwanaland in the Palaeozoic (Du Toit 1922; Wegener 1929; Crowell & Frakes 1970). However, drift alone could not account for Palaeozoic glaciations, because there were long stretches of Palaeozoic time when Gondwanaland was situated over the South Pole yet no large ice sheets existed.

Associated chemical sediments: banded iron- and manganese-formations and cap carbonates

Sedimentary iron- and manganese-formations (BIFs) were the focus of worldwide mineral exploration for post-World War II reconstruction. Major deposits in SW Brazil (Dorr 1945; Almeida 1946; Walde *et al.* 1981), NW Canada (Ziegler 1960; Young 1976; Yeo 1981), Namibia (Martin 1965*a*) and Australia (Whitten 1970) turned out to be intimately associated with Cryogenian glaciogenic diamictites. None is closely associated with volcanics. They are the only large-scale BIFs in the stratigraphic record younger than the Palaeoproterozoic (1.9 Ga) deposits around the Superior craton of North America. Martin (1965*b*) suggested that 'this peculiar combination of sediments' might be attributed to 'oxygen deficiency in stagnating bottom waters caused by an ice cover' (p. 116).

The presence of a generally rather thin but remarkably continuous layer of carbonate, directly above Neoproterozoic diamictites, was occasionally noted by early workers (David 1907*b*; Norin 1937; Robert 1940; Mawson 1949*b*). These descriptions now become more specific, noting that the 'capping dolomite' is a distinctive pink or less commonly cream colour, invariably laminated or thin-bedded, and persists over basement highs where the diamictite cuts out (Dow 1965; see also Biju-Duval & Gariel 1969; Dunn *et al.* 1971; Rankama 1973; Deynoux & Trompette 1976; Plummer 1978; Williams 1979). Williams (1979) was the first to report stable isotope values (measured by Karlis Muehlenbachs) for 'cap dolostones' (and the first to use the term), from the Kimberleys of Western Australia. He notes that their $\delta^{18}O$ is similar to other Neoproterozoic dolostones, but their $\delta^{13}C$ is moderately depleted. Assuming the dolomite to be primary or early diagenetic, he concludes that 'abrupt climatic warming at the close of late Precambrian glacial epochs is implied', consistent with the $\delta^{18}O$ data (Williams 1979).

Climate models and the 'white Earth' instability

The geological literature in this period was fixated on the use or misuse of glaciation for the division of Neoproterozoic time (e.g. Harland 1964*a*; Crawford & Daily 1971; Dunn *et al.* 1971; Rankama 1973; Schermerhorn 1977). The cause of glaciation was barely mentioned – Harland (1964*a*) suggested diminished solar forcing and Williams (1972, 1974, 1975) appealed to large variations in the Earth's orbital obliquity (see below).

Climate physicists, however, made a startling discovery when they used radiative enery-balance equations to calculate surface temperatures as a function of latitude and variable solar forcing, with simple parameterizations of meridional heat transport and feedbacks due to snow, ice and clouds. When they reduced solar irradiance by a few percent, surface temperatures fell below freezing everywhere due to runaway snow and ice-albedo feedback (Eriksson 1968; Sellers 1969, 1990; Budyko 1969; see also North 1990). Although decadal to millenial solar variability is only *c.* 0.1%, solar luminosity is thought to have slowly increased by 25–30% since 4.5 Ga due to the production of helium in the Sun's core. The energy-balance calculations were therefore assumed to be in error because they predicted that the Earth should have been permanently frozen: irradiance roughly 25% higher than present is required to overcome the albedo of an ice-covered planet, which reflects over 60% of the sunlight it receives. This problem was a stimulus for the nascent science of climate modelling. However, the 'white Earth' instability (Wetherald & Manabe 1975) turned out to be a robust feature not only of simple energy-balance models but also of most atmospheric general-circulation models.

In 1981, three planetary scientists proposed a solution to the 'white Earth' problem (Walker *et al.* 1981). They appealed to the geochemical cycle of carbon (which is not accommodated in physical climate models because of its timescale of *c.* 10^6 years). CO_2 is supplied to the ocean and atmosphere by metamorphic-volcanic outgassing and is consumed by silicate rock weathering. The latter is temperature-dependent because of the direct effect of temperature on reaction kinetics and also because moisture in wet regions, where most weathering occurs, increases with temperature (Clausius–Clapeyron relationship). The temperature-dependence of weathering provides a negative climate feedback (abiotic Gaia) that acts as a natural thermostat. If global temperatures rise (or fall) for any reason, CO_2 is consumed at a faster (or slower) rate, thereby limiting the temperature change. In effect, the feedback slowly adjusts the atmospheric CO_2 concentration to maintain a balance between CO_2 sources and sinks. As solar luminosity increased over geological time, atmospheric CO_2 concentration adjusted downwards, keeping surface temperatures within the range suitable for life. This simple hypothesis not only went a long way towards solving the 'faint young Sun' paradox (Sagan & Mullen 1972), it rationalized long-term climate change while explaining why such changes were limited in magnitude.

In the penultimate paragraph of their classic paper, Walker *et al.* (1981) suggested that silicate weathering feedback had been responsible for averting a 'white Earth'. They added, however, that 'if a global glaciation were to occur, the rate of silicate weathering should fall nearly to zero, and carbon dioxide should accumulate in the atmosphere at whatever rate it is released from volcanoes. Even the present rate of release would yield 1 bar of carbon dioxide in only 20 Ma. The resultant large greenhouse effect should melt the ice cover in a geologically short period of time' (Walker *et al.* 1981). If a 'white Earth' did occur, it would be self-reversing. Accordingly, its occurrence in deep time could not be ruled out *a priori*. This, in a nutshell, is the climatological concept behind the 'Snowball Earth' hypothesis (Kirschvink 1992).

