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STRATIGRAPHIC, GEOCHRONOLOGICAL, AND PALEOMAGNETIC CONSTRAINTS UPON THE NEOPROTEROZOIC CLIMATIC PARADOX

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ABSTRACT. Of the many models that have been proposed to account for the enigmatically widespread and apparently low-latitude Neoproterozoic glaciogenic deposits, three are widely considered: (1) the Phanerozoic archetype of glaciated polar regions and mid-latitudes only, (2) the "Snowball Earth" model with globally synchronous glaciations, and (3) the high-obliquity hypothesis. These models respectively predict purely high-to-moderate paleolatitudes, all paleolatitudes, and preferentially low paleolatitudes of glacial deposits. To distinguish among these alternatives, I present a thorough compilation of the Neoproterozoic glacial deposits and their current age constraints, avoiding intercontinental correlations in almost all cases. In this conservative view, paleomagnetic data are relevant only if directly measured upon the glaciogenic deposits or conformable units, or if the glaciogenic formations are precisely dated enough for application of equally well dated paleomagnetic poles from the same craton.

The primary conclusion to be drawn from this compilation is that very few of the deposits have reliable paleomagnetic constraints. Of that subgroup, however, low latitudes are more common than one would expect if randomly drawn from a uniform distribution on the sphere. Not a single high-paleolatitude (poleward of 60°) deposit has been documented convincingly. Both the "Snowball Earth" hypothesis and the high-obliquity model are permitted by the present paleomagnetic dataset. The Phanerozoic archetype fails to account for robust determinations of near-Equatorial paleolatitude from several Neoproterozoic glaciogenic deposits. If a non-uniformitarian model such as the high-obliquity hypothesis is correct, then its transition to the Phanerozoic archetype must have occurred rapidly, near the beginning of Cambrian time. Alternatively, if the Snowball Earth model is correct for Precambrian time then the lack of tropical glaciations since 550 Ma may be fortuitous or may indicate secular changes in the boundary conditions or processes governing surficial conditions on planet Earth.

INTRODUCTION

The widespread, global distribution of Neoproterozoic glacial deposits was recognized nearly 40 ys ago, and it remains the subject of great debate among students of Precambrian geology. Distributed on all seven of the present continents (fig. 1) and at more than one level in many sedimentary basins, they are a nearly ubiquitous occurrence in Neoproterozoic sedimentary successions, commonly associated with apparently low-latitude lithological indicators such as carbonate rocks. At first glance these features might suggest Neoproterozoic ice ages of greater severity than their Phanerozoic counterparts. If the deposits were synchronous, then the glaciated continents could not all fit inside the polar regions, implying the unsettling presence of tropical ice sheets in perhaps one or more globally engulfing ice ages (Harland, 1964; Kirschvink, 1992; Hoffman, Kaufman, and Halverson, 1998; Hoffman and others, 1998a). If the deposits were diachronous (Crawford and Daily, 1971; Crowell, 1983), then drift of various continents in and out of the polar belts would necessarily be rapid to achieve the abrupt

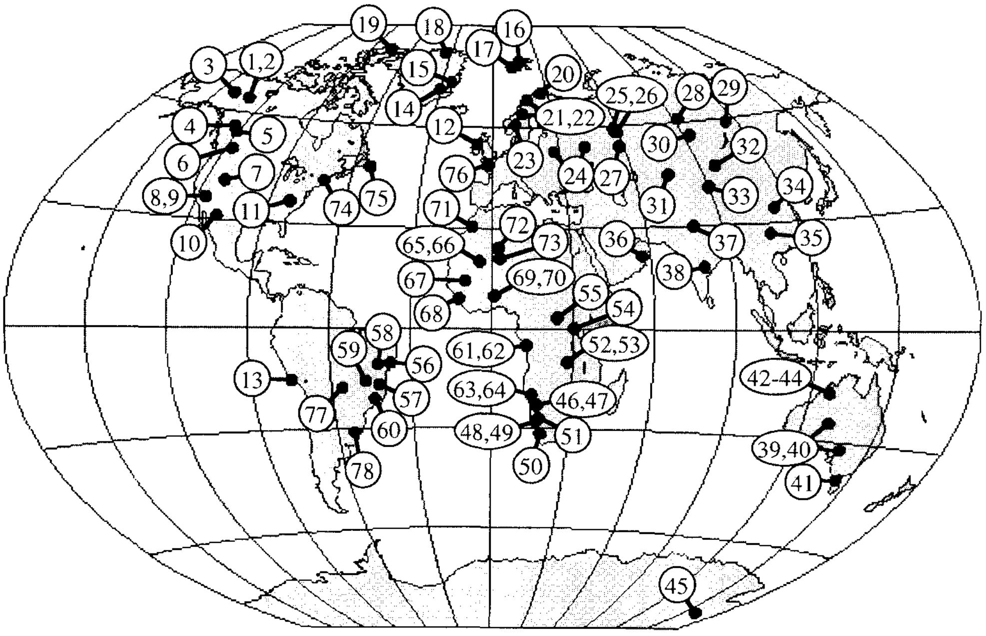

Fig. 1. Present distribution of alleged Neoproterozoic glaciogenic deposits. The numbering scheme is used throughout the text and tables.

climatic changes implied by the warm-cold-warm lithological transitions (Fairchild, 1993).

Paleomagnetism allows us to quantify the depositional paleolatitudes of ancient glacial deposits. Early studies of glacial deposits from southern Norway (Harland and Bidgood, 1959) and East Greenland (Bidgood and Harland, 1961a,b), which seemed to suggest low paleolatitudes of deposition, reinforced the notion of a climatic paradox. Those early studies do not meet present standards of reliability, but they have been superseded by other paleomagnetic results of Neoproterozoic glacial units, many of which were obtained during the last decade and of higher quality. Concurrently, the deposits have become better constrained in age. After a superficial step backward in this regard (late 1980s), when high-precision U-Pb dating techniques demonstrated inconsistencies in Rb-Sr analyses from sedimentary rocks and thereby rendered most if not all those earlier results suspect, many of the relevant successions are now being dated with apparently more reliable chronometers.

The earlier paleomagnetic data, however unreliable when viewed in hindsight, caused sufficient controversy to prompt a host of hypotheses accounting for the low-latitude Neoproterozoic glacial deposits. In the logical order of questioning all the data, the deposits themselves come first. Many of the criteria used to allege a glacial origin of a diamictite (used throughout this paper in the non-genetic sense of an unsorted clastic sedimentary rock with a fine-grained matrix enclosing clasts of a wide range in size) can also be accounted for by a debris-flow origin. Schermerhorn (1974) and Eyles (1993) stated this view, each discussing many of the deposits case-by-case. Advocates for glacial origins of the majority of Neoproterozoic diamictites included Chumakov (1981a, 1985) and Hambrey and Harland (1985). Finally, Rampino (1994) suggested that many of the alleged glacial deposits may have been generated instead by extraterrestrial impacts, which can produce diamictites, striated clasts, and striated pavements. Detailed sedimentological and stratigraphic work, however, should be able to differentiate impact deposits from glaciogenic successions (Reimold, 1998; Williams, 1998); a few examples are given in the detailed descriptions below.

If some deposits are truly glaciogenic and apparently associated with warm lithological climate indicators, then we may question the reliability of those warm indicators. Eyles (1993) suggested that carbonates and glaciogenic deposits together may be explained by continental rifting, with carbonates deposited in restricted basins "starved" of clastic input. Although not widespread, high-latitude carbonate and evaporite rocks coexist with glaciogenic sediments today (Byørlykke, Bue, and Elverhøi, 1978; Walter and Bauld, 1983). A close examination of the pre-, syn-, and post-glacial Neoproterozoic carbonate rocks in the North Atlantic region, however, determined that the immediately pre- and post-glacial carbonates were probably deposited in warm water and thus climatic fluctuations between "balmy" and "icy" conditions occurred quite rapidly (Fairchild, 1993). Of course, the problems still remain, how the suite of glacial deposits was distributed on the planet, and what processes were responsible for the rapid fluctuations in climate.

Another way to question the need for a climatic paradox is by critical evaluation of the low-latitude paleomagnetic data. An early example of this is by Stupavksy, Symons, and Gravenor (1982) who showed that the Port Askaig Tillite was remagnetized during the Caledonian orogeny and that the low paleolatitudes previously determined on that formation represented a middle Paleozoic rather than Neoproterozoic magnetic remanence. Crowell (1983) suggested that this might be a general phenomenon of paleomagnetic studies from Neoproterozoic glacial units. On the other hand, Chumakov and Elston (1989) summarized arguments in favor of tropical latitudes for nearly all the major continental blocks during Neoproterozoic time. As new paleomagnetic results arose during the early part of this decade, however, new apparent polar wander loops emerged that suggested poleward excursions during Neoproterozoic time (reviewed by Torsvik and others, 1996). These data were used by Meert and Van der Voo (1994) to conclude that within uncertainty, all the Neoproterozoic ice sheets were located outside the tropics. Their view was contested (Williams, Schmidt, and Embleton, 1995) based on a rather robust determination of near-equatorial paleolatitude for a widespread glacial deposit in southern Australia (the Elatina Formation; see below). One of the main purposes of this paper is to update and extend these previous critical evaluations of the relevant Neoproterozoic paleomagnetic data.

Assume for the sake of this logical train of thought that at least some truly glaciogenic deposits have reliable enough estimates of low depositional paleolatitude to cause concern. To help explain this phenomenon Schermerhorn (1983) and Eyles (1993) emphasized the continental-rift or unstable-platform tectonic settings in generating high-altitude glaciers in tropical latitudes. This is an important issue, and much of Schermerhorn's (1974) lengthy discussion deals with the problem of local versus widespread distributions of individual glacial deposits. Glaciers occur near the Equator today, although almost entirely above 4800 m elevation (Haeberli and others, 1989) Obviously, small alpine glaciers will have little bearing on global paleoclimatic trends but can leave a substantial record in reworked marine detritus. Schermerhorn (1983) hypothesized that higher atmospheric CO_2 levels (necessary to combat the "faint young Sun paradox"; Sagan and Mullen, 1972) would have effected a greater adiabatic lapse rate in the Precambrian troposphere, enhancing development of abundant moderate-to-high-altitude glaciers at low latitudes. This issue is discussed below in cases where active rifting or orogeny appears to have created topographic highs bearing local rather than regional masses of ice; however, in some instances the Neoproterozoic glaciogenic deposits are spatially and temporally distinct from substantial tectonism. Finally, for interpreting deposits with apparently ice-rafted debris but no evidence for direct or adjacent upslope ice contact, Crowell (1983) noted that modern icebergs can carry glaciogenic debris for over 3000 km (30° latitude) before melting or overturning, and this may lead to low-latitude deposits of merely polar and mid-latitude glaciation. Again, this process can explain the occurrence of low-latitude, dropstone-bearing pelagic sediments but cannot account for low-latitude glaciogenic sediments of a more proximal nature, as are common in the Neoproterozoic record.

If reliable paleomagnetic data indicate low depositional latitudes for one or more demonstrably low-elevation glaciers, then we may consider explaining the data by a non-axial geomagnetic field or a non-dipole field. In these cases, paleomagnetic latitude may not equate with geographic latitude. At least as far back as early Paleozoic time, the geomagnetic and rotational axes appear to coincide; first-order paleomagnetic reconstructions are consistent with lithological climatic indicators like the Gondwanaland glacial deposits (Crowell, 1983; Caputo and Crowell, 1985; Smith, 1997) and carbonates, evaporites, and coals of Laurentia and Laurussia (Witzke, 1990; Van der Voo, 1993, p.20). Also, detailed magnetostratigraphic studies of Neoproterozoic-Cambrian (Kirschvink, 1978a) and Mesoproterozoic (Idnurm, Giddings, and Plumb, 1995) sedimentary successions seem to indicate a self-reversing geomagnetic field with characteristics similar to the present geodynamo. One can postulate such a geodynamo that is nonaxial, but such a hypothesis is rather *ad hoc* (note that the internal fields of Uranus and Neptune, with highly inclined dipolar axes, probably do not serve as good analogues to Proterozoic Earth; see Williams, 1994). Subsidiary non-dipolar components of the geomagnetic field may bias the distribution of observed paleolatitudes from an expected model, and an anomalous abundance of shallow magnetic inclinations (that is, low paleolatitudes) seems to characterize the Paleozoic and Precambrian paleomagnetic database (Kent and Smethurst, 1998). This important observation, recently paired with a plausible geodynamic mechanism (Bloxham, 2000), will be discussed further in my concluding remarks. Lastly, inclination shallowing during compaction of sediments can bias the paleolatitudinal estimate to an apparently lower latitude. This process is not universally observed, however, and can be avoided in paleomagnetic studies that sample across a range of lithologies. Other possible sources of systematic error in paleomagnetism are discussed by Butler (1992).

Assuming that the paleomagnetic data are robust and unbiased and record a Neoproterozoic axial-geocentric magnetic dipole field when averaged over $\sim 10^4$ yr (with perhaps subsidiary higher-order axial components), then we can consider models to explain sealevel glaciers at low-latitudes. Three of the models discussed above imply greatly different predictions of glacial deposits in time and space (fig. 2). First, the Pleistocene-analog model (Meert and Van der Voo, 1994), conforming to the Phanerozoic archetype, allows a slightly more severe climate to generate Neoproterozoic continental ice sheets to 25° latitude, a value generated by a computer circulation model whose parameters included the expected 6 percent less solar luminosity in Neoproterozoic time (Crowley and Baum, 1993). The Pleistocene-analog model predicts that no continental ice sheets should have been located at significantly lower latitudes than ~25°. In addition, many of the deposits should be asynchronous.

Besides the lower solar luminosity, other factors which may have enhanced a Pleistocene-like glaciation during Neoproterozoic times are CO_2 drawdown from chemical weathering of a supercontinent (Young, 1991) or the uplifted shoulders of continental rifts (Eyles and Young, 1994; Young, 1995). Although a general relationship between supercontinents, or breakup stages thereof, and glaciation through Earth history is apparent at the scale of ~100 my, detailed timing of glaciogenic sedimentation with respect to rift-drift transitions in individual basins show great variance (Young, 1997), detracting from the universality of the model. In addition, many of the Neoproterozoic glaciogenic deposits appear unrelated to supercontinental rifting (Powell, 1995). None-

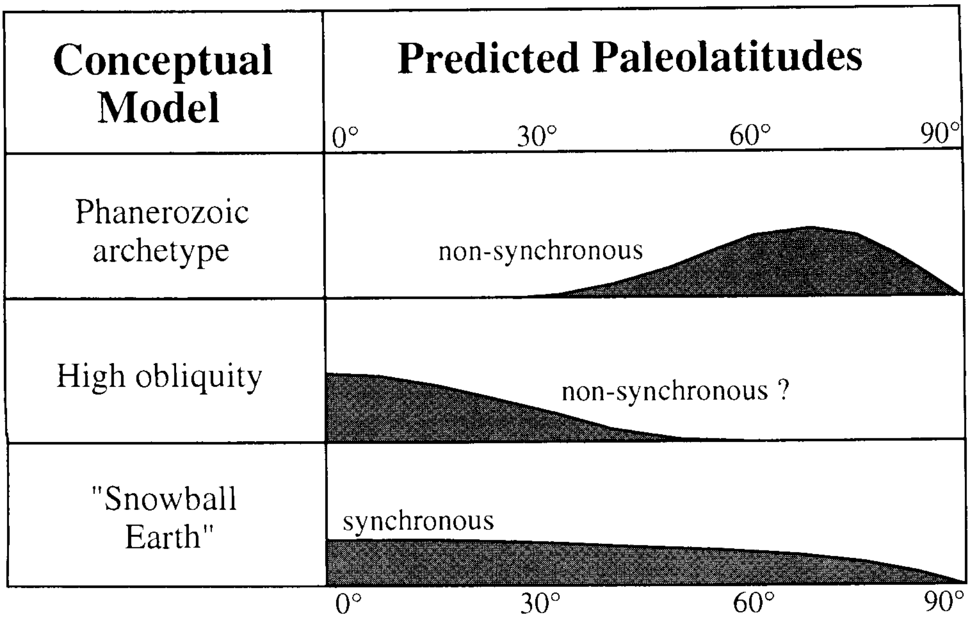

Fig. 2. Approximate paleolatitude distributions of glacial deposits predicted by three conceptual models, assuming a uniform distribution of continents over the globe. The Phanerozoic archetype predicts purely polar glaciation (fewer deposits directly near the poles because of smaller proportional surface area); the Snowball Earth model predicts deposits at all paleolatitudes, with an increased abundance near the Equator because of greater proportional surface area in those regions; and the high-obliquity model predicts the greatest concentration of glaciers at the Equator. Adapted from McWilliams (1977).

theless, long-lived supercontinents may help induce glaciation because they should be driven to the equator via true polar wander; this will enhance chemical weathering as well as increase the planetary albedo (Evans, 1999). Further, supercontinental fragmentation creates passive continental margins that act as efficient repositories for burial of carbon within sedimentary rocks, effecting long-term removal of atmospheric CO_2 (Hoffman and others, 1998a).

The second model is that of global refrigeration (Harland, 1964) recently dubbed the "Snowball Earth" (Kirschvink, 1992; Hoffman, Kaufman, and Halverson, 1998; Hoffman and others, 1998a; Hoffman and Schrag, 2000). Hypothetical variations of this model could range from extremely severe, with the entire oceans frozen over for perhaps ~10 my (Hoffman, Kaufman, and Halverson, 1998; Hoffman and others, 1998a; but see also Jacobsen and Kaufman, 1999), to relatively mild, whereby the latitudinal range of continental ice sheets simply advanced from the polar regions into the equatorial belt, without substantial equatorial sea ice. Obviously, surface temperatures above freezing must have existed locally or temporarily (at the beginning or end of the severe "snowball" model) to allow deposition of the primarily subaqueous Neoproterozoic glaciogenic sediments. The main opposition to the mild variation of the model is that first-order energy-balance models (Budyko, 1969; North, 1975) and computer-generated circulation models (GCMs; see app. A of Wetherald and Manabe, 1975) suggest that if continental ice sheets were to exist on the Equator, then the ice-abedo climatic positive feedback would prevent subsequent recovery from the "ice catastrophe." When the CO_2 cycle was considered, it was shown that a negative feedback between temperature and silicate weathering could apparently stabilize low-latitude continental ice sheets (Marshall, Walker, and Kuhn, 1988). That model was flawed, however, because atmospheric

CO_2 replenishment due to volcanic outgassing is non-instantaneous; a more likely scenario would involve global freezing, concomitant shutdown of the CO_2-weathering feedback, and gradual buildup of greenhouse gases over ~30 my, melting the ice (Caldeira and Kasting, 1992).

In recent opposition to the severe variation of the "snowball" model, Jenkins and Frakes (1998) and Jenkins and Scotese (1998) argued that a more sophisticated GCM, including parameters such as increased Earth rotation, lower solar insolation, and an equatorial continent, failed to generate tropical ice sheets. Nonetheless, as Hoffman and others (1998b) noted, this result was due to a prescribed reduction of equatorial sea-surface temperatures by only 2 °C. Refinement of these models led Jenkins and Smith (1999) to accept the possibility of a Snowball Earth. Given the uncertainties inherent in these relatively simple climate models, it is perhaps better to take the paleolatitudinal distribution of Neoproterozoic glacial deposits at face value and develop conceptual models accordingly.

Another common objection to the Snowball Earth hypothesis is the widespread misconception that a long-lived global glaciation would necessarily leave a stratigraphic record of drastic reduction in eustatic sealevel (Williams, 1999). If continents were situated in low latitudes, then polar to moderate-latitude sea ice would be floating, and in the severe variation of the model, a complete cessation of snowfall would effect rapid ablation of continental ice sheets at all latitudes; hence, any eustatic drawdown leading into the "snowball" state would be short-lived (Walsh and Sellers, 1993; Hoffman, Kaufman, and Halverson, 1998; Hoffman and others, 1998a). Hectometer-scale, glacially related valley incision observed in many Neoproterozoic successions, summarized by Christie-Blick, Sohl, and Kennedy (1999), could appear to contradict the Snowball Earth hypothesis. Nonetheless, Hoffman and Schrag (1999) noted that the continental ice ablation could be balanced in part by sublimation of tropical sea ice and its accretion in mountainous regions. Thus, glacial valley incision could still occur during the "snowball" stage, where ice accretion in these alpine regions proceeded beyond gravitational stability and began to flow; yet such incision would be related only secondarily to eustatic drawdown.

The Snowball Earth model permits glaciation at all paleolatitudes and therefore cannot be tested by paleomagnetic means alone. One testable prediction of the global refrigeration model is some degree of synchroneity among deposits, at least for times of alleged low-paleolatitude ice sheets. However, it is important to note the hierarchy of glacial deposits and consider how much "synchroneity" is really necessary (Chumakov, 1981a). Obviously, individual glaciogenic deposits even in different areas of the same basin are unlikely to be precisely synchronous. Chumakov (1981a) suggests that individual diamictite-bearing formations may by coeval within basins and even throughout regions but cautions against intercontinental correlations without independent stratigraphic evidence (see 'Methods', below).

Third, the high-obliquity hypothesis (Williams, 1975) predicts a preponderance of low paleolatitudes for glacial deposits. At any planetary obliquity >54°, the poles will receive more annual sunlight than any point on the equator. This can be visualized most easily for the extreme case of 90° obliquity, where each pole would experience 6 months of continuous sunlight followed by 6 months with none at all–the torrid summer months would almost certainly eliminate the possibility of any high-latitude glaciers. The high-obliquity hypothesis would therefore be most compatible with diachronous glaciation as continents moved through tropical latitudes (Kröner, 1977). It should be noted, however, that high obliquity would create an inherently mild climatic regime, even for the tropics; in the conceptually simple endmember case of 90° obliquity, equatorial regions would receive zenithal mid-day sunlight at the vernal and autumnal equinoxes, gradually changing to low-angle 24-hr sunlight during the solstices. For analogy on the present Earth, imagine travelling completely around an opposite pair of meridians, throughout 1 yr so that each pole is attained at the respective summer solstice. The fairly moderate temperature changes one would experience during such an adventure are approximately equivalent to the expected annual variations for tropical latitudes in the high-obliquity scenario.

Indeed, Hoffman and Maloof (1999) agree that survival of snow through semi-annual intervals of high insolation is a more important factor for building ice sheets than the amount of accumulation of snow during the intervals of low insolation. G.E. Williams (1999) has countered with the argument that such survival of equatorial snowcover during the semi-annual passages through the equinoxes could be assisted by its high reflectivity. Computer-based GCMs of a high-obliquity Earth and 5 percent reduced solar luminosity show extensive equatorial snow and ice cover during the solstice months but neglect to indicate whether this pattern would persist through the high-solar-incidence days around the equinoxes (Oglesby and Ogg, 1998). Whether or not glaciers could grow in the tropical regions of a high-obliquity Earth probably depends greatly on the annual patterns of precipitation, which are likely to be very different from those to which we are accustomed. One thing is certain, equatorial zones of a high-obliquity Earth would experience dramatic seasonal changes, for which there is the sedimentological evidence of possible frost-wedge polygons in the Neoproterozoic record of South Australia, Scotland, and Mauritania (Williams, 1993).

Alternative, more imaginative explanations for the Neoproterozoic ice ages are possible, for example, the proposed equatorial ice ring (Sheldon, 1984). This model and others are described in Eyles (1993, p.68). Arguments against those hypotheses are reviewed by Chumakov and Elston (1989) and Williams (1994). Figure 2 shows only the predictions of the three endmember models; presumably, they should be distinguishable by paleomagnetism of the glacial deposits and coeval or stratigraphically adjacent rocks. Early attempts to determine the spatial distribution of Neoproterozoic glaciations in this manner used greatly different interpretations of slightly different datasets to arrive at opposite conclusions of deposition in strictly low latitudes (Piper, 1973) or strictly high latitudes (McElhinny, Giddings, and Embleton, 1974). Most of the paleomagnetic data cited by those reviews have since become obsolete. After a period of 15 to 20 yrs, new data were obtained, but the controversy remained: Chumakov and Elston (1989) determined a preponderance of low glacial paleolatitudes, whereas Meert and Van der Voo (1994) found only moderate-to-high paleolatitudes.

The aims of this paper are (1) to present a comprehensive, global update of the stratigraphy and geochronology of Neoproterozoic glacial deposits, and (2) to summarize the most reliable paleomagnetic estimates of their depositional paleolatitudes. This is the first such collection of stratigraphic and geochronological data since the Hambrey and Harland (1981) compendium and its companion manuscript (Hambrey and Harland, 1985), and a wealth of information has appeared since. The paleomagnetic data are subjected to a high level of scrutiny, without concern for the low number of studies that pass the strict quality filter outlined below.

METHODS

Among all earlier attempts at a comprehensive review of paleomagnetic constraints on the Neoproterozoic glaciogenic deposits (Piper, 1973; McElhinny, Giddings, and Embleton, 1974; Meert and Van der Voo, 1994), the general approach entailed construction of apparent polar wander (APW) paths for the various cratons, assignment of numerical ages to the glacial deposits when direct constraints were not available, and then estimation of paleolatitudes by interpolation from these APW paths. In many instances, however, the glaciogenic rocks were not and still are not precisely dated, as is discussed case-by-case below. Thus, for example, Meert and Van der Voo (1994) needed to assume general ages of ~720 ("Sturtian") or ~600 Ma ("Marinoan") for many of the deposits despite allowable age ranges of 50 to 100 my. With this technique it is possible that the paleolatitude determined from a given age on a continent's APW path may be totally inapplicable to the actual time of glaciation. Given the uncertainty in APW paths for most of the Neoproterozoic cratons, the dearth of precise numerical ages for many of the glacial deposits, and the fact that many of the paleomagnetic constraints come directly from the glacial units or conformable rocks in the sedimentary successions, I opt for a "stratigraphic" rather than "chronometric" approach in this paper.

First, I review every alleged Neoproterozoic glacial deposit that I could find from the existing literature. Most are described in the tome edited by Hambrey and Harland (1981). A few entries from that volume are omitted because the evidence for true glaciogenesis is unconvincing–these are usually described as "tilloids" occurring sporadically in active tectonic or volcanic settings. Note that space prohibits my justification that the various deposits are indeed glaciogenic; the reader is referred to Hambrey and Harland (1981) or its more concise synopsis (Hambrey and Harland, 1985). Besides, that is not my primary task: as the interpretation of some of these units has vacillated between glacial and nonglacial and could do so again in the future, I want to include as many potential candidates as possible. Where appropriate, I have included more recent stratigraphic summaries and arguments for or against glacial origin of the various deposits. Notably, for many of the deposits within the cratons constituting "western" Gondwanaland, I include page references from the excellent synthesis by Trompette (1994); that book contains many additional references to the stratigraphy and tectonics of each region. Schermerhorn (1974) and Eyles (1993) present global summaries that discount the model of low-elevation, continent-scale ice sheets for most of the deposits.

