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A fundamental Precambrian–Phanerozoic shift in earth's glacial style?

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Abstract

It has recently been found that Neoproterozoic glaciogenic sediments were deposited mainly at low paleolatitudes, in marked qualitative contrast to their Pleistocene counterparts. Several competing models vie for explanation of this unusual paleoclimatic record, most notably the high-obliquity hypothesis and varying degrees of the snowball Earth scenario. The present study quantitatively compiles the global distributions of Miocene–Pleistocene glaciogenic deposits and paleomagnetically derived paleolatitudes for Late Devonian–Permian, Ordovician–Silurian, Neoproterozoic, and Paleoproterozoic glaciogenic rocks. Whereas high depositional latitudes dominate all Phanerozoic ice ages, exclusively low paleolatitudes characterize both of the major Precambrian glacial epochs. Transition between these modes occurred within a 100-My interval, precisely coeval with the Neoproterozoic–Cambrian "explosion" of metazoan diversity. Glaciation is much more common since 750 Ma than in the preceding sedimentary record, an observation that cannot be ascribed merely to preservation. These patterns suggest an overall cooling of Earth's longterm climate, superimposed by developing regulatory feedbacks involving an increasingly complex biosphere.

Keywords: Glaciation; Phanerozoic; Precambrian; Paleozoic; Tectonics; Paleoclimate; Climate feedbacks

1. Introduction

At a first qualitative glance, the geological principle of uniformitarianism appears fairly well applied to climatically sensitive sedimentary deposits of the last 200 My, within the post-Pangean era. Modern lateritic and bauxitic soils are concentrated deep within the tropics where precipitation and temperature are highest (Thomas, 1994); ancient red, hematitic sediments are also concentrated toward the paleo-equator (Nicholas and Bildgen, 1979). Evaporites form in modern desert belts, which are centered around 30° North and South latitudes because mean tropospheric circulation there traverses the descending portions of Hadley cells (Warren, 1999); this trend continues back into Pangean times (Parrish et al., 1982). Carbonate sediments are concentrated within today's tropics, and Mesozoic–Cenozoic carbonate rocks occupy a similarly restricted latitudinal range (Ziegler et al., 1984), despite numerous independent lines of evidence for warmer global climate in the Cretaceous (Barron and Washington, 1982). Consistent with a largely ice-free global climate, Mesozoic sedimentary rocks lack abundant evidence for continental glaciation (Frakes

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et al., 1992, p. 74–77; Eyles, 1993, p. 156–159; Chumakov and Frakes, 1997; Price, 1999).

These conclusions are highly robust because the continents can be reconstructed to their late Pangean paleogeography with great precision, using several independent techniques. As we venture further back in time, the only demonstrated quantitative method for reconstructing continental positions is paleomagnetism. Landmark studies in the 1950s and 1960s, already exploring this young technique to some of its broadest capabilities, showed that the ancient distributions of climatically sensitive rocks throughout the Phanerozoic Eon remained broadly consistent with modern latitudinal trends (Irving, 1956; Opdyke, 1962; Irving and Briden, 1962; Briden and Irving, 1964). As better chronostratigraphic and paleomagnetic constraints upon those rocks have become available during recent years, there have been few systematic attempts to reproduce or revise the initial conclusions. The motions of individual continents or regions have been estimated by lithological climate indicators (e.g., Caputo and Crowell, 1985; Scotese and Barrett, 1990; Witzke, 1990), with qualitative allusions to general paleomagnetic support (Scotese and Barrett, 1990; Van der Voo, 1993). Likewise, many paleomagnetically derived reconstructions have included maps of lithological climate indicators (e.g., Kent and Van der Voo, 1990; Torsvik and Van der Voo, 2002). One recent study quantitatively investigated the glacial and paleomagnetic records of Paleozoic Gondwanaland (Smith, 1997). Yet none of these more recent studies has been undertaken on a global scale, systematically and quantitatively evaluating the distributions of one or more climate indicators.

This study investigates the depositional latitudes of glacial deposits through all of Earth history. It is motivated by a recently quantified paleoclimatic conundrum, that paleomagnetic data from glaciogenic deposits of the Neoproterozoic Era indicate a predominance of near-equatorial determinations (Evans, 2000). Most surprisingly, not a single, reliable depositional latitude poleward of 60° is indicated. One of the most reliable results, from the Elatina Formation and related units in the Flinders Ranges of South Australia, has been reproduced by several groups during 15 years of intense scrutiny and implies glaciers at sea-level, almost precisely on the paleoequator (reviewed by Evans, 2000).

Only three proposed explanations of these data have survived to date: the snowball Earth (Kirschvink, 1992; Hoffman et al., 1998), the “soft” snowball or “slushball” model (Hyde et al., 2000), and the high-obliquity hypothesis (Williams, 1975a, 1993; Williams et al., 1998). Each of the models faces specific challenges, the details of which cannot be described fully in this contribution. Let it suffice that any successful model of Neoproterozoic low-latitude glaciation must include a corollary of why climate zones appear to be distributed “normally” throughout much if not all of the Phanerozoic Eon, and when and how the dramatic transition transpired. Results of the present study underscore the rapidity and intriguing timing of the fundamental transition between Earth’s Precambrian and Phanerozoic glacial styles, and point toward specific changes in processes or boundary conditions that could have caused it.

2. Methods of analysis

The present analysis seeks to quantify glacial latitudes through time, using a common method of measurement for both recent (Neogene) and more ancient environments. Ancient deposits are assigned paleolatitudes according to a recent synthesis of the Gondwanaland Paleozoic apparent polar wander (APW) path (McElhinny et al., 2003). Although other APW paths have been proposed for that continent (e.g., Schmidt et al., 1993; Grunow, 1999; Torsvik and Van der Voo, 2002), the new path has included the most updated tectonic constraints upon autochthoneity of the paleomagnetically sampled areas, in particular eastern Australia. In addition, it will be shown below that this Gondwanaland APW path generates a consistent pattern of glacial latitudes through Paleozoic time. This study also considers individual paleomagnetic constraints on Paleoproterozoic and Archean glaciogenic formations, for which reliable APW paths are generally not available.

The global distributions of modern and ancient glaciogenic deposits can be quantified by binning occurrences of similar age into equal-area grid cells and computing histograms of latitudes from the present and paleomagnetic coordinate systems. For a meaningful comparison, modern deposits must be restricted to glaciomarine occurrences, for they most closely mimic the likely preservation potential of ancient sedimentary environments. Anderson (1983) compiled the global record of Miocene–Pleistocene glaciomarine sediments, including units containing dropstones or ice-rafted debris from melted icebergs. According to this global map of such deposits, points have been selected from continental shelves at a spacing of 2° latitude by 5° longitude. This array has then been binned into a set of equally spaced points occupying the centers of triangles in icosahedral geodesic grids, using the computational technique created by Moore (1998).

For the Devonian–Permian and Ordovician–Silurian intervals, biostratigraphic age constraints on widespread glaciogenic sedimentary rocks (Hambrey and Harland, 1981) are generally adequate to allow paleolatitudinal determinations according to paleomagnetic APW paths from the various continents. Most Paleozoic glaciogenic deposits lay on the Gondwanaland supercontinent, and a precise set of reconstruction parameters plus a recently updated APW path are used for estimating glacial paleolatitudes on that landmass (McElhinny et al., 2003). Paleolatitudes for the other glaciated Paleozoic terranes are fairly well established, as discussed case-by-case in the next section.

Precambrian glaciogenic deposits cannot be binned together in the aforementioned manner because the absence of precise biostratigraphic ages precludes assumption of coeval deposition among geographically disparate formations. Consequently, each paleolatitude must be found from either a direct determination on the glaciogenic unit itself, or a conformably adjacent formation; or through application of a paleomagnetic pole derived elsewhere on the same continent, given a precisely demonstrated synchrony between the glacial deposit and the paleomagnetic result. Evans (2000) employed this analysis to the world's 80 or so previously identified glaciogenic deposits, and deduced paleolatitudes according to three qualitative levels of reliability—both in the sense of the paleomagnetic result and the confidence with which a truly glacial origin could be ascribed to each deposit. The present study uses a similar methodology to determine the paleolatitude distribution of Paleoproterozoic and Archean glaciogenic formations, far fewer in number than their Neoproterozoic counterparts.

3. Results

3.1. Miocene–Pleistocene

Geodesic binning of Miocene–Pleistocene glaciomarine sediments on submerged continental shelves produces latitude histograms shown in Fig. 1. Three grid spacings have been utilized, corresponding to the 10-ν, 16-ν, and 24-ν icosahedra (Moore, 1998). These create arrays of points at angular distances of approximately 3.8°, 2.5°, and 1.7°, respectively. As shown in Fig. 1, the general trends are not sensitive to these various grid spacings. The binned points correspond well with the general distribution of deposits mapped by Anderson (1983), although some points have been shifted off the continental shelf or onto exposed land due to the computational adjustments inherent in the method. Histograms are presented for both North (Fig. 1a) and South (Fig. 1b) hemispheres. Each histogram is, as expected, biased toward high latitudes, with a decrease in extremely polar occurrences simply due to decreasing polar surface areas of a latitude–longitude coordinate system. Geographic peculiarities of the present world are evident: wide Arctic Ocean repositories for glaciomarine sediment at 75–80°N are complemented by minimal preservation potential on the elevated East Antarctic icecap at equivalent southerly latitudes, and a large proportion of exposed boreal land at about 60°N is balanced by the narrow Antarctic continental margin and Southern Ocean at about 60°S. The Southern distribution (Fig. 1b) tails as low as 10–20° latitude; this represents the western coast of South America, where north-flowing ocean currents have led icebergs to drop glaciomarine debris far from their source areas (Anderson, 1983). Although these deposits do not indicate a proximal glacial source, they could potentially be preserved in the longterm rock record (in an accretionary complex) so they must be included in this comparative analysis.

3.2. Late Devonian–Carboniferous–Permian

Pre-Neogene glaciogenic rocks are enumerated in the seminal tome by Hambrey and Harland (1981). Some of the deposits' characteristics and age constraints presented in that summary remain valid today and are simply included in Table 1 without further discussion; however, there are many new items of

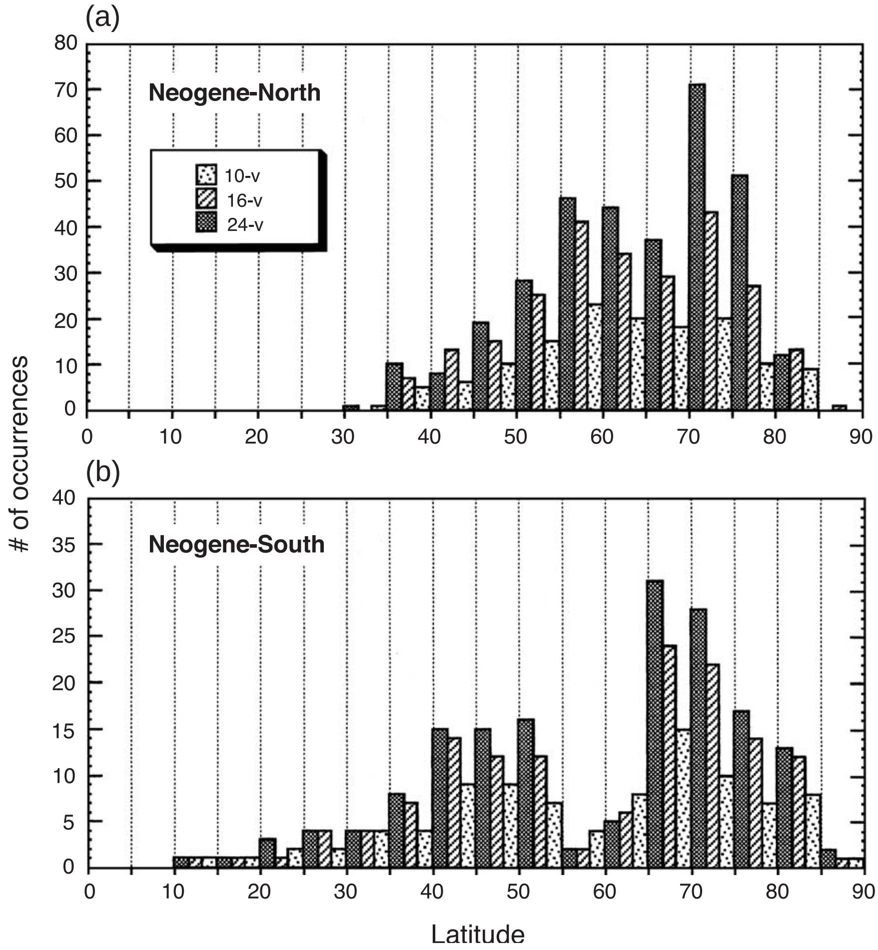

Fig. 1. Histograms of geodesic arrays (binning procedure using 10-ν, 16-ν, and 24-ν icosahedra after Moore, 1998, and described in text) representing the distribution of Miocene–Pleistocene glaciomarine deposits, compiled from Anderson (1983). Panel (a) shows the Northern hemisphere; panel (b) shows the Southern hemisphere.

information whose consequences for the purpose of this study are described in detail hereafter. The late Paleozoic interval witnessed numerous glacial advances and retreats, mainly occurring during Late Carboniferous–Early Permian time. However, local glaciation is recorded as early as Late Devonian, and ice persisted locally into the Late Permian (see below).