Walker *et al.* (1981) made no reference to ancient glaciation, Neoproterozoic or otherwise. This was not because they were unaware of geology. On the contrary, the 'faint young Sun' problem was brought into focus for Walker by his participation in the Precambrian Palaeobiology Research Group (PPRG) organized by micropalaeontologist J. William Schopf. (Third author Kasting and palaeomagnetist Joe Kirschvink were PPRG participants later in the decade.) By 1981, Harland's great infra-Cambrian glaciation had fallen off the radar screen. Palaeomagnetism, on which high hopes were pinned, had encountered problems. The foremost was the susceptibility of sediments to low-temperature chemical remagnetization (e.g. McCabe & Elmore, 1989). Overcoming these problems would require time-consuming stepwise chemical and thermal demagnetization, studies of magnetic mineralogy, 'field tests' to constrain the age of remnant magnetic components, and more sensitive magnetometers.

Earth's Pre-Pleistocene Glacial Record volume

In 1981, Cambridge University Press published *Earth's Pre-Pleistocene Glacial Record* (Hambrey & Harland 1981), a product of Project 38 (Pre-Pleistocene Tillite Project) of the International Geological Correlation Programme. This sought-after volume contains 58 unified descriptions of Neoproterozoic glaciogenic formations worldwide, and is a tribute to Harland's vision and Hambrey's dedication. It was synthesized in Harland (1983) and Hambrey & Harland (1985). An additional global synthesis of Neoproterozoic glaciogenic deposits was published by Chumakov (1981).

1982–1997: the gathering storm

There were many developments during this period – burgeoning information regarding glaciomarine processes and deposits, widespread use of carbon isotopes as a tool for correlation, acquisition of reliable palaeomagnetic constraints on palaeolatitudes of proximal glaciomarine deposits, increased awareness of the sedimentological peculiarites of post-glacial 'cap' carbonates, and growing interest in causal mechanisms for low-latitude glaciation.

Glaciomarine deposits are more likely than their terrestrial counterparts to be preserved in the geological record because they are less susceptible to erosion, but ice-proximal processes in the marine environment are difficult to study because they occur underwater and often under ice. Nevertheless, much has been gleaned from geomorphic and seismic surveys of polar continental shelves backed up by drilling programmes (e.g. Elverhøi 1984; Powell 1984, 1990; Alley *et al.* 1989; Barrett 1989; Boulton 1990; King *et al.* 1991; Powell & Domack 1995). A growing emphasis on glaciomarine processes and deposits is reflected in many books written during this period (e.g. Molnia 1983; Drewry 1986; Dowdeswell & Scourse 1990; Anderson & Ashley 1991; Hambrey 1994; Benn & Evans 1998; Anderson 1999). Neoproterozoic glacial geology benefited enormously from these developments, although the complexity of facies architecture poses a challenge given the limited three-dimensional exposure of ancient deposits and weak or non-existent constraints on rates of accumulation.

Beginning in the mid-1950s, oxygen isotopes have been used as a basic tool for stratigraphic correlation of Quaternary marine carbonates and for estimating seawater temperatures and global ice-sheet volumes. Unfortunately, Neoproterozoic carbonates do not preserve their original mineralogy, and their oxygen isotope compositions are seriously compromised by aqueous diagenesis. However, carbonate rocks buffer the isotopic composition of the relatively low concentration of carbon in aqueous solution. Consequently, the carbon isotopic compositions of carbonate rocks resist alteration. Beginning in the mid-1980s, carbon isotopes were increasingly used for stratigraphic correlation of Neoproterozoic carbonates and to a lesser extent for organic-rich shales (Knoll *et al.* 1986; Knoll & Walter 1992; Kaufman *et al.* 1997). Studies show that Neoproterozoic glaciogenic formations are typically bracketed by negative isotopic excursions. The excursions are reproduceable and approximately five times larger in magnitude than spatial or depth variations in the modern ocean. Because atmospheric CO_2 concentrations were likely higher in the Neoproterozoic than today because of the dimmer Sun, a larger reservoir of dissolved inorganic carbon should have damped spatial and depth variations of $\delta^{13}C$ in Neoproterozoic oceans. Therefore, the observed isotopic excursions are most easily explained as secular variations. Their regular occurrence directly before and after glaciogenic sedimentation suggests that glaciation was roughly synchronous in many areas, even if not every isotopic excursion is associated with glaciation. For those who accepted these basic principles, the arguments for diachronous glaciation receded as the body of Neoproterozoic carbon isotope data grew.

The first robust palaeomagnetic support for glaciation at sea level in low palaeolatitudes came from the Elatina Formation in South Australia (Embleton & Williams 1986). The formation is a glaciomarine unit comprised of reddish sandstone, siltstone and diamictite (Lemon & Gostin 1990; Williams *et al.* 2008), and a 10-m-thick interval of rhythmically laminated siltstone has been the subject of repeated palaeomagnetic studies. The palaeopole determined by Embleton & Williams (1986) implies a palaeolatitude of *c.* 5° and, although no field test was performed, the result is considered reliable because the natural remanent magnetization is stably carried by detrital haematite. At the time of the study, the laminations were considered to be annual varves, and the rhythmic bundles of 12–14 laminae were thought to represent sunspot cycles (Williams 1981; Williams & Sonett 1985). Later, they were reinterpreted as tidal rhythmites (Williams 1989), supporting a glaciomarine origin for the associated diamictites. Field tests were subsequently carried out using soft-sediment folds (Schmidt *et al.* 1991; Schmidt & Wilhlliams 1995) and polarity reversals (Sohl *et al.* 1999), which confirmed the primary nature of the remanent magnetization and the palaeolatitude of 7.5° or less. A somewhat larger palaeolatitude of *c.* 15° was obtained in a recent palaeomagnetic study (Raub & Evans

2006) of the cap dolostone (Nuccaleena Formation) above the Elatina Formation, implying a certain amount of inclination flattening during compaction of the Elatina siltstone, but this hardly alters the thrust of the result because temperatures are broadly uniform across the tropics.