I have numbered and grouped the alleged glaciogenic units by paleocontinent or craton and discuss cases where the specific cratonic association is yet unclear. I then review the available ages, primarily from isotopic dating but secondarily by other stratigraphic means (see following paragraphs), of all the deposits. To some degree, lithostratigraphic correlations are necessary for limiting the deposits into a reasonable number of entries (for more thorough "splitting" of deposits, see Chumakov, 1981a), but too much exuberance in "lumping" can lead to prejudice favoring the Snowball Earth model of synchroneity. In an effort to be as unbiased as possible, I adopt a very conservative stance at correlation–*only intrabasinal or intracratonic correlations are accepted at face value.* Thus, terms such as "Varangian" or "Sturtian", usually cited as postulated global ice ages (for example, Kennedy and others, 1998) at ~600 and ~750 Ma, respectively, are not used here except in describing the type localities. Eyles (1993) and Young (1995) emphasized that many of the glaciogenic deposits occur in the basal parts of continental rift successions; given the discontinuous basin geometry expected in that environment, correlations can be difficult even among strata deposited along the same cratonic margin.

While recognizing the potential for carbon- and strontium-isotopic correlation among separate sedimentary basins (Magaritz, Holser, and Kirschvink, 1986; Knoll and others, 1986; Ripperdan, 1994; Kaufman and Knoll, 1995; Kaufman, Knoll, and Narbonne, 1997; Pelechaty, 1998; Saylor and others, 1998; Jacobsen and Kaufman, 1999; Walter and others, 2000), I acknowledge that the database is continuously growing and encourage readers to draw their own correlations of isotopic excursions. Kaufman, Knoll, and Narbonne (1997) present stable-isotopic evidence for five temporally distinct glacial episodes within the Neoproterozoic record of northwest Canada, Svalbard, and Namibia; if their correlations of Namibian strata are correct and if the isotopic trends truly indicate global, secular changes in seawater geochemistry, then that estimated number of ice ages should be considered a minimum due to the fragmentary nature of the stratigraphic record. Emerging Neoproterozoic correlation tools such as sequence stratigraphy (for example, Christie-Blick, Dyson, and von der Borch, 1995) and sulfur-isotope trends within sedimentary sulfides (Gorjan, Veevers, and Walter, 2000) are likewise omitted here.

These omissions are justified for two reasons. First, because many of the paleomagnetic constraints come from sedimentary rocks within Neoproterozoic basins, there is little need to apply extrabasinal ages to glacial deposits via correlation. For example, even if it could be well demonstrated by carbon-isotope or sequence stratigraphy that the type Marinoan (southern Australia) and Varanger (northern Norway) glacial deposits were precisely synchronous, the best paleomagnetic results are directly from those units themselves, obviating any need for such precise intercontinental correlation. Exceptions to this general rule, for example the undated and paleomagnetically unconstrained upper Tindir Group in Alaska and Yukon Territory which could be assigned a paleolatitude via correlation with the better constrained Rapitan Group, are described case-by-case below.

Second, postulated Neoproterozoic supercontinental reconstructions (for example, Hoffman, 1991; Dalziel, 1991; Powell and others, 1993) are still being tested by geological and paleomagnetic means; thus paleomagnetic data should not be extrapolated from one craton to another. For example, a reliable paleomagnetic pole from Baltica should not be used to constrain Laurentia's paleogeography, even though the two cratons are commonly juxtaposed in hypothesized Neoproterozoic supercontinental configurations (for example, Gower and Owen, 1984; Torsvik and others, 1996; Dalziel, 1997).

From many examples, the Rb-Sr method has been proven to give inaccurate ages of Precambrian and Early Paleozoic rocks. Rb-Sr results from igneous rocks are preferable to those from shales, and a well defined Rb-Sr isochron of mineral separates is preferred over a suite of whole-rock determinations. The most reliable and consistent Eocambrian geochronometers, however, appear at present to be U-Pb ages from zircon, baddeleyite, or other U-bearing accessory minerals in igneous rocks (for example, Compston and others, 1992; Heaman, Le Cheminant, and Rainbird, 1992; Bowring and others, 1993; Grotzinger and others, 1995). I take license, therefore, in discarding Rb-Sr ages that I feel are unreliable; the reader can assess my judgment in each case. Recent U-Pb-zircon data from Eocambrian sedimentary successions in Siberia and Namibia have pinpointed a numerical age of 543 ±1 Ma for the Precambrian-Cambrian boundary (Bowring and others, 1993; Grotzinger and others, 1995). With a few exceptions of very simple discoid forms, the Ediacaran fauna seem to occur during a brief interval immediately prior to the Cambrian, 30 my at the maximum (Grotzinger and others, 1995); therefore, Ediacaran fauna can provide an indirect numerical age estimate even in undated sedimentary successions. I have not relied on other Neoproterozoic biostratigraphic schemes, such as those based on acritarchs or stromatolites, because these remain to be tested by independent means such as numerical dating. Nevertheless I have indicated notable biostratigraphic age estimates for some of the deposits.

Regarding paleomagnetic reliability, I include for discussion all results from units that lie within unconformity-bounded stratigraphic packages containing the glacial deposits. If the chronometric ages of the glacial deposits are tightly constrained, then I also include extrabasinal paleomagnetic results from equally tightly constrained igneous units within the same craton or block. Many of the paleomagnetic studies have not demonstrated a primary remanence, and I provide a tentative interpretation of each magnetization age. Table 1 presents a compilation of Neoproterozoic diamictites and alleged tillites, their conservative age constraints, and applicable paleomagnetic results with interpreted magnetization ages. Uncertainties in paleolatitude are calculated (A) from the limits on magnetic inclination determined by the α_{95} error field of the dataset, (B) by the A_{95} of a mean of virtual geomagnetic poles, or (C) by use of either *dp* or *dm*

TABLE 1

Paleomagnetically determined paleolatitudes of alleged Neoproterozoic glaciogenic deposits

Paleocontinent	Related paleomagnetic study							
Glaciogenic deposit	Age*	Formation; age**	Code	1234567	Q	λ' (°)†	Magn. age‡	Source
Laurentia								
1. Rapitan Gp.	E / 755±18	d	Y	0110010	3	74 +12/-11[a]	Cretaceous	Morris, 1977
(Fe)		d	C2	0111010	4	76 +12/-11[a]	Cretaceous	Park, 1997
		d	X	0110001	3	14 ± 3[a]	diagenetic	Morris, 1977
		d	R3	0011011	4	04 ± 6[a]	diagenetic	Park, 1997
		d	Z	0110000	2	08 ± 2[a]	primary?	Morris, 1977
		d	R2	0011010	3	**06 +8/-7**[a]	primary?	Park, 1997
2. Ice Brook Fm.	E / 755±18	Risky, E	RI_A	1010011	4	46 +16/-12[a]	primary?	Park, 1995
6. Toby Fm.	720-740	Franklin; 723±3	comb.	1111110	6	**08 ± 4**[b]	primary	Christie and Fahrig, 1983; Palmer and others, 1983
9. Johnnie Fm.	C / ~1100	d		0100011	3	**01 ± 4**[a]	primary?	Gillett and Van Alstine, 1982
10. Florida Mtns.	O / 503±6	basement; 504±10		1111011	6	**03 ± 8**[b]	primary	Geissman and others, 1991
11. Appalachians	~700 / 758±12	Franklin; 723±3	comb.	1111110	6	**20-21 (±4)**[b]	primary	Christie and Fahrig, 1983; Palmer and others, 1983
12. Port Askaig Fm.	595±4 / 806±3	d	comb.	0100010	2	03 +2/-1[a]	Ordovician	Urrutia-Fucugauchi and Tarling, 1983
15. Tillite Gp.	C / ~1000	d		0001010	2	08 +11/-9[a]	Camb-Ord	Bidgood and Harland, 1961a,b
Baltica								
20. Vestertana Gp.	E / Pt3	d		0011100	3	**33 +12/-14**[a]	primary?	Torsvik, Lohmann, and Sturt, 1995
23. Moelv Fm.	C / Pt3	d		0000110	2	11 +16/-12[a]	Silurian	Harland and Bidgood, 1959
		Nyborg; V		0011100	3	27-31 (±17)[c]	primary?	Torsvik, Lohmann, and Sturt, 1995

24. East European	551±4 / Pt3	Nyborg; V		0011100	3	34-46 (±17)[c]	primary?	Torsvik, Lohmann, and Sturt, 1995
27. South Urals (Fe)	E / Pt3	d	comb	0001000	1	26 ± 7	pre-Perm	Danukalov, Commissarova, and Mikhailov, 1982
Eastern Asia								
28. Chingasan (Fe)	E / Pt3	d	table	0000010	1	11 +15/-12[a]	???	Khramov, 1984
32. Tsagaan Oloom	C / 732-777	Bayan Gol; C	2-pol	1011010	4	44 ± 5[a]	Paleozoic	Evans and others, 1996
33. Tarim (3 levels)	C / Pt3	d		0111111	6	**08 ± 8**[b]	primary?	Li and others, 1991
34. Luoquan	C / Pt3	d	undoc	0???1?0	?	50-65	???	Mu, 1981
		d		0110110	4	32 ± 3[a]	Mesozoic?	Piper and Zhang, 1997
35. Chang'an /	C / 820	lower Sinian; 748	Z1	1111111	7	30-40 (±12)[b]	primary	Evans and others, 2000
Nantuo (Fe)	C / 748±12	d		0000100	1	12 +12/-10[a]	Mesozoic?	Zhang and Zhang, 1985
		d	A1	0110111	5	37 +6/-5[a]	primary	Zhang and Piper, 1997
		lower Sinian; 748	Z1	1111111	7	**30-40 (±12)**[b]	primary	Evans and others, 2000
East Gondwanaland								
36. Mirbat	E / 723 +16/-10	d		0010100	2	09 ± 4[a]	primary?	Kempf and others, 2000
37. Blaini	E / Pt3	d		0000010	1	57 +10/-9[a]	Carb-Perm	Jain, Klootwijk, and Goswami, 1981
39. Sturtian (Fe)	E / 777±7	d	MT1	0100110	3	06-21 (±6)[c]	primary?	McWilliams and McElhinny, 1980
		d	MT2	0100110	3	53-76 (±17)[c]	Mesozoic	McWilliams and McElhinny, 1980
		d	MT3	0000110	2	13-33 (±8)[c]	Camb-Ord	McWilliams and McElhinny, 1980
40. Lower	E / 650	d	Y1	0000100	1	47-58 (±26)[c]	Cenozoic?	McWilliams and McElhinny, 1980
Marinoan		d	Y2	0000100	1	63-76 (±14)[c]	Mesozoic?	McWilliams and McElhinny, 1980
		d	VGP	0110100	3	05 ± 2[b]	primary	Embleton and Williams, 1986
		d	VGP	0111100	4	03 ± 1[b]	primary	Schmidt, Williams, and Embleton, 1991
		d		0011110	4	**03 +3/-4**[a]	primary	Schmidt and Williams, 1995
		d		0111110	5	**09 +3/-4**[a]	primary	Sohl, Christie-Blick, and Kent, 1999

TABLE 1 *(continued)*

Paleocontinent		Related paleomagnetic study						
Glaciogenic deposit	Age*	Formation; age**	Code	1234567	Q	λ' (°)†	Magn. age‡	Source
42. Walsh	Pt3 (Sturtian?)	d		0011111	5	**45 +14/-11**^{a}	primary?	Li, 2000
43. Egan	Pt3 (E?)	E: Pertatataka		1011111	6	**21 ± 8**^{c}	primary	Kirschvink, 1978b
		Lower Marinoan		0111110	5	**13 ± 3**^{c}	primary	Sohl, Christie-Blick, and Kent, 1999
Kalahari								
46. Blaubeker/Court	Pt3 (Kaigas?)	d	NBX	0000100	1	13 +11/-10^{a}	Camb-Ord	Kröner and others, 1980
51. Schwarzrand	539±1 / 543±1	d	many	N.A.		N.A.	N.A.	Kröner and others, 1980; Meert, Eide, and Torsvik, 1997
		Sinyai; 547±4		1110111	6	**38 ± 3**^{c}	cooling	Meert and Van der Voo, 1996
Congo-São Francisco								
63. Chuos/Varianto (Fe)	550 / 758±4	d	NQ1	1100100	3	17 +15/-11^{a}	Cambrian?	McWilliams and Kröner, 1981
		d	NQ2	1100110	4	21 +14/-10^{a}	Camb-Ord?	McWilliams and Kröner, 1981
		Mbozi; 755		1110101	5	**10 ± 5**^{c}	primary	Meert, Van der Voo, and Ayub, 1995
64. Ghaub Fm. (±Fe)	550 / 746±2	d	DC1	0000110	2	19 +38/-21^{a}	Camb-Ord?	McWilliams and Kröner, 1981
		d	DC2+3	0100110	3	48 +24/-16^{a}	Recent?	McWilliams and Kröner, 1981
West Africa and environs								
57. Jequitaí / Macaubas (±Fe)	~600 / 906 ±2	overlying strata	BGC	0110100	3	39 ± 4^{b}	Cambrian	D'Agrello-Filho and others, 2000
65. Jbeliat Gp. / "triad"	Ord. / 680 (E?)	d	LPO	0110100	3	06 ± 2^{a}	Carb-Perm	Perrin, Elston, and Moussine-Pouchkine, 1988
		overlying strata	CO7-8	0010111	4	07 +6/-5^{a}	???	Perrin, Elston, and Moussine-Pouchkine, 1988
		C: Ntonya (522±13)		1100101	4	**65-81 (±2)**^{b}	primary	Briden, McClelland, and Rex, 1993
		E: Sinyai (547±4)		1110111	6	**23-38 (±5)**^{c}	cooling	Meert and Van der Voo, 1996

66. Basal Atar Gp.	650 / Pt1	overlying strata	19	0010110	3	03 +5/-4^{a}	Carb-Perm	Perrin, Elston, and Moussine-Pouchkine, 1988; Perrin and Prevot, 1988
		underlying strata	12	0110100	3	12 +2/-1^{a}	Carb-Perm?	Perrin, Elston, and Moussine-Pouchkine, 1988
67. Mali Gp/Bakoye (Fe)	C / 661±6	(see item #65)						
71. Série Verte	620 / 696	Adma diorite; 620		1110000	3	70 ± 16^{b}	primary?	Morel, 1981
72. Série Pourprée	519? / 556?	(see item #65)						
Avalonia - Cadomia								
74. Squantum Mbr.	C / 596±2	d		0111010	4	**55 +8/-7**^{a}	primary?	Wu and others, 1986
75. Gaskiers Fm.	565 / 606	Marystown Gp; 608		1111001	5	34 ± 6^{a}	pre-Carb	Irving and Strong, 1985
		Marystown Gp; 608	abstr	1111011	6	**31 +10/-8**^{a}	pre-Carb	McNamara and others, 1997
Amazonia - Rio Plata								
77. Puga et al. (Fe)	E? / 623±15?	d		0000000	0	24 ± 7^{a}	Cambrian?	Creer, 1965

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* Min/max; numerical ages in Ma; sources and uncertainties cited in text; underlined values indicate the interpreted closer constraint.

** See text for complete description.

† Superscript codes denote method of calculation: (^{a}) using $\pm\Delta I$ converted to pole space; (^{b}) using A95 of a mean of virtual geomagnetic poles (VGPs); (^{c}) using dp or dm, whichever more appropriate, of a pole derived from afar. All values rounded to the nearest whole number.

‡ Interpreted magnetization age based on field tests (see text).

Abbreviations: Fm. = Formation; Gp. = Group; (Fe) = associated with iron- or manganese-formation; Pt1 = Paleoproterozoic; Pt2 = Mesoproterozoic; Pt3 = Neoproterozoic; V = Vendian (*sensu lato*); E = diverse Ediacaran fossils; C = Cambrian fossils; d = direct study of glacial deposit; Q and 1-7 = reliability scale of Van der Voo (1990), with 1 = rock age well constrained, 2 = sufficient number of samples and good grouping, 3 = adequate demagnetization procedures, 4 = field stability tests, 5 = structural continuity with a craton, 6 = dual polarity, and 7 = no similarity with younger paleopoles; λ' = paleolatitude; abstr = published in abstract only, comb. = combined results; table = available in tabulated form only.

(whichever is more appropriate) on poles derived from extrabasinal regions. Paleolatitude confidence limits derived from the first method could be approximated more easily by the paleopole parameter *dp*, but the approximation falters when the uncertainties are large, which is the case for several of the studies discussed herein. Note that these methods all overestimate the true paleolatitude uncertainties, which require cumbersome numerical methods of computation (Demarest, 1983).

STRATIGRAPHIC, GEOCHRONOLOGICAL, AND PALEOMAGNETIC CONSTRAINTS

Laurentia and Environs

Hypothesized as the center of the Neoproterozoic supercontinent Rodinia (McMenamin and McMenamin, 1990; Hoffman, 1991; Dalziel, 1991), Laurentia includes the autochthonous Precambrian elements of North America plus Greenland, the northernmost British Isles, eastern Svalbard, and probably suspect basement inliers within the Andean orogen (Hoffman, 1988; Dalla Salda, Bossi, and Cingolani, 1992; Dalziel, 1997). As such, it contains a widespread record of Neoproterozoic glaciation recorded as deposits within the developing proto-Cordilleran, proto-Appalachian, and proto-Caledonian (East Greenland) rifted margins. The following discussion begins with the Mackenzie Mountains and then proceeds anticlockwise around Laurentia.

Mackenzie Mountains (1, 2).–Two glacial episodes have been identified within the Windermere Supergroup of the Mackenzie Mountains of northwest Canada. Within the Rapitan Group, three glaciogenic units are recognized: diamictites of the Mt. Berg and Shezal Formations enclose a unit of rhythmic shales of the hematitic, dropstone-bearing Sayunei Formation (Eisbacher, 1981a; Yeo, 1981). Age constraints on the Rapitan Group are problematic, due to the lack of direct cross-cutting relations with dated igneous units. For example, undated basalts among the unconformably underlying Little Dal Group of the Mackenzie Mountain Supergroup, which could provide a maximum age for the Rapitan Group, may be coeval with intrusive bodies (Morris and Park, 1981; Narbonne and Aitken, 1995) dated at 779 Ma (unpublished data cited in Heaman, Le Cheminant, and Rainbird, 1992) and 778 ±2 Ma (Jefferson and Parrish, 1989); alternatively, they may be consanguineous with the Franklin-Natkusiak episode (Ross, Bloch, and Krouse, 1995) dated at 723 +4/-2 Ma (Heaman, Le Cheminant, and Rainbird, 1992). A more direct upper age constraint is provided by 755 ±18 Ma on a granitic cobble within the Sayunei Formation (Ross and Villeneuve, 1997). Because of differing views regarding stratigraphic correlations between the Mackenzie Mountains and regions containing Franklin-Natkusiak outcrops, the 723-Ma igneous activity has been cited alternatively as a maximum (Young, 1992a; Ross, Bloch, and Krouse, 1995), a minimum (Link and others, 1993), or approximately coeval (Narbonne and Aitken, 1995) age constraint on the lower Rapitan Group sediments. In this paper I recognize the 755-Ma maximum age but do not use the 723-Ma age as a direct constraint on the Rapitan glacial deposits. This leaves a minimum age largely unconstrained except for the diverse Ediacaran assemblages occurring ~2 km higher in the section (Narbonne and Aitken, 1995).

A recently discovered glaciogenic horizon has been named the Ice Brook Formation (Aitken, 1991a,b), occurring above simple discoid megafossils but below the more diverse Ediacaran fauna (Hofmann, Narbonne, and Aitken, 1990; Narbonne and Aitken, 1995). Like the Rapitan Group glacial deposits, the real temporal constraints are between 755 ±18 Ma and Ediacaran (~550-543 Ma; Grotzinger and others, 1995). As listed in table 1, however, I prefer an age somewhat closer to Ediacaran for the Ice Brook Formation and one closer to ~750 Ma for the Rapitan Group. These estimates are broadly compatible with a recent interpretation of carbon-isotopic results from the Mackenzie Mountains as summarized by Kaufman, Knoll, and Narbonne (1997).

Paleomagnetic study of these rocks has a long history. Morris (1977) identified three magnetic components in the Rapitan Group, two of low paleolatitude, one of high. No field tests were performed for constraining the ages of magnetization. Based on order of removal during sequential demagnetization experiments, he concluded that the low-paleolatitude 'X' direction gave a primary remanence held by coarse-grained detrital hematite, and that the high-paleolatitude 'Y' direction represented a diagenetic remanence carried by fine-grained hematitic pigment. The 'Y' pole is similar to the Cretaceous segment of North America's APW path (Van der Voo, 1993), and significant tectonic deformation of that age in the Mackenzie Mountains (Eisbacher, 1981c) suggests the possibility that this component was acquired during deformation (Park, 1997; see below). The 'Y' component was found to fail a fold test on mudstones downwarped beneath a dropstone (Sumner, Kirschvink, and Runnegar, 1987; abstract only) in the Sayunei Formation, consistent with either a diagenetic or Cretaceous remanence. If Morris' (1977) model were correct, then Laurentia would have drifted very rapidly during Rapitan deposition. Finally, the 'Z' component was hypothesized by Morris to be a thermal overprint; similarity of its pole with the Cambrian-Ordovician APW path for North America (Van der Voo, 1993) would imply an early Paleozoic age for that remanence, although no tectonic events of that age are evident in the Mackenzie Mountains.

More recently, Park (1997) conducted an extensive paleomagnetic study of the Mt. Berg and Sayunei Formations. The magnetizations of Park's sites were complex, and more than half of his data was discarded. Among the accepted samples, anomalous declinations of a pre-Rapitan component–widespread throughout the Mackenzie Mountains–suggested that one of the sampling localities had rotated ~130° about a vertical axis after Rapitan deposition; other localities appeared to have undergone minimal post-Rapitan rotations. These rotations, inferred to have occurred shortly after Rapitan deposition (Park, 1997), were restored prior to the final evaluation of results. After these considerations, the data showed three broad groupings similar to those found by Morris (1977); however, the two low-inclination groups were less distinct in Park's dataset. The high-inclination (C, similar to Y of Morris) was determined to group best upon partial restoration of bedding, suggesting a synfolding, and hence Cretaceous, remanence. For the two low-inclination groups Park chose an order of magnetization opposite to that suggested by Morris, with group '2' (similar to Morris' Z) as a direction acquired during Sayunei deposition and group '3' (coaxial with Morris' X) gained during early hydrothermal alteration "during final rifting in Shezal time" (Park, 1997). Both groups 2 and 3 pass fold tests at the 99 percent level, although the success of these tests may have been accentuated by the visual assignment of more ambiguous data to the various directional groups (ambiguities honestly noted by Park). Although there may be other interpretations of the large and complex Rapitan paleomagnetic dataset, I follow the interpretation of magnetization ages from Park (1997) in table 1. The preferred paleolatitude for Rapitan deposition is thus 06 +8/-7°.

The Ice Brook Formation has no direct paleomagnetic constraints, and the stratigraphically nearest units that have undergone such study are the conformably overlying Blueflower and Risky Formations (Park, 1995). In that paper, Park describes four magnetic components, all considered secondary except perhaps the 'A' pole for the Risky Formation, listed here in table 1. That result derives from only 11 samples, however, and its main strength is dissimilarity to any Phanerozoic poles; further work is needed for verification. The age of the Ice Brook Formation is too uncertain to apply an extrabasinal pole for estimation of its paleolatitude.

British Columbia and the U.S. Cordillera (3-10).–Many sub-Cambrian, rift-related, diamictite-bearing sedimentary sequences occur locally throughout the Canadian and United States Cordillera. Described as the "diamictite and volcanic succession" (Link and others, 1993), the glacial rocks are indeed commonly (though not universally) associated with mafic or bimodal volcanism. Some diamictites are better dated than others, but most are generally considered as correlative with the Rapitan Group in the Mackenzie Mountains (Link and others, 1993; Ross, Bloch, and Krouse, 1995). Of course, substantially diachronous rifting across the Cordillera could negate these correlations (noted by Knoll, Blick, and Awramik, 1981). To compound matters, thermal subsidence of the passive margin did not occur until near the Proterozoic-Cambrian boundary (Bond and others, 1985), thus allowing a large degree of freedom in assigning absolute ages to each of the glacial deposits within perhaps greatly diachronous rift sequences.

The northernmost rift-related, allegedly glaciogenic unit of the Canadian and United States Cordillera forms part of the upper Tindir Group (Allison and others, 1981), a diamictite-dropstone-bearing succession similar to the Rapitan Group, even including local iron-formation (Young, 1982). Strontium isotopic data for carbonates above the glaciogenic units show low values consistent with a pre-"Vendian" age for the glacial deposits (Kaufman, Knoll, and Awramik, 1992). This contrasts with K-Ar determinations ranging from 644 ±18 to 532 ±11 Ma (reported in Van Kooten and others, 1997) on dikes presumably emplaced prior to upper Tindir Group deposition (Young, 1982); those dikes appear in map pattern to be part of the same swarm, however, so the 100-my diachroneity in K-Ar ages is unlikely to be real, and the ages are not used in this paper. Because of similarity between the lower Tindir Group and other Neoproterozoic "B" successions in the region (Rainbird, Jefferson, and Young, 1996), the partly glaciogenic upper Tindir Group is also likely to be mid-late Neoproterozoic in age, correlative with the Rapitan Group.

Farther south in the Canadian Cordillera, the Mount Lloyd George Diamictites (Eisbacher, 1981b) are only locally exposed but may correlate with diamictite lenses (possibly glaciogenic) in the Misinchinka Group (Evenchick, Parrish, and Gabrielse, 1984; Evenchick, 1988). Age constraints for these units are broad, for example in the Deserters Range, where the Misinchinka Group nonconformably overlies 728 +9/-7 Ma basement gneiss (Evenchick, Parrish, and Gabrielse, 1984) and is succeeded by fossiliferous Lower Cambrian strata (Evenchick, 1988). The Toby Formation (southeastern British Columbia and northeastern Washington State; Aalto, 1981), is also possibly glaciogenic. In Canada, it rests nonconformably atop gneiss with a late-stage leucocratic phase dated at 736 +23/-17 Ma (discordant U-Pb zircon age; McDonough and Parrish, 1991), and in the United States, it lies directly below volcanic rocks dated at 762 ±44 Ma (preliminary Sm-Nd mineral and whole-rock isochron; Devlin, Brueckner, and Bond, 1988). These ages overlap within uncertainty and suggest, assuming that all the "Toby" diamictites are indeed correlative with each other, that this alleged glacial episode occurred at ~720 to 740 Ma.