Beginning with Africa, the Lukuga and Lutôe Groups (Hambrey and Harland, 1981, codes A3 and A4, respectively) have been combined into a continuous repository, the Congo basin (Visser, 1997). This basin also includes correlative glaciogenic deposits in the Central African Republic (Visser, 1997; Crowell, 1999). Along the middle stretch of the Zambezi River, the Siankondobo Formation (A6) has been studied in detail by Nyambe and Utting (1997). Glaciogenic deposits of the Dwyka Group (A8/A9) in southern Africa are now categorized in terms of three depositional sequences (Visser, 1997). The second deglaciation sequence is represented by fossiliferous marine transgressive deposits, containing the diagnostic *Eurydesma* fauna of early Sakmarian ("Tastubian") age (Visser, 1997). The deglaciation sequences have now been dated precisely by U–Pb methods on ashbeds (Bangert et al., 1998; Stollhofen et al., 1999). Authors of those studies accepted previous timescales and thus assigned entirely Carboniferous ages to the Dwyka, in conflict with the direct biostratigraphic constraints (see also Key et al., 1998). I accept both the fossil evidence and the precise age control to suggest instead that the Tastubian transgression, found widely across Gondwanaland (Dickins, 1985, 1996; Wopfner and Cassyhap, 1997; Wopfner, 1999), occurred at 297 ± 2 Ma, older than previously thought but consistent with a new calibration of the Carboniferous–Permian boundary at about 301 Ma (Rasbury et al., 1998). In Tanzania, correlative glaciogenic strata have been identified in the Idusi Formation (Wopfner and Diekmann, 1996; Wopfner, 1999), not recognized by Hambrey and Harland (1981). Palynological data suggest that the Sakoa Group in southern Madagascar (A11) is older, with deglaciation completed by latest Carboniferous time (Wescott and Diggens, 1997). Such an older age is shared by glaciogenic rocks of the Northern Wadi Malik Formation in Egypt and Sudan (Klitzsch and Squyres, 1990; Wycisk et al., 1990). On the Arabian peninsula, Westphalian–Sakmarian glaciogenic deposits extend across Oman (Braakman et al., 1982; Levell et al., 1988) and into Yemen (Kruck and Thiele, 1983; El-Nakhal, 1984) and southwestern Saudi Arabia (C11).

Late Paleozoic glaciogenic rocks in Antarctica are found in Heimefrontfjella and along the entire length of the Transantarctic Mountains and Ellsworth Mountains (reviewed by Barrett, 1991; Collinson et al., 1994; Isbell et al., 1997). Microflora from Heimefrontfjella (B9), once adjacent to the Karoo basin in southern Africa, verify correlation of glaciogenic strata between the two regions in the earliest Permian (Lindström, 1995). Other Antarctic glaciogenic deposits are rather loosely constrained in age, generally ascribed to the Carboniferous–Permian interval but without more precise estimates (e.g., Miller and Waugh, 1991 (B11); Matsch and Ojakangas, 1992 (B14)).

Among the Gondwanaland-derived terranes now residing in Asia, the Himalayan margin of India (C12–16), the east-central Indian "Gondwana" basins (C17), and the Tengchong–Burma–Malay terrane (C18) bear a glacial record (Wopfner, 1996; Wopfner and Cassyhap, 1997). In Kashmir, the Golabgarh Formation (C13) has now yielded an Asselian age (Gaetani and Garzanti, 1991). The second of two pulses of late Paleozoic glaciation recorded in southern Tibet (C15/16) is likewise Asselian (Garzanti and Sciunnach, 1997). The earlier pulse in that region (Rakyang Formation), of unknown total lateral extent, dates from Visean to early Namurian (ibid.). Recently described Lower Permian glaciogenic units in the Tengchong and Baoshan terranes of southwest China (Wopfner, 1996, 1999; Yang, 1998; Wang et al., 2001; Ueno, 2003) probably correlate with similar, coeval deposits from Burma and the Malay Peninsula (C18).

Virtually every late Paleozoic basin in Australia records a glacial influence. Although Dickins (1985, 1996, 1997) has maintained that the ice age in this region is primarily limited to the Asselian Epoch, several authors have presented palynological evidence that at least several basins began recording glaciation in the Late Carboniferous. In particular, the glaciogenic Kulshill Group of the Bonaparte Gulf basin (D2) straddles the Carboniferous–Permian boundary (Wopfner, 1996, 1999). Glaciogenic influence in the Canning basin (D3), although more directly generated at its margins than toward the center (Eyles and Eyles, 2000), began as early as latest Carboniferous but reached a peak in the earliest Permian (O'Brien et al., 1998). The same age interval is determined for glaciogenic formations in the Carnarvon, Perth, Collie, and Officer basins (D4–D7) in southwestern Australia (reviewed by Eyles et al., 2002), as well as the Joe Joe Group (D9) in Queensland, and the Wynyard Formation (D15) in Tasmania (Jones and Truswell, 1992). An earlier onset of glaciation in eastern Australia has been proposed by Powell and Veevers (1987), citing Namurian fossils from the Spion Kop Conglomerate and correlative rocks (D10). In addition, the Seaham Formation in the southern Tamworth foldbelt (D11) contains *Levipustula* brachiopods, indicating a Namurian age (Dickins, 1996) that has recently been calibrated by U–Pb zircon chronometry at ca. 325–315 Ma (Roberts et al., 1995). Spectacular glacial pavements in South Australia preserve Early Permian continental ice-flow directions (Bourman and Alley, 1999), whereas in some eastern Australian basins an abundant glaciomarine dropstone record (Eyles et al., 1997) persists locally into the Late Permian (Dickins, 1996).

Documentation of the South American late Paleozoic glacial record has increased dramatically since the synthesis by Hambrey and Harland (1981). Major

Table 1
Paleomagnetic latitudes of Earth's pre-Mesozoic glacial record[a]

Code	Unit name	Region	Env.	Age	λ (°N)	ϕ (°E)	Rot.	λ'(°)
Late Devonian–Carboniferous–Permian								
Africa–Arabia								
A2	N'Khom Gp	Gabon	c	C2-P1	01	010	1	– 32
A3/A4	Lukuga/Lutôe	E. Congo	c	C2-P1	– 05	025	1	– 48
A5	Tilítica Gp	lower Zambezi	c	C2-P1	– 16	033	1	– 60
A6	Siankondobo	middle Zambezi	c	C2-P1	– 18	028	1	– 56
A7	Levanga Fm	Limpopo R.	c	C2-P1	– 22	031	1	– 60
A8/A9	Dwyka Gp	Kaoko/Kalahari	c/m	C2-P1	– 22	022	1	– 51
A10	Dwyka Gp	Karoo	c/m	C2-P1	– 30	026	1	– 56
A11	Sakoa Gp	Madagascar	c	C2	– 24	043	11	– 66
–	Idusi Fm	SW Tanzania	c	C2-P1	– 10	034	1	– 59
–	N. Wadi Malik	Egypt–Sudan	c/m	C2	18	026	2	– 41
A33	Teragh Fm	Aïr, Niger	c	D3-C1	19	008	0	– 58
C11	Wajid/Haushi	Arabia	c	C2-P1	19	050	3	– 46
Antarctica								
B9	Beacon Spgp	Heimefrontfjella	c	P1	– 74	350	10	– 63
B10	Gale Fm	Pensacola Mts	c	P1	– 84	303	10	– 72
B11	(many names)	S. Transant. Mts	c	P1?	– 84	165	10	– 83
B12	Metschel Fm	S. Victoria Land	c	C2-P1?	– 78	160	10	– 81
B13	(unnamed)	N. Victoria Land	c	C2-P1?	– 72	163	10	– 76
B14	Whiteout Fm	Ellsworth Mts	m?	C-P?	– 79	275	12	– 65
Southern Asia								
C12	Tobra Fm	Salt Range	c?	P1	33	072	7	– 53
C13	Golabgarh Fm	Kashmir	m	C2?-P1	33	076	7	– 57
C15/16	Rangit Fm	Nepal	m	C2-P1	28	090	7	– 72
–	Rakyang Fm	S. Tibet	m	C1	28	087	7	– 67
C17	Talchir Fm	E. India	c/m	C2-P1	22	084	7	– 71
C18	(many names)	Burma–Malay	m	C2-P1	13	098		– 42[b]
–	(several names)	SW China	m	C2-P1	25	099		– 47[b]
Australia								
D2	Keep Inlet	Bonaparte Gulf b.	c?	C2-P1	– 15	129	8	– 56
D3	Grant Gp	Canning basin	c/m	C2-P1	– 20	124	8	– 61
D4	Lyons Fm	Carnarvon basin	c/m	C2-P1	– 25	115	8	– 65
D5	Nangetty Fm	Perth basin	c/m	C2-P1	– 29	115	8	– 67
D6	Stockton Fm	Collie basin	c	C2-P1	– 33	116	8	– 73
D7	Paterson Fm	Officer basin	c	C2-P1	– 26	126	8	– 67
D8	(many names)	S. Australia	c	P1	– 31	137	8	– 73
D9	Joe Joe Gp	Queensland	c/m	C2-P1	– 24	148	8	– 59
D10	Spion Kop Cgl.	N. Tamworth	c	C1-2	– 31	151	8	– 58
D11	Seaham Fm	S. Tamworth	c	C1-2	– 33	152	8	– 58
D12	Macdonald's Cr.	N. Sydney basin	c	P1	– 33	150	8	– 70
D13	Talaterang Gp	S. Sydney basin	c/m	P1	– 35	150	8	– 72
D14	(several names)	Victoria	c/m	P1	– 37	144	8	– 76
D15	Wynyard Fm	Tasmania	c/m	C2-P1	– 42	146	9	– 72
South America								
G6a	Batinga Fm	Sergipe–Alagoas	c	C2-P1?	– 10	323	4	– 31
G6b	Pimenta Bueno	Rondônia	c	C1?	– 12	298	4	– 37
G6c	Jaurú Valley	Mato Grosso	c	C1	– 16	301	4	– 41
G7	Itararé Subgp	Paraná–Chaco b.	c/m	C2-P1	– 24	308	4	– 36
G8	(many names)	Tarija basin	c/m	C1(-C2?)	– 18	296	4	– 37
G9	Charata/Ordoñez	Chaco basin	c	C2-P1	– 28	300	5	– 36
G12/13	(many)	Paganzo basin	c/m	C1-2	– 31	293	5	– 32
G14	Sauce Grande	Sierra d.l. Ventana	m	C2-P1	– 39	300	5	– 45

Table 1 (*continued*)

