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1 Geodynamic controls on glaciation in Earth history

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Abstract

The geological record of glaciation is sporadic. It begins with poor and fragmentary evidence from Archean rocks but there is unequivocal evidence of glaciation in the Paleoproterozoic of North America, Scandinavia and possibly South Africa, western Australia and India. There follows a long Mesoproterozoic non-glacial period, between about 2.0 and 1.0 Ga with no well-constrained evidence of glaciation but there is a return to sporadic glacial conditions from about 800 Ma to the Cambrian. Neoproterozoic glaciations were very widespread and evidence is preserved on all the present day continents. Some palaeomagnetic evidence suggests that Neoproterozoic glaciation may have taken place at low paleolatitudes but new data supports earlier concerns with regard to rapid plate motions and low latitude 'overprinting' in the Cambrian. 'Warm' climate strata in many Neoproterozoic glacial successions appear to be detrital in origin.

In Phanerozoic times, glaciation is reported from Ordovician successions in Africa, possibly Brazil and Arabia; evidence of Silurian and Devonian glaciation is largely limited to South America. The most significant Phanerozoic glaciation took place in the Permo-Carboniferous, between 350 and 250 Ma, across a large area of the Gondwanan supercontinent. There is no direct geologic record of Mesozoic glaciation but small ice masses may have developed in the interiors of landmasses at high latitudes (*e.g.* Antarctica, Siberia). Small-scale fourth-order cycles of sea-level change recorded on several carbonate and siliclastic shelves at this time are unlikely to be of glacio-eustatic origin.

The earliest Late Tertiary glaciation is recorded from Antarctica about 36 Ma; glaciation in the northern hemisphere was initiated at about 6 Ma with large continental ice sheets developing after about 3 Ma.

Global tectonic cycles of supercontinent amalgamation and dispersal are a first-order control on glaciation in Earth history. These cycles control the long-term composition of the Earth's atmosphere and have created alternating 'icehouse' and 'greenhouse' climates through geologic time. Plate tectonic activity results in amalgamation and dispersal of continental crust, which in turn controls the amount of continental freeboard. The elevation and planetary distribution of continental crust also affects such things as albedo and oceanic circulation, both of which play important roles in controlling surface temperatures. Crustal thickening in both extensional and collisional setting may have played a role in initiating 'adiabatic' glaciations. Active tectonism initiates large sediment fluxes from the continents, the weathering of which may draw down atmospheric CO_2 allowing the growth of larger ice masses controlled in part, by relatively weak 'Milankovitch' astronomical rhythms.

Recent work identifies the importance of palaeogeography in further controlling planetary temperatures and the tendency for very rapid global change involving abrupt changes in ocean–atmosphere coupling.

Introduction

The major questions that need to be answered in relation to Earth's glacial record are as follows:

(1) Why does glaciation occur? Is there a single cause or a series of separate or combined factors that can bring about cold climatic conditions on the planet?
(2) Why is the record of glaciation sporadic?
(3) What is the explanation for glaciation allegedly extending into marine basins at apparent equatorial latitudes in the Neoproterozoic?
(4) Why are many Neoproterozoic glacial deposits associated with other sedimentary rocks that appear to suggest warm climatic conditions?

The geologic record, although imperfect and incomplete, remains the only primary source of information concerning the long-term climatic history of the planet. In this paper we present a brief review of the Earth's glacial record and attempt to place it in the context of the geodynamic evolution of the planet. In doing so we hope to provide tentative answers to some of the questions posed above. We would emphasize that space does not permit a detailed treatment of all glacial occurrences. An exhaustive descriptive summary of Earth's glacial record can be found in Hambrey and Harland (1981). The more recent interpretive summaries include those by

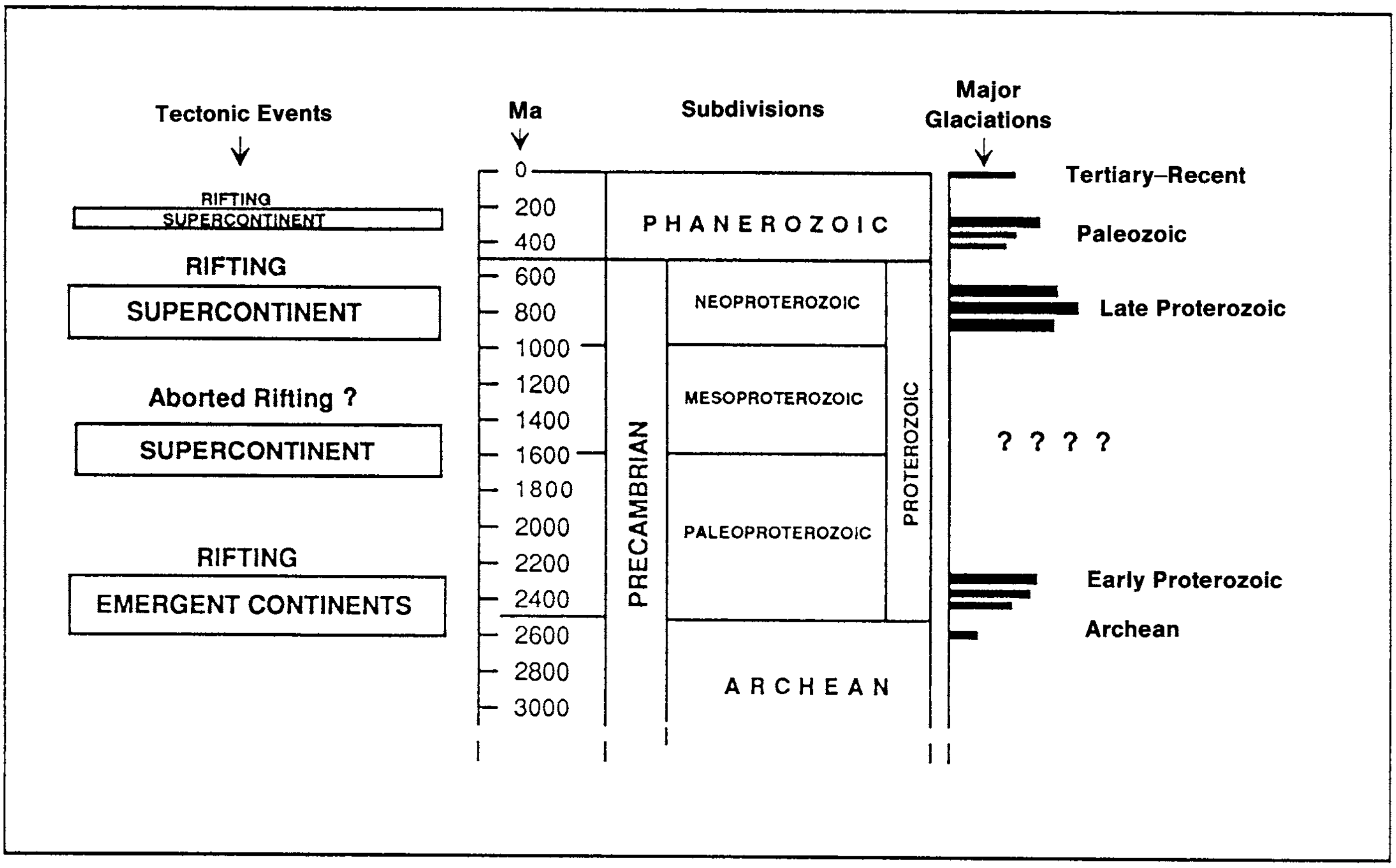

Fig. 1.1. Distribution of major glaciations throughout geologic time (subdivisions after Plumb, 1991). Note the association between some periods of supercontinentality and some glaciations. See text for discussion.