Continuing south along the Cordilleran fold-thrust belt, a metadiamictite exists within the amphibolite-grade Big Creek roof pendant of the Idaho batholith; stratigraphically, the metadiamictite is immediately adjacent to a bimodal metavolcanic suite dated at 699 ±3 Ma (U-Pb SHRIMP described in abstract only; Evans and others, 1997). Due to the metamorphic grade in this region, the sedimentary origin of the metadiamictite remains unclear. Farther south in Idaho, the better preserved Pocatello Formation contains volcanic units and diamictite members of similarly uncertain origin (Link, 1981). Smith and others (1994) interpreted a "Sturtian"-equivalent age for the Pocatello Formation based on highly enriched $\delta^{13}C$ and depleted $^{87}Sr/^{86}Sr$ values for overlying carbonates, but because of the possibility of alteration in these rocks the evidence is more suggestive than conclusive.

Evidence for glacial influence is more convincing in Utah, where diamictites are present in the Formation of Perry Canyon, Mineral Fork Formation, Sheeprock Group, and Horse Canyon Formation (Blick, 1981). Although Oberbeck and others (1994) found a single shocked-quartz-bearing clast out of 62 analyzed specimens collected from Sheeprock-Group-derived colluvium, and cited this as critical evidence for an impact-origin of the Dutch Peak diamictite within the Sheeprock Group, detailed stratigraphic studies of this formation and its nearby correlatives establish a more likely glaciogenic origin (Link and others, 1993; Link, Miller, and Christie-Blick, 1994).

In the Death Valley region of eastern California, the Kingston Peak Formation (Miller, Wright, and Troxel, 1981) contains several levels of diamictites and is associated with volcanic rocks and sedimentary iron-formation (Miller, 1985), the latter association providing compelling evidence for lithostratigraphic correlation with the Rapitan and upper Tindir Groups. The Kingston Peak Formation lies within the upper part of the Pahrump Group, separated from fossiliferous Lower Cambrian sediments by several disconformities. It is most likely younger than 1.07 to 1.09-Ga sills which may immediately postdate sedimentation of the basal Pahrump Group (Heaman and Grotzinger, 1992).

Link and others (1993) and Ross, Bloch, and Krouse (1995) propose a coeval episode of ~720-Ma rift-related glaciation throughout the Canadian and United States Cordillera. If this is correct, then a combined 723-Ma paleomagnetic pole (08°S, 336°E) from the Franklin diabase and Natkusiak basalts of northern Canada (Palmer and others, 1983; Christie and Fahrig, 1983) could be used to determine a range of paleolatitudes from as high as 18° in the Yukon (upper Tindir and Rapitan Groups) to as low as 04° in Utah (Mineral Fork Formation). Paleolatitudes corresponding to this pole are shown across Laurentia in figure 3. Although the lack of age control for most of the glaciogenic units precludes application of this result with absolute confidence, there is greater certainty for the relatively well dated Toby Formation (at 08 ±4° paleolatitude), as indicated in table 1.

Examples of younger, possibly Ice Brook-correlative glaciogenic deposits also occur in the Canadian and United States Cordillera. The locally exposed, probably glaciogenic Mount Vreeland Formation, formerly thought to be equivalent in age to the basal Windermere or Rapitan glacial episode (Eisbacher, 1985), is considered more recently to correlate with the younger Ice Brook Formation, based on a significant thickness of pre-Mount Vreeland strata within the Windermere succession of that area (McMechan, 1990; Link and others, 1993; Ross, Bloch, and Krouse, 1995). Such a consensus may be unwarranted, given the plausibility of large sedimentary thickness variations along strike ~1000 km between basal Windermere rift basins. Nonetheless, metadiamictites in the Deserters Range and Mount Vreeland areas are closely overlain by conspicuous carbonate units that may correlate with Ediacaran-bearing strata from the upper Miette Group (Hofmann, Mountjoy, and Tietz, 1985; Evenchick, 1988; McMechan, 1990); such correlations would support extrapolation of the Ice Brook glacial episode to east-central British Columbia. Unfortunately, the absence of precise and direct age constraints in this region renders depositional paleolatitudes unconstrained. A younger "Vendian" age of glaciation may be indicated by *Bavlinella faveolata* acritarchs from the Mineral Fork Formation, but such findings may instead merely indicate a broader, older stratigraphic range for those fossils (Knoll, Blick, and Awramik, 1981). Indeed, an Ice Brook-equivalent sequence boundary has been postulated for terminal Neoproterozoic successions in Utah and Idaho (Link and others, 1993). Finally, in the Death Valley region, rocks mapped as the Kingston Peak Formation include two diamictite horizons separated by an angular unconformity (Walker, Klepacki, and Burchfiel, 1986); the upper Kingston Peak strata may correlate with the Ice Brook Formation (Prave, 1999a). An even younger diamictite unit has recently been described as canyon fill of a major incision into the Rainstorm Member of the Johnnie Formation (Charlton and others, 1997; Abolins and others, 1999; abstracts only), lying stratigraphically between the

Kingston Peak Formation and fossiliferous Lower Cambrian sedimentary rocks. The diamictite facies, only locally expressed, is immediately overlain by a conspicuous "cap" carbonate with depleted $\delta^{13}C$ at several localities (M. Abolins, personal communication). The Rainstorm Member has a direct paleomagnetic determination from non-diamictite-bearing outcrops in southern Nevada (Gillett and Van Alstine, 1982), which when calculated for the Death Valley region (36.5°N, 243.5°E) gives a paleolatitude of 01 ±4° (table 1); the result may be a primary magnetization, but it is also similar to Middle Cambrian-Ordovician poles (Torsvik and others, 1996) and hence may be an overprint of that age.

In the Florida Mountains of southwestern New Mexico, a small exposure of diamictite and dropstone-bearing shale is suggested to be glaciogenic based on striated and faceted clasts of exotic origin (Corbitt and Woodward, 1981). It lies unconformably between an underlying granitic-gneissic basement and overlying fossiliferous Lower Ordovician sediments and thus was originally considered to be Neoproterozoic-Cambrian in age. However, a U-Pb zircon age of 503 ±6 Ma from the basement granite requires an Upper Cambrian age for the diamictite, which must have been deposited following rapid unroofing of the crystalline basement (Evans and Clemons, 1988). Paleomagnetic results from the ~500-Ma crystalline rocks (Geissman and others, 1991) yield a pole (06°N, 349°E) similar to other Middle-Late Cambrian paleopoles for Laurentia (Torsvik and others, 1996; Kirschvink, Ripperdan, and Evans, 1997), all implying a low paleolatitude (03 ±8°) for the diamictite, whether glaciogenic or not.

Southern and Central Appalachians (11).–In the southern Appalachians, probable glaciogenic strata occur in rift-related volcanic-clastic successions atop crystalline rocks in several parautochthonous regions. The Konnarock Formation (Rankin, 1993) contains both diamictites and dropstone-bearing rhythmites (Schwab, 1981; Miller, 1994) lying paraconformably between the 758 ±12 Ma rhyolitic Mount Rogers Formation (U-Pb zircon; Aleinikoff and others, 1995) and the Vendian-Lower Cambrian Chilhowee Group (Walker and Driese, 1991). Occurring 50 km to the south, the Grandfather Mountain Formation contains pebbly mudstone diamictites (Schwab, 1981) immediately above rhyolitic flows dated at 742 ±2 Ma; the formation nonconformably overlies a 765 ±7 Ma granite (Fetter and Goldberg, 1995). In Virginia, the Mechum River Formation and correlative units have recently been interpreted as glaciogenic and indicative of at least temporary occurrence of glacial ice at sealevel (Bailey and Peters, 1998). Those units rest nonconformably on Grenvillian basement and are cut by dikes similar in chemistry to the ~570-Ma Catoctin volcanic suite (Aleinikoff and others, 1995). Furthermore, Bailey and Peters (1998) state that the glaciogenic portions of the Mechum River Formation are, stratigraphically, slightly below rift-related volcanic rocks (Tollo and Hutson, 1996) dated at 700 to 730 Ma (U-Pb zircon; Tollo and Aleinikoff, 1996). These relations are all consistent with a possible age range of 700 to 750 Ma for the allegedly glaciogenic units of the southern and central Appalachians; in table 1, I quote the more conservative maximum age estimate of 758 ±12 Ma. Using this relatively precise age range, the ~720-Ma Franklin-Natkusiak paleomagnetic pole yields a 20° to 21° (±4°) paleolatitude for the central Appalachian glacial deposits, as shown in figure 3.

Note that Neton and Driese (1993) rejected a glacial association for the upper Grandfather Mountain diamictites. Also, in an attempt to identify the Konnarock Formation as an impact-related ejecta blanket, Rampino (1994) cited undulose extinction and lamellae in quartz grains; however, he admitted that those features are "somewhat atypical" for shocked deposits.

Northern British Isles (12).–The Port Askaig Tillite, actually a sequence of 17 glacial cycles and containing as many as 47 individual diamictite horizons (Spencer, 1971, 1981), forms a distinctive stratigraphic marker throughout the Dalradian Supergroup exposures of northern Ireland and Scotland (Max, 1981; Harris and others, 1994). Lying

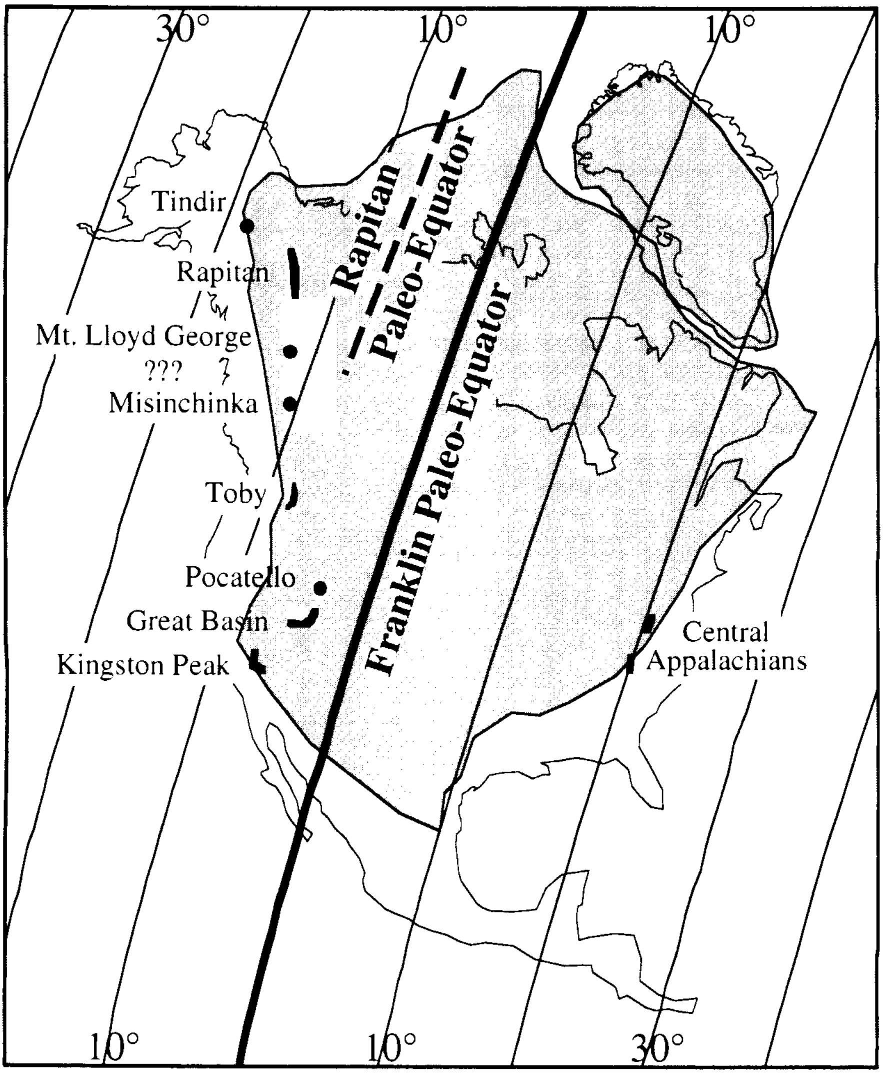


Fig. 3. Paleolatitudes of early syn-rift glaciogenic strata along the Laurentian margins, according to a combined pole for the Franklin diabase-Natkusiak basalts (723 Ma) at 08°S, 336°E (Palmer and others, 1983; Christie and Fahrig, 1983). Note that the Rapitan Group has its own paleolatitude determination of 06 ±7° (Park, 1997), and a portion of the corresponding paleo-Equator is shown as the dashed curve. The Laurentian craton is shaded; Greenland is restored to the Canadian Shield according to Bullard, Everett, and Smith (1965). Deposits are queried where ages or correlations with the Rapitan Group are especially uncertain.

north of the Iapetus suture, this glaciogenic formation was traditionally considered to have been deposited on the Laurentian margin. Within the last few years, however, a debate has developed over this issue, as well as related issues such as the various numerical ages of and correlations among multistage deformation in the Scottish Highlands, and correlation across the Great Glen fault (see Rogers and others, 1989; Rogers and Pankhurst, 1993; Robertson, 1994; Tanner and Leslie, 1994). In short, although Bluck and Dempster (1991) suggested that the Dalradian block was a Gondwanaland-derived terrane, this view was contested by Winchester (1992), Soper (1994a,b), and Soper and England (1995). The debate continues in earnest (see Tanner and Bluck, 1999; Soper, Ryan, and Dewey, 1999), extending regionally into northern Scotland and Ireland: those favoring Neoproterozoic orogeny in the Scottish Highlands, thus allowing an exotic status relative to Laurentia (Bluck, Dempster, and Rogers, 1995; Dempster, Hudson, and Rogers, 1995; Dempster and Bluck, 1995; Noble, Hylsop, and Highton, 1996; Dempster, Hudson, and Rogers, 1997; Friend and others, 1997; Bluck, Dempster, and Rogers, 1997; Rogers and others, 1998; Vance, Strachan, andJones, 1998; Highton, Hyslop, and Noble, 1999; Prave, 1999b) versus those interpreting purely Paleozoic orogeny for the Highlands and adjoining regions, thus permitting their Laurentian-marginal affinity throughout the Neoproterozoic to early Paleozoic (Soper, 1995; Tanner, 1995; Soper and Harris, 1997; Evans and Soper, 1997; Soper, Harris, and Strachan, 1998; Friedrich and others, 1999; Millar, 1999).

The Port Askaig Tillite, at the base of the Argyll Group (Harris and others, 1994) pre-dates the 595 ±4 Ma Tayvallich volcanics (Halliday and others, 1989). A maximum age for the glacial horizon, based on shear-related pegmatites that intrude the pre-Dalradian "basement" but are not observed in the Argyll Group, was first suggested at about 750 Ma (Rb-Sr on syntectonic muscovite; Piasecki and van Breemen, 1983) and has since been refined to 806 ±3 Ma (U-Pb on syntectonic monazite; Noble, Hylsop, and Highton, 1996). However, direct cross-cutting relations between the pegmatites and the glaciogenic strata are lacking. I tentatively accept the U-Pb age as a maximum estimate for the Port Askaig Tillite, which in any case might be closer in age to the overlying Tayvallich volcanics because of close stratigraphic proximity to metazoan ichnofossils (Brasier and McIlroy, 1998).

Other allegedly glaciogenic horizons exist within the Dalradian succession. At the top of the upper Dalradian succession, the partly glaciogenic Macduff Slate (Hambrey and Waddams, 1981), recently considered to be latest Proterozoic in age (Stoker, Howe, and Stoker, 1999), has been reaffirmed as Ordovician (Molyneux, 1998). The Kinlochlaggan Boulder Bed, previously considered as an older glaciogenic deposit within the lower part of the Dalradian or Grampian Group (Treagus, 1981), has recently been correlated with the Port Askaig Tillite based on detailed mapping near its type locality (Evans and Tanner, 1996) In the ensuing discussion, Treagus (1997) suggested correlation of the Kinlachlaggan Boulder Bed with the basal Vendian Smalfjord Formation in Norway (see below) and hence the Port Askaig Tillite with the Mortensnes Formation. Evans and Tanner (1997) replied that the Port Askaig Tillite spans both Norwegian levels, of which the Kinlachlaggan Boulder Bed represents only the upper. As an alternative, however, they suggest the possibility that the Kinlachlaggan Boulder Bed may correlate with a lonestone-bearing unit at the top of the Grampian Group, indicating a pre-Varanger glacial horizon in Scotland. Another variation of regional correlations is suggested by Prave (1999b), who reiterates earlier interpretations of a glacial origin for the Loch na Cille Boulder Bed, lying not far above the Tayvallich Volcanics. These suggestions await further study and, if validated, could upset previous correlations of the North Atlantic region that are influenced strongly upon the notion of a single "Varanger" or "Vendian" glacial episode (Hambrey, 1983; Soper, 1994a; see below).

The Port Askaig Tillite has been the subject of numerous paleomagnetic studies. Tarling (1974) determined a southerly low-inclination direction confirmed by the larger dataset of Urrutia-Fucugauchi and Tarling (1983) and judged by these authors to represent a primary magnetization. To the contrary, Stupavksy, Symons, and Gravenor (1982) showed that diamictite clasts carried the same magnetic direction, indicating a remagnetization which they considered to be Early Ordovician. In light of subsequent clarification of Laurussia's Paleozoic apparent polar wander path (Torsvik and others, 1996), however, the data are also consistent with a post-tilting remagnetization of Middle-Late Ordovician age, a conclusion favored by Trench and others (1989).

New geological and isotopic data have revealed another possible glaciogenic deposit in Scotland, older than the Port Askaig or its equivalents: a re-evaluation of the "Torridonian" Stoer Group as showing glaciogenic sedimentary features (Davison and Hambrey, 1996). New Pb-Pb data from carbonate strata within this sequence yield an age of 1199 ±70 Ma, interpreted as early diagenetic but probably indistinguishable within error from a depositional age (Turnbull, Whitehouse, and Moorbath, 1996). The new data, along with similarly aged Rb-Sr determinations on boulders within the Torridonian succession (~1.2 Ga; Moorbath and others, 1967) and on underlying gneisses (~1.0-1.2 Ga with no analytical uncertainties stated; Cliff and Rex, 1989), suggest a short interval of time between crystalline cooling/exhumation and initial sedimentation of the Torridonian succession; this is corroborated by similar paleomagnetic poles from the crystalline and Stoer Group rocks (Torsvik and Sturt, 1987; Piper and Poppleton, 1991). Those studies showed a pre-Torridon Group (pre-1.0 Ga using the isotopic ages from Turnbull, Whitehouse, and Moorbath, 1996) paleolatitude of 09 ±4° for the Stoer Group, interpreted by the authors as a post-depositional or early diagenetic remanence. Taken together, these results would seem to extend the record of apparently low-latitude glacial deposits to the Mesoproterozoic Era. Nonetheless, the Stoer paleopole reconstructed to North America by Bullard, Everett, and Smith (1965) is similar to 1.1-Ga poles from the Lake Superior and Grand Canyon regions (summarized by Link and others, 1993); hence the Stoer magnetization could be a pre-Torridon, 1.1-Ga overprint. Also, a glaciogenic origin of the Stoer deposits is subject to debate (Stewart, 1997; Davison and Hambrey, 1997; Young, 1999).

Arequipa massif (13).—Included here for the sake of continuity and completeness, the diamictic Chiquerío Formation in the Arequipa massif of coastal Peru is lithologically similar to the Port Askaig Tillite, with which it may have been contiguous during deposition (Dalziel, 1994). Formerly called the Faro Member (Shackleton and others, 1979) or Justa Member (Cobbing, 1981) of the Marcano Formation, its age is poorly known in the absence of correlation; it rests nonconformably atop "Grenvillian" gneisses (Dalziel, 1994; ages of 970-1200 Ma from similar gneisses ~300 km farther to the southeast, Wasteneys and others, 1995) and is intruded by granites with K-Ar hornblende cooling ages of ~440 Ma (unpublished data cited in Shackleton and others, 1979). Further study is now being undertaken to determine better the age and cratonic association of this allegedly glacial unit (I.W.D. Dalziel, personal communication). No paleomagnetic results have been determined for the Chiquerío Formation.

East Greenland (14, 15).—Glaciogenic deposits are exposed in various localities throughout East Greenland, entirely within the Caledonide foldbelt and therefore of questionable autochthoneity to Laurentia. In Gåseland-Paul Stern Land (Phillips and Friderichsen, 1981) and Charcot Land (Henriksen, 1981), respectively at ~70.5° and 72° north latitude, diamictites (the Støvfanget and Tillit Nunatak Formations) nonconformably overlie Paleoproterozoic crystalline rocks (Higgins, 1995). No minimum ages are available from isotopic studies or biostratigraphy, but these regions are involved in the pre-Devonian Caledonian orogeny. Although Higgins (1995) considered the Gåseland-Paul Stern Land and Charcot Land exposures as autochthonous windows of the Laurentian foreland, Manby and Hambrey (1989) viewed these regions as allochthonous.

Farther to the east, the Tillite Group is better exposed, better dated, and better studied (Higgins, 1981; Fairchild and Hambrey, 1995). Lying atop the Eleonore Bay Supergroup (Sønderholm and Tirsgaard, 1993) with slight regional discordance (Higgins, 1995), the Tillite Group contains a continuous succession of two diamictite horizons, the Ulvesø and Storeelv Formations, separated by the interglacial Arena Formation (Hambrey and Spencer, 1987). The succession continues upward into Cambrian strata containing Pacific-realm (Laurentian) trilobites. A loose maximum age constraint is provided by ~1000 Ma Rb-Sr ages on crystalline clasts within the diamictites, but a "Varanger" or Vendian age is commonly assumed, based on gross lithological similarities with glaciogenic deposits of Finnmark and/or eastern Svalbard (see below) as well as acritarch assemblages (Vidal, cited in Sønderholm and Tirsgaard, 1993). Correlations between East Greenland and eastern Svalbard are particularly striking and could suggest direct paleogeographic juxtaposition (see below), hence the ~940 Ma maximum age for glaciogenic deposits in the latter region may also apply to the Tillite Group. Such a conclusion finds support in recent U-Pb zircon ages of 950 to 900 Ma from the Krummedal Group that underlies the Eleonore Bay Supergroup (Strachan, Nutman, and Friderichsen, 1995; Thrane and others, 1999, abstract only), although the nature of the contact between the two successions is debatable (G.R. Watt, personal communication).

Because the Tillite Group seems to have been derived from a source region to the present west (Higgins, 1981), and because the Støvfanget and Tillit Nunatak Formations rest nonconformably on crystalline basement, it is tempting to correlate all these deposits as marking the transition from (eastern) basinal to (western) cratonic facies of the same glacial episode. However, they occur in quite different stratigraphic settings; the Gåseland-Paul Stern Land and Charcot Land deposits rest directly on crystalline basement, whereas the Tillite Group lies atop a ~10 km sedimentary succession. Therefore, I do not associate the Støvfanget and Tillit Nunatak Formations with the "Varanger" glacial epoch until further supporting evidence is found. Similarly, assignment of these exposures to correlate stratigraphically beneath the Eleonore Bay Supergroup exposures farther north (Wenk, 1961) is tenuous. Assigning a numerical age to these deposits for the purpose of assessing paleolatitudes based on a Laurentian APW path is hazardous in any case, given the scarcity of Laurentian paleomagnetic data for the interval 700 to 600 Ma. Direct paleomagnetic measurements of the Tillite Group include early studies by Bidgood and Harland (1961a,b), who found, from a limited number of samples, an apparently pre-fold (that is, pre-Devonian) direction similar to Cambrian-Ordovician directions from autochthonous North America (Van der Voo, 1993; Torsvik and others, 1996). The data show considerable scatter, indicating probably incomplete isolation of magnetic components and/or spurious results. More recently, Bylund and Abrahamsen (1997, abstract only) reported only post-folding directions from the Tillite Group.

Nordaustlandet, Ny Friesland, and Olav V Land (16).–Well preserved successions of glaciogenic-bearing strata are preserved in the Hecla Hoek "geosyncline" straddling westernmost Nordaustlandet and northeast Spitsbergen (Harland, Hambrey, and Waddams, 1993; Harland, 1997; and references therein). In the latter region, two glacial horizons occur within the Polarisbreen Group: the Petrovbreen Member of the Elbobreen Formation, and the thicker, overlying Wilsonbreen Formation. In Nordaustlandet only the thicker (upper) glacial unit is represented as the Sveanor Formation, but its correlation with the Wilsonbreen Formation is unequivocal (Hambrey, Harland, and Waddams, 1981). Both sections continue upward conformably into Lower Cambrian strata, and an older age constraint is provided only by a 939 ±8 Ma date for the Kontaktberget Granite lying several unconformities below the Polarisbreen Group (Gee and others, 1995). A Varanger- or lower Vendian-equivalent age for the Polarisbreen Group is generally assumed, as well as correlation with the Tillite Group in East Greenland (Hambrey, 1983; Harland, Hambrey, and Waddams, 1993; Fairchild and Hambrey, 1995; Harland, 1997). The northeast Svalbard-East Greenland equivalence of glacial units appears to be as robust as any Neoproterozoic interbasinal correlation, and a "Vendian" age is probably broadly accurate based on microflora (Knoll and Swett, 1987); nevertheless, specific correlation between the northeast Svalbard (or East Greenland) diamictites and the type-Varanger glacial units on the Baltic shield is not substantiated precisely. Note that Kennedy and others (1998) suggested that the Elbobreen diamictites may correlate with the older ("Sturtian") of their two proposed Neoproterozoic ice ages.

As is usually the case for Neoproterozoic glacial deposits in the North Atlantic region, the imprecise age of Hecla Hoek glaciogenic formations, coupled with the uncertain reconstruction to a Vendian craton (for example, three pre-Carboniferous tectonic domains of Svalbard; Harland and Wright, 1979) precludes indirect assignment of paleolatitudes from extrabasinal Vendian rocks of Laurentia or Baltica.

Western Spitsbergen (17).–Numerous meta-diamictites and lonestone-bearing phyllites occur in discontinuous outcrop along the western coast of Spitsbergen. Different workers have assigned local names for the various units, but they invariably can be correlated into one glacial episode containing several cycles, correlated with the better preserved exposures of "Vendian" glaciogenic strata to the northeast (Harland, Hambrey, and Waddams, 1993; Harland, 1997). The younger of these cycles is overlain by sedimentary rocks yielding "Vendian" microfossils (Knoll, 1992). Metamorphic grade in western Spitsbergen is higher than in eastern Svalbard; consequently, paleomagnetic investigations of Neoproterozoic rocks are less likely to be successful in this region and have not yet been described. Harland (1997) suggested close paleogeographic proximity between the west Spitsbergen Vendian region, including thick diamictites and associated volcanism, and the Pearya terrane on northernmost Ellesmere Island in Canada (Trettin, 1987, 1998; see below).