Code	Unit name	Region	Env.	Age	λ (°N)	ϕ (°E)	Rot.	λ'(°)
Late Devonian–Carboniferous–Permian								
South America								
G15	Tepuel Gp	Patagonia	c/m	C1-2	– 43	289	6	– 39
G17	Lafonian Fm	Falkland Islands	c/m	C2-P1?	– 52	301	13	– 64
–	Sernambi/Faro	Amazon basin	?	C1	– 04	304	4	– 39
–	Poti Fm	Parnaíba basin	c	C1	– 06	316	4	– 50
G18	Curuá Fm	Amazon basin	m	D3	– 04	304	4	– 67
G19	Cabeças Fm	Amazon basin	c/m	D3	– 06	316	4	– 73
Arctic regions								
C8	(unnamed)	E. Siberia	m	C2-P1	64	150		+43[c]
C9	(many names)	E. Siberia	m	P2	64	145		+59[c]
–	Trold Fiord Fm	Ellesmere Island	m	P2	78	245		+41[d]
Late Ordovician–Silurian								
Africa–Arabia								
A1	Endaga Arbi Fm	N. Ethiopia	c	O-S	14	039	2	– 53
A12	Upper 2nd Bani	Tindouf basin	?	O3-S1	30	354	0	– 78
A13	Tichit Gp	Taoudeni basin	c/m	O3	18	350	0	– 80
A14	Waterfall Fm	Sierra Leone	m	O3-S1	10	348	0	– 74
A15	Tamadjert Fm	Hoggar/Tibesti	c/m	O3-S1	23	009	0	– 82
–	(many names)	E. Sahara	c	O3-S1	22	025	2	– 67
C19	Zarqa/Ammar	Arabia	c	O3-S1	26	042	3	– 53
A17	Pakhuis Fm	W. Cape belt	c/m	O3	– 33	020	1	– 34
Iapetan–Rheic terranes								
E7–E9	(many names)	Armorica	m	O3	49	005		– 39[e]
F9	Stoneville Fm	Exploits subzone	m	O3-S1	49	305		– 20[d]
South America								
G20	Nhamundá Fm	Amazon basin	c/m	S1	– 05	299	4	– 80
G22	Cancañiri/Zapla	Bolivia	m	O3-S1	– 19	294	4	– 47
G23	R.Ivaí/Vila Maria	Paraná basin	c/m	O3-S1	– 21	307	4	– 54
–	Ipu/Tacaratu Fm	Parnaíba basin	c/m	S1	– 06	318	4	– 69
–	Don Braulio Fm	Argentine Precord.	m	O3-S1	– 31	291	5	– 37
Cambrian–Early Ordovician								
F7	Halifax Fm	Nova Scotia	m	O1	45	295	14	– 74
F17	Florida Mts	New Mexico	ng?	Cm3	32	252		– 01[d]
A29	Schwarzrand	S. Namibia	c	Pt3-Cm1	– 27	018	1	– 46
A19	Jbeliat/Fersiga	Taoudeni basin	c/m	(Pt-Cm)	18	350	0	30–70[f]
Neoproterozoic[f]								
D16	Egan Fm	Kimberley	m	~560	– 19	127		21
–	Johnnie Fm	Basin and Range	m	~560	36	243		01
F19	Squantum Fm	Boston basin	m	595–550	42	289		55
F20	Gaskiers Fm	Avalon	m	607–565	47	307		31
D21	Elatina Fm	S. Australia	c/m	650–550	– 32	138		03–09
C33	Tereeken Fm	Tarim	c/m	Pt3	41	090		08
E12	Vestertana Gp	N. Norway	c/m	Pt3	70	028		33
D16	Walsh Fm	Kimberley	c	Pt3	– 17	126		45
A29	Chuos Fm	N. Namibia	m	~750	– 20	015		10
C33/35	Nantuo Fm	S. China	c/m	~750	28	113		35–40
F11	Rapitan Gp	Mackenzie Mts	c/m	≤ 770	64	230		06
F13	Toby Fm	Omineca belt	m	~750	50	243		08
F18	(many names)	Central Appalach.	m	740–720	37	279		20–21

(continued on next page)

Table 1 (*continued*)

Code	Unit name	Region	Env.	Age	λ (°N)	ϕ (°E)	Rot.	λ'(°)
Paleoproterozoic								
A31	Makganyene	South Africa	c/m	≥ 2220	−27	026		11
D23	Meteorite Bore	Pilbara, Australia	m	≤ 2450	−22	116		~05
E35	Sariolian	Fennoscandia	c?	≤ 2450	64	032		07–27
F30–33	Huronian	Great Lakes region	c/m	2450–2200	47	248		03

Locality codes from Hambrey and Harland (1981): A, Africa; B, Antarctica; C, Asia; D, Australia; E, Europe; F, North America; G, South America. Env.(environment): m, marine; c, continental; ng, nonglacial. Ages: C(1,2), Carboniferous (Lower, Upper); Cm(1,3), Cambrian (Lower, Upper); D3, Upper Devonian; P(1,2), Permian (Lower, Upper); Pt, Proterozoic; Pt3, Neoproterozoic; O(1,3), Ordovician (Lower, Upper); S(1), Silurian (Lower); Precambrian numerical ages in Ma. Rot., Gondwanaland total-reconstruction rotations (0–11 from McElhinny et al., 2003; 12–13 from Grunow et al., 1991; 14 from Lottes and Rowley, 1990): 0 (NW Africa fixed), 1 (09.3, 005.7, −7.8), 2 (19.2, 352.6, −6.3), 3 (26.2, 011.2, −14.2), 4 (53.0, 325.0, +51.0), 5 (48.8, 324.9, +52.8), 6 (−49.2, 144.2, −54.1), 7 (26.7, 037.3, −69.4), 8 (−28.1, 293.2, +52.1), 9 (−24.1, 294.3, +51.7), 10 (−12.4, 326.2, +53.3), 11 (−14.9, 277.6, +15.7), 12 (−51.2, 101.3, −77.7), 13 (−45.3, 349.2, 156.3), 14 (61.3, 343.2, +52.8). λ, present latitude; ϕ, present longitude; λ', paleolatitude.

[a] Except where otherwise noted, paleolatitudes are from numerically calibrated Gondwanaland mean poles in McElhinny et al. (2003): Pt3-Cm1 (550 Ma); O1 (480 Ma); O3, O3-S1, O-S (455 Ma); S1 (425 Ma); D3, D3-C1 (360 Ma); C1 (340 Ma); C1-2, C2 (320 Ma); C2-P1, C-P (300 Ma); P1 (280 Ma); P2 (260 Ma).

[b] From Huang and Opdyke (1991); see text.

[c] Interpolated from Khramov and Ustritsky (1990).

[d] Laurentian paleolatitudes from Van der Voo (1993).

[e] Armorican paleolatitudes from Tait et al. (1995).

[f] Paleolatitudes (absolute-value), ages, and reliability summarized in Evans (2000).

reviews by Crowell (1983), Caputo and Crowell (1985), Eyles (1993), França et al. (1995), Eyles et al. (1995), and López-Gamundí (1997) have provided a tectonostratigraphic context for the various basins. Large glacially influenced regions in the northern part of the continent include the Parnaíba, Amazonas, and Solimões basins (Caputo and Crowell, 1985), unrecognized among earlier compilations. Recent palynological data from the Amazon basin imply an early Visean age for the glaciogenic Faro Formation (Loboziak et al., 1998), confirming earlier estimates (Caputo, 1985). Other Early Carboniferous glacial records are found in the Pimenta Bueno basin (G6b) in Rondônia, the Tarija basin (G8) in Bolivia–Argentina (Caputo and Crowell, 1985; López-Gamundí, 1997), the Paganzo basin (G12/13) in west-central Argentina (Powell and Veevers, 1987; González, 1990; Limarino and Gutierrez, 1990; Eyles et al., 1995; López-Gamundí, 1997), and the Tepuel basin (G15) in Patagonia (Powell and Veevers, 1987; González-Bonorino, 1992; López-Gamundí, 1997). A post-glacial transgression across these regions was inhabited by *Levipustula* brachiopods, which López-Gamundí (1997) has used for correlation of "Glacial Episode II," following an earlier ice age in the Late Devonian (see also Caputo, 1985; Caputo and Crowell, 1985). López-Gamundí (1997) has included in "Glacial Episode III" the following units dating from the Carboniferous–Permian transition: Itararé Group (G7) of the Paraná basin, Sauce Grande Formation (G14) of the Sierra de la Ventana, and Lafonian Formation (G17) of the Falkland Islands. Also included in this correlation are diamictites of the Chaco basin (G9), in contrast to earlier estimates of a mid-Carboniferous age (Hambrey and Harland, 1981; Caputo and Crowell, 1985; Eyles et al., 1995) but supported by subsequent palynological data (Winn and Steinmetz, 1998). As with the correlative and depositionally connected Dwyka basins of southern Africa, the Late Carboniferous–Early Permian ice ages of South America were followed by a *Eurydesma*-bearing transgression (González, 1997; López-Gamundí, 1997; López-Gamundí and Rossello, 1998). Northwest-directed ice-flow directions recorded throughout the Paraná basin testify to the intimately linked basin dynamics of late Paleozoic South America and southern Africa (Gesicki et al., 2002).

The record of Carboniferous–Permian ice extends to the northern regions of Pangea, now preserved in Siberia and Canada. The Siberian deposits can be grouped into two alleged glacial intervals, Carboniferous–Permian (C8) and Upper Permian (C9). The latter units have been recently attributed not to glacial influence, but merely to seasonal coastal ice (Ziegler et al., 1997). Similar claims could be made regarding the Upper Permian dropstones in southern Ellesmere Island, recorded by Beauchamp (1994).

The entire global record of Devonian glaciation is toward the end of the period (Famennian), localized in northern Africa (A33) and northern South America (G18–19; Caputo, 1985; Caputo and Crowell, 1985). The age of the African unit may extend into the Early Carboniferous. These regions should have lain close to the pole if the McElhinny et al. (2003) paleomagnetic synthesis is accepted. Note that the Furnas Formation of the Paraná basin, previously considered as Devonian in age (G32), has subsequently been assigned to the Early Silurian (Caputo, 1985; Caputo and Crowell, 1985).

Some of the late Paleozoic deposits described in Hambrey and Harland (1981) are omitted here. These include one with age constraints so poor that a paleolatitude is inestimable (Jinkeng Ridge, North China; C10), one of Neoproterozoic age but misassigned to the late Paleozoic (Blaini, Lesser Himalaya; C14) and those originally considered to be of nonglacial origin (several deposits, Chile; G10–11).

Ages of the selected deposits have been categorized into the brackets chosen by McElhinny et al. (2003) for determining mean Gondwanaland pole positions (Fig. 2). Arrays of points at $1\times1^\circ$ spacing (1° latitude by 5° longitude in the Transantarctic Mountains) have been chosen to represent the spatial extents of the glaciogenic basins, as was done for the Miocene–Pleistocene glaciomarine sediment, above. All Gondwanaland deposits of the same age have been restored to NW African coordinates according to the rotation parameters listed in McElhinny et al. (2003), except for those from the Ellsworth Mountains and Falkland Islands, for which the rotation parameters given by Grunow et al. (1991) are used. The rotated arrays of points are then binned into icosahedral geodesic grids of the 10-ν, 16-ν, and 24-ν spacings (Moore, 1998), as was done above for the Miocene–Pleistocene group.

Glaciogenic rocks from Sibumasu (Burma-Malay, Baoshan, and Tengchong terranes) are directly overlain by basalts, which have yielded robust paleomagnetic constraints (Huang and Opdyke, 1991). These results show a consistency of magnetic inclination but a wide discrepancy in declination, likely the result of vertical-axis rotation of the sampled regions near the Yunnan–Burmese syntaxis of the India–Asia collision. The present reconstruction accepts the inclination data for paleolatitude of the northern sector of

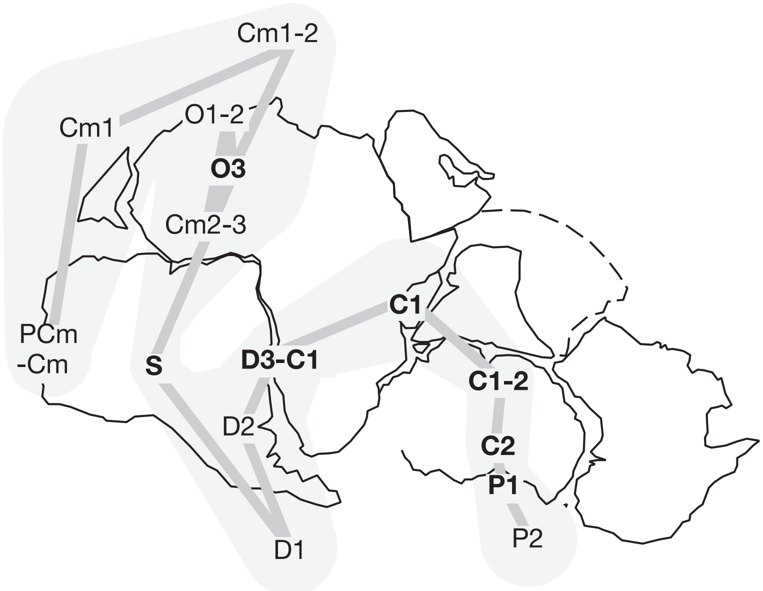

Fig. 2. Gondwanaland Paleozoic APW path, after McElhinny et al. (2003). Width of the shaded region indicates the degree of uncertainty in the paleomagnetic means. Abbreviations as in Table 1; boldface indicates ages of widespread glaciogenic deposits.

Sibumasu, and arbitrarily aligns the terrane parallel to the northern Australian continental margin, as supported by tectonostratigraphic and biogeographic evidence (Nie et al., 1990). The terrane is then included with Gondwanaland for binning.