Crowell (1982), Deynoux (1985b), Veevers (1990), Worsley and Nance (1989), Young (1991a), Frakes *et al.* (1992) and Chumakov (1992). A detailed treatment of the relationship between glaciation and geodynamic setting can be found in Eyles (1993); Socci (1992) provides a thorough overview of likely global geochemical controls.

Any attempt to depict the distribution of glaciogenic rocks in Earth history is frustrated by a lack of precise and reliable geochronological data from the largely sedimentary successions in which they occur. A second problem is inadequate discrimination between diamictites formed by direct glacial activity (tillites) and those formed by other processes. As pointed out by Schermerhorn (1974) this problem is compounded by the fact that many glaciogenic successions were deposited in tectonically active settings and had complex histories sometimes involving both glacial processes and deposition by sediment gravity flows.

Figure 1.1 shows the distribution of glaciogenic rocks based on radiometric age dating of the past 2500 Ma. The age of most ancient glacial deposits is very poorly constrained and Figure 1.1 probably underestimates the frequency and duration of glaciation in Earth history.

Late Archean–Paleoproterozoic glaciations

The preserved record of glaciation is confined to the last half of geologic history when four major glacial eras (Chumakov, 1985) occurred over a period of about 2.5 Ga (Fig. 1.1). The dearth of evidence of glaciation in Archean times is puzzling because of the widely held belief that the sun's luminosity was much less (about 70% of the present value) during the early part of Earth history (Sagan and Mullen, 1972; Kasting and Toon, 1989; Gilliland, 1989; Caldeira and Kasting, 1992). It has been estimated (Kasting *et al.*, 1984) that under the present atmosphere, such a weakly radiant sun would have resulted in surface temperatures on Earth about 30 °C lower than at present. At such temperatures liquid water could not have existed on the surface of the planet but preservation of abundant waterlaid Archean rocks clearly indicates that this was not the case. The simplest resolution of what has been called the 'faint young sun paradox' (Kasting, 1987) is to invoke an enhanced greenhouse effect due to a much higher CO_2 content in the early atmosphere (Hart, 1978). If this interpretation is correct, then the paleoclimatic controls in the early part of Earth history differed greatly from those today. If it is accepted that solar luminosity was much weaker than at present, then cooling could easily have been induced by drawdown of atmospheric CO_2.

The Archean Witwatersrand Supergroup of South Africa, deposited between 3000 and 2700 Ma (de Wit *et al*., 1992), contains diamictites and other rocks that have been interpreted as products of glaciation (Wiebols, 1955; Harland, 1981) but a non-glacial origin is also possible (Kingsley, 1984; Tainton and Meyer, 1990). Von Brunn and Gold (1993) report glacialastic strata from the Pongola Sequence of Southern Africa dated between 3100 and 2800 Ma. Dropstones in Archean rocks associated with the Stillwater Complex in western USA have also been ascribed to glacial transport (Page, 1981). Apart from these small occurrences the Archean record is barren of glaciogenic deposits.

In the Early Proterozoic (Paleoproterozoic in the new terminology of Plumb (1991)) there is evidence of glaciation in several regions of North America, and in Karelia, South Africa, Australia, India and Scandinavia (e.g. Okajangas, 1988). Early Proterozoic glacial deposits are widespread but most again are poorly dated. The most clearly exposed deposits are those of the Gowganda Formation in Ontario, Canada dated to about 2300 Ma (Young and Nesbitt, 1985; Figs. 1.2, 1.3). Other, poorly constrained deposits occur in the Northwest Territories (Padlei Fm; older than 2.1 Ga; Patterson and Heaman, 1991), several occurrences in Michigan (2100–2000 Ma) and in the Black Hills (2560–1620 Ma). Deposits of similar age occur in southern Africa (Griquatown Basin; *c.* 2300 Ma), India (Bijawar Group, *c.* 1815 Ma) and Australia (Meteorite Bore Member, *c.* 2500–2000 Ma). Diamictites, some of which may be glaciogenic (Ojakangas, 1988), occur in the Baltic shield and in east-central and northern former USSR.

The most widespread and most convincing glaciogenic deposits of this era are found in North America. The Gowganda Formation, which forms part of the Huronian Supergroup deposited between about 2.5 and 2.2 Ga (Krogh *et al*., 1984), is perhaps the best known (Miall, 1985; Young and Nesbitt, 1985) but similar rocks have also been described (Young, 1970) from Michigan, Chibougamau in northern Quebec, the Hurwitz Group in the Canadian Northwest Territories and from Wyoming (Fig. 1.2). In some Paleoproterozoic successions there is evidence of multiple glaciations. For example, the Gowganda Formation (Fig. 1.3) is the youngest of three diamictite-bearing formations, all of which may be glaciogenic. Thus there is evidence in the Huronian Supergroup (and also possibly in the Snowy Pass Supergroup of Wyoming; Houston *et al*., 1981) of three major glacial intervals. Geochemical and sedimentological data (Fig. 1.3) suggest that glacial episodes in the Huronian Supergroup were separated by periods of intense chemical weathering (Nesbitt and Young, 1982).

In both areas, the tectonic setting has been interpreted in terms of a rifting continental margin (Karlstrom *et al*., 1983; Zolnai *et al*., 1984; Young and Nesbitt, 1985; Goodwin, 1991). Emergence of large areas of continental crust in the late Archean–Early Proterozoic (Taylor and McLennan, 1985) may have been a critical factor in reduction of global temperatures to the point where glaciation could occur. The quantitatively most important weathering reaction on the surface of the planet (Nesbitt and Young, 1984) is that between carbonic acid (CO_2+H_2O) and labile minerals of the continental crust (particularly plagioclase felspars). This reaction at present accounts for about 80% of the drawdown of CO_2 from the atmosphere (Houghton and Woodwell, 1989). Any tectonic process leading to emergence and subaerial exposure of significant regions of crustal material should therefore result in increased rates of drawdown of atmospheric CO_2. Under the conditions of weak solar luminosity inferred for the early part of Earth history, such a reduction in CO_2 may have caused a sufficiently strong 'antigreenhouse' effect to bring about the first glaciations.