Cap carbonates

Arguably the first person to fully appreciate that the basal Ediacaran (Marinoan) cap carbonate was not only 'an important stratigraphic marker' but also 'a perplexing and paradoxical lithostratigraphic unit' (Table 2.2) was Aitken (1991), a leading Rocky Mountain carbonate stratigrapher. Concerning the cap dolostone atop the younger glaciogenic unit (Stelfox Member of the Icebrook Formation) in the Mackenzie Mountains of NW Canada, Aitken (1991) stressed that the aphanitic microcrystalline dolostone was not deposited as fine-grained mud, but as silt and sand-sized peloids. Ubiquitous subparallel lamination is defined by oscillatory variation in peloid size. His photographs show well-sorted, coarse-grained peloids in layers only one or two peloids deep. Where coarse peloids form the base of a layer, they commonly rest on facets, apparently resulting from abrasion. The layers are not laterally continuous, and low-angle cross-lamination is ubiquitous. The lamination is clearly mechanical in origin, interspersed locally with microbially stabilized masses of micropeloidal stromatolite (James *et al.* 2001).

Aitken (1991) argued that the intrastratal 'tepees' for which the unit was informally named (Eisbacher 1981) are not true peritidal tepees as defined by Assereto & Kendall (1977; see also Kendall & Warren 1987), which result from layer-parallel expansive growth associated with cementation under alternating vadose and phreatic conditions. The cap 'tepees' are not associated with breccias or void-filling cements, as are peritidal tepees, and in plan view their crestlines are always linear and parallel, not polygonal like true tepees (Assereto & Kendall 1977). Bundles of laminae may exhibit onlap or offlap relations on 'tepee' flanks. These metre-scale trochoidal bedforms were later analysed as giant wave ripples (Allen & Hoffman 2005).

Large-scale stromatolites, within which are spaced, micrite- or cement-filled tubes or gutters that maintain a palaeovertical ('geoplumb') orientation irrespective of the inclination of laminations in the host stromatolite, were described from cap dolostones in California (Cloud *et al.* 1974; Wright *et al.* 1978), Namibia (Hegenberger 1987) and later in Alaska, Brazil, Canada and Mongolia (Fig. 2.3).

Aitken (1991) identified crystal fans and pseudo-stromatolites in limestone overlying the cap dolostone, correctly interpreting them as pseudomorphic after aragonite sea-floor cements. His photograph of 'blades of calcite' is from a unique layer of sea-floor barite ($BaSO_4$) cement, variably calcitized, which occurs in the top <10 cm of the cap dolostone for >150 km along a strike southeastward from the location of Aitken's (1991) photograph. Thus, cap dolostones and overlying limestones display a panoply of idiosyncratic features (Fig. 2.3) that generally occur in a consistent vertical sequence (Hoffman *et al.* 1987).

Aitken (1991) offered no genetic explanation for cap carbonates. Eyles (1993) suggested that they represent detrital rock 'flour', as proposed earlier for carbonate layers *within* glaciogenic diamictites (Fairchild 1983). A detrital origin could hardly

Table 2.2. *Idiosyncratic sedimentary and early diagenetic features in cap dolostones*

Palaeocontinent	Cap dolostone (ref.)	SCC	TPB	DGB	PEL	LAC	GWR	TBS	SFB
Amazonia	Mirassol d'Oeste (1)	–	–	–	√	√	√	√	–
Arabia	Hadash (2)	–	–	–	–	–	–	–	–
Arctic Alaska	Nularvik (3)	–	–	–	√	√	√	√	–
Australia	Mount Doreen (4)	√	–	–	√	√	√	–	√
Australia	Nuccaleena (5)	√	–	–	–	–	√	–	–
Australia	Cumberland Creek (6)	–	–	–	√	√	√	–	–
Australia	Lower Ranford (7)	–	–	–	–	√	√	–	–
Baltica	Lower Nyborg (8)	–	–	–	–	–	–	–	–
Congo	Keilberg (9)	√	–	–	√	√	√	√	–
Congo	Calcaire Rose (10)	–	–	–	–	–	–	–	–
Congo	C1 (11)	–	–	–	–	–	–	–	–
India	Upper Blaini (12)	–	–	–	–	–	–	–	–
Kalahari	Dreigratberg (13)	–	–	–	√	√	√	√	–
Laurentia	Ravensthroat (14)	√	–	–	√	√	√	√	√
Laurentia	Noonday (15)	–	–	–	√	√	√	√	–
Laurentia	Lower Canyon (16)	–	–	–	–	√	–	–	–
Laurentia	Lower Dracoisen (17)	–	–	–	√	√	√	–	–
Laurentia	Cranford (18)	√	√	–	–	√	–	–	–
Laurentia	Hard Luck (19)	√	√	–	–	–	–	–	–
Mongolia	Ol (20)	√	–	–	√	√	√	√	–
Mongolia	Baxha (21)	–	–	√	–	–	–	–	√
South China	Lower Doushantuo (22)	√	√	√	√	–	–	–	–
Tarim	Lower Zhamoketi (23)	–	–	–	√	–	–	–	–
West Africa	Oued Djouf (24)	√	√	–	–	–	–	–	–
West Africa	Amogjar (25)	–	√	√	√	–	–	–	–
West Africa	Mid Sud-Banboli (26)	√	√	√	√	√	–	–	–

Abbreviations: SCC, sheet-crack cements; TPB, tepee breccia; DGB, diagenetic barite; PEL, peloids; LAC, low-angle cross-laminae; GWR, giant wave ripples; TBS, tube biostrome; SFB, seafloor barite.