Of all the glaciogenic formations thus far described, only those in eastern Siberia and northern Canada represent the late Paleozoic northern hemisphere. Those groups of deposits were produced at intermediate to subpolar latitudes, according to summarized paleomagnetic data from the Omolon massif (Khramov and Ustritsky, 1990) and Laurentia (Van der Voo, 1993). The remainder of formations represent a rich diversity of glacial and tectonic environments of the ancient southern hemisphere. To first approximation, the centers of Gondwanaland glaciation track the position of the South pole as the continent drifted across latitudes (Caputo and Crowell, 1985). Except for the Spion Kop Conglomerate and Seaham Formation in the Tamworth belt of eastern Australia, Early Carboniferous (Namurian) glaciogenic units are confined to South America, deposited at fairly low latitude in active tectonic settings. Such an environment is similar to that which hosts moderately low-latitude, glacially influenced marine sedimentation today, where reworked Andean glaciogenic debris is deposited on the steep continental slope.

Late Devonian to Early Permian glaciogenic deposits, representing the most severe of the Paleozoic ice ages (Dickins, 1985; Powell and Veevers, 1987; González-Bonorino and Eyles, 1995), were widely distributed across moderate to polar southern latitudes (Fig. 3). Once again, the various grid spacings of the binning algorithm do not produce appreciably different results; this is because all of the grid resolutions are lower than the sample resolution of 1° or less spacing (see Moore, 1998, p. 970). For comparison with Precambrian deposits, a fourth histogram of occurrences simply by region or name (after Hambrey and Harland, 1981; corresponding to rows in Table 1), is also shown for each time interval. Peaks are similar only in the broadest sense to the geodesically binned modes, indicating the utility of the equal-area binning technique.

An intriguing pattern emerges when different time slices of the late Paleozoic glaciations are contrasted. The onset of glaciation, recorded mainly in Late Devonian basins of northern South America, is characterized by high-latitude deposits. By the Early Carboniferous, the locus of glaciogenic deposition shifted to moderate latitudes as the pole moved

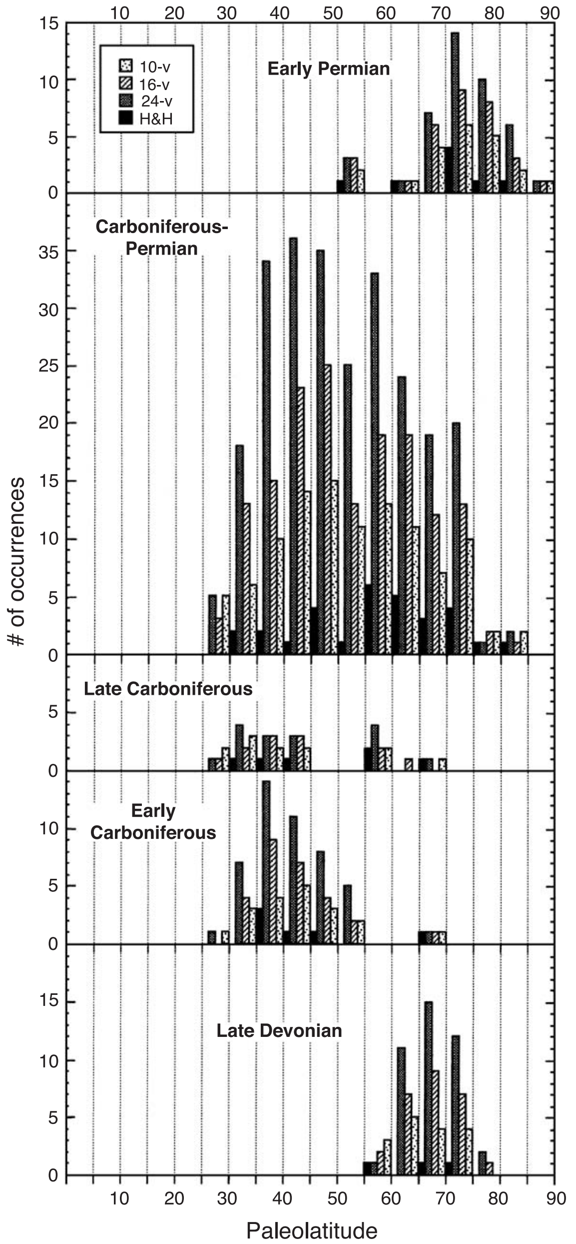

Fig. 3. Histograms of paleolatitudes for late Paleozoic (Late Devonian–Early Permian) glaciogenic deposits. Although arranged temporally, the vertical axis is not to temporal scale. Symbols as in Fig. 1, but also: H&H = binning by entry in the compilation of Hambrey and Harland (1981).

rapidly away from those basins (Fig. 3). This could divulge the growth of a large ice cap in Early Carboniferous time, as previously postulated by González-Bonorino and Eyles (1995). Such an early (Visean) onset of significant ice cover is consistent with independent records such as eustatic oscillations in paleo-tropical carbonate successions (Smith and Read, 2000; Wright and Vanstone, 2001), although it is slightly older than most interpretations of seawater isotope variations (Bruckschen et al., 1999; Mii et al., 1999, 2001; Saltzmann, 2003). The last waning stages of Gondwanaland glaciation, in the Permian, were restricted once again to high paleolatitudes (Fig. 3). Thus, the broad trends of late Paleozoic glaciogenic deposits across Gondwanaland appear to indicate an expanding ice cap between Late Devonian and Early Carboniferous time, followed by contracting glaciated regions in Late Carboniferous to Permian time. The details of this first-order pattern will no doubt prove more complex as higher-resolution data are obtained.

It should be noted that central and southern Africa had already passed over the pole prior to deposition of the Dwyka Formation and its correlatives at the end of the Carboniferous. Therefore, the abrupt and widespread appearance of those deposits cannot be attributed solely to changes in latitude. Powell and Veevers (1987) and Eyles (1993) stressed the importance of tectonic factors in generating glaciers and preserving their debris. Nonetheless, subtropical latitudes of northwest Africa and the Amazon–Parnaíba region during this interval *can* explain why those areas were no longer glaciated during Late Carboniferous–Early Permian time; they had crossed a critical latitude threshold as the pole retreated farther toward eastern Australia (Fig. 3). Paleolatitude poleward of about 30° appears to be a necessary—but not sufficient—condition for the accumulation of glaciers in late Paleozoic time.

3.3. Late Ordovician–Silurian

Compared with the late Paleozoic glacial era, that of the Ordovician–Silurian interval is much more spatially and temporally restricted (Brenchley et al., 1994; Sutcliffe et al., 2000). At the time of previous compilations, glaciogenic deposits of that age were described only in the Appalachian–Variscide belt, northern Africa, and South America, with an isolated but well known occurrence in South Africa (Hambrey and Harland, 1981; Hambrey, 1985). Several newly recognized deposits can now be added to this list, along with new constraints on the previously known units. In northern Africa, the Endaga Arbi "Tillite" in northern Ethiopia (A1) is now regarded as Ordovician in age (Saxena and Assefa, 1983), possibly correlative with the widespread latest Ordovician glaciogenic deposits that sweep across northern Africa from Mauritania and the Hoggar massif (A13–A15; Deynoux, 1985; Ghienne and Deynoux, 1998; Paris et al., 1998; Underwood et al., 1998; Ghienne, 2003) through Jebel Uweinat in southwestern Egypt (Vaslet, 1990), into Arabia (C19; McGillivray and Husseini, 1992) and Jordan (Abed et al., 1993; Powell et al., 1994; Amireh et al., 2001). A particularly good graptolite record has been described from the Hodh of Mauritania, allowing precise dating of postglacial shales within the uppermost Ordovician Period (Underwood et al., 1998). Omitted from the north African entries in Hambrey and Harland (1981) is the Ajua Formation in Ghana (A16), whose age may be Neoproterozoic rather than mid-Paleozoic (Crowell, 1999). The southern African record of Late Ordovician glaciation is limited to a single outlying deposit, the Pakhuis Formation (A17), lying directly below an Ashgillian shale lagerstätte (Gabbott et al., 1998).

As with the Late Devonian–Permian ice age, the Ordovician–Silurian glacial interval has been increasingly recognized in South America during the last two decades. Caputo and Crowell (1985) reviewed this record among the entire continent, correlating all known lower Paleozoic glaciogenic units into a single episode coinciding with the Ordovician–Silurian boundary. Subsequent work in the northern Brazilian basins (Amazon, G20; and Parnaíba) has shown that glaciation there is represented by three intervals, all early Silurian in age (Grahn and Caputo, 1992; Caputo, 1998). In the southern part of the continent, including the Paraná basin (G23; Grahn and Caputo, 1992), Tarija basin (G22; França et al., 1995; Gagnier et al., 1996; Crowell, 1999), and San Juan province (Sanchez et al., 1991; Buggisch and Astini, 1993), glaciogenic units are older, laid down near the Ordovician–Silurian boundary.

Ordovician–Silurian glaciogenic deposits are locally abundant in the Appalachian–Variscide belt, belonging to terranes of the Iapetan and Rheic oceans (see Keppie and Ramos, 1999). Within the Bohemian Massif of central Europe, the Saxo–Thuringian "Lederschiefer" (E7) represents the Ashgillian glaciation (Storch, 1990), correlative with the diamictite-bearing Kosov Formation in the Prague basin (Brenchley and Storch, 1989) and numerous glaciomarine formations in Iberia (E9; Young, 1988, 1990; Robardet and Doré, 1988; Robardet et al., 1990; Brenchley et al., 1991). Long (1991) has discounted a glaciogenic origin for the Cosquer Formation, in western Brittany, but other outcrops of the Ashgillian "Feuguerolles" (E8) in Normandy retain at least an indirect glaciomarine influence (Doré et al., 1985).

Among the Iapetan terranes in Maritime Canada, controversy surrounds the alleged glaciogenicity of most of the deposits considered by Hambrey and Harland (1981). These deposits are considered here regardless, for at least an update of their tectonic and chronological settings. The Late Ordovician to Early Silurian Stoneville Formation (F9), of the Badger Group in the Exploits Subzone of central Newfoundland (Williams et al., 1995), lies northwest of the recently recognized Dog Bay Line and is thus interpreted to have been deposited at the Laurentian active margin (Williams et al., 1993; Mac Niocaill et al., 1997). To the southeast of the Dog Bay Line, a glaciogenic origin for diamictite bodies resting atop (apparently) the Middle Ordovician Davidsville Group (F8) has been challenged by Long (1991). The poor age constraints and questionable depositional setting for these units preclude their use in this study as indicators of broad trends in paleoclimate. British Iapetan deposits considered by Hambrey and Harland (1981) and Hambrey (1985) as possibly glaciogenic, are likewise omitted from this analysis. The Macduff Formation of Banffshire (E5; Stoker et al., 1999) may be Neoproterozoic in age (A. Prave, personal communication) rather than Ordovician as reported (Molyneux, 1998). In Connemarra, Ireland, the mid-Ordovician Maumtrasna Formation (E6) is probably non-glaciogenic (Graham, 1987).

As with the late Paleozoic group of deposits, the Ordovician–Silurian glaciogenic formations have been mapped into $1 \times 1°$ arrays of points, grouped by age, rotated to NW African coordinates (McElhinny et al., 2003), and binned into icosahedral geodesic grids (Moore, 1998). A direct paleomagnetic determination on Late Ordovician rocks of the Bohemian Massif (Tait et al., 1995) provides a paleolatitudinal constraint that is probably valid, to first-order, for all of the Variscan terranes constituting the Armorica microplate (see Keppie and Ramos, 1999). The Exploits arc had probably accreted to Laurentia by Late Ordovician time (Williams et al., 1993; Mac Niocaill et al., 1997), so the Stoneville Formation paleolatitude is constrained by the Laurentian apparent polar wander path (Van der Voo, 1993, p.78).

According to the Gondwanaland paleomagnetic APW path from McElhinny et al. (2003) and the Armorican result from Tait et al. (1995), the Late Ordovician–Early Silurian glaciation would appear to be restricted to polar-moderate latitudes of the southern hemisphere (Fig. 4). The Late Ordovician trend is bimodal, due to an abundance of mid-paleolatitude deposits in southern South America and South Africa (Table 1). The South American deposits are largely marine, sourced from active tectonic uplands (Ramos, 2000). The sole tropical outlier is the Stoneville Formation, whose depositional paleolatitude of about 20° is anomalously low but within the range of Neogene glaciomarine deposits sourced via equatorward currents from distant regions (Anderson, 1983; see above). It should be noted that the glaciogenicity of this unit is also disputed (Williams et al., 1995).