How can we explain the 'anomalous' association of glacial deposits with sedimentary rocks that are highly weathered? This has been ascribed by Young (1991) to a negative feedback mechanism. Glacial cover inhibits surface weathering and results in increased albedo which contributes to a general lowering of surface temperatures. All of these factors contribute to a decrease in chemical weathering, so that CO_2, which is constantly being supplied from volcanic sources, can once more build up, leading eventually to the destruction of the ice sheets and initiation of a 'warm' period. Such 'warm' periods could only have occurred when concentrations of CO_2 were sufficiently high to counteract the inferred weak sun. This interpretation is in keeping with the strong chemical depletion observed in Huronian formations between the glacial units (Nesbitt and Young, 1982) and elsewhere in Archean and Early Proterozoic sedimentary rocks (Reimer, 1986; Eriksson *et al*., 1990). Thus negative feedback may provide an explanation for multiple glaciations separated, at least in the early part of Earth history, by sedimentary rocks carrying evidence of intense chemical weathering.

Mesoproterozoic non-glacial epoch

Perhaps even more puzzling than the glacial periods themselves is the apparent absence of glaciation on the planet between about 2.0 and 1.0 Ga (Fig. 1.1). Chumakov and Krasil'nikov (1992) report possible Riphean glacial deposits from the Jena area of the former Soviet Union but their age is uncertain. Tectonic syntheses such as that of Hoffman (1989) suggest that the period between about 2.0 Ga and 1.8 Ga was a time of growth and aggregation of continents (at least in the Laurentian shield). A supercontinental configuration has been proposed by many (Windley, 1977; Piper, 1978; Hoffman, 1989) for the ensuing period. According to ideas outlined by Worsley *et al*. (1984) and Nance *et al*. (1988) such a supercontinental configuration should have provided suitable conditions for glaciation by the mechanism of drawdown of atmospheric CO_2 as a result of enhanced weathering reactions on the high standing supercontinent. Hoffman (1989), noting widespread magmatic activity (including unusually abundant anorthosites and rapakivi granites) during the period from about 1.8 to 1.3 Ga, proposed a unique stage in the thermal evolution of the planet, invoking a 'mantle superswell' that caused

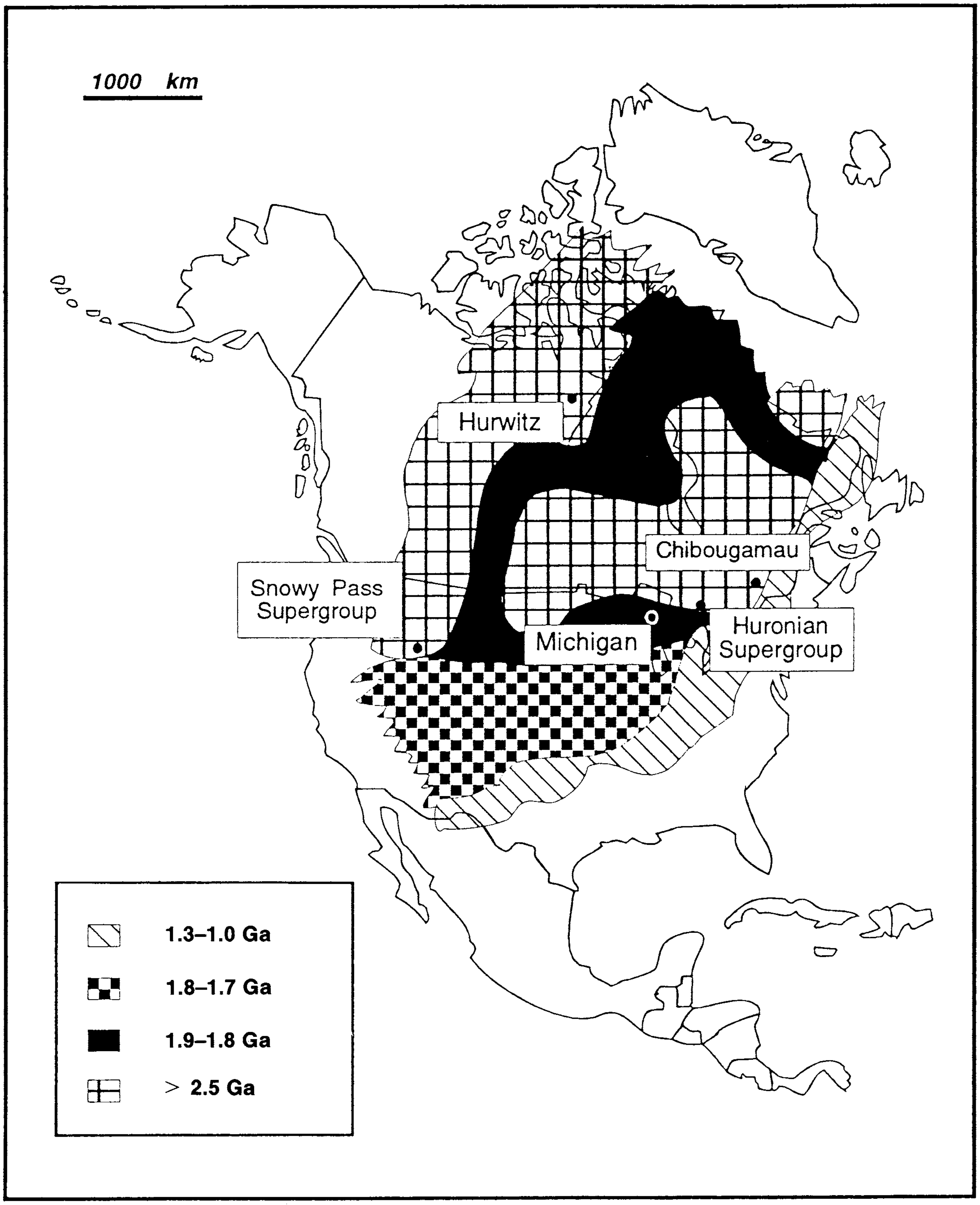

Fig. 1.2. Sketch map to show the location of Early Proterozoic glaciogenic rocks in North America in relation to major tectonic provinces (after Hoffman, 1988) the ages of which are shown at bottom left. Note that most occurrences are at or close to the margins of Archean cratons.