References: (1) Nogueira *et al.* (2003); (2) Allen *et al.* (2004); (3) Macdonald *et al.* (2009*b*); (4) Kennedy (1996); (5) Plummer (1978); (6) Calver & Walter (2000); (7) Corkoran (2007); (8) Edwards (1984); (9) Hoffman *et al.* (2007); (10) Cahen & Lepersonne (1981); (11) Cahen (1950); (12) Kaufman *et al.* (2006); (13) Macdonald *et al.* (2011); (14) James *et al.* (2001); (15) Corsetti & Grotzinger (2005); Corsetti & Kaufman (2005); Wright *et al.* (1978); (16) Hambrey & Spencer (1987); (17) Halverson *et al.* (2004); (18) McCay *et al.* (2006); (19) F. A. Macdonald, pers. comm.; (20) Macdonald *et al.* (2009*a*); (21) F. A. Macdonald & D.S. Jones, pers. comm.; (22) Jiang *et al.* (2006); (23) Xiao *et al.* (2004); (24) Bertrand-Sarfati *et al.* (1997); (25) Shields *et al.* (2007); (26) Nédélec *et al.* (2007).

account for the systematic variations in $\delta^{13}C$ observed in cap dolostones (Kennedy *et al.* 1998; Hoffman *et al.* 2007). Kaufman *et al.* (1993) and Grotzinger & Knoll (1995) related cap carbonates to the overturn of alkalinity-charged, isotopically depleted, bottom water after a lengthy period of ocean stagnation. This suggestion requires that primary production be maintained in the absence of upwellings, which is the overwhelmingly predominant nutrient flux in the modern ocean. Upwellings would not be necessary if organic matter did not settle, but then alkalinity would not build up in the deep nor would a vertical isotopic gradient develop. This and the physical implausibility of prolonged (>10 000 years) ocean stagnation (in the absence of an ice cover) make the overturn hypothesis unconvincing, despite its popular appeal.

Kennedy (1996) proposed that cap carbonates represent non-skeletal analogues of carbonates deposited (in low latitudes) during times of Quaternary deglaciation according to the 'coral reef' hypothesis (Berger 1982; Opdyke & Walker 1992). This results from the hypsometry of ocean basins. When sea level is lowered due to ice-sheet growth, the area of shallow shelves and inland seas is disproprotionately reduced. These areas are favoured for carbonate burial because the sediment never encounters undersaturated deepwaters. When these areas are lost, all carbonate production occurs in the open ocean and the alkalinity of deepwaters increases. Upon deglaciation, alkalinity-rich waters flood the shelves, depositing carbonate and releasing CO_2 to the atmosphere. The process can account for much of the glacial–interglacial variation in pCO_2 (Opdyke & Walker 1992) and thus provides a positive climate feedback. This mechanism must have contributed to cap carbonates, but the estimated average thickness of carbonate produced (Ridgwell *et al.* 2003) is only approximately one-tenth of the average thickness of cap dolostones of 18.5 m (Hoffman *et al.* 2007, Table 2.1). The coral reef hypothesis does not explain why cap carbonates are only associated with the ultimate deglaciation, and do not accompany glacial–interglacial cycles within the glacial period where these are recognized (Leather *et al.* 2002; Allen *et al.* 2004). If cap carbonates were eroded during glacial readvances, they would be present as clasts in glacial deposits, which they are not.

Nearly 70 papers concerning cap carbonates appeared in the decade after 1997. This reflects the central but contentious role they have assumed in the ongoing controversy over the nature of Cryogenian glaciation. Arguably they are the most richly enigmatic horizons in the entire stratigraphic record.

Causative theories for Neoproterozoic glaciation

Walker *et al.* (1981) provided the conceptual basis for a self-reversing global glaciation from the point of the initial 'white Earth' (ice-albedo) instability. They did not speculate on how the climate ever reached that point because they credited silicate-weathering feedback with its avoidance. However, a number of causative theories (not mutually exclusive) were proposed between 1971 and 1997 to explain Neoproterozoic glaciation (Table 2.3).

Carbonate burial. Roberts (1971, 1976) was struck by the fact that the Cryogenian glaciation was preceded by thick successions of shallow-water carbonate strata in the North Atlantic (Greenland, Svalbard and the British Isles), western North America and central Africa. He postulated a period of anomalous carbonate burial in which atmospheric CO_2 was sequestered, producing an 'anti-greenhouse' effect. He did not give any reason why this might have occurred. It would require a rise in the global rate of silicate weathering relative to the rate of CO_2 outgassing (see below; *Continental distribution* and *Continental break-up*).

Table 2.3. *Causative theories for Neoproterozoic glaciation*

Astronomical theories	
1. Solar variation	Harland (1964*a*)
3. Large obliquity	Williams (1972, 1975), Jenkins (2000), Donnadieu *et al.* (2002)
5. Ice-ring collapses	Sheldon (1984)
13. Impact ejecta	Bendtsen & Bjerrum (2002), Fawcett & Boslough (2002)
4. Giant molecular clouds	Pavlov *et al.* (2005)
Oceanographic theories	
2. Carbonate burial	Roberts (1971)
7. Ocean stagnation	Kaufman *et al.* (1993), Grotzinger & Knoll (1995)
10. Hypsometric effect	Kennedy (1996), Ridgwell *et al.* (2003)
11. Organic burial	Kaufman *et al.* (1997)
Geodynamic theories	
6. Continental distribution	Marshall *et al.* (1988), Worsley & Kidder (1991), Kirschvink (1992), Schrag *et al.* (2002)
9. Continental break-up	Eyles (1993), Young (1995*a*), Donnadieu *et al.* (2004*a*)
15. Basalt weathering	Goddéris *et al.* (2003)
16. True polar wander	Li *et al.* (2004)
Biological theories	
8. Biocatalysed weathering	Carver & Vardavas (1994), Lenton & Watson (2004), Kennedy *et al.* (2006)
12. Methane destruction	Pavlov *et al.* (2000, 2003), Catling *et al.* (2001), Claire *et al.* (2006)
14. Methane substitution	Halverson *et al.* (2002), Schrag *et al.* (2002)