3.4. Cambrian–Early Ordovician

Earliest Paleozoic glaciogenic deposits are few in number, and commonly controversial in character or age. The most convincing in terms of a glacial influence is the so-called "triad" in the Taoudeni basin of West Africa (A19; Bertrand-Sarfati et al., 1995). As discussed at length in Evans (2000), however, the age of this unit is highly uncertain, and any terminal Proterozoic–Cambrian age is possible. Assignment of depositional paleolatitudes is made even more problematic by the fact that the West African sector of Gondwanaland was drifting rapidly during that interval (Evans, 1998). Within the uncertainties in age, paleolatitudes in the center of the basin could have been anywhere between about 30° and 70° (Evans, 2000).

The Meguma Zone of southern Nova Scotia, Canada, restores against northwest Africa in pre-Pangean reconstructions, and contains Ordovician

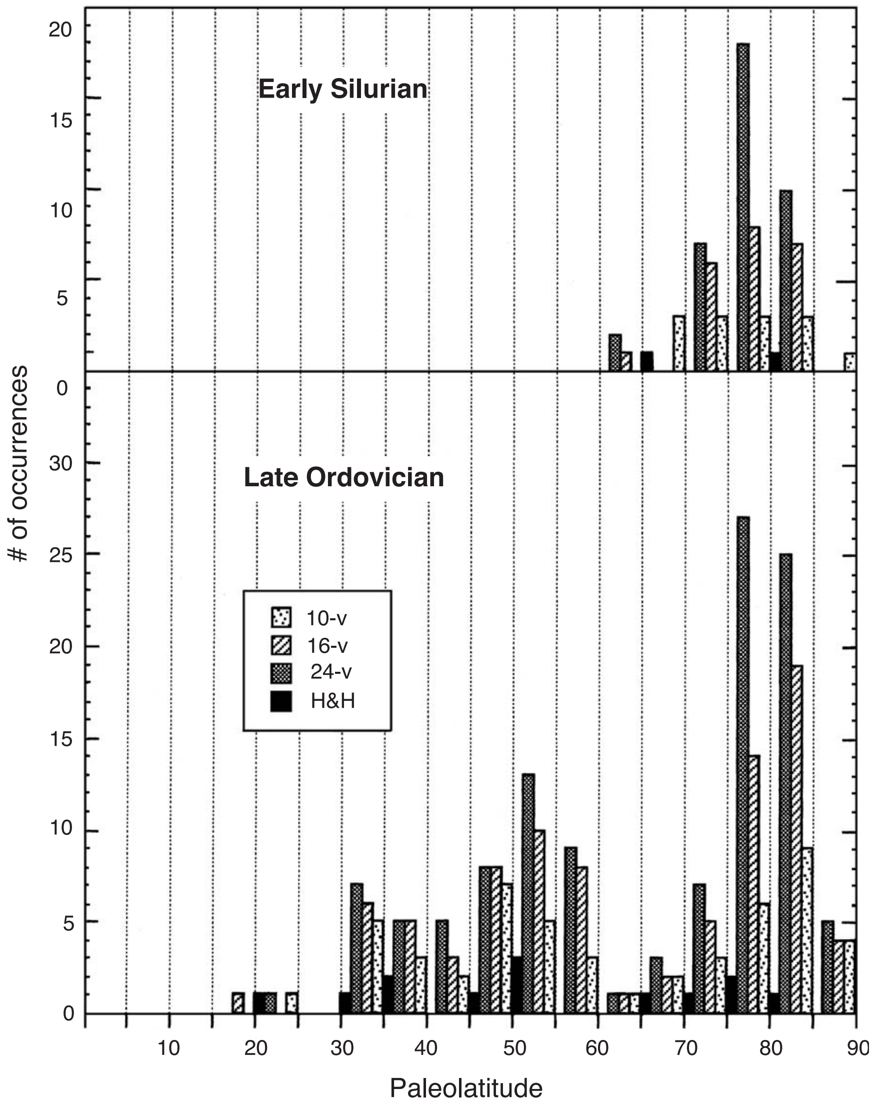

Fig. 4. Histograms of paleolatitudes for middle Paleozoic (Late Ordovician–Early Silurian) glaciogenic deposits. Symbols as in Fig. 3.

glaciogenic deposits (Schenk, 1995). Named the Rockville Notch Formation of the Halifax Group (F7), it has yielded a surprising Tremadocian age from acritarchs (W.A.M. Jenkins, in Schenk, 1995, p. 270)—too old to correlate with the Hirnantian ice age, yet too young to correlate with the "triad." In this case, restoration of Nova Scotia adjacent to Morocco according to Lottes and Rowley (1990) and utilization of the Early Ordovician pole from McElhinny et al. (2003) generates a paleolatitude of about 75°. It may be no surprise that this unit, deposited in an active tectonic setting with near-polar latitude, is the only convincing representative of Early Ordovician glaciation on Earth.

In Namibia, southwestern Africa, the terminal Proterozoic–Cambrian Schwarzrand Subgroup of the Nama Group contains two stratigraphic levels with alleged periglacial features (A29). Both are bracketed by precise U–Pb zircon ages between 550 and 543 Ma (Grotzinger et al., 1995), thus the

Precambrian–Cambrian boundary pole for Gondwanaland by McElhinny et al. (2003) is appropriate. These units were deposited within an active foreland-basin setting (Gresse and Germs, 1993) at moderate latitudes.

Other reputed Cambrian glaciogenic deposits are not well documented. In the Florida Mountains of southwestern New Mexico, an unnamed diamictite body (F17) lies between Ordovician sediments and an underlying crystalline complex with ages as young as 503 ± 6 Ma (Evans and Clemons, 1988). Direct paleomagnetic study on the basement granites (Geissman et al., 1991) produces results that are concordant with other Late Cambrian poles from Laurentia (Van der Voo, 1993), indicating a near-equatorial paleolatitude (Table 1). This would seem to create the only clear violation of the high-latitude Phanerozoic glacial paradigm, but a glaciogenic origin for the diamictite has not been demonstrated convincingly. In Bolivia, the Limbo Group (G24) is reported to have a partly glaciogenic origin, but its age is highly uncertain (Aceñolaza et al., 1982), preventing an indirect paleolatitudinal assignment from the Gondwanaland apparent polar wander path.

In summary, Cambrian–Early Ordovician glacial paleolatitudes are not well constrained. The most reliable paleomagnetic determinations on the most reliably glaciogenic of the formations generate moderate or polar paleolatitudes. A single near-equatorial result from southwestern New Mexico, demands further investigations regarding its glaciogenicity. The wide range of allegedly glaciogenic latitudes for this interval of time contrasts sharply with the consistently polar-centered spatial distribution of Earth's ice ages from the Late Ordovician to the present (Figs. 5 and 6), an elegant confirmation of both a uniformitarian (*s.l.*) climate regime and an axial-centric geomagnetic field during the last 450 million years (Smith, 1997).

3.5. Neoproterozoic

Depositional latitudes of Neoproterozoic glaciogenic deposits have been extensively reviewed by Evans (2000), and details will not be recapitulated here. As noted above, because age control upon these units is generally so poor and Neoproterozoic continental reconstructions are so uncertain (see Wingate et al., 2002), geodesic binning of outcrops is not very useful. Also, the limited number of even moderately constrained deposits obviates any level of latitudinal precision less than 10° for each determination. Therefore, the Neoproterozoic glacial paleolatitude histogram (Fig. 7) appears much less precise and complete than its Phanerozoic counterparts. Nevertheless, the chief observations are the abundance of near-equatorial glacial occurrences and the complete absence of any glaciogenic formation deposited poleward of 60° (Evans, 2000). This conclusion is in strong contrast to the previous analysis by Meert and Van der Voo (1994), whose assumptions of glacial ages and supercontinental reconstructions negated the lowest-latitude, direct paleomagnetic determinations upon the glaciogenic deposits.

3.6. Archean–Paleoproterozoic

Only a handful of alleged glaciogenic deposits remain from the first half of Earth history. Despite their limited number, however, glaciogenic units are present among most of the well preserved sedimentary successions of early Paleoproterozoic age (2.5–2.2 Ga), perhaps suggestive of an originally more widespread extent. The most famous of these is the Huronian Supergroup of the Great Lakes region in North America, which contains three glaciogenic horizons (F30–33; Young, 1983; Miall, 1983, 1985; Young and Nesbitt, 1985; Mustard and Donaldson, 1987; Ojakangas, 1988; Morey, 1996; Menzies, 2000). The uppermost level, the Gowganda Formation, and the directly overlying hematitic Lorrain Formation have yielded a paleomagnetic latitude of 03°, interpreted as primary by the authors of that study (Williams and Schmidt, 1997). In support of that inference are generally concordant paleomagnetic results from a hematitic paleosol directly beneath the sandstone (Schmidt and Williams, 1999), and generally consistent paleomagnetic results from 2.5–2.2 Ga mafic units of the Superior craton (reviewed by Buchan et al., 1998). However, as many as six paleomagnetic remanence components have been isolated from the Huronian glacial succession (Williams and Schmidt, 1997), and recent data have cast some doubt on the interpretation of the 03°-paleolatitude component as primary (Hilburn et al., 2002).

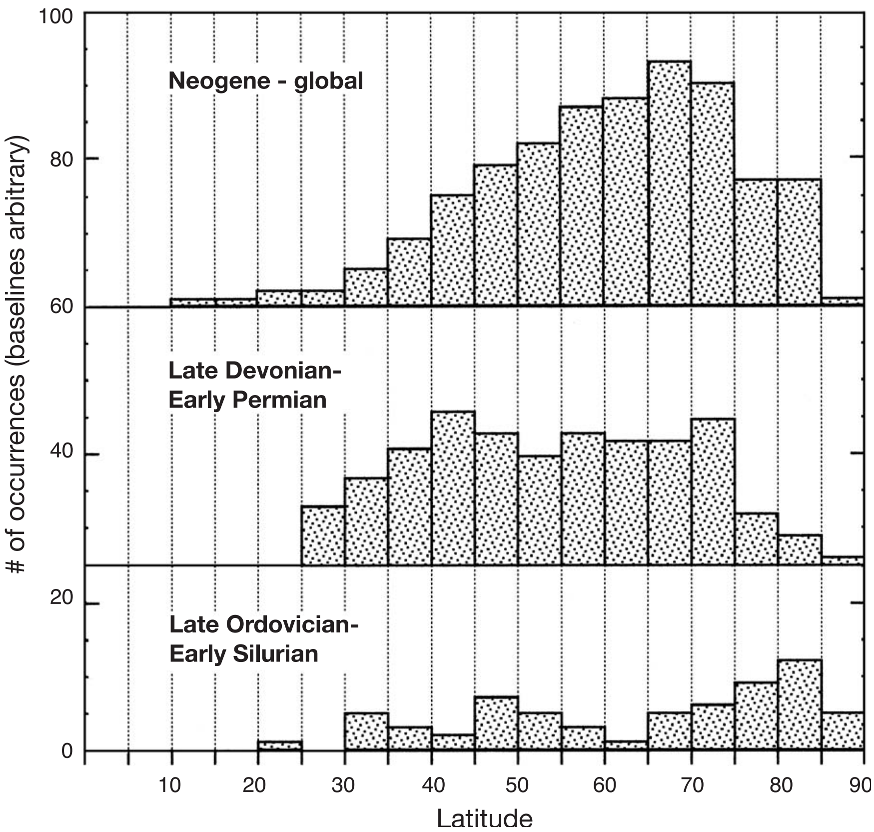

Fig. 5. Summary of paleolatitudes for three broad groups of Phanerozoic glaciogenic deposits, using 10-ν icosahedral geodesic binning and display at 5° latitude intervals.

No reliable palaeomagnetic results are available for other Paleoproterozoic glaciogenic deposits in North America, such as in the Snowy Pass Supergroup of southeastern Wyoming (F28; Karlstrom et al., 1983; Houston et al., 1992) or the Hurwitz Group, Nunavut, Canada (F27; Aspler and Chiaranzelli, 1997). Nonetheless, the two mafic dike swarms most closely bracketing the depositional age of the Hurwitz Group, the older Kaminak suite (2.45 Ga; L. Heaman, unpublished, cited in Heaman, 1997) and the probably slightly younger Tulemalu swarm (2.19 Ga; L. Heaman, unpublished, cited in Tella et al., 1997), have both yielded low paleolatitudes, interpreted as primary by the authors but without complete substantiation, of 02–20° (Christie et al., 1975; Fahrig et al., 1984). Diamictites in the Black Hills, South Dakota (F29) are of uncertain origin and very poorly constrained in age.