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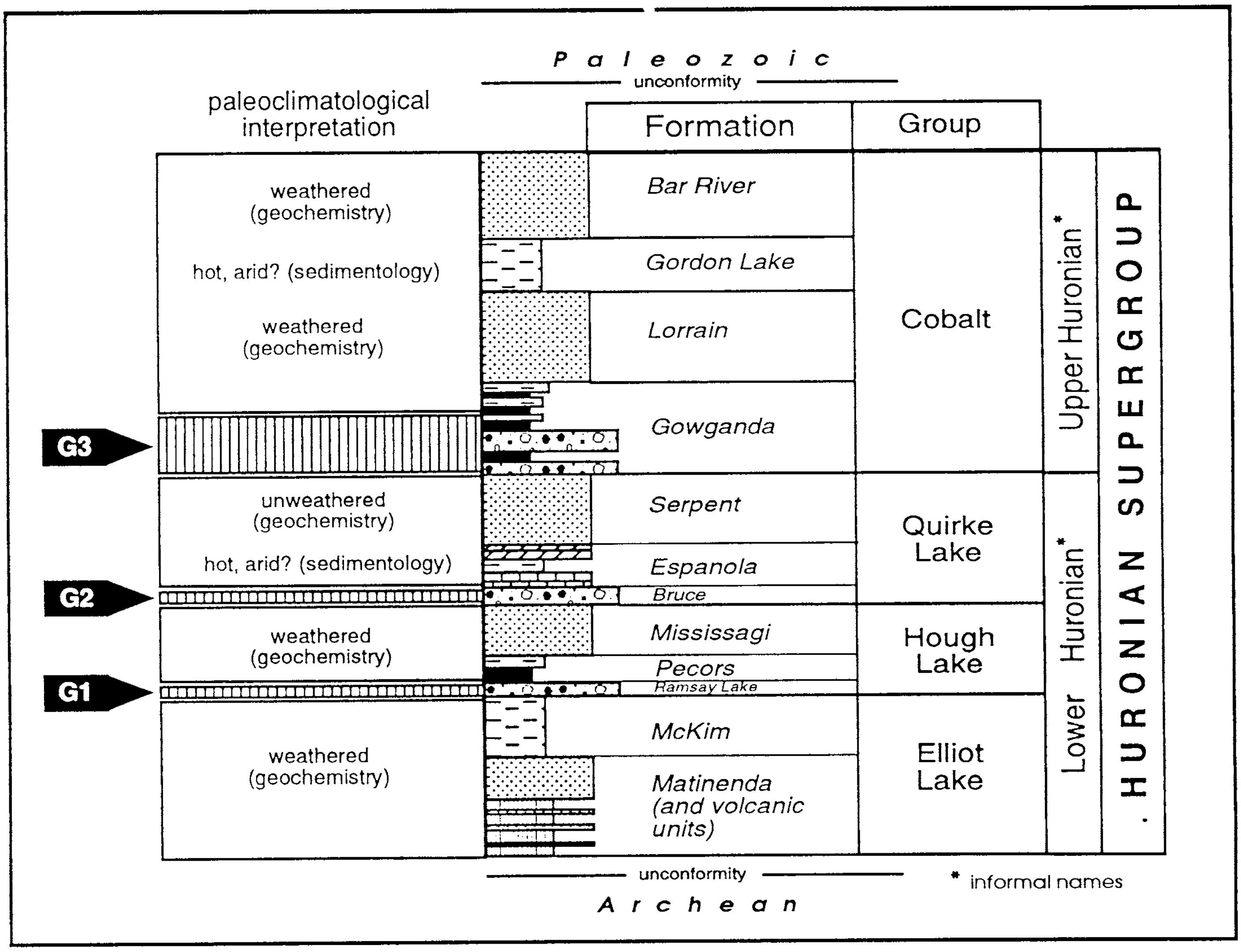

Fig. 1.3. Schematic representation of the stratigraphic succession of the Paleoproterozoic Huronian Supergroup (maximum thickness about 12 km) together with inferred paleoclimatic changes. Note the three major glacial periods (G1–3) separated by warm and/or humid periods of intense chemical weathering. These fluctuations are attributed to changes in atmospheric CO_2 content, ascribed to negative feedback (see text). Abbreviations in the stratigraphic column are as follows: fine dots, sandstone; large dots, diamictites; dashes, siltstones; black, mudstones; horizontal bricks, limestones; inclined bricks, dolostones; vertical hatching, volcanic units in the basal part of the Huronian. See text for discussion.

elevated temperatures beneath the first extensive supercontinent. Outgassing of CO_2 during this period of extensive magmatism and aborted continental fragmentation (Emslie, 1978) may have outstripped drawdown by weathering, causing the Earth to enter a long period of greenhouse-induced warm climate.

Neoproterozoic glaciations

Overview

Late Proterozoic glaciogenic deposits are known from all the continents. They provide evidence of the most widespread and long-ranging glaciation on Earth, extending from about 1.0 Ga to just before the Cambrian. Most regions display evidence of several glaciations, preceded, separated and followed by periods of relatively warm climate. This alternation of glacial and 'warm' climatic conditions, on a scale of tens of millions of years, is similar to that observed in the Paleoproterozoic, and could likewise have resulted from negative feedback, related to weathering reactions on a high-standing supercontinent (Young, 1991a, b). One of the most puzzling aspects of Late Proterozoic history is that glaciers appear to have descended to sea level in low paleolatitudes (Chumakov and Elston, 1989). Many explanations for these anomalous occurrences have been proposed. Williams (1975) suggested that the obliquity of the ecliptic was much higher than today and argued that this would have led to preferential glaciation at low latitudes as inferred from paleomagnetic study of Late Proterozoic glaciogenic sediments in South Australia (Embleton and Williams, 1986). Williams also suggested that this model provides an explanation for evidence of seasonality of low latitudes. Others (Stupavsky *et al.*, 1982; Crowell, 1982) have considered the paleomagnetic data to be suspect and some of the reported low paleolatitudes to be spurious.

Recent work by Chumakov and Elston (1989) and Embleton and Williams (1986) tends to support an equatorial positioning for many Late Proterozoic glacial deposits but new data also show the probable importance of high latitude positioning during the Neoproterozoic followed by very rapid plate motions to low latitudes in the Cambrian when the original remanent magnetization was overprinted (Meert and Van der Voo, 1992). Young (1991a) proposed that a high standing Neoproterozoic supercontinent (Windley, 1977; Piper, 1978; Khramov, 1983; Hoffman, 1989, 1991; Moores, 1991; Dalziel, 1991) centred over the equator would have provided ideal conditions for withdrawal of atmospheric CO_2, reducing the greenhouse effect and leading to glaciation. This model finds support in recent theoretical climate modelling (Worsley and Kidder, 1991) which emphasizes the critical role of continental positioning and paleogeography. Deynoux (1985b) pointed out that an emergent supercontinent would also lead to increased planetary albedo which could provide a positive feedback contributing to a build-up of ice volumes. There is an urgent need for further paleomagnetic studies of Neoproterozoic glacial strata and related deposits.

Timing and tectonic setting

The notion of a globally correlative Neoproterozoic glaciation was proposed by Harland and Bidgood (1959) in order to explain the apparent relationship of glacial deposits with warm climate indicators and low depositional paleolatitudes as suggested by paleomagnetic data. Roberts (1971, 1976) suggested that a world-wide episode of dolomite deposition at the end of the Proterozoic triggered a drawdown of atmospheric CO_2, favouring glaciation.