Large orbital obliquity. The Australian sedimentologist George E. Williams always had an eye on 'the big picture' and possessed an early appetite for planetary orbital mechanics. Impressed by the prevalence of varves (i.e. rhythmic lamination, supposedly seasonal) and other seasonal indicators such as polygonal sand wedges associated with Neoproterozoic glaciogenic strata (Chumakov 1968; Spencer 1971), he proposed that the Earth's obliquity (i.e. the angle between the equatorial and ecliptic planes) oscillated between large (e.g. Neoproterozoic) and small (e.g. Phanerozoic) values (Williams 1972). When the angle was large, the seasonal cycle was strong everywhere; when small, the seasons did not differ greatly in low latitudes (except in strongly monsoonal areas). In addition, as the angle increased, the meridional insolation gradient declined and actually reversed (i.e. greater annual insolation at the poles than at the equator) whenever the angle exceeded 54°. Williams (1972, 1974, 1975) inferred that when the obliquity was very large, low latitudes were subject to glaciation preferentially. As low latitudes cover more area than high latitudes, there should be more ice (and higher planetary albedo) during periods of low-latitude glaciation than when ice is limited to the poles. Accordingly, glacial periods occur when obliquity is large and when there is strong seasonality everywhere (contrary to the Wegener–Milankovic hypothesis that ice-sheet growth is favoured by cool summers and low seasonality). Under large obliquity, glaciation at the poles is precluded by intense summertime insolation.

The large obliquity hypothesis was strengthened by the discovery of 2.5-m-deep sand wedges associated with the Elatina Formation (Williams & Tonkin 1985), which, combined with palaeomagnetic evidence (Embleton & Williams 1986), indicates strong seasonality at low palaeolatitude (Williams 1993; but see Maloof *et al.* 2002). In the Phanerozoic, deep periglacial sand wedges occur only in middle and high latitudes, where seasonal forcing is strong (Lachenbruch 1962). With large obliquity, seasonality is strong at all latitudes.

Like any theory, large obliquity was not without difficulties. Although a large (and chaotic) obliquity could have been imparted on the early Earth by stochastic accretion (Laskar & Robutel 1993), obliquity cannot oscillate between large and small angles because of 'entrapment' at small angles by gravitational attraction between the Moon and the Earth's equatorial bulge (Laskar *et al.* 1993). Therefore, the theory was revised such that obliquity was persistently large until Ediacaran times, when it was rapidly reduced to small angles before the Cambrian (Williams 1993, 2000). This proved difficult to replicate in models or justify from orbital theory (Néron de Surgy & Laskar 1997; Williams *et al.* 1998; Hoffman & Maloof 1999; Pais *et al.* 1999; Donnadieu *et al.* 2002; Levrard & Laskar 2003). Moreover, large obliquity over most of Precambrian time is at odds with a growing body of palaeomagnetic evidence that pre-Ediacaran evaporites and carbonate-dominated successions formed at latitudes inconsistent with a reversed climatic gradient (Park 1994; Buchan *et al.* 2001; Evans 2006; Maloof *et al.* 2006). However, decisive falsification of the large obliquity hypothesis, glaciation at high palaeolatitude, has yet to be observed.

Ice-ring collapses. Richard P. Sheldon was a sedimentary phosphorite specialist with the United States Geological Survey who developed an 'outlandish hypothesis' (Sheldon 1984) for Neoproterozoic glaciation following a field excursion in northern Mongolia, where phosphorites, carbonates and glacial marine strata occur in close stratigraphic proximity. He speculated that ice rings denser than those of Saturn orbited the Archaean Earth. In the Proterozoic, sublimation thinned the rings and they became more discrete. As the Moon pulled away from the Earth, their orbits slowly decayed and they successively entered the atmosphere. As each ring approached the Earth, its shadow caused an ice age. Upon entering the atmosphere and before the next ring approached, the shadow was lost and greenhouse gases were added, causing a warm interglacial. Once the final ring had fallen, the tropics were shadowless for the first time. Ocean overturning driven by the steepened climatic gradient led to phosphogenesis and organic diversification. Sheldon (1984) noted that seasonality would be greatest in glacial times, consistent with Williams' (1993) observation, because the rings cast shadows only on the winter hemisphere. Again, this is contrary to the Wegener–Milankovic theory of ice ages, which posits that ice sheets grow when summers (not winters) are cool and seasonality is weak (Köppen & Wegener 1924; Milankovic 1941).

Continental distribution. Atmospheric $p\mathrm{CO_2}$ will adjust downwards in response to an equatorward migration of the continents, because weathering rates are greatest in the tropics (Marshall *et al.* 1988; Worsley & Kidder 1991). Global cooling is therefore consistent with sedimentological and palaeomagnetic evidence for an unusual preponderance of low-latitude continents in the Neoproterozoic, 'a situation that has not been encountered at any subsequent time in Earth history' (Kirschvink 1992). Equatorward displacement of the continents also changes the planetary albedo: the tropical ocean is a strong absorber of sunlight, unlike land areas (minimally vegetated in the Neoproterozoic) and the fog-bound polar oceans, which are relatively good reflectors (Kirschvink 1992). Moreover, a positive albedo feedback would result from the enlargement of tropical land area that would accompany any glacioeustatic fall in sea level (Kirschvink 1992).

A palaeomagnetist, Kirschvink (1992) accepted the low-latitude result for the Elatina glaciation (Embleton & Williams 1986) but rejected the large-obliquity hypothesis because of the co-occurrence of Neoproterozoic glacial deposits and thick carbonate-dominated successions. If the meridional climatic gradient were reversed, the carbonate belts would move from low latitudes to the poles, 'where the glaciers (in Williams' model) should not encounter them' (Kirschvink 1992). He was therefore compelled to assume that if ice sheets reached sea level close to the equator, higher latitudes must have been glaciated as well, wherever ablation did not exceed precipitation. He envisaged an ocean largely covered by pack ice, perhaps with 'warm tropical "puddles" in the sea of ice, shifting slightly from north to south with the seasons' (Kirschvink 1992). At the 1989 PPRG meeting at UCLA (Maugh 1989), he proposed that the 'white Earth' disaster had actually occured in the Neoproterozoic, perhaps repeatedly, and that each pan-glacial period was abruptly terminated when the slow build up of CO_2 reached a critical level, as deduced by Walker *et al.* (1981). He attributed the deposition of banded iron-formation during Neoproterozoic glaciation to ocean stagnation and deepwater anoxia, as previously proposed by Martin (1964*b*), and he predicted that the abrupt climate switches accompanying glaciation and deglaciation should be represented in widely separate areas by lithologically similar strata, a result of the global scale of the climatic fluctuations (Kirschvink 1992). He called the glacial state a 'Snowball' Earth (Maugh 1989) to highlight the central role of planetary albedo in the phenomenon. It is an evocative description of how the late Cryogenian Earth might have looked from outer space.