A low depositional latitude of 11° is obtained from the Ongeluk lavas that conformably overlie the Makganyene glaciogenic formation (A31) of South Africa (Evans et al., 1997). These data have contributed to a "Snowball Earth" interpretation of the diamictites, lavas (2222 ± 13 Ma; Cornell et al., 1996), and overlying Fe–Mn formations (Kirschvink et al., 2000). Correlative or slightly older strata in the northeastern part of the craton, the Timeball Hill Formation, contain diamictites and rhythmites interpreted as representing a glaciogenic setting associated with volcanism (Eriksson et al., 1994, 1995). Older possible glaciogenic deposits, perhaps by as much as 200 My, are found in the Duitschland Formation, a localized diamictite-bearing unit of the northeastern Transvaal (Bekker et al., 2001).

Early Paleoproterozoic glaciogenic deposits are found across the southwestern margin of the Pilbara

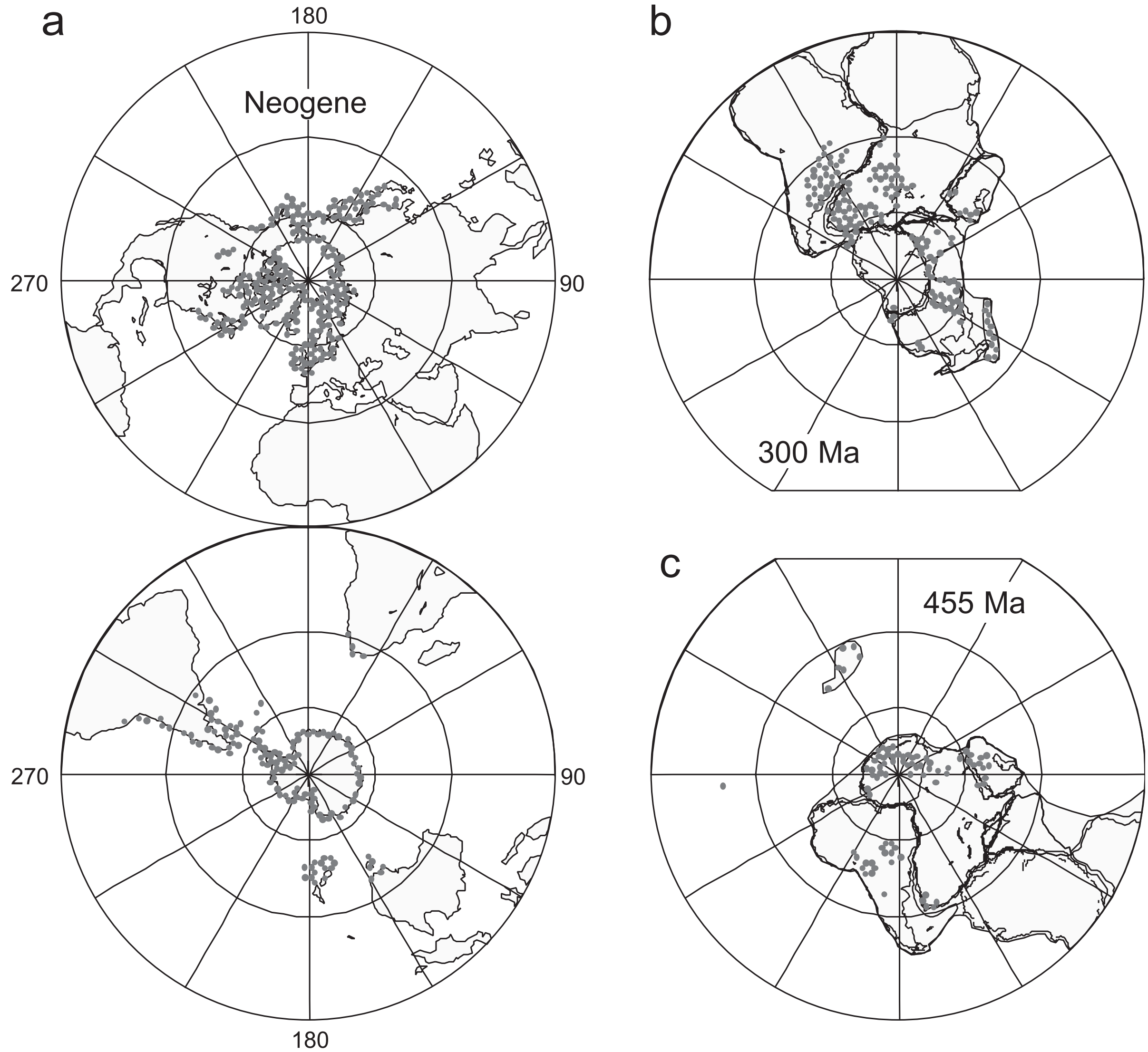

Fig. 6. Polar projections of icosahedrally binned glacial occurrences from (a) Neogene, (b) the Carboniferous–Permian boundary, and (c) latest Ordovician time. Panel (a) includes projections of both North and South poles, and includes only glaciated continental shelves for the sake of comparison with the geological record. Panels (b) and (c) show only the Southern hemisphere, where the great majority of Paleozoic glaciogenic deposits were laid down.

craton in northwest Australia (D23). First known from only a few localities in the southern part of the outcrop belt (reviewed by Krapez, 1996), the strata recording glaciogenic influence are now represented by more convincing deposits in the northern part (Martin, 1999). Although Martin (ibid.) interprets all the deposits to be coeval, cutting across previously established lithostratigraphic boundaries, there remains the possibility of more than one glacial level with sporadic preservation. In either case, the diamictite-bearing Boolgeeda Formation rests conformably upon the 2.45-Ga Woongarra Rhyolite (Barley et al., 1997). That unit bears an eruptive paleolatitude of about 05° from the equator, demonstrated as primary through a positive conglomerate test in the overlying diamictite (Evans, 2002, in preparation).

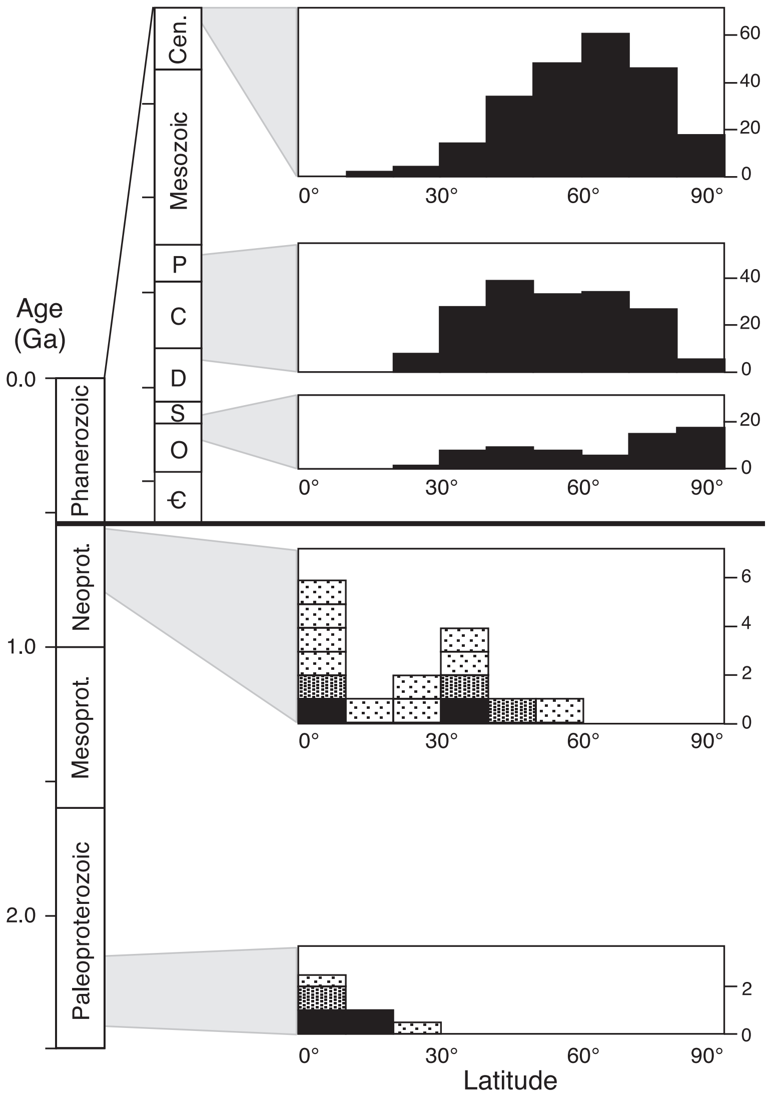

Fig. 7. Depositional paleolatitudes of Earth's glaciations through time, displayed at 10° latitude intervals for comparison across the Proterozoic–Phanerozoic boundary. The Neoproterozoic histogram is from Evans (2000); in both Proterozoic panels, darker shading indicates greater reliability in the paleomagnetic constraint and glaciogenicity of the deposit.

Paleolatitudes for other Paleoproterozoic glaciogenic rocks are less certain. The Sariolian diamictites in Fennoscandia (E35), widely considered to be glacially derived (Marmo and Ojakangas, 1984; Ojakangas, 1988; Strand and Laajoki, 1993), may have been deposited at 27° or 07° latitude, depending on the paleomagnetic component (D vs. D′) chosen to represent intrusion of the immediately underlying 2.45-Ga magmatic episode (Mertanen et al., 1999). Younger, still Paleoproterozoic, Fennoscandian diamictites (E33–34) are less convincingly glaciogenic, less well dated, and less constrained paleomagnetically. The same could be said for the Udokan diamictites near Lake Baikal (C39). Williams and Schmidt (1996) ruled out any glacial contribution to the tilloids at the base of the Vindhyan Supergroup (C36–37), now known to be older than 1630 Ma (Ray et al., 2002; Rasmussen et al., 2002) rather than Mesoproterozoic as earlier thought.

Evidence for Archean glaciation is rare. The best-documented example is in the Pongola–Witwatersrand basin of the Kaapvaal craton in southern Africa (A32), where diamictites of the Odwaleni Formation (Delfkom Formation of Beukes and Cairncross, 1991) are found both in South Africa (von Brunn and Gold, 1993; Young et al., 1998) and Swaziland (Gutzmer et al., 1999) in the upper part of the stratified succession, the Mozaan Group. Age is constrained by a high-precision U–Pb zircon determination on underlying Nsuze volcanism (2985 ± 1 Ma; Hegner et al., 1994) and less precise Sm–Nd mineral isochron (2871 ± 30 Ma; Hegner et al., 1984) and Rb–Sr whole rock ages (2875 ± 40 Ma; Layer et al., 1988) from the intruding Usushwana Complex. Paleomagnetic directions from the diamictite at Klipwal, South Africa (N. Nhleko, N.J. Beukes, J.L. Kirschvink and D.A.D. Evans, unpublished and in preparation), are similar to those from the Usushwana Complex (Layer et al., 1988), suggesting the possibilities of either a primary magnetization and little APW motion between Mozaan and Usushwana time, or a regional overprint during intrusion of the latter complex.

The paleolatitudinal and even depositional constraints on other alleged Archean glaciogenic deposits are still less clear. Tilloids in the Karelian craton have been considered of tectonic or volcanogenic rather than glacial origin (E36–37). Diamictites in the contact-metamorphosed footwall of the Stillwater Complex, Montana (F34), may be glaciogenic but have very poor geochronological and no direct paleomagnetic constraints. Recently, Modie (2002) has proposed a glaciogenic origin for diamictites and lonestone-bearing shale units within the Nnywane Formation within the Derdepoort outlier near the Botswana–South Africa border. That succession, mainly volcanic, was dated at 2782 ± 5 Ma with a possible depositional/eruptive paleolatitude of 65 ± 18° (Wingate, 1998). Nonetheless, a glaciogenic origin for the Nnywane Formation could be questioned on the basis of mere volcanic terrain instability producing the diamictites, and volcaniclastic ejecta generating the shale-enveloped lonestones.

In summary of the early Precambrian glacial record, the Paleoproterozoic ice ages appear to be exclusively low-latitude (Fig. 7). The most reliable paleomagnetic constraints on volcanic units lying conformably adjacent to convincingly glaciogenic deposits are from Western Australia (Woongarra/Boolgeeda, 05° paleolatitude) and South Africa (Ongeluk/Makganyene, 11° paleolatitude). The Gowganda/Lorrain paleolatitude of 03° is considered here of moderate reliability, pending further scrutiny. Lastly, selection of a magnetic remanence component within the Burakovka intrusion will determine whether the associated Sariolian glaciogenic deposits were deposited at 27° or 07°; this uncertainty is depicted on Fig. 7 as a split entry with half heights. The pattern of strictly low-latitude glaciation established for the Neoproterozoic glacial intervals (Evans, 2000) thus appears to hold true for their older counterparts. No reliable paleolatitudinal estimates are yet available for convincingly glaciogenic deposits in the Archean Eon.