North Greenland (18).–A glacially derived diamictite-bearing sequence occurs near Independence Fjord as the Moræneso Formation (Clemmensen, 1981; Higgins, 1986; Collinson and others, 1989), bracketed in age between unconformably overlying Cambrian dolostone and unconformably underlying dolerite sills dated at 1230 ±20 Ma (Rb-Sr whole-rock; Kalsbeek and Jepsen, 1983). Paleomagnetic results from the underlying dolerite sills and associated basalts suggest that the Independence Fjord region is, after restoration of Baffin Bay, autochthonous with respect to Laurentia (Abrahamsen and Van der Voo, 1987). No paleomagnetic data have been obtained for the Moræneso Formation directly, however, and the poor age control of that unit precludes any application of extrabasinal paleomagnetic results to the North Greenland glaciogenic deposits.

Pearya (19).–A pre-Ordovician sedimentary succession lies within structurally complex regions of northernmost Ellesmere Island, apparently atop 1.0- to 1.1-Ga crystalline basement of the Pearya allochthonous terrane (Trettin, 1987). Within this succession, a diamictite serves as a stratigraphic marker throughout the terrane. Although tentatively considered to be at least partially glacial in origin, the diamictites lack conclusive evidence for glacial transport and are only broadly similar to other Neoproterozoic glaciogenic deposits of the North Atlantic region (Trettin, 1987), with the possible exception of western Spitsbergen (Harland, 1997; see above). The Pearya terrane is more fully described by Trettin (1998).

Baltica

Finnmark (20).–Virtually unmetamorphosed Neoproterozoic autochthonous sedimentary cover of the Baltic shield occurs along its northern margin (Finnmark), in northeasternmost Norway and northwesternmost Russia. Earlier identifications of possibly glaciogenic strata on Sredni and Rybachi peninsula (Chumakov, 1981b) are dismissed in more recent stratigraphic studies (Siedlecka, 1995). Thus glaciogenic strata in this region are limited to Norway, within the Vestertana Group and equivalent Alta Formation farther west (Edwards and Føyn, 1981). Representing the "type" deposits of the Varanger ice age, the Vestertana Group contains two diamictite-dominated units (Smalfjord and Mortensnes Formations) separated by interglacial shale, sandstone, and minor occurrences of a thin dolomitic "cap" carbonate (Nyborg Formation). The classic locality of a striated pavement beneath the Smalfjord diamictite at Bigganjar'ga (Reusch,

1891) was recently attributed to entirely non-glacial debris-flow mechanisms (Jensen and Wulff-Pedersen, 1996), leading the latter workers to question the existence of the entire Varanger glacial episode. The lower Vestertana Group, however, contains a variety of glaciogenic facies from several horizons in many localities around the region, best attributed to glacial/interglacial sedimentation (Edwards, 1984, 1997).

The precise age of this succession is poorly constrained. Pringle (1973) obtained two nearly parallel Rb-Sr isochrons from Nyborg shales, computed ages from each, and weight-averaged the two results to obtain 653 ±23 Ma (recalculated using the decay constant of Steiger and Jäger, 1977). As these data are from whole-rock analyses alone and may be derived from a combination of mineral inheritance and secondary processes, the age should not be considered reliable by present standards (Faure, 1986, p.130-131). Subsequently, Dallmeyer and Reuter (1989) dated a 1 to 2μm size fraction of detrital mica grains from the Nyborg Formation by stepwise ^{40}Ar/^{39}Ar and found monotonic increases in apparent ages from 637 to 783 Ma over a range of constant K/Ca. They interpreted these data as resulting from a ~635-Ma diagenetic isotopic disturbance within >783-Ma grains. It is not clear, however, whether their model of episodic Ar-loss during a single event is unique or whether the isotopic disturbance truly occurred during Nyborg diagenesis. Roberts and others (1997) report equally problematic illite Rb-Sr data from the Nyborg shales. More robust age constraints are paleontological. Ediacaran fauna occur ~200m above the Mortensnes Formation (Farmer and others, 1992), and microfossils from the unconformably underlying Tanafjord Group suggest an "Early Vendian" age (Vidal, 1981). In an unconventional interpretation, Kennedy and others (1998) suggested that the Smalfjord Formation might correlate with the "Sturtian" glacial interval (>700 Ma). Recognizing the potential revisions in acritarch biostratigraphy as new isotopic data from Neoproterozoic sedimentary successions emerge, I cite only a "Neoproterozoic" maximum age limit for the Finnmark glaciogenic units in table 1.

The Vestertana Group has direct paleomagnetic constraints, all from the interglacial Nyborg Formation. In a more general study of several Vendian-Cambrian formations from northern Norway, Bylund (1994) found a wide range of directions, many interpreted to be Caledonian overprints. As no field stability tests were employed (other than comparison of dispersion before and after minor tilt corrections) and because the best-grouped data came from a wide stratigraphic range, estimation of paleolatitudes from that study is uncertain. In a more concentrated study, however, a definitive pre-fold magnetic component was isolated from the Nyborg Formation, yielding a moderate paleolatitude (Torsvik, Lohmann, and Sturt, 1995). Although the resulting pole (24°N, 089°E with large uncertainty) falls near Early Ordovician results from Baltica (Torsvik and others, 1996) and the fold test probably only provides a pre-Silurian (Caledonian) age, the authors considered the magnetization to be primary. If so, then Vestertana Group paleolatitudes were 33 +12/-14° during interglacial deposition (table 1); further, if other glaciogenic units in Baltica are correlative then moderate paleolatitudes would prevail throughout the craton's Vendian ice age (fig. 4).

Scandinavian Caledonides (21-23).–Farther to the south along the foreland of the Caledonide orogen, several allegedly glaciogenic associations of diamictites and lonestone-bearing shales occur within the autochthon and lower nappe stack. Reviewed in the Hambrey and Harland (1981) volume, they constitute the following formations, from north to south: Sito and Vakkejokk (Strömberg, 1981), Långmarkberg (Thelander, 1981), Lillfjället (Kumpulainen, 1981), and Moelv (Byørlykke and Nystuen, 1981). Traditionally, these units were correlated with each other and with the Vestertana Group based on stratigraphic positions closely below fossiliferous Cambrian sediments (for example, Kumpulainen and Nystuen, 1985; Vidal and Nystuen, 1990; Vidal and Moczydlowska, 1995).

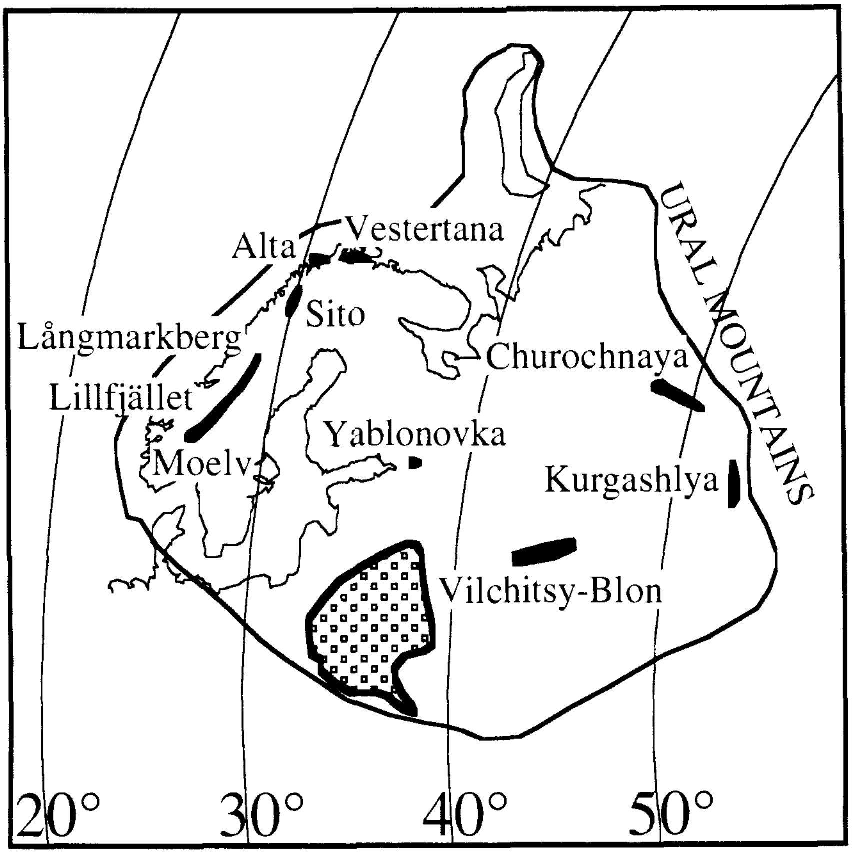


Fig. 4. Paleolatitudes across Baltica determined by the Nyborg paleopole (Torsvik, Lohmann, and Sturt, 1995). Note that the paleolatitudes carry an uncertainty of ~13°. Approximate areal extents of Neoproterozoic glacial deposits are shown in black shading or hatchures; note that they may not all be coeval (see text).

Numerical age constraints on these units are not straightforward to interpret. Within the Särv nappe, the Lillfjället Formation is cut by the Ottfjället dolerites which have yielded a variety of isotopic ages. The commonly cited age of 665 ±10 Ma is a mean of the three youngest whole-rock ages selected from a suite of previous K-Ar analyses, which are consistent with subsequent (poorly defined) ^{40}Ar/^{39}Ar results of 640 ±80 Ma from plagioclase of a single sample (Claesson and Roddick, 1983). Those authors note, however, that excess Ar is common in the Ottfjället dikes, so the 665-Ma age should be treated as a maximum. The Ekre shale gradationally overlies the Moelv Formation and yields an unpublished whole-rock Rb-Sr age, attributed to E. Welin, of 617 Ma (Rankama, 1973; age recalculated according to new decay constants of Steiger and Jäger, 1977) or quoted as 612 ±18 Ma (Vidal and Nystuen, 1990; Vidal and Moczydlowska, 1995), which I consider to be unreliable until verified by other isotopic data.

The Vakkejokk breccia (within the Torneträsk Formation of the Dividalen Group) occurs directly above shale containing *Phycodes* (or *Treptichnus*) *pedum* (Jensen and Grant, 1998), implying an Early Cambrian age that is thus demonstrably younger than the Vestertana Group glaciogenic units of Finnmark (predicted by Føyn and Glaessner, 1979). Existence of two temporally distinct ice ages (assuming the Smalfjord-Mortensnes doublet represents a single interval of glaciation in the broad sense) in the Scandinavian

Caledonides could complicate regional lithostratigraphic correlations. Some of the less-constrained units may correlate with the Lower Cambrian Vakkejokk breccia rather than the basal Vendian Smalfjord-Mortensnes succession in Finnmark.

Baltica's early Vendian glacial paleolatitudes can be extrapolated from the Nyborg paleomagnetic data (Torsvik, Lohmann, and Sturt, 1995), given that those pre-fold magnetizations are indeed primary and that the glaciogenic formations are indeed correlative (fig. 4). Paleomagnetic results from the Ottfjället dolerites showed only Silurian-Devonian remagnetizations (Bylund, 1980), as would be expected from their greenschist-amphibolite metamorphic grade. Torsvik and Rehnström (1999, abstract only) report moderate-paleolatitude results from the Torneträsk Formation. Until these data are reported in full, Baltica's Early Cambrian paleolatitudes remain controversial (Torsvik, Smethurst, and Meert, 1998; Evans, Ripperdan, and Kirschvink, 1998).

East European platform (24).–Vendian diamictites exist in numerous paleodepressions across the East European platform, known only from boreholes throughout the region and a few outcrops along the Dniester River in Podolia-Moldavia (Aren, 1981; Chumakov, 1981c-g). Some of these units (for example, the Vilchitsy and Blon Formations) may correlate with each other representing the "Laplandian Horizon", whereas others (for example, Podolian outcrops) may be nonglacial but secondarily derived from that horizon (Aksenov, 1990). In some boreholes, two suites of glacial strata are separated by an erosional disconformity (Chumakov, 1981c,d), reminiscent of the Smalfjord-Mortensnes package in Finnmark.

All across the East European platform, the volcanogenic Volhyn "Horizon" occurs directly above the glacial succession and below Ediacaran-bearing sediments (Aksenov, 1990). The Slawatycze Formation, penetrated by boreholes through the Lublin slope of eastern Poland, provides an age of 551 ±4 Ma (SHRIMP U-Pb on zircon; Compston and others, 1995); the dated bed occurs ~200m above a diamictite unit within the same borehole without any apparent intervening stratigraphic breaks (Aren, 1981). If the sub-volcanic diamictites of the Lublin slope correlate in a single broad glacial episode with the other Vendian tillites on Baltica, then the Varanger ice age occurred shortly before 550 Ma. This conclusion would seem to be inconsistent with the age of 665 ±10 Ma for the Ottfjället dolerites (Claesson and Roddick, 1983) which cut the supposedly Varanger-equivalent diamictites in the central Caledonides, but as mentioned above, the Ottfjället age should be treated as a maximum. Of course, the various diamictites and tilloids on the East European platform may be diachronous, representing several glacial intervals, but testing synchroneity *versus.* diachroneity will require further high-resolution stratigraphic or geochronological work. Paleomagnetic estimates of the East European platform's glaciogenic strata or conformable units unfortunately do not exist except in brief mention with inadequate documentation (for example, Chumakov, 1981c).

Ural Mountains (25-27).–Several levels of glacial deposits occur within the northern (Chumakov, 1981h), central (Chumakov, 1981i), and southern (Chumakov, 1981j) Ural Mountains. The general stratigraphy of these regions was summarized by Becker (1990), who described a total of three distinct glacial episodes within the lower Vendian. The episodes are represented, from oldest to youngest, by the (1) Tany and lower Vil'va Formations in the central Urals, (2) Koyva Formation in the central Urals, and (3) Churochnaya, Staryye Pechi, and Kurgashlya Formations of the northern, central, and southern Urals, respectively. All three levels are associated with deposits of sedimentary iron-formation. Widespread volcanism, characteristic of the postglacial Volhyn "Horizon" on the East European platform, is largely absent along the Uralian margin of Baltica; hence, lithostratigraphic correlation of any one of the three Uralian glacial levels with the allegedly single occurrence on the Vendian craton is tenuous. Note, however, that Chumakov (1981i) described no unambiguously glaciogenic features for the lower two levels (Tany and Koyva Formations) of the central Urals, so perhaps there is only one truly glacial level (the uppermost one) in the Uralian chain.

Becker (1990) cites Ediacaran fossils occurring above the uppermost level of correlated glacial units. Upper Riphean stromatolites occur in carbonate units unconformably below the lowest glacial level. Direct paleomagnetic studies of the upper level, the Kurgashlya Formation and equivalent ferruginous strata (Bakeevo Formation) on the western slope of the Bashkir anticlinorium in the southern Urals have produced paleopoles (Danukalov, Commissarova, and Mikhailov, 1982) that fall generally near the middle Paleozoic apparent polar wander path for Baltica (weighted mean of 01°N, 197°E). Positive fold tests for these and other Vendian formations in the vicinity (P. Mikhailov, personal communication) nevertheless only constrain the magnetization ages to pre-Permian, based on the age of deformation across the Bashkirian foreland (Brown and others, 1997). If primary, the Kurgashlya and Bakeevo results would imply paleolatitudes of ~26 ±7° for the upper Uralian glacial deposits. Of course, if both the Kurgashlya-Bakeevo and Nyborg (Torsvik, Lohmann, and Sturt, 1995) poles are primary, then the substantial apparent polar wander (APW) separating the two results would support some diachroneity of glacial deposits across Baltica. If on the other hand the Kurgashlya diamictites do in fact correlate with the Varanger glaciation, then the combined data would imply very rapid APW between the interglacial and postglacial interval. Such motion is consistent with previously determined rapid rates of APW for Vendian-Cambrian time, whether the cause be due to faster between-plate rates (Meert and others, 1993; Torsvik and others, 1996), true polar wander (Kirschvink, Ripperdan, and Evans, 1997; Evans, 1998a), or both (Meert, 1999).

East Asia

Siberian craton and environs (28-30).–The Neoproterozoic Siberian craton and its margins were long considered largely devoid of glacial deposits, but recent studies are reporting new instances of possibly glacial strata. In the northern Yenisey Range, the lithostratigraphic correlations show several depositional troughs, blanketed by first a terrigenous succession and followed by latest Proterozoic to early Cambrian dolostones (Khomentovsky, 1990). Within the Teya-Chapa trough in the northern part of the range, the Chivida Formation of the Chingasan Group contains "boulder-pebble mudstones" that have been considered glaciogenic by some authors and non-glacial by others (see references in Postel'nikov, 1981). What is particularly intriguing about these deposits is their possible equivalence to ore-grade iron formation of the Nizhny Angara Formation occurring in the southern part of the Yenisey Range (Khomentovsky, 1990). Temporal equivalence of the Chivida tilloids and the iron-formation could strengthen the hypothesis of a glacial influence, given the general association of Neoproterozoic iron-formations with glaciogenic deposits (see Kirschvink, 1992). The diamictites could also have been debris-flow or volcaniclastic deposits, with or without minor glacial contribution (Postel'nikov, 1981).

Notably, Khomentovsky (1990, 1997) concluded that the Chingasan Group is pre-Yudomian and that the Yudomian of Siberia correlates with the Vendian of Baltica; hence, if the Chivida tilloids are glacial then they do not correlate with the "Laplandian" ("Varanger") of the Russian platform and the Ural Mountains. The Ediacaran fossil *Cyclomedusa davidi* Sprigg was found from the Ostrov Formation, which is a platformal deposit overlapping the various sub-basins of the Teya-Chapa trough; this provides a firm minimum age for the Chingasan Group (Khomentovsky, 1990). Upper Riphean clasts within the diamictites provide a maximum age (Postel'nikov, 1981). A paleomagnetic determination of ~10° paleolatitude for the Oslyanka Group, which contains the Nizhny Angara iron-formation (Vlasov and Popova, 1968), is of zero reliability on the Q-scale (Van der Voo, 1990), but an updated tabulation by Khramov (1984) combined poles from several studies to yield a slightly better result (Q = 1) with apparently dual magnetic polarity (table 1). The listed pole (16°S, 115°E) is similar to Ordovician and Silurian poles for Siberia (Torsvik and others, 1995; Smethurst, Khramov, and Torsvik, 1998), and may either be primary or an overprint of possibly early Paleozoic age. Paleomagnetic results from the Kandykskaya Suite in the eastern Aldan Shield (summarized by Smethurst, Khramov, and Torsvik, 1998) would indicate tropical paleolatitudes for the Chingasan Group if both successions are coeval.

Other possibly glaciogenic formations occur along the southern margin of the Siberian craton. In the Patom region, Chumakov (1981k) reported several levels of tilloids within the Teptorgo and overlying Patom Groups, the latter of which was referred to as the Dzhemkukan tillite by Khomentovsky (1997), who correlated it with the Chingasan Group within the Baikalian System. Khomentovsky also mentioned glaciogenic strata within the Dalnyaya-Taiga Group, but it is unclear from the limited information available to me, whether this is simply another name for the tilloids described by Chumakov (1981k). Khomentovsky (1997) also cited maximum and minimum ages for the Baikalian deposits at 850 and 690 to 630 Ma, but these are of unknown reliability. Within the Sayan foldbelt, the late Neoproterozoic Darkhat-Khubsugul rift allegedly contains a glacial level comprising tillite and dropstone-bearing shale (Ilyin, 1990).

East of the Verkhoyansk foldbelt, the Omolon massif contains little-deformed Neoproterozoic clastic and carbonate successions overlying crystalline basement (Zonenshain, Kuzmin, and Natapov, 1990). Although Zonenshain and coworkers (*1990,* p.126) mention "Vendian tillites" at the base of the sedimentary succession, no glaciogenic rocks are reported in the more lengthy description by Komar and Rabotnov (1977).

Kazakhstan (31).–Diamictites of probably nonglacial origin exist throughout Kazakhstan and northern Kyrgyzstan, deposited in active tectonic settings. Two levels are known in central Kazakhstan, as the upper Baykonur (Kheraskova, 1981a,b) and lower Satan (Kheraskova, 1981c) Formations. Possibly equivalent strata occur near Lake Balkhash as the Kapal (Kheraskova, 1981d) and Shopshoky (Kheraskova, 1981e) Formations. Farther south, tilloids are either correlated with the Baykonur Formation or named the Dzhetym tilloid (Korolev, Maksumova, and Sagyndykov, 1981), and to the east in the Dzhungar Alatau Range they occur at two levels within the Tyshkan Group (Korolev, 1981), possibly correlatable with Chinese "tillite" deposits on the so-called Yili microcontinent (Chen and others, 1999; location 1 of Wang and others, 1981). Many of these deposits are associated with iron-formation, usually in the form of hematitic shales. Active tectonic settings during deposition is indicated by volcanic units within the sedimentary successions. A Vendian age is possible for all the deposits, which commonly underlie Cambrian strata and locally contain clasts of fossiliferous Riphean carbonates or volcanics dated at 800 to 850 Ma (of unknown reliability; quoted by Kheraskova, 1981a,c). The southernmost occurrences rest nonconformably upon granitic rocks dated at 660-665 ±60 Ma (also unknown reliability; U-Pb results quoted by Korolev, Maksumova, and Sagyndykov, 1981).

All the deposits occur within the southwesternmost imbrications of Altaid foldbelt, palinspastically restored to a common segment of the Siberia-marginal Kipchak arc envisioned by Sengör and Natal'in (1996; their paleotectonic units 1,3,4,6). Although no undoubted glaciogenic features have been described from these units, their common association with iron-formation–as is the case for the more likely glaciogenic Uralian and Siberian diamictites (see above) which perhaps were deposited on the opposite side of the same backarc basin (Sengör and Natal'in, 1996)–may suggest a minor glacial influence. In any case, glaciers in this region would most likely be small alpine entities with lesser importance for assessing global paleoclimate.

Western Mongolia (32).–Two possibly glaciogenic diamictite levels are described from Neoproterozoic strata of the Zavkhan basin in western Mongolia, within the basal

Maikhan Uul Member of the Tsagaan Oloom Formation (Lindsay and others, 1996). This deposit is bracketed in age between unconformably underlying volcanic deposits of the Dzabkhan Formation (unpublished isotopic ages of 732-777 Ma, cited by Lindsay and others, 1996) and paraconformably overlying deposits of lowermost Cambrian age. Carbon- and strontium-isotope stratigraphy from higher units in the Mongolian sections suggests that the Tsagaan Oloom diamictites are older than "Vendian" age (Brasier and others, 1996). In addition, Brasier, Green, and Shields (1997) correlated a prominent sequence boundary from higher in the section, immediately below the abrupt strontium-isotopic rise, to the so-called "Varangerian" ice age. From the Lower Cambrian part of the overlying succession (Bayan Gol Formation), a pre-fold magnetic remanence suggested a paleolatitude of ~44°, but that result may be a Silurian-Devonian overprint (Evans and others, 1996). The Tsagaan Oloom glacial deposits are therefore still unconstrained in paleolatitude.

Tarim (33).–Sinian (terminal Proterozoic) glacial deposits are common within the Tarim block, mainly along the southern margin of the Tian Shan which bounds the block to the north (locs. 2-3 of Wang and others, 1981; Chen and others, 1999). Gao and Qian (1985) also describe an occurrence in the western Kunlun Mountains. At the best exposed area near Quruqtagh, three diamictite levels have been recognized and interpreted as tillites (Gao and Qian, 1985). From bottom to top, they are named the Beiyixi, Tereeken or Altungol, and Hangelchaok Formations. Any of these levels may be missing from other localities (Wang and others, 1981; Lu and Gao, 1994). The uppermost Hangelchaok Formation is disconformably overlain by Lower Cambrian strata and overlies Vendotaenid-bearing units, suggesting a latest Neoproterozoic age (Wang and others, 1981). The lower two glacial levels are commonly correlated with those of the South China craton (for example, Li, Zhang, and Powell, 1996; see below), but direct dating of the Tarim deposits is lacking. Allegedly, a fourth glacial level may lie within the pre-Sinian Palgang Group at Quruqtagh, but further study is needed to verify its alleged glacial origin (Gao and Qian, 1985). Note that the first extensive study of the Quruqtagh region (Norin, 1937) identified only the Tereeken and Hangelchaok levels as likely to be glaciogenic, plus possibly an intervening varved horizon with lonestones; the basal Beiyixi Formation was interpreted as a fluvial arkose.

A magnetostratigraphic study of the Tarim Sinian revealed a two-polarity remanence with apparently stratabound polarity zones broadly consistent among several stratigraphic sections at Wushi (Li and others, 1991). The section spans two "tillite" horizons (correlating with the Beiyixi and Tereeken Formations, Gao and Qian, 1985) with little change in the measured paleolatitude of about 8°. The invariance of this result through great stratigraphic thickness and an angular unconformity, thus plausibly through a long span of time, implies that either the Tarim block experienced little latitudinal or rotational motion during that interval, or that the paleopole represents a magnetic overprint, perhaps during Late Cretaceous time based on similarity to poles of that age (Zhao and others, 1996).

North China block (34).–In contrast to the multiple occurrences in Tarim, only one glacial level is present in any one location within the North China block; that level is called the Luoquan, or locally, the Zhengmuguan Formation (Wang and others, 1981; Mu, 1981; Guan and others, 1986; Zheng and others, 1994). The age of this level is debatable, ranging from ~750 Ma (that is, correlated with the Nantuo or Gucheng tillite; see below) to Early Cambrian (references in Xiao and others, 1997). The unit disconformably overlies probably Mesoproterozoic sedimentary rocks (Xiao and others, 1997) and is disconformably overlain by Early Cambrian strata (Piper and Zhang, 1997). Based on microfossil occurrences in the Luoquan and adjacent units and biostratigraphic correlations elsewhere within China, Guan and others (1986) and Yin and Guan (1999) consider the Luoquan Formation to have been deposited near the Proterozoic-Cambrian boundary.

Mu (1981) reports paleomagnetic data of unknown quality for the Luoquan Formation, which would indicate paleolatitudes of ~50 to 65° for the glaciated North China block if reliable. The corresponding pole position is similar to an interpolated path between Early Triassic and Late Jurassic poles for North China (Van der Voo, 1993; Zhao and others, 1996) and hence may be attributed to an overprint related to the Triassic collision between North and South China blocks along the adjacent Qinling-Dabie Shan orogen (Yin and Nie, 1996). Piper and Zhang (1997) describe paleomagnetic results from the Luoquan-equivalent Fengtai Formation in Anhui Province; they assign a syn-compaction age to the characteristic magnetization, based on dual polarity, a failed conglomerate test, apparently synfolding remanence from downwarped beds beneath lonestones in the diamictite, and similarity to the direction from overlying Cambrian rocks. However, the fold tests are inconclusive because of the small sample size, and the paleopole (in geographic coordinates) is similar to previous Mesozoic results from the North China block.

Occurring primarily in North Korea, the Pirangdong Series of the terminal Proterozoic Kuhyon System contains numerous occurrences of pebble-bearing phyllite, schist, and limestone; although lacking truly diagnostic glaciogenic features, this stratigraphic unit has been correlated with the basal Sinian "tillite" of the North China block (Ri and Om, 1996). It occurs in the same general region as the tilloid-bearing Kuken sedimentary succession, briefly described by Harland (1981), and the two studies may actually be referring to the same rocks using different stratigraphic names.