Impact ejecta. Impact ejecta figured in discussions of Neoproterozoic glaciation in two completely different ways. They were proposed as an alternative emplacement mechanism for diamictite, implying that the low-latitude and carbonate-associated glacial problem was a chimera (Oberbeck *et al.* 1993; Rampino 1994). This was a small reenactment of the pebbly mudflow challenge of 1957–1974. Impact ejecta aprons do in fact share many diagnostic features with glaciogenic diamictites (e.g. unsorted debris, faceted and striated erratic stones, outsized dropstones). However, continuous ejecta blankets extend only about one crater-radius from the crater rim, regardless of crater size (Melosh 1989). Therefore, an improbably large flux of big impacts (roughly one Cretaceous–Tertiary sized impact every 100 ka for 200 Ma) would be required to account for the distribution of Neoproterozoic diamictites, given the low probability of their exposure at the Earth's surface today.

It was later proposed on the basis of modelling experiments that large impacts could have caused Neoproterozoic glaciation (Bendtsen & Bjerrum 2002; Fawcett & Boslough 2002). The models suggest that short-term (<10 year) shielding of sunlight by ejecta from a Cretaceous–Tertiary sized impact could cause the ocean surface to freeze over if seawater was as cold as today, unlike the Cretaceous ocean, which was $8–12^{\circ}$ warmer than present.

Ocean stagnation. As the patterns of C and Sr isotope variations associated with Neoproterozoic and Phanerozoic glaciations are distinct (Kaufman *et al.* 1993), the cause of glaciation might also be different. Kaufman *et al.* (1993) and Grotzinger & Knoll (1995) attempted to genetically link the isotopic patterns, glaciation and the associated Fe- and Mn-ore formations and 'cap' carbonates in a conceptual model involving prolonged ocean stagnation followed by overturn. They hypothesized that when the ocean was stagnant, organic export and subsequent respiration in anoxic deep waters would progressively enrich the surface waters in ^{13}C, while simultaneously creating an isotopically depleted, bicarbonate-enriched reservoir at depth (Kaufman *et al.* 1993; Grotzinger & Knoll 1995). In their model, the attendant transfer of CO_2 from the atmosphere to the deep ocean contributes to global cooling. Ultimately, the growth of sea ice triggers the formation of cold, saline, deep water, and the meridional overturning circulation is reestablished. Upwelling of alkalinity-laden deepwater leads to the precipitation of isotopically depleted cap carbonates and the simultaneous release of CO_2 to the atmosphere, which melts the ice (Kaufman *et al.* 1993; Grotzinger & Knoll 1995).

The parsimony of the model temporarily outshone its flawed foundation. Short of covering the ocean with ice, physical stagnation (as distinct from dynamic stratification) on geological timescales is implausible because the energy driving the upwelling flux (the rate-limiting step in the overturning circulation) derives from winds and tides (Wunsch 2002). Winds and tides are not so easily turned off. Moreover, if the surface ocean was ever deprived of nutrients upwelled from depth, primary production would crash, depriving the deep of its alkalinity pump and the isotopic gradient of its driver. To its credit, the overturn hypothesis at least lined up all the key observations in the same viewfinder: glaciation, isotopes, iron formations and cap carbonates.

Biocatalysed weathering. Carver & Vardavas (1994) constructed a time-dependent model of the Earth's mean surface temperature controlled by the geochemical cycle of carbon (Walker *et al.* 1981). They parameterized CO_2 outgassing as having declined rapidly before and more slowly after 3.5 Ma. The solar flux rose almost linearly, and the consumption of CO_2 through silicate weathering was shaped by the rapid growth of 90% of the present continental crust between 3.5 and 2.0 Ga, and by stepwise increases in weathering efficiency (i.e. the rate of CO_2 consumption for any given pCO_2) resulting from biological colonizations of the land (Schwartzman & Volk 1991). The first colonization (microbial) is assumed to have occurred before 3.5 Ga. The second (organisms unspecified) supposedly occurred between 1.2 and 0.7 Ga, and the third in the middle Phanerozoic in response to the rise of vascular plants. The observed enrichment in ^{13}C of most pre-Ediacaran carbonates (e.g. Knoll *et al.* 1986) is cited in support of the middle (and largest) step in weathering efficiency (Carver & Vardavas 1994), although the isotopic pattern requires an increase in fractional *organic* burial as opposed to overall carbon burial. Increased Neoproterozoic biocatalysed weathering receives some support from micropalaeontology (Horodyski & Knauth 1994), molecular divergence analysis (Heckman *et al.* 2001) and clay mineralogy (Kennedy *et al.* 2006), but remains controversial. The modelled temperature curves (Carver & Vardavas 1994) feature two relatively cold intervals in Earth history, early Palaeoproterozoic in response to continental growth and mid-Neoproterozoic in response to the prescribed increase in biocatalysed weathering.

Continental break-up. Supercontinents tend to be dry because most land is far from the ocean. Upon fragmentation, new continental margins are created and land overall is brought closer to the source of moisture. As silicate-weathering is catalysed by moisture as well as by temperature (Walker *et al.* 1981), continental break-up should result in lower pCO_2 and a colder global climate. Cooling will be most severe if rifting occurs in the tropics, and least so for rifting in the polar regions where chemical weathering rates are low. Continental break-up in tropical Pangaea was followed by the coolest period (Late Jurassic–Early Cretaceous) of the Mesozoic era (Frakes *et al.* 1992); break-up in the polar North Atlantic by the warmest period (Early Eocene) of the Cenozoic (Zachos *et al.* 2001). This is consistent with Cryogenian cooling, related to the break-up and dispersal of equatorial Rodinia beginning *c.* 800 Ma (Li *et al.* 2008). These ideas have recently been tested with a simplified atmospheric general circulation model coupled to a model of the geochemical cycle of carbon (Donnadieu *et al.* 2004*a*, *b*).