The association between diamictites of supposedly glacial origin and warm climatic indicators such as dolostones led Schermerhorn (1974, 1977) to question the glaciogenic nature of many diamictites and to emphasize their active tectonic setting. Subsequent research (reported in Hambrey and Harland, 1981), stimulated in part by Schermerhorn's controversial review, has confirmed the glaciogenic nature of many Neoproterozoic diamictite-bearing successions. In addition, however, work also stresses the importance of active tectonic settings and the deposition of non-glacial debris flows and thick turbidite successions (N. Eyles, 1990). Schermerhorn (1977) also emphasized the role of regional tectonics as a means of depleting Late Proterozoic atmospheric CO_2 levels thereby providing a background to Late Proterozoic glaciation.

The notion of a widespread, global late Proterozoic glaciation has been weakened by new tectonic models (Hoffman, 1991) and by new data regarding the nature of allegedly 'warm' strata interbedded within glaciclastic successions (Fairchild, 1993). A key factor in the origin and timing of Neoproterozoic glaciation is the history of accretion and fragmentation of the Neoproterozoic supercontinent. In turn, the structural setting is central to understanding the sedimentology of the resulting glacial deposits. Hoffman (1991) proposed that break-up of a Late Proterozoic supercontinent (cf. Moores, 1991 and Dalziel, 1991) resulted in the 'break out' of Laurentia and fan-like rotation of the other component parts. The initial break-up occurred along the Proto-Pacific margin of Laurentia after 750 Ma and is recorded by thick passive margin deposits in western North America, Australia and China. An early episode of plate tectonic activity is indicated by the presence of a thick succession of fine-grained siliciclastic deposits and carbonates (Wernecke Supergroup of Delaney, 1981) that were deformed and thrust onto the North American craton (Clark and Cook, 1992) at about 1.2 Ga. The Neoproterozoic break-up in western North America (Hayhook 'orogeny' of Young *et al.*, 1979) appears to have begun at about 750 Ma near the transition between the Mackenzie Mountains Supergroup and the glaciogenic Rapitan Group (Fig. 1.4). The exact timing of break-up has remained elusive but it now appears that there were two episodes of rifting as pointed out by Ross (1991). The first may be contemporaneous with deposition of the Rapitan Group at 750 Ma (Stewart, 1972; Young *et al.*, 1979) with a second episode near the base of the Cambrian (Bond and Kominz, 1984). Early opening of the Proto-Pacific Ocean is a prerequisite of the tectonic model of Hoffman (1991) for Pan-African amalgamation of Gondwana. Glacial deposits are a prominent (correlative?) component of Neoproterozoic strata in western North America (e.g. Eisbacher, 1985; Ross, 1991; Young and Gostin, 1991; Young, 1992) suggesting a close relationship between rifting and glaciation (see below).

Younger rifting (<650 Ma) along the Proto-Atlantic (Iapetus) margin of Laurentia is also associated with widespread glacial strata now scattered around the margins of the North Atlantic Ocean in Scotland, Scandinavia, Greenland and Spitsbergen. In most of these areas there is clear evidence of glacial deposition in extensional tectonic settings (e.g. C. Eyles, 1988; see below). New U-Pb zircon dates from granites in the Blue Ridge of North Carolina and Tennessee, however, have given ages of 740 to 760 Ma (Su *et al.*, 1992). The significance of these dates is that the granites are thought to be closely associated with the Mount Rogers volcanic centre in Virginia and could be evidence of an early episode of rifting of Laurentia from the Neoproterozoic supercontinent. A relationship between rifting and the preservation of glaciogenic strata of the Mount Rogers volcanic centre is examined by Miller (this volume).

In contrast, glaciogenic deposits of the Pan-African belt now preserved around the margins of the North Atlantic Ocean are young (they post-date *c.* 550 Ma) and were deposited in an active plate margin setting. Diamictites form a prominent stratigraphic component of basin fills in eastern North America, North Africa and Europe and are closely associated with voluminous volcanic sediments; local glaciation of volcanic cordillera is indicated in many areas (e.g. Socci and Smith, 1987; N. Eyles and C. Eyles, 1989; N. Eyles, 1990). Elsewhere, in North Africa, glaciation covered a huge (2×10^6 km^2) foreland region of the West African platform (Taoudeni Basin; Deynoux and Trompette, 1976; Deynoux, 1985a,b) and is described elsewhere in this volume.