South China block (35).–At most two glacial levels have been identified within any single section of the cratonic and geosynclinal successions of South China, but correlations among them are not universally straightforward (fig. 5; see Li, 1998). In type section of the lower Yangtze Gorges, only one diamictite exists, named the Nantuo Formation. Underlying the diamictite conformably or with a subtle disconformity, the Liantuo Formation is a fining-upward siliciclastic-volcanic unit dated at 748 ±12 Ma and rests nonconformably upon the 819 ±7 Ma Huangling granite (U-Pb SHRIMP; Ma, Lee, and Zhang, 1984). Slightly farther southeast at Gucheng, Hubei, two diamictite levels are separated by the Mn-bearing Datangpo Formation (Wang and others, 1981). Because a siliciclastic unit underlies the lower of the two diamictites (called Gucheng), that unit is correlated with the type Liantuo Formation (Wang and others, 1981). In southeastern Guizhou Province, two diamictite horizons are commonly assigned to different ice ages; the lower Chang'an unit is separated from the upper diamictite (generally named or correlated with "Nantuo") by the volcanic-sedimentary Fulu Formation. Liao (1981) and Wang (1986) correlated the Fulu and Liantuo Formations based on their common occurrence disconformably below diamictites correlated with the Nantuo Formation (fig. 5B).

Alternative stratigraphic correlation schemes are possible. Recognizing the similarity between the Fulu Formation, which contains hematitic iron-formation near its base, and the Mn-bearing Datangpo Formation in southern Hubei, Lu and others (1985) considered the Chang'an diamictite as an equivalent of the Gucheng Formation (fig. 5C). This correlation is problematic because it rejects the Liantuo-Fulu equivalence long favored by stratigraphic syntheses of South China (Wang and others, 1981; Liao, 1981; Wang, 1986). Nonetheless, it is favored by Li, Zhang, and Powell (1996) and Li (1998).

The alternative correlations carry different implications for the absolute ages of the Sinian glacial levels. In the scheme of Lu and others (1985) and Li, Zhang, and Powell (1996), the 748-Ma Liantuo age provides a maximum limit for all Sinian glaciations in South China (fig. 5C). According to the correlations by Wang and others (1981), Liao (1981), and Wang (1986), however, the 748-Ma date is a maximum for the Nantuo but a minimum for the Chang'an deposit (fig. 5B). In the latter case, a maximum age for the Chang'an glaciation is provided by ~820-Ma post-orogenic granites (Li, 1999) of the so-called Jinningian movement (Wang, 1986) or Sibao orogeny (Li, 1998), unconformably overlain by relatively undeformed Sinian strata (Li, 1999). To be conservative, I have listed the latter age as a maximum constraint for the Chang'an deposits in table 1. In both models, the 748 ±12 Ma date is a maximum for the Nantuo level.

The earliest attempts at paleomagnetism of the lower Sinian deposits produced high paleolatitudes; conversely, subsequent studies showed consistently low paleolatitudes (summarized by Evans and others, 2000). Recent work, however, is converging upon a moderate paleolatitude for the Sinian glacial deposits. An unpublished Sinian paleolatitude of 40° ±7° quoted by Meert and Van der Voo (1994) is consistent with the fully documented study by Zhang and Piper (1997); their most stable, hematitic, two-polarity direction yields a paleolatitude of 37° +6/-5° for strata immediately underlying (Chengjiang Fm.) and overlying the "Nantuo" diamictite in Yunnan Province (probably a correlative of the Nantuo *s.s.* as defined in the Yangtze gorges). Their result is broadly consistent with paleomagnetic results from other laboratories on the type locality of the Liantuo Formation, and all the data are incorporated into a mean lower Sinian pole yielding paleolatitudes of 30° to 40° for the South China block (Evans and others, 2000). The combined pole is interpreted to be primary based on stratigraphically consistent magnetic polarity zonations and a soft-sediment fold test. Close similarity of the Liantuo (dated directly at 748 ±12 Ma) and Chengjiang-Nantuo paleopoles, support the suggestion that both the Chang'an and Nantuo Formations are ~750 Ma in age, and that the paleomagnetic results apply to all the Sinian glacial deposits on the South China block, regardless of the correlations (fig. 5).

East Gondwanaland

Arabian peninsula (36).–The Huqf Supergroup crops out in southeastern Oman and is also encountered in numerous boreholes across the eastern Arabian peninsula. Although no glaciogenic features have been reported from the type section of the Huqf Supergroup, diamictite levels are present in a correlative unit of the basal Huqf succession farther south (Ghadir Manqil-1 borehole and Mirbat region) as well as in outcrop farther north, where the dropstone-bearing Mistal or Ghadir Manqil Formation is exposed in the Jebel Akhdar tectonic window beneath the Semail/Hawasina nappe stack of the Oman Mountains (Gorin, Racz, and Walter, 1982; Hughes Clarke, 1988). Brasier and others (2000) report a U-Pb zircon age of 723 +16/-10 Ma from ash beds within the glaciogenic Ghubrah Member of this formation. They also report overlying glaciogenic deposits of the Fiq Member that are separated from the former by a cap carbonate, suggestive of a distinct ice age altogether. The allegedly separate, upper glaciogenic horizon occurs conformably below Cloudina-bearing sediments (Conway Morris, Mattes, and Menge, 1990; Brasier and others, 2000), indicating a minimum Ediacaran age.

In the first published paleomagnetic study of the Huqf Supergroup, Kempf and others (2000) have found a high-unblocking-temperature component that indicates low paleolatitudes for the Mirbat Sandstone, containing glaciogenic features and correlated with the Abu Mahara Formation. It is uncertain whether the Mirbat Sandstone would correlate with the Ghubrah or Fiq Member, so age constraints on the paleomagnetically sampled unit are between ca.730 and 550 Ma (Brasier and others, 2000). Meaningful results were obtained from only 10 samples among two sites, with no field tests to constrain the age of magnetization. Furthermore, the resulting paleomagnetic pole is similar to previously determined Early Cambrian paleomagnetic poles from Gondwanaland. For these reasons, the Mirbat result is listed in table 1 as possibly primary but is excluded from the final compilation of the most reliable paleomagnetic constraints.

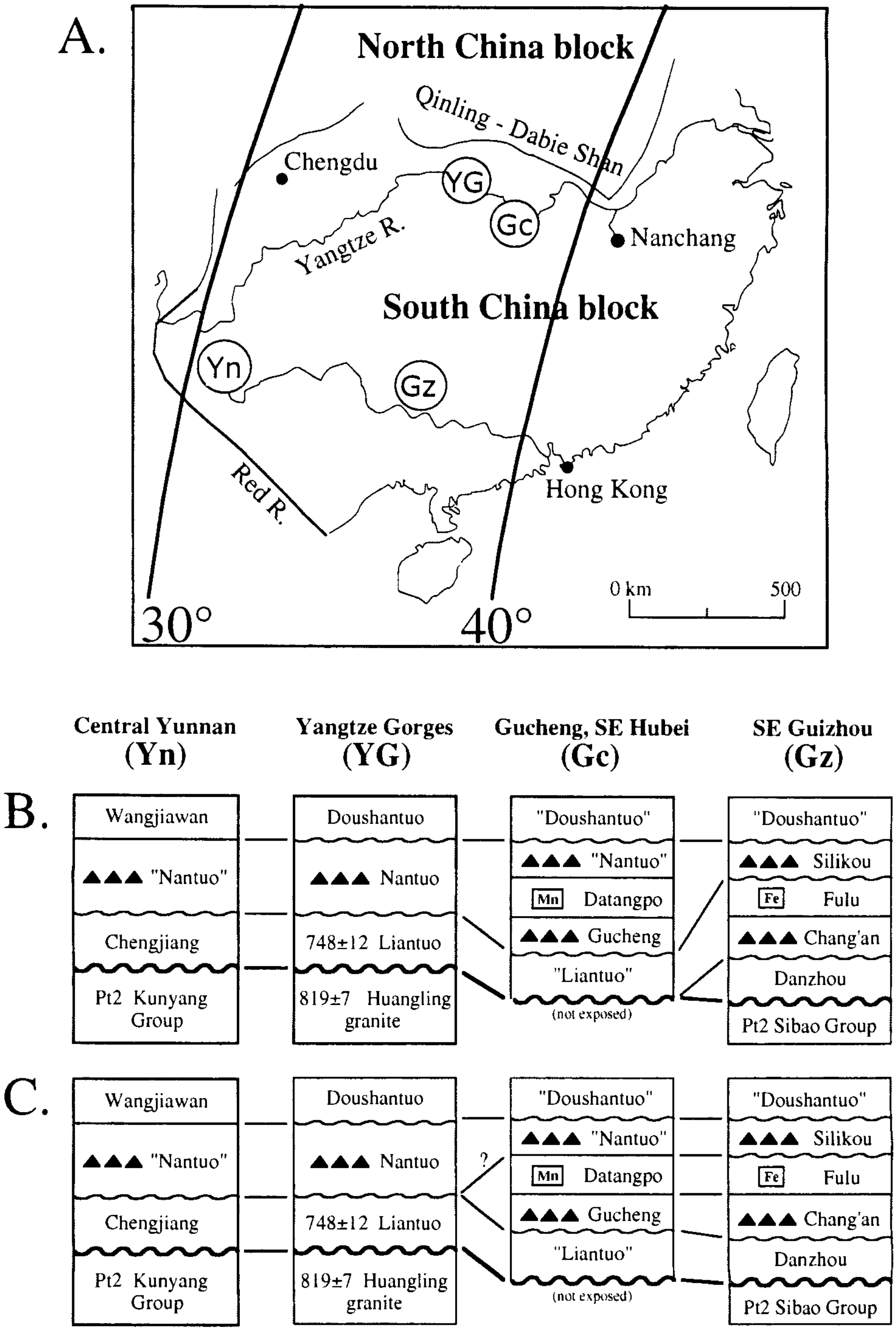


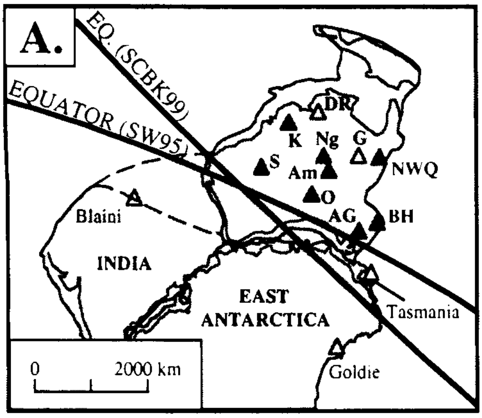
Fig. 5. Alternative correlation schemes for Neoproterozoic (Sinian) volcanic-sedimentary successions on the South China block, adapted from Evans and others (2000). (A) Location map for stratigraphic columns, with paleolatitudes according to the new lower Sinian paleomagnetic pole (Evans and others, 2000). Yn = Yunnan Province; YG = Yangtze Gorges, western Hubei Province; Gc = Gucheng, southeast Hubei Province; Gz = southeast Guizhou Province. (B) Correlation scheme of Liao (1981) and Wang (1986). (C) Alternative correlation by Lu and others (1985), adopted by Li, Zhang, and Powell (1996). In (B) and (C), dark wavy contacts indicate profound stratigraphic breaks such as angular unconformities or nonconformities; light wavy curves show slight disconformities on the regional scale (may not actually represent significant hiatuses); straight contacts are conformable. Quoted names indicate occurrences outside of the type localities. Triangles depict glaciogenic deposits. "Mn" and "Fe" depict formations with bedded manganese and iron-formation, respectively. Ages in Ma (from Ma, Lee, and Zhang, 1984).

India (37, 38).–The Blaini Formation, occurring within the Lesser Himalaya, was long mistaken to correlate with the Carboniferous-Permian Talchir Boulder beds of the Gondwana system; subsequent paleontological discoveries of the Blaini and overlying Krol-Tal succession generated overwhelming evidence in favor of a Neoproterozoic age (Singh, 1979; Mathur and Shanker, 1990). I tentatively follow the regional correlations by Valdiya (1995) in assigning a probable Neoproterozoic (rather than Mesoproterozoic) age for the Blaini Formation, although maximum age limits are poor. A glaciogenic origin of the Blaini diamictites has been debated, but the unit has many features that would suggest at least indirect glacial influence (summarized by Brookfield, 1987). Possibly correlative units are the newly named Manjir Formation of the Haimanta Group in the Higher Himalayan Crystalline belt (Frank and others, 1995) and the Tanakki diamictite in northern Pakistan (Brookfield, 1994), both with overlying "cap" dolostones.

A paleomagnetic study of the Blaini and Krol succession was hampered by the assumption of a Carboniferous-Permian age for the glaciogenic strata (Jain, Klootwijk, and Goswami, 1981). Given such limitations, that study assigned magnetization ages (in the absence of a fold test) through liberal use of vertical-axis rotations and comparisons with the Carboniferous-Recent apparent polar wander path for India. Obviously, if a Neoproterozoic age had been considered, more freedom would have existed in choosing viable magnetization ages; unfortunately, incomplete presentation of the data prohibit a thorough reevaluation of the results. For the sake of completeness I include a weighted mean of results from the three stratigraphic units containing the stated "primary" direction in table 1.

Although not cited as glaciogenic by recent sedimentological studies, the Chanda Limestone of the Penganga Group in the Godavari basin in east-central India contains some unusual characteristics in common with Neoproterozoic glaciogenic strata. First, the formation calls attention to itself by its banded manganese ores (Bandopadhyay, 1996), similar in texture and genesis to many of the Neoproterozoic iron-formations which are in turn commonly associated with glaciogenic deposits (for example, Rapitan and Urucum deposits as described above and below; Gutzmer and Beukes, 1998). Second, Chaudhuri and others (1989) describe "varve-like" units within the limestone, and Bandopadhyay (1996) describes several horizons of "matrix-supported chaotic polymictic conglomerate" with a variety of clast lithologies, some "exotic" to the basin. Although Bandopadhyay (1996) prefers an active-rift setting for these unquestionably deep-water deposits, they deserve at least mention as containing true diamictites (in the purely descriptive sense) and invite further study. Although unpublished and probably unreliable Rb-Sr glauconite ages of 775 ±30 and 790 ±30 Ma are reported for the Chanda Limestone (Chaudhuri and others, 1989), its micofossil assemblages support a probable Neoproterozoic age (Bandopadhyay, 1989).

Australia (39-44).–Neoproterozoic glaciogenic rocks are widespread throughout Australia (Dunn, Thomson, and Rankama, 1971): on or near the Kimberley block, within the "Centralian Superbasin" (Savory, Officer, Amadeus, Ngalia, and Georgina structural basins; Walter and others, 1995), and in the Adelaide "geosyncline" (fig. 6). Most of these areas contain two glacial levels, commonly considered to represent an older "Sturtian" and a younger "Marinoan" episode (using the chronostratigraphic terminology), and each level may contain glacial-interglacial stratigraphic repetition (see Preiss and Forbes, 1981; Brookfield, 1994). From present south to north they are described below, followed by a discussion of possible correlations and relevant paleomagnetic data. Note that recent correlations (see below) have indicated the possibility of two discrete glacial levels within the Marinoan (terminal Proterozoic) chronostratigraphic interval; therefore, one cannot simply refer to a "Marinoan" ice age when constructing correlations, even within Australia. What in common usage has been termed the



B. southern Australia

	Adelaide region	Broken Hill	
Umberatana Gp.	Yerelina Subgroup	Teamsters Creek	M
	Yudnamutana Subgroup	Yancowinna Subgroup	St
Burra Gp. 777 ±7			
Callana Gp. 802 ±10			

C. central Australia

Savory	Officer	Amadeus	Ngalia	Georgina	NW Qnslnd	
Boondawari	Lupton	Olympic	Mount Doreen		Little Burke	M
	↑ (?) Chambers Bluff	Areyonga	Naburula	Yardida		St

D. Kimberley region

Coats and Preiss, 1980

Walsh	Egan	Moonlight Valley/Fargoo ↓	M
Landrigan		(?)	St

Plumb, 1981 — **Plumb, 1996; Corkeron et al., 1996**

Egan			M2
Walsh	Landrigan	Moonlight Valley/Fargoo	M1

Fig. 6. Correlation of Neoproterozoic glaciogenic deposits of Australia. (A) Reference map including possibly correlative deposits in other sectors of East Gondwanaland (restored according to Powell and Li, 1994), with lower Marinoan paleo-equators according to Schmidt and Williams (1995), and Sohl, Christie-Blick, and Kent (1999). Solid (open) triangles represent deposits that are confidently (tenuously) correlated with the lower Marinoan glaciation. AG = Adelaide "Geosyncline", Am = Amadeus basin, BH = Broken Hill region, DR = Daly River basin, G = Georgina basin, K = Kimberley region, Ng = Ngalia basin, NWQ = northwest Queensland, O = Officer basin, and S = Savory basin (B) Correlations within southern Australia (Coats and Preiss, 1987). (C) Correlations within the "Centralian Superbasin" of central Australia (Walter and others, 1995). (D) Alternative correlations within the Kimberley region of Western Australia. In panels (B)-(D), St = Sturtian glacial epochs, M = Marinoan, M1 = lower Marinoan, M2 = upper Marinoan.

"Marinoan" glacial interval, I refer to as the "lower Marinoan" interval. Finally, Australian lithostratigraphic names commonly include the term "Tillite" even if the rocks may be non-tillitic, merely glacially derived deposits; below, I follow the lithostratigraphic nomenclature without regard to precise sedimentological origin.

The Sturtian and lower Marinoan glacial ages of the Umberatana Group are recorded by many discontinuous occurrences of glaciogenic sediments in the Adelaide "Geosyncline" in South Australia (Coats, 1981; Coats and Preiss, 1987; Preiss, 1993; fig. 6B). The older, Sturtian episode is associated with iron-formation and commonly comprises two levels of diamictites. Local names of Sturtian diamictites are the Pualco and Bolla Bollana Tillites (lower level) and the Merinjina and Appila Tillites (upper level). In addition, the Bibliando and Hansborough Tillites are included in the Sturtian interval. The younger, lower Marinoan glacial episode is more continuously exposed across a variety of depositional facies. On the Stuart shelf to the west, the Whyalla Sandstone fills structures resembling frost wedges in a subaerial setting (see Williams, Schmidt, and Embleton, 1995). Glaciomarine deposits include the Elatina Formation and, farther east, the Mount Curtis and Pepuarta Tillites. The deposits are commonly overlain by a fine-grained "cap" carbonate, the Nuccaleena Formation or its equivalents. At a higher level and contemporaneous with the terminal Proterozoic Pound Subgroup containing diverse Ediacaran fossils (Preiss, 1993), the Billy Springs Formation contains diamictites and lonestone-bearing clastic sediments that were suggested to represent a third interval of glaciation in the Adelaide geosyncline (DiBona, 1991); this hypothesis is tenuous, however, as a non-glaciogenic origin is compatible with both the sedimentology and the regional stratigraphic facies (Jenkins, 1995; N. Christie-Blick, personal communication).

In the Broken Hill area, part of the Willyama inlier east of the Adelaide geosyncline, two diamictite levels bracket the Torrowangee Group: the basal Yancowinna Subgroup and the terminal Teamsters Creek Subgroup (Tuckwell, 1981). The lower of these occurs in fault-controlled basins with coarse, mainly locally derived detritus. The upper, Teamsters Creek Subgroup contains two levels of more uniformly deposited diamictite, containing a greater proportion of exotic clasts. Direct age constraints on the Torrowangee Group are lax, somewhere between Paleo- to Mesoproterozoic (age of basement) and the Cambrian (age of overlying sedimentary rocks). Following Coats and Preiss (1987), Young (1992b) correlated the Yancowinna and Teamsters Creek deposits with the Sturtian and (lower) Marinoan, respectively, of the Adelaide geosyncline.

Farther south, on King Island and in Tasmania, sporadic outcrops of possibly Neoproterozoic diamictites exist (Jago, 1981). The best preserved of these is the Cottons Breccia on King Island, where a glaciogenic origin is possible but not conclusive. The deposit contains the acritarch *Bavlinella faveolata*, widely considered diagnostic of the Vendian; however, this species was also found in the Mineral Fork Formation of the western United States, possibly ~700 to 750 Ma (see above). The Cottons Breccia is overlain conformably or disconformably by thin, laminated dolostone which may correlate with the Nuccaleena "cap" of the lower Marinoan in the Adelaide geosyncline (Calver and Walter, 2000). Alternatively, thick basaltic and pyroclastic rocks higher in the section suggest a correlation with the Chambers Bluff tillite in the Officer basin, a possible Sturtian equivalent (Coats and Preiss, 1987). In Tasmania, the diamictic Julius River Member (formerly "Trowutta Breccia"; Jago, 1981) of the Black River Dolomite has a similar association with overlying volcanic rocks and has recently been assigned to the Sturtian interval based on carbon-isotope stratigraphy (Calver, 1998). In both areas, the alleged glaciogenic deposits unconformably overlie higher-grade rocks of the Wickham Orogeny, whose granitoids have been dated at 760 ±12 Ma and 777 ±7 Ma (SHRIMP U-Pb on zircon; Turner, Black, and Kamperman, 1998); however, the stratigraphic relationships on King Island are less certain (Bottrill and others, 1998).

Farther to the north, a vast region of central Australia contains three major structural basins (Officer, Amadeus, and Georgina) and subsidiary outliers (for example, the Ngalia basin) separated by east-west trending horsts which expose earlier Proterozoic crystalline basement. Neoproterozoic sedimentary deposits of the basins have been correlated into a "Centralian Superbasin" (Walter and others, 1995; Walter and Veevers, 1997), following earlier correlations of lithology (Preiss and Forbes, 1981) and tectonic subsidence (Lindsay, Korsch, and Wilford, 1987). As in the Adelaide region, glaciogenic deposits can be grouped into two episodes coinciding with the Sturtian and (lower) Marinoan ice ages (fig. 6C). The Officer basin contains an apparently Sturtian representative, the Chambers Bluff Tillite (Preiss, 1993; Lindsay and Leven, 1996), although the correlation is tenuous because of lithological disparities between the Officer and Adelaidean basins. For example, the Wantapella Volcanics, overlying the Chambers Bluff Tillite with apparent disconformity, have no direct lithostratigraphic equivalent in the Umberatana Group of the Adelaide geosyncline (Preiss, 1993). In the western Officer basin, the sporadically exposed Lupton and Turkey Hill Formations are interpreted as glaciogenic, possibly correlative with either Sturtian or (lower) Marinoan ice ages (Townson, 1985). A diamictite unit with overlying cap carbonate has been penetrated by borehole, correlated with the Lupton Formation, and interpreted as a probable (lower) Marinoan equivalent (Grey and others, 1999).

In the Amadeus basin of central Australia, the Sturtian glaciation is represented by the Areyonga Formation, and the lower Marinoan ice age is recorded by the Olympic Formation (Wells, 1981). In the Ngalia basin, a structural outlier north of the Amadeus basin, equivalent strata are named the Naburula and Mount Doreen Formations, respectively (Wells, 1981). Within any other individual basin, only one interval is represented by true glaciogenic deposits. A correlative of the Areyonga Formation in the southern Georgina basin is the Yardida Tillite and its equivalents, lying within distinct pre-Paleozoic grabens (Walter, 1981). Farther to the east, the Little Burke Tillite of the Mount Birnie beds is believed to correlate with the (lower) Marinoan glaciation based on their distinctive cap carbonate (Plumb, 1981b). On the westernmost side of the Centralian Superbasin, in the so-called Savory basin, the Marinoan interval includes the glaciogenic Boondawari Formation (Walter and others, 1994; Bagas and others, 1999).

In the Kimberley block and along its margins, several diamictites and overlying cap carbonates occur in discontinuous exposure (fig. 6D). Each of the glacial formations except one (Fargoo) rests locally upon a striated pavement. Initial mapping of the Kimberley region led to its original intrabasinal correlations described by Plumb (1981a), with a Marinoan glacial episode represented by the Egan Formation, and a Sturtian episode represented by the Walsh, Landrigan, and Moonlight Valley/Fargoo Tillites. The latter pair of glaciogenic units would thus document several glacial cycles within the Sturtian interval. An alternative correlation proposed by Coats and Preiss (1980) and accepted by Brookfield (1994) and Kennedy (1996), assigned the Landrigan (and perhaps Fargoo) diamictites to the Sturtian, and the Walsh, Egan, and Moonlight Valley diamictites to the Marinoan interval. Recently, Plumb (1996) and Corkeron and others (1996) have favored the first set of correlations but have shifted the entire stratigraphic package younger relative to the Sturtian and Marinoan deposits of central and southern Australia (fig. 6D). Thus, there may be no Sturtian glacial unit in the Kimberley region (except perhaps the Fargoo Tillite), and the Egan Formation would solely represent a younger, only locally developed, glaciation occurring within the later stages of the Marinoan interval (Grey and Corkeron, 1998). Nonetheless, Li (2000) presented paleomagnetic evidence that at least the Walsh Tillite cannot be coeval with the lower Marinoan deposits in South Australia (see below).

The Fargoo and Moonlight Valley glacial deposits are definitely Neoproterozoic in age, lying unconformably above sedimentary rocks of the Victoria River basin (largely late Mesoproterozoic in age; Plumb, 1981) and unconformably below Lower Cambrian sedimentary strata. To the northeast, the Uniya Tillite occupies a similar stratigraphic position along the western margin of the Daly River basin (Needham and Stuart-Smith, 1984). The other Kimberley glacial deposits are unconformable upon late Paleoproterozoic rocks of the Kimberley basin but are thought also to be younger than development of the Victoria River basin (see Plumb, 1981). Coats and Preiss (1980) and Plumb (1981a) report Rb-Sr ages generally in the range of 600 to 700 Ma, with large uncertainties, from Neoproterozoic shales throughout the region; I consider all these figures to be unreliable by present standards.