4. Discussion

The transition between the Precambrian and Phanerozoic glacial paradigms appears to have been quite abrupt. The youngest near-equatorial, Proterozoic glacial deposits may be found in northern Australia (Grey and Corkeron, 1998) and western North America (Abolins et al., 1999), in close stratigraphic proximity to basal Cambrian sediments. Their ages are likely in the range of 550–580 Ma (ibid.; Evans, 2000). West Africa contains glacially derived rocks that are probably Early Cambrian in age, although correlations across the craton are tenuous (Culver et al., 1988; Bertrand-Sarfati et al., 1995); depending on their precise ages, they were deposited at mid- to high-paleolatitudes (Evans, 2000). As noted above, the diamictite in the Florida Mountains of New Mexico—almost certainly Early Ordovician in age and deposited at low paleolatitudes—presents a conundrum if a glaciogenic origin is confirmed. This problematic unit notwithstanding, the general trend of glacial paleolatitudes would appear to indicate a fundamental global climatic transition, from equatorial-dominated to polar-centered ice ages, within 50–100 My of the Proterozoic–Cambrian "explosion" of metazoan diversity (Grotzinger et al., 1995).

The stark contrast between distributions of Proterozoic and Phanerozoic glaciogenic deposits (Fig. 7) demands an explanation: is the modal shift merely coincidentally coeval with the widespread divergence of multicellular organisms at the Neoproterozoic–Paleozoic transition? At broadly the same interval of time (beginning in the mid-Neoproterozoic) ice ages increased dramatically in abundance, following a three-billion-year sedimentological record practically devoid of glacial influence. Why was most of Precambrian paleoclimate equable but punctuated by ice ages penetrating deep within the tropics? And why has Phanerozoic glaciation been commonplace and of mild severity? The following discussion places these questions in context by reviewing the various geological factors that have been proposed to explain Earth's longterm record of glaciation. The recent synthesis on factors affecting atmospheric CO_2 levels (Boucot and Gray, 2001) contains many more references than can be included here. Proposed factors can be placed into two broad categories: changes in paleoclimatic boundary conditions such as continental positions plus tectonic effects influencing the carbon cycle and hence atmospheric greenhouse gases, and secular changes in climate trends involving longterm geophysical and biogeochemical evolution.

4.1. Boundary-condition effects

Processes that contribute to the presence or absence of glaciation at the 100-My timescale are considered here as due to changing boundary conditions such as paleogeography, regional and global tectonics, supercontinental episodes, and possible extraterrestrial factors. Considering first the paleogeographic factors, the absence of any well documented Gondwanaland glaciation when that continent lay off the South pole (e.g., Early–Middle Cambrian, Late Silurian–Middle Devonian, and post-Late Permian) would appear to support the hypothesis of Caputo and Crowell (1985) that land at the pole itself is necessary for continental ice sheets to grow at high latitudes. The effect is not merely one of preservational bias toward continental rather than oceanic settings; the South pole remained near enough to Gondwanaland in the nonglacial times (Fig. 2) that any ice ages should have been expressed in adjacent alpine regions if they had existed at all. Nonetheless, not all times of polar land were ice ages, so other mechanisms must be at work.

Energy-balance and general-circulation modeling has indicated a strong sensitivity of icecap growth to paleogeographic boundary conditions (e.g., Crowley et al., 1987, 1993). These models provide an attractively simple equivalence between glaciation and immediate proximity of the South pole to Gondwanaland's coastline, as long as a simple APW path similar to that of Morel and Irving (1978; see also Scotese et al., 1999) is accepted. More convoluted APW paths, such as the one accepted here (McElhinny et al., 2003), eliminate this one-to-one correspondence by creating pole-crossing points of the Gondwanaland coastline in Late Silurian and Middle Devonian times, when no glacial deposits are yet reported. There does appear to be a correspondence of glaciation specifically when the South pole lay 1000–1500 km inland of the coastline, a fact apparent to Crowley and Baum (1992; their Figs. 4 and 5) but not explicable by their geographical analysis alone. One prominent exception to the trend is the Middle–Late Cambrian (nonglacial) pole position nearly coincident with its Late Ordovician (glacial) counterpart.

Idiosyncracies of paleogeography can alter ocean circulation patterns and thenceforth climate (e.g., Kennett, 1977), and creation or destruction of such "gateways" has been proposed for initiation of the Paleozoic ice ages (Brenchley et al., 1994; Smith and Read, 2000). In particular, the Carboniferous closure of the Tethyan–Pacific oceanic gateway as evidenced by the Hercynian orogeny in Europe should have profoundly affected global circulation (Wilson et al., 1994; Saltzmann, 2003). Atmospheric circulation can also be fundamentally affected by expansive orogenic plateaus (Ruddiman and Kutzbach, 1989). Such factors are difficult to incorporate into any general model of longterm paleoclimate, especially for Precambrian time, when paleogeographies are known to first-order at best.

Regional tectonic factors can also affect climate. Eyles (1993) summarized the tectonic environments accompanying glaciogenic deposits, be they emplaced in foreland basins or rifts. He showed that many of the ancient glaciogenic formations were deposited in active tectonic settings, and suggested tectonism as a controlling influence on glaciation at least on the regional scale. Powell and Veevers (1987) invoked uplift of mountains along the Panthalassan margin of Gondwanaland as the trigger for the late Paleozoic glaciation, although González-Bonorino and Eyles (1995) demonstrated difficulties with this simple model. It is true that individual deposits appear influenced by local or regional-scale tectonics, but the existence of ice ages in general must have other causes. Otherwise, every foreland basin or rift in Earth history would have been glaciated! Contrary to these ideas, Villas et al. (2003) have recently proposed that passive-margin carbonate sedimentation during Late Ordovician time was the driver of subsequent glaciation, because of carbon sequestration in the sedimentary piles. That model will be challenged to explain the uniqueness of the latest Ordovician ice age among several other early Paleozoic intervals of widespread carbonate deposition. Another problem is that the transgression allowing carbonate burial and alleged cooling would also have substantially decreased tropical albedo, creating an important warming factor that would need to be overcome by any effects of carbon sedimentation.

Orogeny exhumes deep-seated rocks. Commonly, although not necessarily (England and Molnar, 1990), this is associated with uplift of mountains, creation of erosional relief, riverine incision, and voluminous physical weathering of silicate rocks. As the sediments are carried downstream to lower, warmer elevations and then into the oceans, chemical breakdown of feldspars ultimately removes CO_2 from the atmosphere and sequesters it into carbonate rocks (see Raymo and Ruddiman, 1992). Such a process will be enhanced if the mountains are located in tropical latitudes, where warmer, wetter climates accentuate the onset of chemical weathering at greater altitudes. It is notable that the three largest Phanerozoic ice ages all correspond broadly with major episodes of continental or arc-continental collision at moderate to low latitudes: Ordovician–Silurian (Taconic–Iapetan), Carboniferous–Permian (Alleghanian–Hercynian–Uralide), and Neogene (Himalayan–Tibetan).

Considering each of the three Phanerozoic icehouse intervals on an individual basis, Raymo (1991; expanding on the ideas of Chamberlin, 1899) proposed a correlation with silicate weathering due to development of collisional orogenic plateaus such as Tibet. She linked each ice age to a rapid rise in the $^{87}Sr/^{86}Sr$ ratio of seawater as recorded in carbonate rocks, noting that radiogenicity of the oceans is ultimately tied to riverine input of weathered continental material. However, this model encounters three difficulties. First, Himalayan source areas, which dominate continental riverine input into the oceans, contain unusually Sr-rich carbonate rocks (Edmond, 1992), whose Sr isotopic ratios are surprisingly radiogenic (Quade et al., 1997). Thus seawater Sr isotopes cannot be used as a proxy for global silicate weathering and should show no direct correlation with climate (see also recent results from Lear et al., 2003). Second, there are many more peaks in the Phanerozoic seawater Sr-isotope curve (Burke et al., 1982) than recognized ice ages or plateau-building orogenic events. Third, longterm (~ 500 My) trends in this curve show an opposite pattern to that established for the Neogene: despite a correspondence of Carboniferous–Permian ice ages to development of the Ouachita–Alleghanian–Hercynian–Uralide orogenic belt, Sr-isotopic ratios of coeval seawater are near their Phanerozoic nadir.

If the Sr-isotopic composition of the oceans is ignored, however, one may still accept Chamberlin's (1899) hypothesis that ice ages are driven primarily by enhanced silicate weathering during times of orogenic maxima. As stated above, there is indeed a temporal correspondence between the Carboniferous–Permian glaciation and collisional orogenic events marking the final amalgamation of Pangea (Powell and Veevers, 1987). Friedmann (1994) summarized evidence for Permian erosional removal of 7 km across a large region of the Appalachian belt; regardless of whether a Tibet-like plateau existed (see England and Molnar, 1990), this could indicate a substantial amount of silicate weathering and concomitant CO_2 removal from the atmosphere. Kump et al. (1999) have recently revived the orogenic-weathering model to account for Ordovician–Silurian glaciation under an otherwise greenhouse-rich atmosphere. One potential problem with that application is a weak link to broader tectonic trends of the early Paleozoic. For example, the Taconic–Grampian orogeny (460–470 Ma; Friedrich et al., 1999), implicated as a CO_2 sink in the model, is but one of several arc-continent collisions among the Iapetan–Rheic realm during early Paleozoic time (Mac Niocaill et al., 1997; Keppie and Ramos, 1999), so why would not these other collisions induce widespread glaciation, beyond the meager record of regional ice in South America during the Early Silurian (Grahn and Caputo, 1992)? Furthermore, weathering yield from these arc-continent collisions should have been dwarfed by the post-Scandian

(400 Ma; Andersen, 1998) orogenic exhumation following Laurentia–Baltica collision; yet the Late Silurian–Early Devonian interval is notable for its *lack* of any known glacial deposits (Frakes et al., 1992).

On the next larger temporal and spatial scales, several models have purported to show a relationship between supercontinents and glaciation according to regular, periodic cycles (e.g., Fischer, 1984; Worsley et al., 1984; Nance et al., 1988; Veevers, 1990, 1994), but Ordovician–Silurian and Neogene ice ages during the alleged "dispersed-continent" stages are not as easily explained as the Carboniferous–Permian icehouse developed during Pangea's final assembly (Young, 1997). Broad statements relating ice ages to supercontinental rifting (Young, 1991) tend to oversimplify the tectonic record; as an example, the Carboniferous–Permian so-called "Gondwana" rift basins cannot be related to Pangea and its breakup, for Pangea was still assembling as those basins were developing (Harris, 1994).

Many of the Proterozoic glaciogenic deposits, however, were indeed deposited within supercontinental rift basins (Young, 1995, 1997), forming the basal sequences to post-Rodinian (Neoproterozoic) and post-Kenorland (Paleoproterozoic) passive margins. In addition, those two supercontinents are noted by their longevity (about 300 and 400 My, respectively, according to peaks in global isotopic ages and dates from abundant mafic dike/basalt suites) relative to ephemeral Pangea (100 My at most). Evans (1998, 1999, 2003), following earlier concepts of supercontinental effects on the mantle, has speculated that the low-latitude breakup positions of both Rodinia and Kenorland were attained via true polar wander. In both cases low-latitude glaciation directly ensued. A predominance of rifting continents in low latitudes, with elevated topography due to regional mantle upwelling, can enhance global albedo and silicate weathering rates: a taphrogenic equivalent to the processes described by Moore and Worsley (1994). Increased lengths of coastline around rifting continents enhances precipitation (hence silicate weathering and glacial recharge) over a greater proportion of continental surface area (Kirschvink, 1992; Otto-Bliesner, 1995). Rapid sedimentation in rift basins can bury carbon efficiently, lowering atmospheric greenhouse gas concentrations (Hoffman et al., 1998). In the special case of almost no landmasses at high latitudes, the lack of a significant silicate-weathering negative feedback (Walker et al., 1981) will amplify any cooling begun through the other processes (Hoffman and Schrag, 2000, 2002; Schrag et al., 2002).