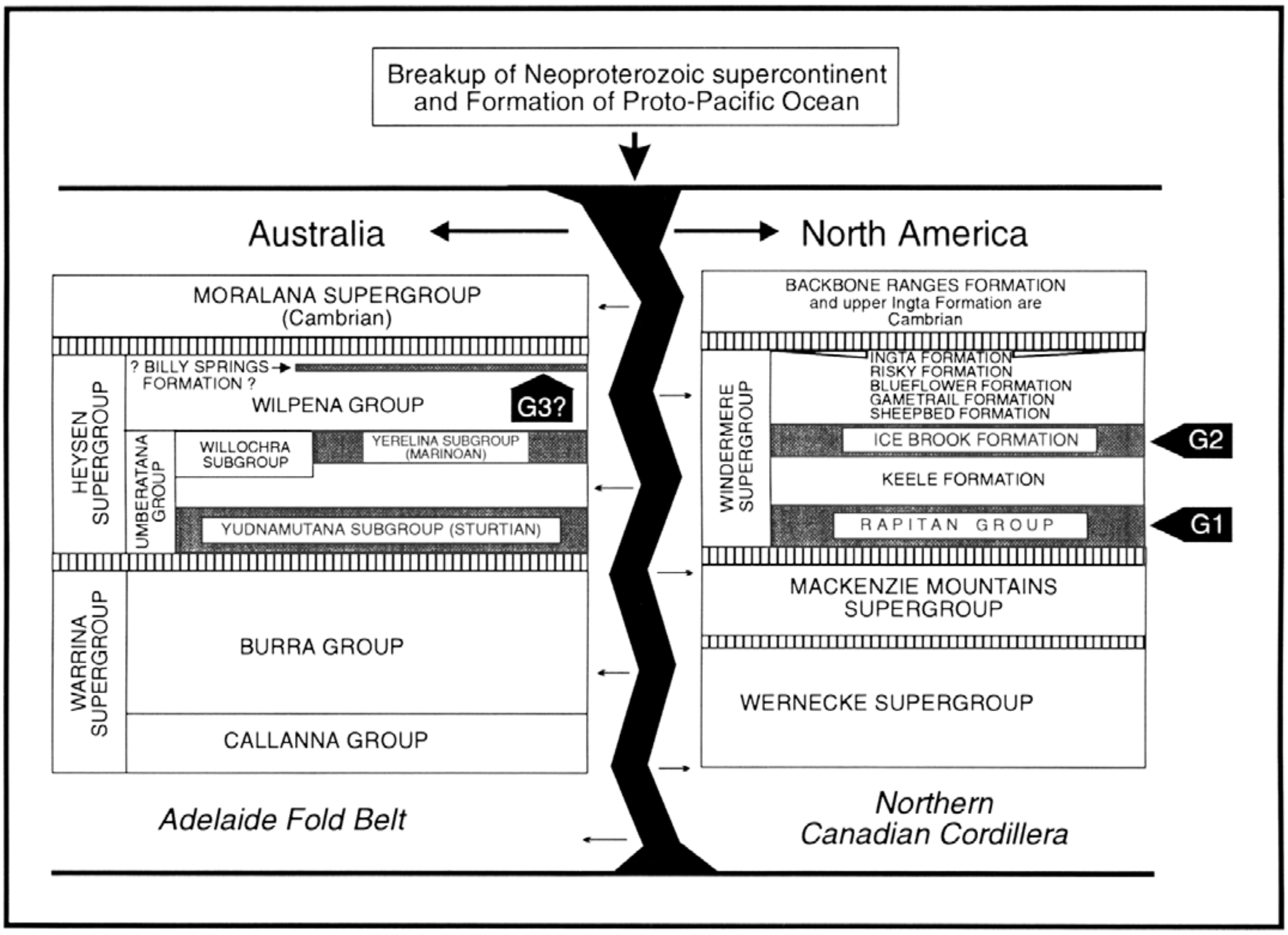

Fig. 1.4. Comparison of Late Proterozoic stratigraphy of South Australia and the northern Canadian Cordillera. Glacial episodes are shown by the grey ornament and are designated G1–3. Black ornament in central part of the figure represents oceanic crust formed by the break-up of the inferred Neoproterozoic Supercontinent. New paleontological data after MacNaughton *et al.*, (1992).

The importance of extensional rift basins

Most Neoproterozoic glacial deposits accumulated as glacially influenced marine strata along rifted continental margins or interiors; phases of crustal extension preceded disintegration of the Neoproterozoic supercontinent starting at around 750 Ma along its western (Pacific) margin. Given that strong, localized uplifts of *many* kilometres can occur along the boundary faults or rifted terrains, a glacial source area and sediment repository are thus provided. Dolomites that underly many glaciclastic strata may record clastic-starved sedimentation in restricted warm-water basins prior to onset of uplift on the basin margins. Dolomitic strata that occur within several glaciclastic successions are detrital in origin reflecting the uplift and glacial erosion of underlying dolomites (Fairchild, 1993). Strong uplift and the development of local ice masses adjacent to rapidly subsiding rifted basins could also provide the uniformitarian key to the occurrence of Late Proterozoic glacial deposits in low to middle paleolatitudes (Schermerhorn, 1977). However, whilst uplift may initiate glaciation on elevated massifs, the deposition of *glaciomarine* sediments indicates cold temperatures at sea level. It is the case however, that paleomagnetic work on which models of low latitude glacial deposition are based is increasingly suspect (e.g. Meert and Van der Voo, 1992).

South Australia/northern Canadian Cordillera

Glaciogenic rocks figure prominently in the Neoproterozoic stratigraphy of southeastern Australia and the northern Canadian Cordillera (Fig. 1.4). The Sturtian glaciogenic succession (*c.* 740 Ma) unconformably overlies rocks of the Burra Group

(Coats, 1981; Coats and Preiss, 1987). In a regional study of the northern part of the Adelaide 'geosyncline' Young and Gostin (1991) described widespread development of a four-fold subdivision of the Sturtian succession into two major diamictite–mudstone sequences, representing two glacial advance–retreat cycles. Rapid thickness and facies changes were attributed to contemporaneous extensional tectonics interpreted by Young and Gostin (1991) as expressions of a rift episode. Strong similarities with the Late Proterozoic succession in the northern part of the Canadian Cordillera led Rowlands (1973), Bell and Jefferson (1987) and Young (1992) to propose stratigraphic correlations between Australian and Canadian Neoproterozoic successions including the glaciogenic Sturtian and Rapitan Group.

A second glaciation known as the Marinoan, for which an age of about 680–690 Ma has been proposed (Coats, 1981) also ocurs in the Adelaide 'geosyncline' (Preiss, 1987; Lemon and Gostin, 1990). It is separated from the Sturtian by a thick succession of sedimentary rocks containing no evidence of glaciation. This glacial phase could correspond to the recently described Ice Brooke Formation (Aitken, 1991) in the northern Cordillera. A third glaciation has been tentatively identified in the Billy Springs Formation of the Pound Subgroup (Fig. 1.4) just below the Cambrian in South Australia (Di Bona, 1991).

The Neoproterozoic glaciogenic successions of southeastern Australia and northwestern North America are remarkably similar and lend support to the idea that Australia/Antarctica was juxtaposed against North America as part of a Neoproterozoic supercontinent and that glaciogenic sediments were preserved in rift basins during continental fragmentation (Bell and Jefferson, 1987; Moores, 1991; Dalziel, 1991; Hoffman, 1991; Ross, 1991).

North Atlantic sector

Late Proterozoic glacial deposits accumulated under extensional crustal regimes around the rifted margins of several cratonic blocks.

Scotland

In Scotland, the Port Askaig Tillite of the Dalradian Supergroup, was first described by Thompson (1871). The succession, about 750 m thick, is seen in strike-parallel outcrops from southwest Ireland to northeast Scotland (700 km) and consists of a relatively simple alternation of massive diamictites conformable with sandstones, minor conglomerates and laminated siltstones, some of tidal origin. Spencer (1971, 1985) emphasized repeated cycles of sedimentation below a grounded ice sheet; Eyles (1988) argued that the geometry and sedimentology of the deposit were inconsistent with subglacially deposited facies and argued for a subaqueous glaciomarine origin. She highlighted the structural setting of the Port Askaig Tillite at the base of a deepening-upward succession of turbidites and basinal facies (Argyll Group; Anderton, 1982) deposited in fault-bounded blocks and basins on the northern margins of what was to become the Iapetus Ocean. Large olistostromes composed of dolomite (e.g. the Great Breccia; Spencer, 1971) record the collapse of fault-scarps breaking up shelfal carbonates. Glacial deposits record sedimentation from meltwater plumes and floating ice; interbedded facies record tidally influenced conditions. Continued seismic activity may be recorded by penetrative, sandstone deformation structures (N. Eyles and Clark, 1985). In contrast, a subaerial, periglacial origin has also been suggested for these structures (Spencer, 1971).