A summary of Australian correlations, subject to revision, is presented in figure 6. Except for the Egan Formation according to recent stratigraphic models (see above), all the glaciogenic deposits can be ascribed to either a Sturtian or a lower Marinoan glacial epoch. The Rook Tuff, in the Callana Group two unconformities below Sturtian deposits, has a U-Pb date of 802 ±10 Ma (Fanning and others, 1986). A slightly tighter constraint of 777 ±7 Ma comes from the Boucaut Volcanics, which may be part of the lowermost Burra Group, unconformably below the Sturtian succession (unpublished U-Pb zircon date by C.M. Fanning; W.V. Preiss, personal communication). Diverse Ediacaran fauna occur in strata several kilometers above the lower Marinoan glaciogenic succession. Several whole-rock Rb-Sr ages on shale, widely cited despite their suspect accuracy inherent to that geochronological method, include 750 ±53 Ma for the Tapley Hill Formation overlying the Sturtian deposits and 676 ±204 Ma on the Woomera shale overlying the Elatina-equivalent horizons (reported by Coats, 1981; and Preiss, 1993). In addition, the Bunyeroo shale, ~1 km higher than the lower Marinoan glaciogenic level but below Ediacaran-bearing strata, has yielded a whole-rock Rb-Sr age of 588 ±35 Ma (cited by Preiss, 1993). Two detrital zircons from the Marino Arkose, lying stratigraphically between the Sturtian and lower Marinoan glacial horizons, have U-Pb dates of ~650 Ma which could provide a maximum age for the Marinoan if no significant Pb-loss had occurred (Ireland and others, 1998). In sum, I accept the recent U-Pb zircon work as providing true constraints such that the Sturtian is younger than 777 ±7 Ma, the Marinoan is younger than ~650 Ma, and both are older than ~550 Ma, the age of diverse Ediacaran radiation (Grotzinger and others, 1995). Presumably, the Sturtian horizon is much closer in age to ~770 Ma than ~550 Ma (table 1).

The Neoproterozoic glaciogenic deposits from Australia have generally yielded low paleomagnetic latitudes. The paleomagnetic study of the Adelaide "geosyncline" by McWilliams and McElhinny (1980) provided many new paleopoles but unfortunately without convincing arguments for primary magnetizations. The Merinjina Tillite yielded three two-polarity directions standing out from the general scatter within the stereoplot; this behavior is reminiscent of the three Nama poles of Kröner and others (1980), which may not be useful as determined by Meert, Eide, and Torsvik (1997; see below). Nevertheless, the three components may in fact suggest discrete magnetizations, with MT2 and MT3 more likely post-folding rather than pre-folding. McWilliams and McElhinny (1980) interpreted those two components as remagnetizations during Tertiary and Ordovician time, and the latter interpretation is bolstered by the similarity of MT3 with the pole from the Cambrian-Ordovician boundary in Queensland (Ripperdan and Kirschvink, 1992). MT1 may be a primary remanence, but it is similar to the poles from the lower Marinoan Elatina Formation and the Precambrian-Cambrian boundary (see below), possibly indicating remagnetization of the Sturtian rocks. Rocks within the Marinoan interval, including the Mount Curtis Tillite, were also studied by McWilliams and McElhinny (1980) who found two antiparallel components, Y1 and Y2, both of which are demonstrably younger than Delamerian (Cambrian-Ordovician) folding and resemble Mesozoic-Cenozoic poles from Australia. Table 1 summarizes these data, showing ranges of paleolatitudes extrapolated from the sampled region in the northern Flinders Ranges to the entire Centralian Superbasin.

More reliable results from the Marinoan were obtained from the Elatina Formation in the western part of the Flinders Ranges. The pioneering study by Embleton and Williams (1986) showed consistently low-paleolatitude results from both outcrop and borehole. Subsequently, Sumner, Kirschvink, and Runnegar (1987, abstract only) performed a fold test on allegedly soft-sediment features (see Williams, Schmidt, and Embleton, 1995; Williams, 1996), indicating that the previously reported remanence was primary. This test was reproduced and published in full by Schmidt, Williams, and Embleton (1991). The preceding data were discounted by Meert and Van der Voo (1994), who suggested two possibilities why the Elatina paleomagnetic directions might not indicate depositional latitude: first, the hand-specimen-scale folds used for the fold test could result from the Delamerian orogeny (Cambrian-Ordovician in age), and thus the Elatina magnetization could be an overprint acquired as late as the Cambrian; or second, even if the direction was primary, the sampled Elatina rhythmites would not have averaged paleosecular variation of the Earth's geomagnetic field; thus the observed paleolatitude could be as much as 20° in error from the true paleolatitude, or even more if the signal was acquired during a geomagnetic excursion or reversal. The first interpretation was met with some contention (Williams, Schmidt, and Embleton, 1995), and subsequent studies have disproved the second option, confirming a long-lived interval of magnetic acquisition through two polarities of directions (Schmidt and Williams, 1995) that define several stratigraphically consistent polarity zones among pre-, syn-, and post-glacial strata (Sohl, Christie-Blick, and Kent, 1999). These later studies show that the Elatina Formation yields a robust and primary paleomagnetic pole, indicating a depositional paleolatitude of 2.7° ±3.7° (Schmidt and Williams, 1995) or 8.6° ±3.4° (Sohl, Christie-Blick, and Kent, 1999; using site means).

As for the northern deposits, if the Egan Formation is equivalent to the Julie Member of the Pertatataka Formation of the Amadeus basin (Corkeron and others, 1996; Grey and Corkeron, 1998) then the paleopole from the latter unit (Kirschvink, 1978b) determines a paleolatitude of 21° ±8° for the younger ice age recorded in the Kimberley area. If the Egan, or any of the Kimberley glaciogenic units, is correlative with the lower Marinoan deposits of the Adelaide geosyncline, then paleolatitudes of 12° to 16° (±3°) can be extrapolated to the Kimberley area from the latter region (Sohl, Christie-Blick, and Kent, 1999). The Walsh Tillite has recently yielded a pre-fold (pre-Cambrian) magnetic remanence indicating a paleolatitude of 45° +14/-11° (Li, 2000). That result is consistent with an earlier, unpublished study of overlying strata with "possibly primary" directions (McWilliams, 1977). Li (2000) inferred from this moderate paleolatitude that the Walsh Tillite could not be coeval with the low-latitude lower Marinoan glaciogenic deposits.

Lastly, Burek, Walter, and Wells (1979) describe stratabound magnetic polarity reversals within the Wonnadinna Dolomite, considered to be correlative with the lower Marinoan cap carbonates of the Centralian superbasin (Wells, 1981); the low paleomagnetic inclinations imply low paleolatitude, although the magnetic declination from these samples is distinct from those of the Elatina Formation in the Adelaide region. McWilliams (1977) obtained more "possibly primary" paleopoles: the Mount Birnie beds and the Cottons Breccia. All his results are generally similar to the Elatina and Pertatataka poles described above, supporting the correlations of those deposits into Marinoan (Elatina-equivalent) or Proterozoic-Cambrian boundary (Pertatataka-equivalent) stratigraphic intervals. Individual poles, however, are unpublished and not very reliable, so they are not listed in table 1.

Antarctica (45).–Possible Neoproterozoic glacial deposits have been described from the central Transantarctic Mountains, in the Goldie Formation of the Beardmore Group (Stump and others, 1988). The alleged glacial units are interbedded with mafic basalts dated at 762 ±90 Ma (Sm-Nd, Borg, Depaolo, and Smith, 1990; using the more conservative error estimates from Storey and others, 1992). To the contrary, Walker and Goodge (1994) found detrital zircons as young as 650 to 700 Ma from the Goldie Formation near the Beardmore Glacier, about 200 km to the southeast of the diamictites and the Sm-Nd sampling locality; furthermore, Goodge (1997) cited an unpublished detrital-zircon age as young as ~585 Ma from the Goldie Formation near the diamictite and Sm-Nd dated localities. Either deposition of the Beardmore Group occurred during a protracted episode of ~200 my, or at least one of the geochronological studies is incorrect. Until this issue is resolved, the diamictite horizons are herein considered to be poorly constrained in age. In any case, a definitive younger age limit is provided by the unconformably (or perhaps conformably; Goodge, 1997) overlying Lower Cambrian Shackleton Limestone (Stump and others, 1988).

No reliable paleomagnetic data exist for the deformed Neoproterozoic rocks of the central Transantarctic Mountains. Poor age control of the Goldie diamictites prohibits application of other poles from East Gondwanaland or from Laurentia via the postulated SWEAT connection (Moores, 1991).

Kalahari Craton

Southern Damara belt (46, 47).–Several discontinuous exposures of diamictites or associated cap carbonates occur within the autochthonous and parautochthonous foreland of the Damara orogen in central Namibia. In the Naukluft area south of Windhoek, an apparently older diamictite, the Blaubeker Formation, is autochthonous (Kröner, 1981); an apparently younger diamictite, part of the Blasskrans Formation, occurs in the Naukluft nappe complex. The latter unit contains the tuffaceous Tsubgaub Member, which has yielded an unpublished whole-rock Rb-Sr age of 620 ±55 Ma (Kröner, 1976; recalculated according to updated decay constants of Steiger and Jäger, 1977). That unit is directly overlain by a distinctive cap dolostone of the lower Tsabisis Formation, which is correlated with a similarly distinctive dolostone in the Gobabis region 600 km to the northeast, the Bildah Member of the Bushmansklippe Formation (Hoffmann, 1990; Kaufman, Knoll, and Narbonne, 1997). No diamictite or tuff is present directly beneath the Bushmansklippe Formation, but an older glaciogenic unit occurs in that region, called the Court diamictite (Martin, Porada, and Walliser, 1985) or the Blaubeker Formation *sensu lato* (Kröner, 1981). Given the questionable nature of the unpublished isotopic age for the Blasskrans Formation, I limit the age of that unit merely to a range older than Ediacaran (from the overlying Kuibis Formation) in table 2.

Within the allochthonous southern margin zone of the Damara orogen, the Naos Formation (formerly called the Chuos Pebbly Schist; Kröner, 1981) is composed mainly of pebble-bearing mica schist with minor metaconglomeratic and metavolcanic lenses (Hoffmann, 1983). Thin units of banded-iron-formation are also present. As a note of dissension, Martin, Porada, and Walliser (1985) found no compelling evidence for direct glacial association within these rocks, although they noted the possibility of remote (alpine) glacial influence and subsequent sedimentary reworking. Through regional stratigraphic synthesis, however, the Naos Formation has been correlated with the Blasskrans Formation and the presumed glacial interval prior to deposition of the Bildah cap carbonate (Hoffmann, 1989).

Gariep belt (48-50).–Resting on crystalline basement of the ~1.0-Ga Namaqua or 1.8-Ga Richtersveld provinces near the mouth of the Orange River (Kröner and Blignault, 1976), the Gariep Group is a mixed volcanic and sedimentary sequence containing two glacial horizons (Kröner, 1981). The lower, Kaigas Formation is overlain by rhyolites of the Rosh Pinah Formation with four single-zircon Pb-evaporation ages regressed to 741 ±6 Ma (Frimmel, Klötzli, and Siegfried, 1996). This minimum age is slightly more restrictive than 717 ±11 Ma on Gannakouriep dikes (Reid and others, 1991) which locally intrude the Kaigas Formation and adjacent strata. Farther to the south, the 780 ±10 Ma Lekkersing granite (Allsopp and others, 1979) may be intrusive to or nonconformably underlying the Gariep Group (Frimmel, Klötzli, and Siegfried, 1996). If part of the basement complex, then the Lekkersing intrusion provides a tight maximum age constraint for the Kaigas diamictite. Otherwise, ~1.0-Ga Namaqua

TABLE 2

Age constraints for paleomagnetically unstudied alleged Neoproterozoic glacial units

Paleocontinent Glacial deposit	Age	*Paleocontinent* Glacial deposit	Age
Laurentia and environs		*Kalahari*	
3. upper Tindir Gp. (Fe)	(Rapitan)	47. Blasskrans / Naos (Fe)	Pt3 (620±55?)
4. Deserters Range	C / 728+9/-7	48. Kaigas Fm.	741±6 / 780?
5. Mount Vreeland Fm.	E / ~750?	49. Numees (Fe)	E / 741±6
7. Idaho / Utah	(Rapitan)	50. Aties (Fe)	Pt3 (Numees?)
8. Kingston Peak (Fe)	C / ~1100		
13. Chiquerío Fm.	~440 / ~1000	*Congo - São Francisco*	
14. Gåseland / Charcot Land	Sil. / Pt1	52. Grand Conglomérat (Fe)	600? / 880-970
16. Polarisbreen Gp.	C / 939±8	53. Petit Conglomérat	600? / 880-970
17. W. Spitsbergen	E / Pt3	54. Tshibangu	739±7 / 962±2
18. Moræneso̸ Fm.	C / ~1230	55. Akwokwo / Bandja	620±10 / ~970?
19. Pearya	Ord. / Pt3	56. Sergipe diamictites	620±10 / ~1000
		58. Bebedouro / Rio Preto (±Fe)	(Jequitaí)
Baltica		59. Ibia / Cristalina (Fe)	(Jequitaí)
21. Vakkejokk	Lower C	60. Carandái	(Jequitaí)
22. Långmarkberg	(Vestertana)	61. W.Congo lower	600 / 920±10
25. North Urals	(South Urals)	62. W.Congo upper	600 / 920±10
26. Central Urals	(South Urals)		
		West Africa and environs	
Eastern Asia		68. Rokel River Gp.	533±7 / Ar3
29. Patom	Pt3	69. Kodjari Fm. (±Fe)	~620 / Pt1
30. Sayan rift	Pt3	70. Tamale Gp.	Pz / ~620
31. West Altaids (Fe)	C / Pt3	73. Tiddiline Gp.	~570 / 615±12
East Gondwanaland		*Avalonia - Cadomia*	
38. Penganga	Pt3?	76. Brioverian (non-glacial?)	C / 584±4
41. King Island / Tasmania	C / 760±12		
44. Landrigan (Fargoo?)	Pt3 (Sturtian?)	*Amazonia – Rio Plata*	
45. Goldie Fm.	C / 762±90	78. Camaquã Gp. (non-glacial?)	Cambrian

Abbreviations as in table 1, but also: Ar3 = Neo-Archean: Pz = Paleozoic.

gneisses constitute the maximum age constraint, although it is very likely that the diamictite is much closer in age to the Rosh Pinah Formation.

Higher in the sequence, the Numees Formation (see Jasper, Stanistreet, and Charlesworth, 1995) contains both schistose diamictite and iron-formation and underlies a cap carbonate thought to correlate with the post-Blasskrans carbonate (Hoffmann, 1989; Saylor, Grotzinger, and Hoffmann, 1995). The association with iron-formation is unusual but not unique among deposits commonly correlated with the "Varanger" or "Marinoan" episode of ~600 Ma (see below). Direct age constraints on the Numees Formation are lax, however; it is younger than the 741-Ma Rosh Pinah volcanic rocks (Frimmel, Klötzli, and Siegfried, 1996) and is overlain with local angular unconformity (Kröner and Germs, 1971) by Ediacaran-bearing quartzites of the Nama Group dated at 548.8 ±1 Ma (Grotzinger and others, 1995). An apparent age of 481 ±20 Ma (Rb-Sr on shale whole-rock) for the Numees Formation is interpreted to date a postdepositional silicification event (Allsopp and others, 1979).

In the Vanrhynsdorp basin midway between the Orange River and Cape Town, a tectonically imbricated Eocambrian sedimentary succession known as the Vanrhynsdorp Group is correlated with the Nama Group and slightly older sedimentary rocks in Namibia (Gresse and Germs, 1993). Near the base of the Vanrhynsdorp succession, the Aties Formation contains members of diamictite as well as lonestone-bearing, laminated iron-formation (Gresse, 1992). These associations suggest correlation with the diamictic and ferruginous Numees Formation in the Richtersveld, consistent with an upper Vanrhynsdorp-Nama equivalence (Gresse and Germs, 1993). In the absence of correlation, the Aties Formation is most likely Neoproterozoic in age, as its sedimentary succession nonconformably overlies 1.0 to1.1-Ga Namaqua gneisses (Gresse, 1992) and is deformed within the terminal Neoproterozoic Gariep-Saldania belt (Gresse and others, 1996).

Summary of correlations and paleomagnetic results.—For correlation of the glacial deposits on the periphery of the southern Damara and Gariep belts (Fig. 7), I follow Hoffmann (1989), who proposed two distinct glacial intervals. The older is represented by the Blaubeker (*s.l.*) and Kaigas Formations; the younger, commonly associated with distinctive "cap" carbonates or iron-formation, includes the pre-Bildah unconformity, Blasskrans, Naos, Numees, and Aties Formations. The first glacial interval is probably not much older than ~741 Ma, constrained in age by overlying Rosh Pinah volcanics and crosscutting Gannakouriep dikes (Frimmel, Klötzli, and Siegfried, 1996). Reliable paleomagnetic data for the pre-740-Ma glaciation are lacking; the Blaubeker Formation was studied directly but yielded a wide scatter of directions that loosely define an apparently post-folding component ("NBX") similar to Cambrian-Ordovician poles from Gondwanaland (Kröner and others, 1980). The second glacial episode lies unconformably below basal Nama Group sediments dated at 549 ±1 Ma (Grotzinger and others, 1995) and is perhaps significantly younger than the ~741 Ma Rosh Pinah volcanics. No paleomagnetic data exist for the younger Neoproterozoic glaciogenic deposits. Note that Saylor and others (1998; followed by Pelechaty, 1998) assign the Blaubeker and Blasskrans Formations as representatives of two distinct post-600-Ma ("Vendian") glaciations, although the absolute ages are derived from non-unique carbon-isotopic correlations and assumed regular sedimentation rates.

Precambrian-Cambrian glaciation? (51).—Glacial pavements have been described at several erosional levels within the Schwarzrand Subgroup of the Nama Group in southern Namibia (Kröner, 1981; Germs, 1995), which are tightly bracketed in age between 548.8 ±1 and 539.4 ±1 Ma (U-Pb on zircon from interbedded ash layers; Grotzinger and others, 1995). On the other hand, Saylor, Grotzinger, and Germs (1995) found no evidence for glacial erosion in more westerly outcrops. Direct paleomagnetic data from the Nama Group have been difficult to interpret. Strata of the Schwarzrand

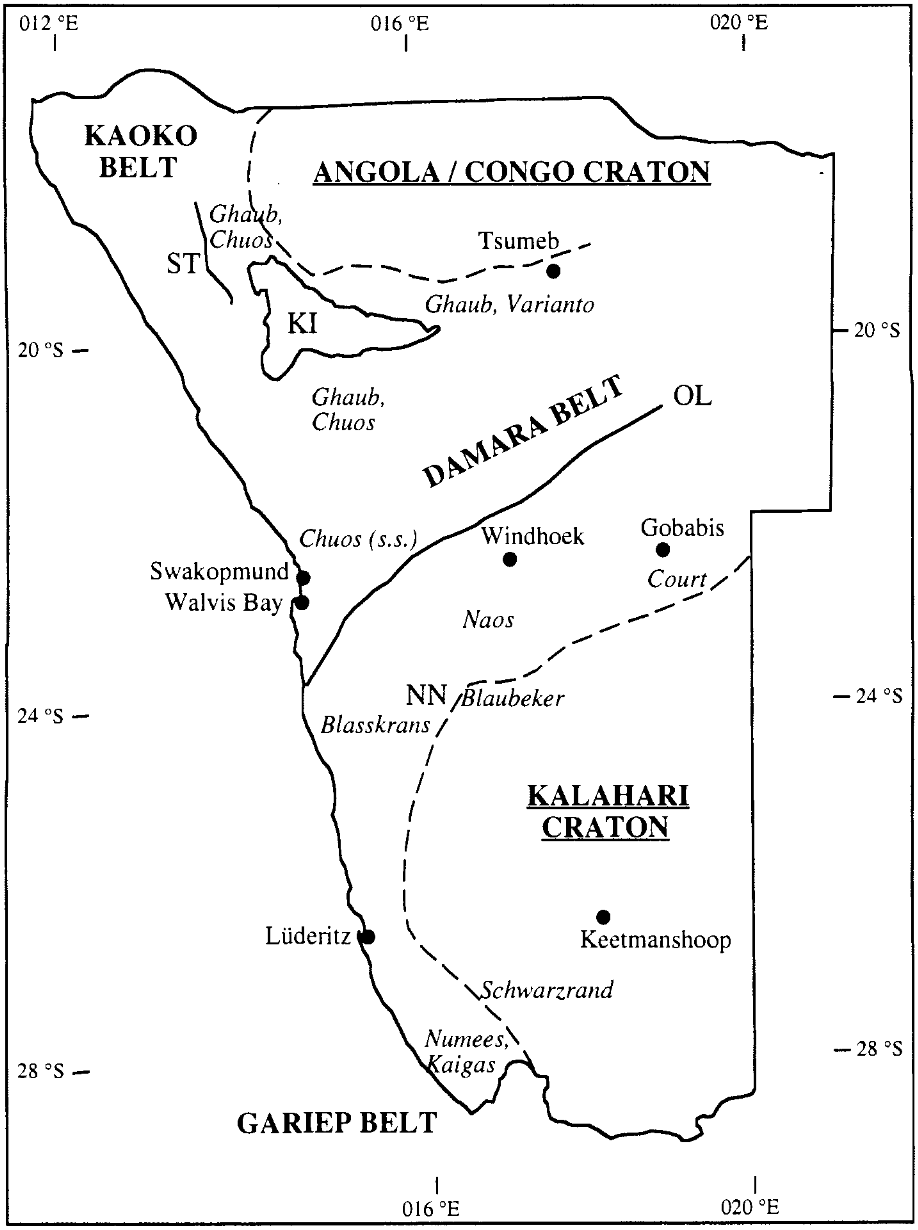


Fig. 7. Location map for Neoproterozoic-Cambrian glaciogenic formations (names italicized) in Namibia The Okahandja lineament (OL) separates the Central and Southern Zones of the Damara orogen. KI = Kamanjab basement inlier; NN = Naukluft nappes; ST = Sesfontein thrust. Dashed curves demarcate the approximate limits of Pan-African deformation related to the Kaoko, Damara, and Gariep belts.

Subgroup were first analyzed paleomagnetically by Kröner and others (1980), who observed three imprecise clusters of bipolar directions. Subsequently, Meert, Eide, and Torsvik (1997) identified no fewer than five groups of magnetizations, four of which are of dual polarity. In both studies, distinguishing among the clusters is commonly difficult because of the high dispersion among the data.

Because of the precise age constraints on the allegedly glaciogenic Schwarzrand erosional surfaces, a more reliable estimate of their paleolatitudes can be applied indirectly via paleomagnetic results from the Sinyai metadolerite intruding the Mozambique belt. Its $^{40}Ar/^{39}Ar$ age of 547 ±4 Ma and paleomagnetic direction (Meert and Van der Voo, 1996) imply a Schwarzrand depositional paleolatitude of 38 ±3°, corroborated by the Australian Precambrian-Cambrian boundary magnetostratigraphic study by Kirschvink (1978a,b). The Sinyai pole is preferred here because it can be applied without the need for reconstructing Gondwanaland to a specific pre-Mesozoic configuration.

Congo/São Francisco craton

Eastern Congo (52-54).–Neoproterozoic glaciogenic strata occur within extensive sedimentary basins as well as isolated occurrences along the eastern margin of the Congo craton (Fig. 8; Cahen, 1982). Straddling southeastern Zaire (Shaba) and Zambia, the Katangan sedimentary succession contains two readily correlatable levels of diamictite, called the Grand Conglomérat and Petit Conglomérat (Cahen and Lepersonne, 1981a) of the Kundelungu Supergroup. Although the lower horizon, the Grand Conglomérat, is more readily ascribable to direct glacial provenance than the upper, Petit Conglomérat, the latter is nearly everywhere capped by a pink dolostone similar to other cap carbonates of glacial deposits throughout the world. In some places, the Grand Conglomérat rests with slight angular unconformity upon older strata of the Roan Supergroup

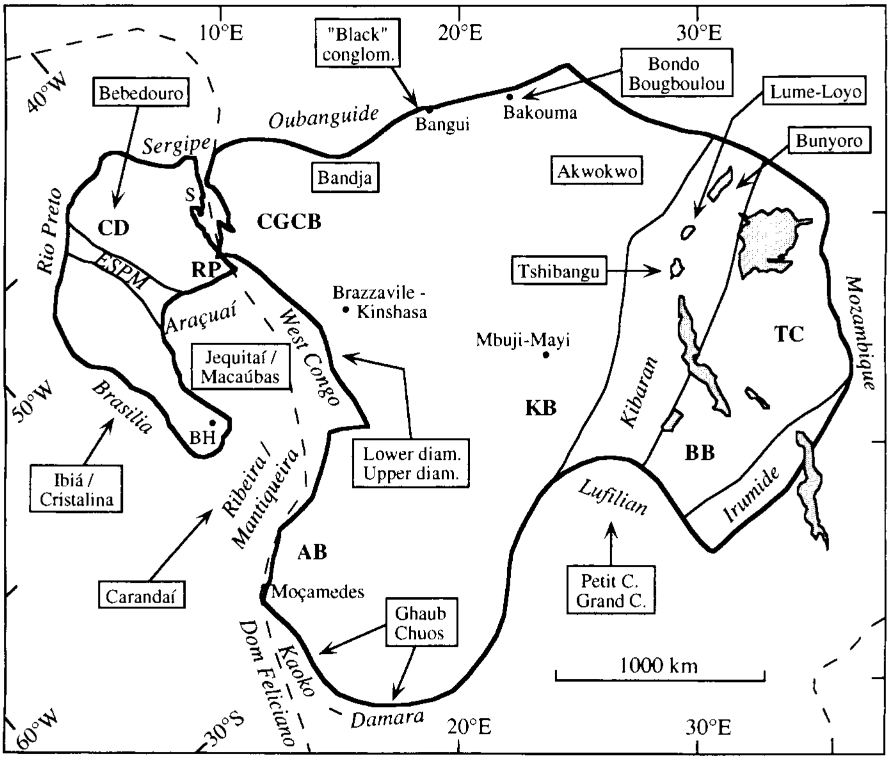


Fig. 8. Location map for alleged glaciogenic deposits of the Congo-São Francisco craton. South America is restored to according to Powell and Li (1994). Dashed curves are present continental coastlines. Geological and geographical regions: AB = Angola block; BB = Bangweulu block; CD = Chapada Diamantina; CGCB = Cameroon-Gabon-Congo block; ESPM = Espinhaço-Paramirim belt; LV = Lake Victoria; RP = Rio Pardo; TC = Tanzania craton. Cities: BH = Belo Horizonte; S = Salvador.

(Cahen and Lepersonne, 1981a); in other localities the transition appears conformable, including one area where Cailteux (1994) describes a gradational contact between the Grand Conglomérat and the underlying, hematite iron-formation bearing, Mwashya Formation. Equivalent units of the Grand and Petit Conglomérat are found within the Luapula Beds on the Bangweulu tectonic block to the northeast (Andersen and Unrug, 1984).