It is therefore difficult to accept the argument that Neoproterozoic glaciation could be explained in terms of a close spatial (as well as temporal) association with rift valleys and rifted margins, without recourse to extreme climate states (Eyles 1993; Young 1995*a*; Eyles & Januszczak 2003). [In Reply to a Comment, Young (1995*b*) wrote that he did not intend to imply a genetic relationship, but then his paper should have been titled, 'Was the *preservation* of Neoproterozoic glacial deposits on the margins of Laurentia related to the fragmentation of two supercontinents?'.] The Late Quaternary glacial maxima were arguably as cold as any time in the Phanerozoic, but moraines on the mountains of the East African rift system do not extend below *c.* 3.5 km above sea level (Osmaston 2004). In the Red Sea rift, only a handful of the highest peaks (>4.0 km above sea level) on the Ethiopean Dome were glaciated at the Last Quaternary Maximum (Osmaston 2004). For Neoproterozoic ice sheets to have reached sea level at the same latitudes implies a drastically colder climate.

Epilogue: the long road to consensus

Figure 2.2 clearly shows that publications concerning Neoproterozoic glaciation 'took off' in the mid-1990s and possibly peaked in 2007. There is little doubt that the galvanizing factor was the Snowball Earth hypothesis (Kirschvink 1992; Hoffman & Schrag 2002). In the first four years after its publication, Kirschvink (1992) had been cited only three times, and favourably only once (Klein & Beukes 1993). However, as soon as it was advocated prominently (Hoffman *et al.* 1998; Hoffman & Schrag 2000) and in the public arena (Walker 2003), the reaction was swift.

This is not the place to review all the papers that have appeared since 1998: more time and more space are needed. Opinions about the nature of Neoproterozoic glaciation are now at a stalemate. Most workers agree that one, and possibly two, Cryogenian glaciations occurred simultaneously on virtually all palaeocontinents (Evans 2000; Halverson 2006; Hoffman & Li 2009; but see Eyles & Januszczak 2003 for a contrary view). The basic argument is that if ice sheets reached sea level in the warmest areas (i.e. close to the palaeoequator and where thick non-skeletal carbonates accumulated), then colder areas must have been frozen as well. This argument makes no *a priori* assumption about correlation, but correlation follows from the premise.

The stalemate concerns the state of the ocean: was it largely ice-covered as assumed in the Snowball Earth hypothesis (Kirschvink 1992; Hoffman & Schrag 2002), or substantially ice-free as posited in the so-called slushball solution (Hyde *et al.* 2000; Peltier *et al.* 2004; but see Bendtsen 2002; Voigt *et al*. 2011)? Because of subduction, direct evidence from the ocean floor has been eliminated, or survives only from post-Cryogenian (Ediacaran), regional-scale glaciation (Kawai *et al.* 2008). Those who favour the snowball model point to its ability to account for synglacial Fe- and Fe–Mn deposits, post-glacial cap carbonates, and isotopic evidence for highly elevated pCO_2 during and after glaciation (Bao *et al.* 2008, 2009). The slushball solution does not account for these features (Pollard & Kasting 2005). Those who favour the slushball solution point to evidence that the snowball hypothesis is incompatible with the fossil record (Knoll 2003; Xiao 2004; Corsetti *et al.* 2006; Moczydlowska 2008) and also with the sedimentology of glacial deposits (e.g. McMechan 2000; Condon *et al.* 2002; Allen & Étienne 2008). The biotic argument presupposes a knowledge of the limits to survival of species. The argument from sedimentology assumes knowledge of when the deposits were formed because the snowball hypothesis predicts that glaciation was 'polar' in character at the beginning and 'temperate' or 'Alaskan' at the end. Studies of modern ice-sheet stability show that outlet glaciers are buttressed by ice shelves (Dowdeswell *et al.* 2000; De Angelis & Skvarca 2003; Nick *et al.* 2009): ice-shelf removal triggers ice-sheet deflation, with potential for ice-sheet collapse. On a Snowball Earth, the ice shelf is global. Ice-shelf removal, the first step in snowball deglaciation, should trigger rapid deflation and collapse of low-latitude ice sheets. Much of the glacial sediment record on continental margins will date from this period. These deposits will bear evidence of open water (e.g. iceberg rafting and dumping, wave ripples) because open water then existed. Sedimentology is thus a clumsy means of testing the snowball hypothesis because of uncertainty over which stage of the glacial cycle is stratigraphically preserved (Hoffman 2005). Moreover, sediment fluxes cannot be estimated because chronology is lacking. Even varve chronology (De Geer 1912), from which the duration of the Holocene was estimated