East Greenland

Vendian 'tillites' occur within the 17 km thick Neoproterozoic section of East Greenland. The 1300 m thick Tillite Group consists of two diamictite horizons variably called Lower and Upper Tillite (or Ulvescö, Storeelv Formations respectively) separated and capped by thin-bedded marine mudstones and sandstones. Massive diamictites with sandstone rafts, associated with thin to medium bedded turbidites with dropstones comprise the dominant glacial facies; as in Scotland, penetrative deformational structures may record ongoing seismicity. The glacial and associated marine strata rest on predominantly dolomitic shallow water sediments of the Eleonore Bay Group (Hambrey and Spencer, 1987). The sudden incursion of glacial clastics can be interpreted as a record of accelerated uplift and subsidence along the border of Laurentia and Baltica. A combination of glacially influenced shelf and slope facies appears dominant; as in many Late Proterozoic successions 'tillite' horizons occur within thick turbidite successions suggesting a mass flow origin; the characteristic record of glacially influenced marine deposition in rapidly subsiding basins (Eyles, 1993).

Similarity of the West Greenland and eastern Svalbard (Wilsonbreen Formation) glacial successions has been emphasized by Spencer (1975), Hambrey (1983) and Fairchild (1993). This suggests close geographic juxtaposition perhaps within the same basin. Strata of western Svalbard, however, are thicker indicating either accelerated subsidence or deposition in another basin. Western Svalbard was juxtaposed against central and eastern Svalbard by later strike-slip movement.

Scandinavia

The importance of block faulting during early rifting of the Iapetus Ocean is strongly apparent in the Neoproterozoic glacial deposits on the west-facing margin of Baltica.

In Finnmark, sedimentation occurred on the western rifted flank of Baltica, adjacent to the Timanian Aulacogen which separated Baltica from the Barents Craton to the north. Thick (450 m) glacial deposits of the Vestertana Group (Smalfjord, Nyborg, Mortensnes Formations; Edwards, 1975) began to accumulate in the Gaissa Basin following late Sturtian uplift of shallow water stromatolitic dolomites (Porsanger, Grasdal Formations); uplift of at least 1500 m can be inferred in the eastern boundary of the Gaissa Basin but nearly 8 km of strata were removed in other areas (Gayer and Rice, 1989). It is highly significant that coeval strata in closely adjacent basins to the west and north (Baltoscandian miogeocline and Timanian Aulacogen) are represented by alluvial fan, fluvial facies and shallow marine facies; these contain diamictites formerly regarded as glacial (Siedlecka and Roberts, 1972) but now interpreted as alluvial fan debris flows (Laird, 1972, Gayer and Rice, 1989). These findings are especially pertinent to the 'tillites' of the Vestertana Group.

Diamictites of the Smalfjord Formation are rich in dolomite clasts and record downslope slumping and sediment gravity flow of coarse-grained sand and gravel facies. Diamictites are interbedded with shallow marine, storm-influenced facies. A fan-delta depositional setting is suggested by Warman (personal communication, 1991). The so-called 'interglacial' Nyborg Formation comprises a thick succession of turbidites probably recording enhanced fault-controlled subsidence. Toward the top, it is interbedded with massive diamictites of the Mortensnes Formation which record glaciomarine sedimentation primarily from suspended sediment plumes and icebergs, with secondary mass flow. Initial extension and rifting of the continental margin is marked by dykes dated at 640 Ma (Beckinsale *et al.*, 1976). These are part of a wider network along the margin of Baltoscandia interpreted as syn-Iapetus rifting intrusions and are correlative with turbidites of the Nyborg Formation.

Broadly similar 'fan-delta' settings as identified in the Vestertana Group of the Gaissa Basin occur in other Norwegian basins along the former Baltoscandian continental margin (e.g. Tietzsch-Tyles, 1989). In central and southern Scandinavia five main basins (Risbäck, Hedmark, Valdres, Engerdalen, Tossasfjallet) record sedimentation in extensional settings along the Baltoscandian margin. Nystuen (1987) recognized a common succession in the western Baltoscandian basins of lowermost diamictites overlain by laminated mudstones with decreasing quantities of ice-rafted debris upward in section. This succession is most easily interpreted as a basinal deep-water assemblage recording the downslope slumping of primary glacial debris and syndepositional subsidence.

The structural setting and sedimentology of the glacial deposits in Scotland, Greenland, Spitsbergen and Scandinavia point to the importance of uplift during regional extension along the paleo-Atlantic continental margin of Laurentia. At about 750 Ma the North Atlantic region of the Neoproterozoic supercontinent was characterized by the widespread development of inter- and intracratonic basins peripheral to an early Iapetus Ocean. Geophysical models show that large-scale extension and thinning of the crust, whether by simple or pure shear, is associated with topographic 'arching' on the margins of the extended terrain (Lister *et al.*, 1991). This may create a glacial source area and a depositional repository (see below; Late Cenozoic Glaciations). Many areas undergoing extension show broad topographic domes separated from the active rift by a fault-bounded topographic escarpment. The modern Red Sea rift and associated updomed Arabian Shield is one example; the Death Valley area east of the Sierra Nevada also shows a marked topographic updoming; the lowest (−86 m asl) and highest (4419 m) points in mainland USA occur in this area (Wernicke, 1985).

Strong localized uplift of rift shoulder flanks and associated crustal blocks resulted in widely-dispersed glacial centers, across a very broad range of paleolatitudes, adjacent to rapidly subsiding Neoproterozoic rift basins (cf. Yeo, 1981; Young, 1989). Most of the diamictites preserved in these basins are not true tillites but glacially influenced marine deposits recording the release of large volumes of mud and coarser debris to the marine environment. Schermerhorn (1975) argued that tectonic differentiation involving complementary uplift and subsidence set the scene for deposition of many Neoproterozoic 'glacial' successions. Schermerhorn's arguments were largely ignored by geologists who felt that he had largely written off *all* Neoproterozoic tillites in favour of non-glacial, tectonically generated mass flows. This is not so, for as Schermerhorn (1975) concluded, 'though the Late Precambrian mixtite formations were laid down in obvious tectonic settings, among them exist glacial deposits or glacially influenced sediments, and one task of future unprejudiced studies will be to determine the proportion of glacial components and the origin and extent of glacial activity within the frame of mixtite deposition'. Because of local tectonic controls, individual diamictite horizons have a limited potential for lithostratigraphic correlations (cf. Chumakov, 1985).

Discussion

Strong tectonic uplift in extensional regimes may explain the widespread and seemingly paradoxical association of diamictites and carbonates in Neoproterozoic strata. Fairchild (1993) showed that carbonates *below* 'glacial' strata in Greenland, Scotland and Spitsbergen record an upward change from limestone to dolostone facies consistent with upward shallowing. The shallowing upward carbonate successions may record strong tectonic uplift accompanying the initiation of an extensional tectonic regime. This uplift is recorded by progressive unroofing of carbonates and associated exposure of basement lithologies and by abundant clastic carbonate within overlying 'glacial' strata.

Carbonates *within* diamictite successions are overwhelmingly of clastic origin (most are dolarenites) and record the glacial erosion of uplifted fault-bounded carbonate massifs and rift shoulders. Fairchild and Spiro (1990) and Fairchild *et al.* (1989) also report primary lacustrine precipitates from Spitsbergen comparable to those of 'arid' saline lakes in present-day Antarctica. *Cap* dolostones are common in Australia, West Africa and North America (Williams, 1979; Deynoux, 1985a). Fairchild (1993) stressed the lack of detailed data and could not discriminate between the effects of diagenesis or primary deposition. The most important variable appears to be the availability of large amounts of reactive carbonate rock flour produced by glacial erosion.