As noted by Unrug (1988), the Grand Conglomérat was probably deposited during an episode of lithospheric extension (rifting?). The diamictite entrains clasts of Roan sedimentary rocks as well as crystalline rocks similar to surrounding exposures of basement; some of these clasts appear to be derived from granites and pegmatites dated at 976 ±10 Ma (Rb-Sr on whole-rock and feldspar separates; Cahen and Ledent, 1979) or 962 ±2 and 968 +33/-29 Ma (U-Pb on columbite; Romer and Lehmann, 1995). Consistent with this is a somewhat poorly defined K-Ar age of 948 ±20 Ma–from a range of 953 ±20 to 870 ±20–on basalts in the uppermost Mbuji Mayi Supergroup, correlated with the Roan Supergroup (Cahen, 1982; Cahen and Snelling, 1984, p.155). These basalts may correlate also with the lower Kundelungu extensional event. A much younger age for the Kundelungu glaciogenic deposits is suggested by a recent U-Pb zircon age determination of 877 ±11 Ma for the Nchanga Granite, whose detritus can be identified as pebbles within Roan conglomerates (Armstrong and others, 1999, abstract only). Likewise, the Zambezi belt contains a metamorphosed supracrustal succession resembling the Roan Supergroup, with basal metarhyolites dated by U-Pb zircon at 879 ±19 Ma (R. Warslaw, ms., cited by Hanson, Wilson, and Munyanyiwa, 1994). A loose minimum age constraint for the Kundelungu Group is provided by post-Lufilian-folding U-mineralization dated at 602 ±20 Ma (cluster of concordant and nearly concordant U-Pb uraninite ages; Cahen and others, 1961; decay constants updated by Cahen and Snelling, 1984, p.132). Such a minimum age, however, would be incompatible with apparent early Paleozoic microfossils found within the upper-Kundelungu-correlative Luapula beds to the northeast (Vavrdova and Utting, 1972); further work should be directed toward this problem.

In eastern Kivu, Zaire, the basal diamictite unit of the Tshibangu Group (upper part of the Itombwe Supergroup; Cahen, Ledent, and Villeneuve, 1979; Villeneuve, 1987) contains clasts that appear to be derived from the ~960 to 970 Ma pegmatites (Cahen and Snelling, 1984, p.202; Romer and Lehmann, 1995). The diamictite is effected by a phase of deformation that is also recognized to the east, in Burundi, where the 739 ±7 Ma Ruvubu syenite belongs to a suite of late- or post-tectonic intrusions found throughout the region (U-Pb zircon age; Tack and others, 1984). Thus the Tshibangu diamictite appears to be older than ~740 Ma. If it is glaciogenic then it may correlate with the Grand Conglomérat as suggested by Cahen (1982); alternatively, it may correlate with the Petit Conglomérat or define a separate glacial episode altogether.

Northern Congo/São Francisco (55, 56).–Farther north around the craton's margin, possibly glaciogenic, lonestone-bearing metarhythmites are described from moderate- to high-grade zones of the central Mozambique Belt in Kenya (Mosley, 1993). An isolated exposure of Bunyoro diamictite in western Uganda (Byørlykke, 1981) has been correlated with the Lume and Loyo Group (Cahen and Snelling, 1984, p.246) and also the Akwokwo "Tillite" (Verbeek, 1970; Trompette, 1994, p.94) of the little-deformed Lindian Supergroup along the northern margin of the Congo craton (Cahen, 1982). These may be older than the Tshibangu diamictite, because the Bunyoro and Lume-Loyo successions are apparently affected by deformation coeval with the sub-Tshibangu unconformity (Cahen and Snelling, 1984, p.205). They may correlate with the Grand Conglomérat (Cahen, 1982), permitting the Tshibangu-Petit Conglomérat correlation suggested above.

From the Lindian Supergroup westward, the Akwokwo diamictite probably correlates with a diamictite at the base of the Bougboulou Group in the Fouroumbala basin (Cahen, 1982; Cahen and Snelling, 1984, p.246), the so-called "black conglomerate" in the Bangui basin (Cornacchia and Giorgi, 1986), and the Bandja "tillitic complex" in the Nola or Sembé-Ouesso region (Cahen, 1982; Cahen and Snelling, 1984, p.246; Poidevin, 1985). The Neoproterozoic succession containing these units can be traced continuously in the subsurface for hundreds of km in the western Cuvette Centrale of Zaire (Daly and others, 1992). For the last step to the West-Congo foldbelt, Cahen (1982), Cahen and Snelling (1984, p.175), Poidevin (1985, p.28), and Trompette (1994, p.220) equate the Bandja diamictite-volcanic association with the lower diamictite or both diamictites in the West-Congolese Supergroup (see #61-62 below). A higher, allegedly "fluvio-glacial" level, overlain by a pink carbonate rock resembling the "cap" of the Petit Conglomérat, also occurs in the Fouroumbala basin near Bakouma and is called the Bondo "series" or "complex" (Bonhomme and Weber, 1977; Cahen and Snelling, 1984, p.245; Poidevin, 1985; Alvarez, 1996).

Numerical ages for these units are not reliably constrained. A Rb-Sr shale isochron yielding an interpreted metamorphic age of 707 ±11 Ma for the Bakouma Group (including the Bondo "series"; Bonhomme and Weber, 1977, using $\lambda^{87}\mathrm{Rb} = 1.42 \times 10^{-11}$ yr^{-1}) could provide a minimum age for the glacial rocks, although the whole-rock Rb-Sr chronometer on shale is suspect without independent verification. In southern Cameroon, granulitic metamorphism is dated at 620 ±10 Ma (concordant U-Pb on zircon; Penaye and others, 1993), which I tentatively extend along the length of the Oubanguide-Sergipe orogen to provide a rough minimum age constraint for its diamictite deposits. A poor maximum age estimate for the Bandja complex is the determination of 971 ±28 Ma (K-Ar recalculated by Cahen and Snelling, 1984, p.177) on dolerite encountered in a borehole; it may be coeval with both the Bandja volcanics and numerous dikes cutting the pre-Bandja substrate.

On the Brazilian side of the dissected east-west trending Oubanguide-Sergipano orogen, possibly correlative diamictites occur within the Sergipano belt and the possibly consanguineous Rio Preto belt (Trompette, 1994, p.217 and 244). Depending on regional stratigraphic correlations within the Sergipano belt, sedimentary successions there may contain one (Trompette, 1994, p.217) or two (D'el-Rey Silva, 1999) diamictite-bearing levels, either of which may or may not be glaciogenic. Nonetheless, D'el-Rey Silva (1999) correlates his two levels with other allegedly glaciogenic deposits across the Congo/São Francisco craton. The Rio Preto region contains a transition of facies from the cratonic equivalents of the Bebedouro Formation and neighboring strata (see #58 below) to correlative basinal deposits. Assuming that the deformation along the northern margin of the São Francisco craton is coeval with that in Cameroon, the 620 ±10 Ma granulite age from the latter region (Penaye and others, 1993) can be used as a minimum age constraint for the allegedly glaciogenic rocks in northeast Brazil. Absence from the Sergipano sedimentary "cover" succession of an otherwise prominent 1.0 to1.1-Ga dike swarm within São Francisco crystalline rocks (Renne and others, 1990) suggests an earliest Neoproterozoic maximum age for the alleged glaciogenic units. In addition, D'el-Rey Silva (1999) reports unpublished U-Pb detrital-zircon ages as young as 810 Ma from his lower diamictite unit.

Recognizing the potential for stratigraphic revisions along the eastern and northern borders of the Congo/São Francisco craton in both central Africa and eastern Brazil, I tentatively follow Cahen (1982), Cahen and Snelling (1984, p.175 and 246), and Trompette (1994, p.94 and 220) in correlating the diamictites together. Specifically, an older glaciation may be represented by the Grand Conglomérat, Akwokwo, basal Bougboulou, "black conglomerate," and Bandja units. A younger glaciation is represented by only the Petit Conglomérat and the Bondo series. Whereas two diamictite levels can be distinguished along the eastern margin of the craton–Bunyoro and Lume/Loyo units affected by a phase of folding that apparently predates the Tshibangu Group–it is unclear how these correlate with the two hypothesized glaciogenic levels found elsewhere in central Africa. If the Tshibangu Group correlates with the Grand Conglomérat (Cahen, 1982), then at least three distinct glacial horizons would appear to exist on the Congo craton. For many of these units, however, a glacial association of the diamictites is based more upon regional correlation of Neoproterozoic stratigraphy than abundant diagnostic lithological features, and thus some of the alleged glaciogenic deposits may be reevaluated as non-glacial according to future study.

São Francisco craton and southern marginal foldbelts (57-60).–Glaciogenic rocks occur within and along all the margins of the São Francisco craton, within the São Francisco Supergroup (Karfunkel and Hoppe, 1988; Trompette, 1994, p.81). Around the Serra do Cabral inlier in the southeastern portion of the craton, the type Jequitaí Formation occurs unconformably above clastic rocks of the Paleo-Mesoproterozoic Espinhaço Supergroup; farther east, a similar unconformable relationship occurs at the base of the glaciogenic Macaúbas Group (Rocha-Campos and Hasui, 1981a; Uhlein, Trompette, and Alvarenga, 1999). The latter succession, commonly associated with iron-formation or manganese concentrations, constitutes an important stratigraphic marker within the Araçuai foldbelt (Trompette, 1994, p.141; Uhlein, Trompette, and Egydio-Silva, 1998) and is most likely correlative with the Jequitaí Formation. Elsewhere within less-deformed regions of the craton, in the Irecê mid-cratonic basin to the north, the Bebedouro Formation occupies a similar stratigraphic position, resting unconformably atop the Chapada Diamantina Group (Rocha-Campos and Hasui, 1981b), which comprises the upper part of the Espinhaço succession (Trompette, 1994, p.78). Farther east, in the Rio Pardo region, the molasse-like and diamictite-bearing Salobro Formation has been tentatively correlated with more convincingly glaciogenic rocks of the São Francisco Supergroup (Rocha-Campos and Hasui, 1981c; Trompette, 1994, p.147).

In the Brasília foldbelt on the western margin of the São Francisco craton, the Ibiá or Cristalina Formation (Rocha-Campos and Hasui, 1981d; Karfunkel and Hoppe, 1988; Trompette, 1994, p.178) occurs as discontinuous exposures among various levels of tectonic imbrication. These sedimentary rocks, deposited in more distal regions from the craton, may be glaciogenic themselves or secondarily derived from proximal glacial sediments. In the Ribeira belt immediately south of the São Francisco craton, the Carandaí Formation of the greenschist-grade São João del Rei Group (Rocha-Campos and Hasui, 1981e) has recently been re-interpreted as glaciogenic and correlative with the Macaúbas Group and Jequitaí Formation, based on associations of diamictites and varve-like rhythmites containing outsized clasts interpreted as dropstones (Karfunkel and Hoppe, 1988). The age of the Carandaí Formation, however, is somewhat uncertain (Trompette, 1994, p.161).

Following Karfunkel and Hoppe (1988), I tentatively correlate all the São Francisco craton-marginal diamictite-bearing units. Of course, it is possible that future stratigraphic work will reveal diachroneity of glacial units, as has been recently demonstrated in northern Namibia (see below). The numerical age of the São Francisco Supergroup was recently reviewed by Trompette (1994, p.86) and Fairchild and others (1996). U-Pb constraints from zircon and baddeleyite within dikes cutting the Espinhaço Supergroup but not the São Francisco Supergroup suggest a maximum age for the latter (including the glacial deposits) of 906 ±2 Ma (unpublished U-Pb on baddeleyite and zircon reported in abstract only; Machado and others, 1989). This contradicts the commonly quoted age of ~950 to1000 Ma for the Jequitaí Formation and correlative glacial rocks of the São Francisco craton (Bonhomme and others, 1982), but most of those earlier determinations are Rb-Sr analyses of argillites, demanding an interpretation of when isotopic closure took place. Indirect evidence from carbon isotopes provides some supporting evidence for a post-900-Ma age for the glacial rocks; very enriched $\delta^{13}C$ from conformably overlying carbonate strata of the Bambuí Group (Iyer and others, 1995) suggests correlation with post-Rapitan (younger than ~750 Ma; see above) carbonates in the Mackenzie Mountains of Canada (Kaufman and Knoll, 1995). Moderately enriched strontium isotopic values from a Bambuí-correlative carbonate unit in the Irecê basin support a late Neoproterozoic age, as well as the possibility that the underlying Bebedouro Formation could be similarly as young (Misi and Veizer, 1998). For a minimum age constraint, the São Francisco Supergroup is involved in Brasiliano orogenesis (Chemale, Alkmim, and Endo, 1993) dated at ~580 Ma from the Ribeira belt (Machado and others, 1996) and at ~600 Ma from the internides of the Brasília belt (Pimentel, Fuck, and Botelho, 1999).

West Congo/Angola (61, 62).–Thick Pan-African sedimentary successions occur within the West Congo foldbelt from Gabon to Angola. Two diamictite horizons are recognized in the northern part of this region (Cahen and Lepersonne, 1981b), whereas farther south there are two additional, albeit less prominent, diamictite levels (Schermerhorn, 1981). The lower diamictite, stratigraphically located at the base of the Haut Shiloango Group, is associated with mafic volcanic rocks and bounded by unconformities. The upper diamictite occurs at the base of the Schisto-Calcaire Group; it is succeeded by a pink or gray, thin but laterally persistent laminated dolostone reminiscent of many Neoproterozoic "cap" carbonates overlying glaciogenic strata. In this case, however, the thin dolostone appears at least locally to be disconformable upon the upper diamictite (Cahen and Lepersonne, 1981b). The amount of glacial contribution to these diamictites has been debated (for example, Vellutini and Vicat, 1983), but a current synthesis of all the data suggests at least a minor glacial input, if not true glaciomarine deposition of the upper diamictite (Trompette, 1994, p.93).

One stratigraphic interpretation groups the two diamictites into a single glacial episode, correlative with the basal São Francisco Supergroup across the Araçuai belt (Trompette, 1994, p.94). The prominence of two horizons, however, with the upper diamictite overlain by a cap carbonate, could also invite correlation with either the Kundelungu Group in Katanga or the Damara Supergroup in northern Namibia (*see* below).

In any case, no paleomagnetic data exist for the West Congolese succession. Possible ages for the two diamictite units range from a maximum of ~920 Ma on the unconformably underlying Mayumbian Supergroup (concordant SHRIMP U-Pb on zircon; Tack, Fernandez-Alonso, and Wingate, 1999) to a poor assortment of minimum ages dating the main phase of the West Congo orogeny, which affects the diamictite horizons: 733 Ma (discordant U-Pb upper intercept of the post-tectonic Noqui granite; Cahen and Snelling, 1984, p.168-169), ~730 to 740 Ma (Rb-Sr "resetting" ages of the Mativa, Yoyo, and other granites; Cahen and Snelling, 1984, p.169), 625 ±20 Ma (whole-rock and feldspar Rb-Sr resetting age on the Paleoproterozoic Vista Alegre pluton; Cahen, Kröner, and Ledent, 1979), or 604 ±58 Ma (lower intercept of strongly discordant U-Pb results from gneiss within the internal zone; Maurin and others, 1991). Trompette (1994, p.153) opts for dating the West Congo orogen at 600 to 620 Ma, primarily using constraints from its Brazilian counterpart, the Araçuai belt. Large specimens of *Obruchevella* filamentous cyanobacteria suggest an age near the Proterozoic-Cambrian boundary for the Schisto-calcaire Group, which conformably overlies the upper diamictite of the Haut Shiloango Group (Alvarez, Chauvel, and Van Viet-Lanoë, 1995).

Northern Damara and Kaoko belts (63, 64).–Neoproterozoic diamictites occur at several levels within the sedimentary prism flanking the southern margin of the Angola/Congo craton in northern Namibia (fig. 7). Proper stratigraphic relations have long been elusive because the most obvious horizon punctuating the carbonate foreland succession was correlated with and named after the Chuos Formation, a biotite-cordierite schist in the deformed Central Zone of the Damara orogen (Hedberg, 1979; Kröner, 1981). Recent mapping of the northern Damara and southern Kaoko belts has determined two widespread glaciogenic levels, the lower correlated with the Chuos Formation and the upper renamed the Ghaub Formation within the Otavi Group (Hoffmann and Prave, 1996). The Ghaub-equivalent horizon can be followed in semi-continuous outcrop around the southern and western margins of the Kamanjab inlier into the parautochthonous Kaoko belt (Hoffmann and Prave, 1996; Hoffman, Kaufman, and Halverson, 1998; Hoffman and others, 1998a). Farther west, across the Sesfontein thrust, the schistose diamictite formerly called "Chuos" but probably correlative with the Ghaub Formation (Hoffman and others, 1994), is associated with iron-formation (Henry, Osborne, and Schermold, 1993; Dingeldey and others, 1994); this ferruginous association is unique among Ghaub-equivalent strata in northern Namibia. Near the Kunene River, "Chuos" diamictites are well exposed, presumably occurring at a level similar to the Ghaub (Kröner, 1981).

Lower in the stratigraphy, diamictites of probable glacial origin occur in several regions of the northern Damara orogen, more commonly associated with iron-formation than the Ghaub-equivalent units (Hoffmann and Prave, 1996). In the type area of the Otavi Mountains, the Varianto Formation is a massive, ferruginous tilloid (Hedberg, 1979; Kröner, 1981; Hoffmann and Prave, 1996). Ferruginous carbonate lenses with local diamictites occur locally within the lowest sedimentary sequence along the parautochthonous southern margin of the Kamanjab inlier. The Fe-rich diamictite has also been recognized in the parautochthonous Kaoko belt to the northwest (Hoffmann and Prave, 1996; Hoffman, Kaufman, and Halverson, 1998; Hoffman and others, 1998a) and at a similar level at the base of the "Ugab" Group west of the Sesfontein thrust (Henry, Osborne, and Schermold, 1993). When considered in light of stratigraphic revision by Hoffmann and Prave (1996), this allochthonous lower diamictite might well be considered correlative of the Chuos Formation *s.s.* (Hoffman and others, 1994). The Chuos Formation of the parautochthon lies unconformably above the Ombombo Subgroup (Hoffmann and Prave, 1996), which contains an ash bed dated at 758.5 ±3.5 Ma (unpublished U-Pb zircon data illustrated in Hoffman, Kaufman, and Halverson, 1998; Hoffman and others, 1998a).

South of the Kamanjab inlier, equivalents of both the Chuos and Ghaub Formations (P.F. Hoffman, personal communication) are found stratigraphically above the Naauwpoort volcanics dated at 746 ±2 Ma and 747 ±2 Ma (Hoffman and others, 1996). Farther south, in the more metamorphosed Central Zone of the orogen, the Chuos Formation (*s.s.*) is widespread and commonly associated with sedimentary iron- and manganese-formation (Henry, Stanistreet, and Maiden, 1986; Badenhorst, 1988; Bühn and Stanistreet, 1993). A minimum age constraint for the Chuos Formation, as well as the higher stratigraphic level of the Ghaub, is provided by late- to post-tectonic Damaran granites dated at 589 ±40 and 546 ±30 Ma (discordant U-Pb zircon from the Omangambo and Otjozondjou plutons; Miller and Burger, 1983), 548 ±31 Ma (whole-rock Rb-Sr data from both plutons combined; Hawkesworth and others, 1983), and 580 ±30 Ma (discordant U-Pb zircon from the "Salem-type" granite at Goas in the central zone of the orogen; Allsopp and others, 1983). From these dates, I apply a conservative minimum age estimate of about 550 Ma for the glaciogenic deposits of the northern and central Damara belt. Note that Kaufman, Knoll, and Narbonne (1997), Hoffman, Kaufman, and Halverson (1998), and Hoffman and others (1998a,b) interpret both the Chuos and Ghaub levels as pre-700 Ma ("Sturtian"), whereas Kennedy and others (1998) interpret the Ghaub horizon to be substantially younger ("Varangian"). Finally, note also that Martin, Porada, and Walliser (1985) reported finding no compelling evidence for a glaciogenic origin of either the Chuos or the subsequently named Ghaub diamictites.

Paleomagnetic constraints.–Direct paleomagnetic measurements on sedimentary units probably correlative with the Varianto and Ghaub glacial deposits in the Kaoko belt (McWilliams and Kröner, 1981) yielded several magnetic components of uncertain ages. Fold tests were either negative or inconclusive so the following discussion refers only to *in situ* results. From the Nosib Group (Varianto equivalent) two stable magnetizations were obtained, one (NQ1) similar to some Early Cambrian poles from Gondwanaland (Meert and Van der Voo, 1997; Evans, Ripperdan, and Kirschvink, 1998), and the other (NQ2) similar to the Cambrian-Ordovician boundary magnetostratigraphic pole from Queensland, Australia (Ripperdan and others, 1992; restored to African coordinates according to Powell and Li, 1994). Ghaub-equivalent and adjacent strata also yielded two distinct components, one (DC1) lying midway between NQ1 and NQ2 with a large error oval enveloping Early Cambrian poles, and the other (DC2+3) most likely post-folding and resembling the present field direction in southern Africa. A subsequent study of the Ghaub Formation in the Otavi Mountains near Tsumeb found a consistent component similar to DC1 residing in clasts of the diamictite, further suggesting that DC1 is an overprint (D.A.D. Evans, unpublished data). Paleolatitudes from these results are computed in table 1 for a reference locality at 20°S, 015°E.

Carbonate rocks of the Bambuí Group are magnetically overprinted throughout the São Francisco craton (D'Agrella-Filho and others, 2000). The corresponding pole position is similar to Early-Middle Cambrian results from Gondwanaland, and the overprint probably is due to large-scale fluid migration associated with late stages of Brasiliano orogenesis. Table 1 includes the "BGC" pole as representative of several overprint directions reported from the Bambuí Group (D'Agrella-Filho and others, 2000).

Indirect constraints on Congo/São Francisco glacial paleolatitudes vary in applicability. Although presented to satisfy such a purpose, the dikes studied by D'Agrella-Filho and others (1990) are older than 1000 Ma and hence pre-date the glacial deposits by >100 my if the pre-glacial date of 906 ±2 Ma (Machado and others, 1989) is correct. From across the Congo craton, paleomagnetism of the post-Ubendian Mbozi complex (Meert, Van der Voo, and Ayub, 1995) may provide a paleolatitudinal constraint on the Chuos Formation. A cooling age of 755 ±25 Ma (K-Ar on biotite; Cahen and Snelling, 1966, p.64-65; recalculated using updated decay constants, Dalrymple, 1979) on syenites which intrude the paleomagnetically sampled mafic phases, indicates that the preserved magnetization may be of similar age to the Chuos glacial deposits. These paleomagnetic results, if primary and of applicable age, would indicate a 10 ±5° paleolatitude for the Chuos Formation (calculated for a reference locality at 20°S, 015°E).

West Africa and Environs

Taoudeni basin, northern region (65, 66).–An extensive Neoproterozoic to Paleozoic cover sequence is exceptionally preserved on the West African craton, mainly in the centrally located Taoudeni basin (fig. 9A). Within this succession are several levels of glaciogenic strata, including a well dated Upper Ordovician unit. Lower in the succession, a distinct glacial and post-glacial (barite-limestone)-chert-bearing sequence has been recognized as the so-called "triad," or Bthaat Ergil Group (Trompette, 1994, p.49-50) or Jbeliat Group (Deynoux and Trompette, 1981), extending with a fairly uniform lithostratigraphy across the Adrar of Mauritania and El Hank (Moussine-Pouchkine and Bertrand-Sarfati, 1997). Traditional correlations from the Adrar of Mauritania to elsewhere along the margins of the West African craton were based on the "triad," but as shown below such correlations may not be warranted in light of recent geochronological data. Several Rb-Sr ages on fine-grained sediments within the Taoudeni succession suggested that the glacial unit was deposited between ~695 and 595 ±45 Ma (Clauer and others, 1982), ages which I consider suspect until verified independently by other isotopic systems.

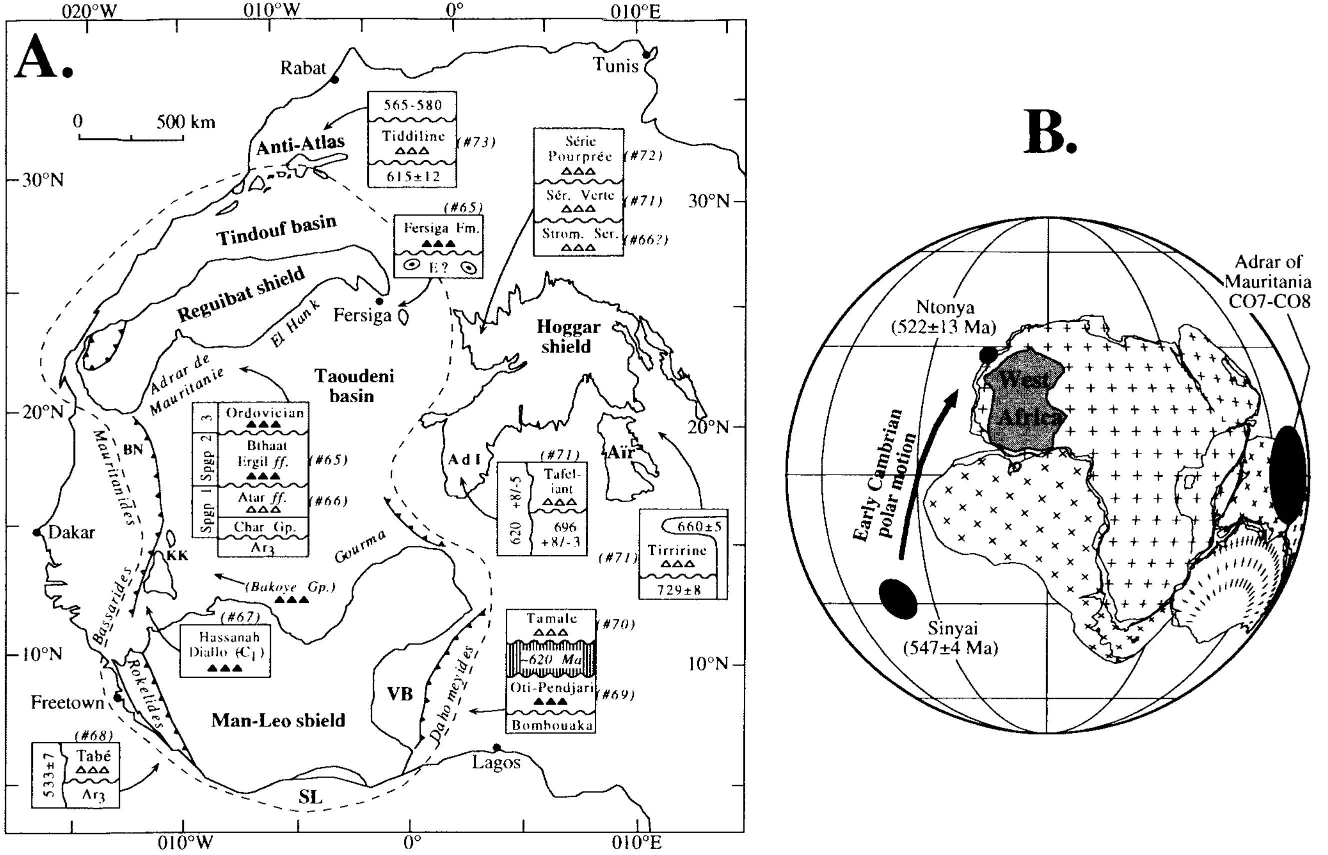


Fig. 9(A) Stratigraphic and spatial context of alleged glaciogenic deposits of western Africa. Symbols as in figure 5, plus: open triangles, less-convincing evidence for glaciogenesis; dashed curve, limits of the West African craton; barbed curves, foreland limits of Pan-African fold-thrust belts. Ages in Ma (see text for details). AdI = Adrar des Iforas; BN = Bou Naga tectonic window; KK = Kayes and Kenieba/Kedougou basement inliers; SL = approximate pre-Cretaceous position of the São Luis cratonic block, presently in northern Brazil; VB = Volta basin.