In the Evans (1998, 2003) true polar wander model for supercontinents, breakup should always occur at low latitude, whereas assembly can occur at any latitude. According to the factors presented above, glaciation should be expected during all breakup intervals of long-lived supercontinents, but only those episodes of supercontinental assembly which entailed continental collisions at (by chance) low latitudes. This model would then predict that assembly of supercontinent Nuna (2.1–1.8 Ga; Hoffman, 1996), which was not accompanied by glaciation, occurred at moderate to high paleolatitudes. Although Nuna's configuration remains highly speculative, truncations of Paleoproterozoic cratons and foldbelts along the northern and western margins of Laurentia suggest a centroid for that supercontinent in the broad vicinity of those truncations. Indeed, the Laurentian APW path for the interval 2.1–1.8 Ga hovers in that region (Buchan et al., 2000). Too little is known about possible Neoarchean glaciations and continental positions to assess a possible relationship between paleoclimate and the assembly of Kenorland. A notable exception to the model of glaciation during supercontinental fragmentation is the Jurassic–Cretaceous interval of Pangean breakup amid an equable global climate. Because Pangea stretched nearly from pole to pole, all of the feedback mechanisms described in the previous paragraph would have been reduced in this instance (Marshall et al., 1988). Nuna may have had a similar paleogeography at breakup, during the ice-free early Mesoproterozoic interval.

Finally, one can turn to possible exogenic forcing of longterm climate. Several workers have noted an apparent 150-My periodicity of cool climates since Neoproterozoic time (Frakes et al., 1992), and speculated upon possible cosmic controls. Early work focused on the periodicity of a slightly eccentric Solar transit around the galactic center (Steiner and Grillmair, 1973), or passage of the Solar System through nodes of galactic flexural response to the Magellanic Clouds (Williams, 1975b). These proposals lose their attractiveness when updated astronomical values (Reid et al., 1999) of solar distance from the galactic center (8.0 ± 0.5 kpc) and tangential velocity of the local standard of rest (LSR; 218 ± 20 km/s) are used to calculate a solar-galactic period of 225 ± 20 My. Neither this value nor the half-period suggested by some models (Williams, 1975b) matches the quasi-periodicity of Earth's ice ages. A more recent analysis (Shaviv, 2002, 2003) incorporates the presently favored density-wave theory for galactic spiral arms, in which each star passes through the spiral arms at a velocity that is small relative to both the star's and the arms' rotation about the galactic center. Shaviv has computed a ca. 150-My periodicity of the Sun's passage through various spiral arms of the Milky Way galaxy, and has postulated that cosmic rays from supernovae within the arms caused greater average cloud cover on Earth, inducing or enhancing broad ice intervals. Such a periodicity, however, is sensitive to the highly uncertain pattern speed of spiral arms; Leitch and Vasisht (1998) used a different estimate to produce a possible correlation of spiral-arm crossings with major biological extinctions, rather than ice ages. Until better estimates for pattern speed are obtained, spiral-arm-crossing correlations can be postulated for any alleged geological periodicity greater than ca. 90 My (see Table 3 of Shaviv, 2003).

In summary, it is difficult to pinpoint a single boundary-condition effect that can explain the long-term appearance and disappearance of ice in the geological record. Continental positions, regional tectonics, oceanic barriers and gateways, and globally averaged sedimentary constituents all probably contributed to climate change (Crowell, 1999; Boucot and Gray, 2001). Scientists working on Phanerozoic ice ages have considered relations with collisional tectonic events and supercontinental assembly, whereas those studying Precambrian glacial deposits have correlated cold climates with supercontinental fragmentation. Location of supercontinents during the various stages of their evolution may be controlled by true polar wander, alternately enhancing or reducing climate feedbacks involving silicate weathering and albedo. Exogenic processes are difficult to relate confidently to glacial episodicity because of large uncertainties in several important cosmic parameters.

4.2. Secular trends

A dramatic shift in glacial modes could have coincided with a similarly dramatic secular change in primary paleoclimatic influences, or an abrupt shift may have occurred if a slow secular change crossed a critical threshold. Possible causes of the fundamental shift shown in Fig. 7 include a "trivial" solution that denies validity of the Precambrian paleomagnetic results, as well as proposed longterm trends in geophysics and global biogeochemistry.

The trivial solution is one whereby the assumptions underlying interpretation of paleomagnetic data from pre-Ordovician rocks are simply discounted. The patterns shown in Fig. 7 are too internally consistent to founder from individual attacks on specific results, which include some of the most reliably determined of any in the paleomagnetic database (reviewed by Evans, 2000). Recent suggestions of enhanced non-dipole geomagnetic field harmonics during ancient times (Kent and Smethurst, 1998; Bloxham, 2000; Torsvik and Van der Voo, 2002) could nonetheless cast doubt on the equivalence of paleomagnetic and rotational paleolatitudes throughout early Earth history. Unfortunately, the proposed influence of non-dipole geomagnetic field components fails in both timing and magnitude to match the observed shift in paleomagnetically determined glacial latitudes. The low-latitude bias in the paleomagnetic database (Kent and Smethurst, 1998) applies to late Paleozoic as well as Precambrian rocks, so any possible geomagnetic transition would have postdated the shift in glacial paleolatitude modes. The magnitude of such proposed non-dipolar components is small, on the order of 10–20% of the dominating dipole (Kent and Smethurst, 1998; Torsvik and Van der Voo, 2002). As Evans (2000) pointed out, the biasing effect of octupolar components to the main field is strongest at mid-latitudes, whereas near-equatorial results would be virtually unchanged by the subtle corrections arising from minor higher orders superimposed upon a dominantly dipolar field. Special geomagnetic gymnastics will be required if nonuniformitarian fields are invoked to explain all the enigmatic features of the Neoproterozoic paleomagnetic database (Evans and Raub, 2003).

Dramatic reduction in planetary obliquity from an initially large value of $>54°$ has been proposed to explain the qualitatively long-recognized shift from Precambrian tropical to Phanerozoic polar ice ages (Williams, 1993, reviewing earlier work). At first impression the quantitative analysis presented here could appear to support that model. One could argue that oceanic and atmospheric circulation patterns would have shifted profoundly during a rapid obliquity transition, thus possibly influencing biological evolution. None of the proposed physical mechanisms for reducing planetary obliquity can achieve the task in as little as 100 My (Williams et al., 1998; Pais et al., 1999), but lack of a known mechanism should not be used to rule out an observationally based hypothesis. Whereas other arguments can be marshaled against the high-obliquity hypothesis (e.g., Néron de Surgy and Laskar, 1997), the distribution of glacial paleolatitudes presently cannot. Under special conditions, exclusively low-latitude Proterozoic glaciogenic deposits (Fig. 7) can be explained by all three models: snowball (Hoffman et al., 1998; Schrag et al., 2002), "slushball" (Hyde et al., 2000), and high-obliquity (Williams, 1993).

Secular trends in the solar output and atmospheric greenhouse effect over the course of Earth history can also produce transitions in glacial modes, which may be abrupt if critical thresholds are crossed. As the Sun has gradually produced greater amounts of helium relative to hydrogen during the course of stellar evolution, its luminosity is expected to have risen as much as 30% since the Archean (Gilliland, 1989). Absence of glaciation through most of Precambrian Earth history is known as the "faint young Sun paradox" (Sagan and Mullen, 1972). The gradual warming trend predicted by this process is opposite to the observed increase in ice age occurrences through the geological record (Fig. 7), and the paradox is usually countered by a presumed thick CO_2 greenhouse atmosphere (Owen et al., 1979). Methane may also have played a major role in greenhouse warming during early Earth history (Rye et al., 1995), and reduction in CH_4 levels during initial stages of atmospheric oxidation could have induced the Huronian ice ages (Pavlov et al., 2000). By late Neoproterozoic time, oxygen levels should have risen sufficiently to preclude methane as a persistently significant greenhouse gas; however, Schrag et al. (2002) invoke brief returns to methane-dominated greenhouse conditions immediately prior to proposed snowball Earth events. Another postulate for overcoming the faint young Sun paradox is a larger Archean hydrosphere relative to flooded cratons (Henderson-Sellers and Henderson-Sellers, 1989); gradual, globally averaged continental emergence would have increased albedo and silicate weathering rates, causing a longterm cooling trend.

Several biogeochemical processes involving metazoans are capable of altering global climate via the carbon cycle. Hoffman et al. (1998) suggested that the onset of deep bioturbation in earliest Cambrian time (Bottjer et al., 2000) could have decreased the efficiency of carbon burial, reducing silicate weathering rates and thus precluding Snowball Earth events from the last 500 My. Another potential feedback related to the Cambrian "explosion" of skeletal animals, is the enlargement of a biospheric reservoir of phosphorus, which limits primary productivity and consequently retains elevated levels of greenhouse gases in the atmosphere (Worsley and Nance, 1989). However, these processes are expected to cause longterm warming, opposing the general trend of cooling that is indicated by increased temporal proportion of ice ages in Phanerozoic relative to Precambrian time.

Evolution of land plants probably had a large effect on global biochemical cycles. In addition to the terrestrial invasion of Devonian forests storing a minor increase of carbon in the biosphere, development of less-reactive lignin from late Paleozoic trees allowed for more efficient burial of organic carbon and led to atmospheric CO_2 drawdown that accompanied the late Paleozoic glaciations (Berner, 1998). In addition, deeply penetrating root systems into soil increased the aggregate reactive surface area for chemical weathering of primary silicates, further enhancing atmospheric CO_2 removal (Algeo et al., 1995; Retallack, 1997; Algeo and Scheckler, 1998). Such processes carry the correct sign of observed temperature change, effectuating longterm cooling, but mid-Paleozoic expansion of terrestrial forests occurred hundreds of millions of years too late to explain the late Neoproterozoic increase in ice-age abundance.

A better temporal correlation may exist with the evolution of fungi and lichens. Members of both groups measurably enhance silicate weathering rates, by factors as great as 10 times or larger, thus reducing atmospheric CO_2 levels and global mean temperature (Schwartzman and Volk, 1989; Boucot and Gray, 2001). Divergence times of major fungal clades correspond approximately to those of animals (Heckman et al., 1999), and the paleontologically best-calibrated molecular-clock estimates of the latter divergences fall within the range of 750–550 Ma (Peterson et al., in press). Therefore, within these broad age limits, the evolution of fungi and lichens could well have contributed significantly to the general cooling of climate since mid-Neoproterozoic time (Fig. 7; Schwartzman and McMenamin, 1993).

Are the broad changes in glacial style through the Neoproterozoic–Cambrian transition merely coincidental, then, with "explosions" of metazoan diversity? Without attempting to identify an exhaustive list of current biological feedbacks on global climate, one may correctly state that they are numerous in both positive and negative signs (e.g., Woodwell and Mackenzie, 1995). Whether the net result of these numerous feedbacks tends to regulate climate (Lenton, 1998), and if so whether such regulation can be ascribed to natural laws (Lovelock and Margulis, 1974; Lenton, 2002) rather than just chance (Watson, 1999; Kirchner, 2002) is actively debated. If Precambrian ice ages with equatorial glacial deposits are interpreted as snowball—or even slushball—Earth events, then amplitudes of climatic oscillations would appear to have diminished through time. Absence of low-latitude glaciation since the Cambrian Period could argue in favor of a net regulatory mechanism on global climate imposed by an increasingly complex biosphere, at least at these long timescales.

5. Concluding remarks

The primary factors influencing Earth's longterm glacial record may thus be summarized as follows. On timescales of 5–50 My, idiosyncracies of paleogeography and ocean/atmospheric circulation may dominate the comings and goings of ice ages. On the timescale of 100–200 My, abundance or absence of continental glaciers is likely due to tectonic or perhaps cosmic controls, although there appears to be no simple periodicity of ice ages, nor a simple relationship with supercontinents. Phanerozoic glacial maxima appear to correlate roughly with continental collisions in low latitudes. Silicate weathering from Precambrian collisions may not have been sufficient to remove the thick Precambrian greenhouse, especially if those collisions occurred at moderate to high paleolatitudes. Contemporaneity of Proterozoic glaciation with low-latitude fragmentation of long-lived supercontinents could result ultimately from true polar wander and the albedo plus weathering effects of continents lying largely or exclusively in the tropics. On a Gy timescale, the fundamental transition (Fig. 7) between low-latitude, possibly globally engulfing, rare ice ages (Precambrian) and polar, restricted, common glacial intervals (Phanerozoic) appears to indicate secular development of biogeochemical fluxes and feedbacks (or feedback enhancements) that respectively cool and regulate the planet's surface temperature. The timing of such a transition during evolution of multicellular life may be no coincidence but rather, indicate causative factors.

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