In summary, the widespread development of clastic-starved, carbonate platforms in restricted Neoproterozoic rift basins may have been followed by the shedding of glacial and clastic carbonate debris from uplifted crustal blocks. The ultimate cause of global cooling may, however, be the drawdown of atmospheric CO_2 by enhanced weathering of the uplifted supercontinent. This may have taken place several times in response to negative feedback as described earlier for the Paleoproterozoic (see above).

Another outstanding problem that may be resolvable by reference to tectonic setting is the 'enigma' of low paleolatitude glaciation first recognized by Harland and Bidgood (1959) and supported by paleomagnetic data from other workers (e.g. Chumakov and Elston, 1989). Glacial deposition in low equatorial paleolatitudes has been used to argue for global refrigeration (Harland, 1975) and for the use of tillites as global chronostratigraphic marker horizons (Chumakov, 1985). The stratigraphic association of tillites and carbonates was cited as evidence of short-lived cold conditions in low latitudes (paleoclimatic crises).

Piper (1982) argued that supposed primary magnetization showing alleged low paleolatitudes is the result of secondary overprinting and this was clearly demonstrated for the Port Askaig Formation (see above) by Stupavsky *et al.* (1982). Concerns with the quality of paleomagnetic data were also voiced by Crowell (1982) and have been borne out by new data (Meert and Van der Voo, 1992). Schermerhorn (1974) had earlier argued that many supposed tillites were non-glacial mass flows and it is true that a wide variety of diamictites deposited under very different tectonic settings have been lumped together as Late Proterozoic tillites (see above). The most compelling evidence for cold climates at low latitude is forthcoming from the Marinoan succession of the Adelaide Geosyncline in South Australia (*c.* 650 Ma). Fine-grained tidal laminites (Elatina rhythmite, member of the Elatina Formation; Williams, 1989) record low inclinations consistent with deposition in a paleolatitude belt between 20 °N and 12 °S of the paleoequator (Schmidt *et al.*, 1991). Coeval strata host permafrost structures recording mean annual temperatures of less than -4 °C.

Embleton and Williams (1986), whilst recognizing that reasonable doubt remains regarding many Neoproterozoic paleomagnetic data, suggested three hypotheses to account for cold climates in low latitudes. The first (global glaciation) was rejected; a 'frozen-over Earth' requires an unrealistic increase in solar luminosity to thaw (e.g. Bahcall and Ulrich, 1988), but changes in atmospheric composition (increasing CO_2) were not considered. A second argument proposes that the present-day situation of an axial geocentric dipole model for the Earth's magnetic field is invalid for the Late Proterozoic. The third hypothesis argues for a considerably increased obiquity of the ecliptic such that areas at low latitudes would receive less solar radiation during an annual cycle than areas at high latitudes. Both these arguments are reliant on more but better geophysical and paleomagnetic data combined with rigorous analysis of the geodynamic setting of supposed glacials.

If the available paleomagnetic data stand the test of time and it is shown that many Neoproterozoic glaciogenic successions were definitely deposited in low paleolatitudes then these data must be accommodated within an appropriate depositional model. One possibility is that, given an atmosphere richer in CO_2 than at present, a high-standing supercontinent that straddled the equator (e.g. Hoffman, 1991) where both rainfall and temperatures are high (Gyllenhaal *et al.*, 1991), would have been particularly susceptible to chemical weathering. Resultant drawdown of atmospheric CO_2, together with a slightly weak sun, could have brought about glaciation even at low latitudes (Worsley and Kidder, 1991). Glaciation could eventually have been terminated by build-up of CO_2 due to negative feedback (Young, 1991a). The resultant sediments would reflect the tropical location of many of the depocentres. In this way, the 'anomalous' occurrence of warm climate indicators such as dolomites between glacial formations can be explained. However, as noted above, dolomites within glaciogenic formations are largely clastic and reflect the cannibalization of earlier-deposited carbonate units during uplift of the shoulders of rifts in which many Neoproterozoic glaciogenic successions are preserved (Schermerhorn, 1974; Yeo, 1981; Young, 1989).

The answer to the enigma of low latitude glaciation lies in further detailed facies studies of Neoproterozoic strata to determine their precise depositional and tectonic setting, paleoclimatic modelling using general circulation models and better age dating to determine correlative or diachronous glaciation. However, before the theory of low latitude glaciation can be accepted, a first and critical requirement is the need for more paleomagnetic investigations of remanence acquisition in Neoproterozoic sediments and the clear demonstration that low latitude magnetization is primary. Geochemical studies, to determine the degree of weathering undergone by 'interglacial' clastic units, could also contribute to resolution of this problem.

Early and Late Paleozoic glaciations

Caputo and Crowell (1985) argued that the pattern of Early and Late Paleozoic glaciation of Gondwana could be explained by the migration of the supercontinent across the South Pole. The Late Ordovician south pole was sited over northern Africa and there is a well-defined record of glaciation across North Africa and Saudi Arabia that identifies an ice sheet similar in size to that of the present day Antarctic ice sheet (Beuf *et al.*, 1971; Deynoux, 1985a,b; Vaslet, 1990). Glacio-eustatic sea-level changes, at about 440 Ma, have been held responsible for a series of well-defined extinction events in marine fauna (Brenchley, 1989; Fig. 1.5).

A major glacial episode at *c.* 440 Ma (Figs. 1.4, 1.6), is recorded in Late Ordovician strata (predominantly Ashgillian) in West Africa (Tamadjert Formation of the Sahara), in Morocco (Tinduf Basin) and in west-central Saudi Arabia, all areas at polar latitudes at this time (Vaslet, 1990). Less well-dated and poorly understood deposits occur in Scotland (Macduff tillite), Ireland (Maumtransa Formation), Normandy (Tillite de Feugeurolles) and Spain and Portugal. Late Ordovician deposits that may record small ice centres in eastern North America are represented by the Halifax Formation of Nova Scotia and Gander Bay tillites of Newfoundland. The sedimentology of these deposits is not well known and is probably non-glacial (Long, 1991). Other well-constrained Ordovician deposits occur in South Africa (Pakhuis Formation).

From the Late Ordovician to the Early Silurian the centre of glaciation moved from northern Africa to southwestern South America. The Cancaniri Formation of Argentina, Bolivia and Peru and Iapo and Furnas Formations of the Parana Basin, Brazil and the Trombetas Group of the Amazonas Basin may record glacially influenced deposition in middle to high latitudes as a result of polar wander (Fig. 1.7) but these deposits are not well-studied as yet (Grahn and Caputo, 1992). From the mid-Silurian to the early Late