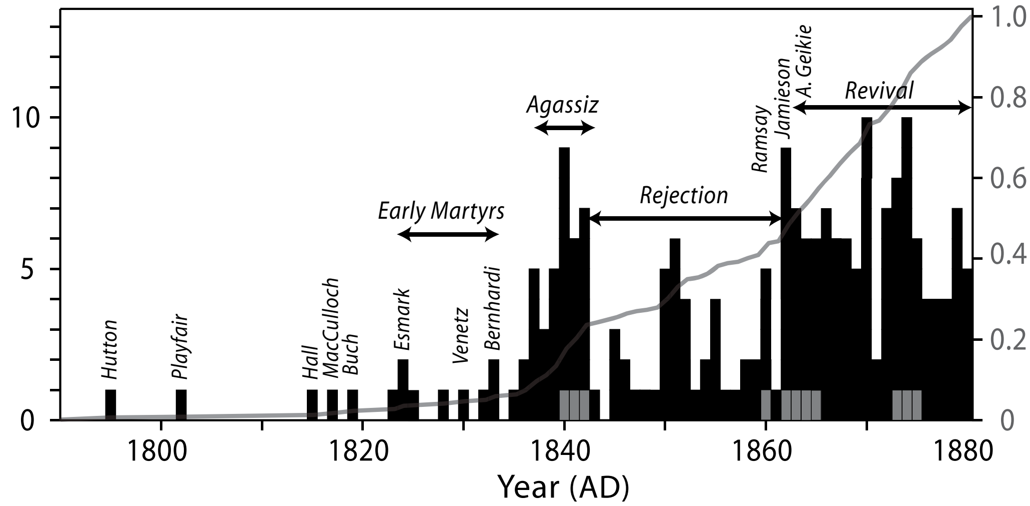

Fig. 2.4. Bibliographic history of the Pleistocene ('Newer Pliocene') glacial controversy. Number of papers (black) and books (grey) per year concerning the 'Great Northern Drift'. *Early Martyrs* refer to the essentially correct but ignored glacial theories of Esmark (1824), Dobson (1925), Venetz (1830) and Bernhardi (1832). *Agassiz* refers to the time of tumult associated with the championing of the glacial theory by Agassiz (1837, 1840, 1842). *Rejection* was a 20-year period of eclipse of glacial theory. *Revival* marks the widespread acceptance of glacial theory following the studies of Ramsay (1860), Jamieson (1862, 1863, 1865) and Geikie (1863).

with a fair degree of accuracy, is closed to us because of the relatively low palaeolatitudes of most Neoproterozoic deposits, where seasonality is weak. The decisive falsifying test is chronometric: any glaciation that is short-lived (<5 Ma) cannot have been a snowball glaciation because of the time required to accumulate enough CO_2 to overcome the 'white Earth's' albedo. It is the duration of the glaciation, not that of particular glacial deposits, that provides the test. The tightest existing constraint on the duration of the end-Cryogenian (Marinoan) glaciation is 19.3 ± 4.2 Ma (Condon *et al.* 2005; Zhang *et al.* 2008). The older Cryogenian (Sturtian) glaciation is virtually unconstrained, whereas the mid-Ediacaran (Gaskiers) glaciation could not have been a snowball glaciation because of its short duration (Hoffman & Li 2009).

If the snowball hypothesis is not falsified geochronologically, how and when might a consensus on the nature of Cryogenian glaciation come about? The history of the Pleistocene glacial controversy (Fig. 2.4) may offer a preview. Agassiz's outspoken advocacy of the glacial theory of Esmark, Venetz and Bernhardi engendered much excitement, but failed to achieve consensus. In fact, with few exceptions (James Smith, Charles Maclaren, Robert Buckland, Robert Chambers and, on-and-off, Charles Darwin), British geologists soon turned hostile to the glacial theory and remained so from 1842 until 1860. One can track the fortunes of the theory through the writings of Charles Lyell (1841, 1845, 1851, 1852, 1855, 1857, 1863, 1865), or the Anniversary addresses of the President to the Geological Society, London (see also Woodward 1907; Davies 2007).

Then, in the first half of the 1860s, everything changed. This was the result of comprehensive regional studies by a new generation of Scottish geologists led by Ramsay (1860), Jamieson (1862, 1863, 1865) and the Geikie brothers (1863, 1874) in what may be termed the 'Scottish glacial revival'. When they began their studies, they had assumed the glacial theory was false because that was the prevailing view. Their own work, systematic and wide-ranging, left them no choice but to conclude that Agassiz had been correct after all. From their time onwards, critics of the glacial theory became the exception rather than the majority.

One is tempted to think that consensus might be reached faster in the early 21st century than in the mid-19th century. This is doubtful. It is clear from the literature that Victorian geologists actually read each others' papers. No one could claim that this is true today.

Will consensus be achieved through some conceptual or technical breakthough? I doubt it: there have been several already and consensus is no closer than at the millenium. Was consensus on Pleistocene glaciation brought about by Jamieson's (1882) solution to the submergence problem, which was that each ice age was accompanied by submergence of the land, followed by slow reemergence after the ice disappeared? This problem (based on marine fossils within tills that are now elevated above sea level) was central to the interpretation of the Drift (e.g. Smith 1836). Agassiz had offered no explanation for it, there being no marine fossils in the Swiss glacial deposits. Geophysicists attributed ice-age submergence to the gravitational 'pull' of a gigantic polar ice cap on the adjacent seas (Adhémar 1842; Croll 1875), a theory easily accommodated by those like Lyell and Murchison who believed that the drift came by way of iceberg-laden floods of Arctic origin. In contrast, Shaler (1847) and Jamieson (1865) suggested, and later demonstrated with geological evidence (Jamieson 1882), that submergence was caused by deflection of the lithosphere under the load of the ice sheet (glacioisostasy). However, Jamieson's (1882) theory of submergence came after broad acceptance of the glacial theory and cannot therefore have been its cause.

Consensus on the glacial theory came about because of the overwhelming weight of evidence of the same kind as that obtained in skeletal form by Agassiz (1842). If consensus on the Snowball Earth hypothesis ultimately rests on the weight of evidence of the kind identified by Kirschvink (1992) (palaeomagnetic evidence for low-latitude glaciation at sea level, syn-glacial sedimentary Fe_2O_3 and MnO_2 ore formations, and global evidence of abrupt climate switching (cap carbonates)) a positive outcome appears more and more probable in the end because that evidence has only become stronger.

Must we also wait for two decades or more for consensus to emerge? Long delays between initial discussion of radical hypotheses and their broad acceptance is a general phenomenon in science (e.g. heliocentrism, biological evolution, continental drift, the big bang), so we must suppose there is a reason for the delays. The simplest one is that no consensus can emerge until after the original antagonists, tethered to hasty judgements, have passed from the scene. Given the lengthening of productive lifespans, we could be in for a long wait.

Whatever the final outcome concerning the state of the Cryogenian glacial ocean, it is apparent that a third climate state needs to be added to the two first recognized by Brooks (1926): 'non-glacial' (no ice sheets), 'glacial' (polar to temperate ice sheets) and 'pan-glacial' (ice sheets at all latitudes).

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