(B) Terminal Proterozoic to Early Cambrian paleomagnetic poles for Gondwanaland, showing rapid poleward motion of the West African craton. The Sinyai metadolerite pole is from Meert and Van der Voo (1996), the Ntonya ring structure pole is from Briden, McClelland, and Rex (1993), and the anomalous Adrar of Mauritania pole, shown only for comparison, is from Perrin, Elston, and Moussine-Pouchkine (1988). Gondwanaland reconstruction in African coordinates, from Powell and Li (1994). Orthographic projection, grid lines spaced 30° apart.

The "triad" or its conformably overlying clastic succession can be traced south from the Adrar of Mauritania along the foreland of the Mauritanide foldbelt. An important older age limit for the glaciogenic succession may be provided by granitoid ring complexes in the Bou Naga tectonic window of the northern part of the orogen, dated at 678 ±8, 686 ±5, and 687 ±8 Ma (Blanc and others, 1992). These igneous complexes nonconformably underlie a sedimentary succession that–although lacking diamictites–has been correlated with the Bthaat Ergil and higher sequence on the autochthon (Dallmeyer and Lécorché, 1990a). Successively farther south along the Mauritanide front, successions correlative to the "triad" are named the Tichilît el Beïda Group and its metamorphic equivalent the Djônâba Group, the Nagara and Bouly Groups, and the so-called "green" series (summarized by Le Page, 1986; Dallmeyer and Lécorché, 1989, 1990b; and Lécorché and others, 1991).

Farther to the northeast, a recent discovery of simple Ediacaran-like discoid fossils from strata unconformably underlying a "triad"-like succession of the Fersiga Group (Bertrand-Sarfati and others, 1995) has apparently provided a late Vendian or possibly Cambrian age for the entire Fersiga-Bthaat Ergil glaciogenic succession (Moussine-Pouchkine and Bertrand-Sarfati, 1997). Note that these fossils lack the complexity of the 550 to 545 Ma "late" Ediacaran forms (Grotzinger and others, 1995), and simple discoid fossils have also been found below "Vendian" glaciogenic deposits in the Mackenzie Mountains (Hofmann, Narbonne, and Aitken, 1990); in addition, chronometric support for a Cambrian "triad" relies on tenuous correlations between the West African craton and the Hoggar shield (Bertrand-Sarfati and others, 1995; see below). Nevertheless, an Early Cambrian age for the "triad" is consistent with fossil findings from the southwestern margin of West Africa, as discussed in the next section.

Older rocks from the Neoproterozoic cover of northern West Africa may also be glaciogenic. Trompette (1994, p.49) describes possible periglacial features from the base of the Atar Group of the Adrar of Mauritania, although Benan and Deynoux (1998) made no mention of these in their more detailed study. Microfossils from this horizon indicate a Late Riphean age, generally supported by Neoproterozoic Rb-Sr ages which, again, I consider unreliable until verified independently. Otherwise, the possible glaciogenic features at the base of the Atar Group are only constrained in age by the angular-unconformably overlying "triad" and the nonconformably underlying Paleoproterozoic rocks of the northeast Reguibat shield (Trompette, 1994, p.23). Along the Mauritanide and Bassaride foreland, the unconformity immediately beneath the "triad" has been related to ~650-Ma "Pan-African I" deformation (for example Dallmeyer and Villeneuve, 1987; see below); this age provides a reasonable younger limit for the Atar Group (table 1).

Direct paleomagnetic study of the Adrar of Mauritania spanned the glacial horizons (Perrin, Elston, and Moussine-Pouchkine, 1988; with references to and interpretations of earlier paleomagnetic studies). The authors interpreted three poles to be primary: (1) from the Char Group directly below the basal Atar Group disconformity, which may contain periglacial features as described above, (2) from a higher level within the Atar Group, and (3) from the redbeds conformably overlying the Bthaat Ergil glaciogenic strata. Samples from the Bthaat Ergil Group itself yielded a direction that Perrin, Elston, and Moussine-Pouchkine (1988) considered to be a late Paleozoic overprint (table 1). The Char Group samples generated a pole roughly antipolar to the late Paleozoic cluster identified by Perrin and Prevot (1988); despite this, both studies considered the Char pole to be primary. I consider the possibility that this pole, too, is a late Paleozoic overprint, although that hypothesis implies normal polarity in temporal proximity to the moderately well established Kiaman reversed superchron. Results from the middle of the Atar Group, although judged by Perrin, Elston, and Moussine-Pouchkine (1988) as probably primary, were likewise rejected as a late Paleozoic overprint by Perrin and

Prevot (1988). These interpretations leave unconstrained the depositional paleolatitude of alleged periglacial features at the base of the Atar Group.

The third paleopole from Adrar de Mauritanie derives from units that conformably overlie the glaciogenic Bthaat Ergil Group and conformably underlie sparsely fossiliferous Cambrian-Ordovician strata (Perrin, Elston, and Moussine-Pouchkine, 1988). This result carries two polarities and is unlike any other accepted Phanerozoic paleopole from Africa or Gondwanaland, both factors favoring a primary magnetization. Its anomalous position relative to other Cambrian paleopoles from Gondwanaland (see Meert and Van der Voo, 1997; Evans, Ripperdan, and Kirschvink, 1998), however, renders it suspect. Perhaps the anomalous pole position indicates a primary magnetization with a substantially older age for the Bthaat Ergil Group than Early Cambrian, but only further work will determine if this is so. I include this direct constraint from the Adrar of Mauritania in table 1, but other poles from the Vendian-Cambrian interval of the Gondwanaland APW path are also listed. Rapid motion of West Africa from moderate to high latitudes is suggested by the Gondwanaland APW path (Powell and others, 1993; Meert and Van der Voo, 1997; Kirschvink, Ripperdan, and Evans, 1998). Because the polar motion is so rapid, a Nemakit-Daldynian versus Atdabanian age of the "triad" can make the difference between deposition in moderate versus high latitudes (fig. 9B). Better constrained ages or more direct and reliable paleomagnetic studies of the West African glacial deposits are required to distinguish among these possibilities.

Taoudeni basin, southwest (67).–Within the Bassaride belt on the Guinea-Senegal border, another "triad" has recently been defined formally within the Mali Group as the glaciogenic Hassanah Diallo Formation and the post-glacial Nandoumari Formation (Culver and Hunt, 1991). The upper unit contains *Aldanella attleborensis* and echinoderm-like fossils suggesting an Early Cambrian age (Culver, Pojeta, and Repetski, 1998; Culver and others, 1988). Note that those authors' preferred age of Atdabanian relied on now-obsolete numerical timescales of the Early Cambrian (see Grotzinger and others, 1995). In addition, a stratigraphical debate arose regarding the ages implied by these fossils; if one assumed that the glacial rocks at the base of the Mali Group were as old as ~600 Ma (Clauer and others, 1982) and deformed by 540 to 560-Ma tectonism (Dallmeyer and Villeneuve, 1987), this required a significant disconformity separating the tillite from its post-glacial "cap" sequence. Such a sequence boundary was indeed described at the base of the Nandoumari Formation (Culver and Hunt, 1991). However, because the entire "triad" is involved in that episode of folding (see Dallmeyer and Lécorché, 1990), then any deformation of the Mali Group *must* be younger than Early Cambrian, in light of the fossils it contains (noted by Trompette, 1996, 1997). In that case, the glacial unit may also be of Early Cambrian age, obviating the need for a major unconformity separating it from its post-glacial sequence. This same conclusion was attained on different grounds by Bertrand-Sarfati and others (1995) as discussed above. Nonetheless, this model requires that the muscovite $^{40}Ar/^{39}Ar$ total gas age of 538 ±1 Ma from the interior of the Bassaride belt (Dallmeyer and Villeneuve, 1987) is perhaps representative of a stage of deformation that pre-dated Mali Group folding in the foreland or is geologically insignificant. In any event, the Mali Group unconformably overlies the deformed Guinguan Group in the Bassaride external thrust belt (Villeneuve, 1982); a mean of 661 ±6 Ma from three $^{40}Ar/^{39}Ar$ plateau dates on metamorphic muscovites of the Guinguan Group thus provides a maximum age constraint for the "triad" in the southwestern Taoudeni basin (Dallmeyer and Villeneuve, 1987).

The Bakoye Group in southwestern Mali contains diamictites as well as units of banded hematite-chert; atop this succession is a cap carbonate succeeded by bedded chert and siltstone of the Nioro Group (Deynoux, Kocurek, and Proust, 1989; Proust and Deynoux, 1994). Similarity between this "triad"-like sequence and those of the northwest Taoudeni basin suggests correlation (Villeneuve, 1988).

Rokelides (68).–Occurring as a lone outlier of sedimentary-volcanic cover within the southwest margin of the West African craton, the Rokel River Group contains an alleged glaciogenic diamictite (the Tibai Member of the Tabe Formation) at its base (Tucker and Reid, 1981). The evidence marshalled in favor of a glaciogenic origin is not overwhelming, consisting of surface textures on quartz grains from diamictites and centimeter diameter outsized clasts in laminated mudstone (Culver, Williams, and Bull, 1980). The Tabe Formation is nonconformable atop Neo-Archean ("Liberian") crystalline rocks and deformed within the Rokelide orogen, which in turn is intruded by the post-tectonic Coyah Granite dated at 533 ±7 Ma (Rb-Sr whole-rock and mineral isochron, Dallmeyer, Caen-Vachette, and Villeneuve, 1987). Although the Rokel River Group is likely to be Neoproterozoic, the enormous uncertainty in age hinders any assessment of paleolatitudes for the Tibai Member, whether glaciogenic or not.

Volta basin and Dahomeyides (69, 70).–On the southeastern margin of the West African craton, the Oti or Pendjari Group contains a basal succession (Kodjari Formation) of diamictite, barite-bearing limestone, and chert, which Trompette (1981) compared to the "triad" of the Taoudeni basin. Correlatives of the Kodjari succession, including glacial deposits and associated ore-grade iron-formation, occur in the Buem nappe and also possibly the Atacora nappe of the adjacent Dahomeyide belt (Trompette, 1994, p.120). Schistosity in the Atacora unit was recently dated at 634 to 607 Ma (^{40}Ar/^{39}Ar on muscovites of the northern region; Attoh, Dallmeyer, and Affaton, 1997), providing a minimum age for the Oti-Pendjari equivalents that is supported by identification of *Chuaria* sp. in post-glacial strata (Amard and Affaton, 1984; although a Cambrian age is also compatible; Amard, 1997).

Periglacial and/or glacial deposits are also cited for basal strata in the unconformably overlying Tamale or Obosum Groups, considered as a molassic succession to the Dahomeyide deformation, of probably Neoproterozoic-Cambrian but possibly middle Paleozoic age (Trompette, 1981; 1994, p.57; Villeneuve and Cornée, 1994). In the internal zone of the Dahomeyide orogen, the Nigerian "schist belts" contain several metadiamictites that may be glaciogenic (Turner, 1983); alternatively, these deposits may have merely a debris-flow origin associated with local syndepositional tectonism (Fitches and others, 1985; Trompette, 1994, p.263). Farther south within the internal zone and across the Atlantic into northeastern Brazil, the Ubajara Group of the Sobral region contains similarly poorly understood diamictite levels; existing Rb-Sr ages on cross-cutting plutons are probably unreliable (Trompette, 1994, p.257-259).

Hoggar shield (71, 72).–Traditional stratigraphic summaries of the Pan-African sedimentary succession within the Pharuside belt along the eastern margin of the West African craton have included a deformed Série Verte unconformably overlain by a molassic Série Pourprée (summarized by Caby, 1987). Five diamictite horizons occur within these successions, one in the Série Verte or correlative successions (Caby and Fabre, 1981a) and the others within the Série Pourprée (Caby and Fabre, 1981b). The Série Verte appears to have been deposited in a magmatic-arc setting (Trompette, 1994, p.124), and its diamictites may or may not be glaciogenic (Caby and Fabre, 1981a). Known as the Tafeliant Series in the central-southern Iforas, its age is bracketed between unconformably underlying quartz diorites at 696 +8/-3 Ma (concordant U-Pb on zircon, Caby and Andreopoulos-Renaud, 1985) and syn- to late-metamorphic diorite at 620 +8/-5 Ma (near-concordant U-Pb on zircon, Caby and Andreopoulos-Renaud, 1989). A possibly–but not necessarily–correlative succession in the eastern Hoggar, the Tirririne Formation (known as the Série du Proche Ténéré in northern Niger), rests nonconformably upon a 729 ±8 Ma granite (Caby and Andreopoulos-Renaud, 1987) and is intruded by sills dated at 660 ±5 Ma (Bertrand and others, 1978; both constraints from U-Pb on zircons). The Tirririne succession contains distinctive conglomerates of disputed glacial origin (Caby and Fabre, 1981a; Boullier, 1991; Trompette, 1994, p.253).

The Série Pourprée is composed of two stratigraphic entities, the lower Tagengan't and the upper Ouellen-in Semmen units (Deynoux and others, 1978). A unit variably correlated with the "triad" of the Taoudeni basin occurs near the top of the Tagengan't series but as discussed by Deynoux and others (1978) several options for correlation are possible. Specifically, the dilemma arises because the "triad" occurs within molassic deposits in the Hoggar shield, whereas it occurs below "molassic" deposits along the Adrar of Mauritania. The numerical age of the Série Pourprée is shrouded in a fog of unpublished results: it nonconformably overlies late- or post-tectonic intrusions dated at 620 to 580 Ma (Rb-Sr and U-Pb using outdated decay constants, no isotopic data presented, Allègre and Caby, 1972), 560 ±10 Ma (1972, cited by Caby and Fabre, 1981b; allegedly using updated decay constants of Steiger and Jäger, 1977), or 556 ±12 Ma (U-Pb, unpublished data of J.R. Paquette, cited by Bertrand-Sarfati and others, 1995); and the Série Pourprée is directly overlain by rhyolites dated at 538 ±30 Ma (whole-rock Rb-Sr, no isotopic data presented, Caby, 1967; recalculated here using updated decay constants) or 519 ±11 Ma (cited by Allègre and Caby, 1972; also recalculated here).

Note that in western Hoggar, unconformably below the Série Verte is a succession called the "stromatolitic series," which may be correlative with the Atar Group in Mauritania (Trompette, 1994, p.123). Like the Atar Group, the stromatolitic series begins with a basal conglomerate with possible glaciogenic features, such as striated and faceted clasts. These correlations between the West African craton and the Pharusian belt, however, may not be warranted as they are separated by a younger, Pan-African suture (Black and others, 1979).

A paleomagnetic study of the Adma diorite/adamellite intrusion, a post-Série Verte pluton dated at 620 +8/-5 Ma (U-Pb zircon, Caby and Andreopoulos-Renaud, 1989), yielded a steep direction indicating a high latitude of magnetic remanence acquisition (Morel, 1981). No field stability tests were performed, however, and the Adma pole is similar to the Early-Middle Cambrian cluster of Gondwanaland poles (Powell and others, 1993; Meert and Van der Voo, 1997; Kirschvink, Ripperdan, and Evans, 1998).

Anti-Atlas Mountains (73).–Although also on the West African craton and thus relevant to the preceding discussion, the Tiddiline Group in the Anti-Atlas Mountains of Morocco contains a diamictite assemblage (Leblanc, 1981) which is lithologically distinct from the "triad" of the Taoudeni basin and thus is discussed separately. Lacking the overlying carbonate and bedded chert of the "triad," the diamictites may only have a minor, if any, glacial contribution in an otherwise active tectonic setting (Leblanc, 1981; Hefferan, Karson, and Saquaque, 1992). The Tiddiline tilloids contain clasts strongly resembling the Bleida granodiorite dated at 615 ±12 Ma (U-Pb, Ducrot, 1979). The Tiddiline succession, paleomagnetically unstudied, unconformably underlies volcanic rocks of the Ouarzazate Formation dated at 578 ±15 and 563 ±20 Ma (Juery, unpublished U-Pb data cited by Odin and others, 1983). A reconnaissance study of the Sarhro Group in the Jebel Sirwa area identified diamictite horizons that are intruded by an allegedly ~699-Ma granite (Gresse and others, 1998, abstract only), which would make it demonstrably older than the Tiddiline tilloids; further work is required to verify a glaciogenic origin for the Sarhro diamictites.

Summary of correlations.–As shown above, direct age constraints for the Neoproterozoic-Cambrian glacial deposits of cratonic West Africa are scarce or one-sided, or well constrained only in active orogenic regions where firm evidence for continental glaciation is lacking (fig. 9A). Combination of unpublished isotopic results from the Hoggar shield with fossils from the Taoudeni basin suggests a Cambrian age for the "triad" and its equivalents within the northern and western areas of the West African craton (Bertrand-Sarfati and others, 1995). This correlation depends on the rejection of several Rb-Sr ages on fine-grained sedimentary rocks from the latter regions. Correlation of these units with the "triad" of the Oti-Pendjari Group is more problematic, as the Kodjari deposits must predate ~620-Ma Dahomeyide deformation. Based on the distinctive lithologic character of the "triad" in the western and northern parts of the craton, as well as the paleontological data, I tentatively extend the correlation by Villeneuve and Cornée (1994) to equate the Bthaat Ergil-Jbeliat, Fersiga, Bakoye, Mali Groups, all equivalent to the Série Pourprée (see below) with an Early Cambrian age (Bertrand-Sarfati and others, 1995). The "triad" of the Kodjari Formation, however, is considered older as required by the isotopic and (limited) paleontologic data. In its place, the basal Tamale Group periglacial features may correlate with the major glaciation elsewhere in West Africa. Thus the "triad" succession and also the "molassic" deposits in West Africa as grouped by Trompette (1994) are probably diachronous, representing different intervals of glacial and post-orogenic sedimentation.

If this correlation is correct, then two older glaciogenic units exist within the West African cover sequences: the base of the Atar Group in the Adrar of Mauritania and the base of the Oti Group in the Volta basin. These may be coeval, but in the absence of reliable data I list them separately in tables 1 and 2. Note that if the "triad" indeed records a synchronous glacial-postglacial transgression across the West African craton, including the Dahomeyides and Volta basin, then Dahomeyide folding of the Oti Group must be Cambrian, in contrast to the $^{40}Ar/^{39}Ar$ data from the Atacora nappe (Trompette, 1996). Further study should be directed toward this problem.

Avalonia-Cadomia

Several alleged Neoproterozoic glacial deposits occur within the exotic terranes of the eastern Appalachian orogen, within the paleocontinental fragment Avalonia (Nance, 1990). Terranes of similar tectonic affinity occur within Europe comprising the Cadomian belt (Nance and Thompson, 1996). The most notable Neoproterozoic glacial occurrences are found in eastern Massachusetts, eastern Newfoundland, Canada, and Normandy-Brittany, France. Described individually below, these units were considered as a group to record marine-debris flows within the active tectonic setting of the Avalonian-Cadomian orogen, with only an indirect, "upstream" glacial influence (Eyles, 1990). Similar features characterize various conglomeratic units in the Bohemian massif (Fiala, 1981) and boreholes from southern Poland (Brochwicz-Lewinski, 1981). Such a distinction between localized alpine glaciers and continental ice sheets is important for assessing the broad issues of Neoproterozoic paleoclimate; this topic is addressed following the detailed stratigraphic reviews.

Boston Basin (74).–The Squantum 'Tillite' member of the Roxbury Conglomerate (Rehmer, 1981) is a glaciogenic unit within the Boston Bay Group of eastern Massachusetts. Based on associations of sedimentary facies throughout the Roxbury Conglomerate, the formation is interpreted to have been deposited as a succession of marine debris flows with significant glacial influence (Smith and Socci, 1990). The Mattapan volcanic rocks yield U-Pb ages of 602 ±3 Ma (Kaye and Zartman, 1980) and 596 ±2 Ma (Thompson and others, 1996); clasts of these tuffs occur in the Squantum member and thus provide a maximum age for the glaciation. The Roxbury Conglomerate is overlain by purple mudstones of the Cambridge Argillite, for which a latest Precambrian age is suggested by microfossil assemblages, including *Bavlinella faveolata* (Lenk and others, 1982). The Squantum member is cleaved to much greater extent than nearby Cambrian formations, thus Rast and Skehan (1990) hypothesized a latest Proterozoic orogenic event to have occurred between deposition of the two successions. I accept this argument as providing a minimum Early Cambrian age limit for the Roxbury Formation (table 1).

A direct paleomagnetic study of the Roxbury Formation has generated a possibly primary paleopole indicating a 55° +8/-7° paleolatitude for the glaciogenic Squantum member (Wu, Van der Voo, and Johnson, 1986). The characteristic magnetic direction found by that study is pre-Alleghanian and could be primary if the conglomerate test upon clasts of Mattapan volcanics is reliable. Unfortunately, the authors admit that any chemical overprints in the Roxbury Formation could affect matrix material without remagnetizing the more impermeable Mattapan clasts; this renders their conglomerate test slightly suspect. Because of the possibility of local vertical-axis rotations, direct paleopole comparisons cannot be drawn between the Roxbury result (Wu, Van der Voo, and Johnson, 1986) and paleomagnetic results from other Avalonian terranes; however, it is noted that Avalonia occupied similar moderate-high latitudes in Llanvirn time (Mac Niocaill, van der Pluijm, and Van der Voo, 1997), and the Roxbury characteristic magnetization could therefore be an overprint of that age.

Eastern Newfoundland (75).–The glaciogenic Gaskiers Formation (Anderson and King, 1981; Myrow, 1995) occurs within a terminal-Proterozoic to Cambrian siliciclastic succession in the Avalon zone of eastern Newfoundland (Conway Morris, 1989). Eyles (1990) described the glaciogenic units in sedimentological detail and concluded that they are composed of slump deposits from an unstable volcanic-arc setting, whose glacial influence was restricted to higher elevations. Nonetheless, Myrow and Kaufman (1999) described a δ^{13}C-depleted "cap" carbonate within the succession, supporting the notion of deposition within an extensive ice age. The Gaskiers Formation is well dated between the nearly immediately underlying Harbour Main Group volcanics dated at 606 +3.7/-2.9 Ma (U-Pb on zircon; Krogh and others, 1988) and the more distantly overlying Mistaken Point Formation containing a low-diversity Ediacaran faunal horizon directly dated at 565 ±3 Ma (U-Pb on zircon with no isotopic data given, Benus, 1988).

Direct paleomagnetic study of the Gaskiers Formation has yielded disappointing results. Gravenor, Stupavsky, and Symons (1982) found a direction that failed a conglomerate test, and in a separate study the formation failed a fold test with a probably Devonian overprint (D. Morgan, cited in Myrow, 1995). Indirectly, the Marystown Group–at 608 +20/-7 Ma (Krogh and others, 1988) equivalent in age to the Harbour Main Group–may have been deposited at 34° ±6° (Irving and Strong, 1985) or 31° +10/-8° paleolatitude (McNamara and others, 1997). The earlier of these results passed a pre-Carboniferous inclination-only fold test but yielded a wide range of declinations presumably due to strike-slip-related vertical-axis rotation among sampling sites (Irving and Strong, 1985). The later study (McNamara and others, 1997) is published in abstract only, and it is tentatively assigned 'Q' ratings in table 1. Although these studies agree upon paleolatitude within uncertainty, another preliminary study found a range of virtual geomagnetic paleolatitudes of 15° ±8° to 32° ±8° from a limited number of volcanic flow-units within the Marystown and Harbour Main successions (Hodych, 1991).

Normandy and Brittany, France (76).–The diamictite-bearing Granville Formation (Doré, 1981) forms part of the upper Brioverian series of the Cadomian orogen in Normandy (Dupret and others, 1990). The upper Brioverian succession is not affected by the otherwise-extensive contact aureole of the adjacent Coutances diorite dated at 584 ±4 Ma, which thus provides a maximum age limit for the Granville diamictites (Guerrot and Peucat, 1990). The upper Brioverian is unconformably overlain by Lower Cambrian sediments (Rabu and others, 1990). Sedimentological studies by Doré, Dupret, and Le Gall (1985) and Eyles (1990) refuted the existence of any glacial influence for the Granville Formation. Paleolatitudinal estimates of the Cadomian belt vary with age: ~0°-10° at about 570 Ma, and ~30°-40° at about 550 Ma (Cadomian paleomagnetic data reviewed by Taylor, 1990). The paleomagnetic results are principally derived from Cadomian terranes in western Brittany or the Channel Islands, and thus an additional uncertainty regarding application of the paleomagnetic data to the Granville Formation is the timing of terrane juxtaposition, perhaps as late as about 540 Ma (Strachan and others, 1989). For these reasons, I do not include these paleomagnetic data in table 1.

Amazonia–Rio Plata

Southern Amazon region (77).–Neoproterozoic cover successions on the Amazonian craton occur mainly within a small area along its southernmost margin, along the arcuate

Paraguay belt (de Alvarenga and Trompette, 1992). Glaciogenic strata can be correlated throughout this belt, primarily in southwestern Brazil (Rocha-Campos and Hasui, 1981f; Rocha-Campos, 1981), and into the Tucavaca-Chiquitos region in Bolivia (Trompette, 1994, p.75). The Brazilian glaciogenic strata occur within the Jangada Group (in northern outcrops), Jacadigo Group (central outcrops), or Puga Formation (southern outcrops), whereas the Bolivian glacial rocks are included within the Boqui Group. The Brazilian examples exist through a range of metamorphic grades from the unmetamorphosed foreland (northwest) to the more internal zone of the orogen (southeast). The Jacadigo Group is associated with sedimentary iron- and manganese-formation, including the banded manganese ores of Urucum (Urban, Sribrny, and Lippolt, 1992). The glacial deposits are conformably overlain by the Araras Formation, a carbonate-rich succession containing fauna of possible Ediacaran affinity (summarized by O' Connor and Walde, 1985). Trompette (1994, p.75) and Trompette, de Alvarenga, and Walde (1998) cite an unpublished age of 623 ±15 Ma on quartz porphyry associated with volcanic rocks lying stratigraphically below the Boqui Group. If the age and correlations proved to be reliable and if the Jangada-Puga-Jacadigo-Boqui glacial deposits are coeval, then they would thus be relatively well constrained in age.

The Jacadigo Group was one of the first targets of paleomagnetic study in the southern hemisphere (Creer, 1965), at a time when the Urucum deposits were thought to be Silurian. This early result is of zero reliability on the modern Q-scale, but it is interesting in that it produces a paleopole near those of Early-Middle Cambrian and Early Ordovician age from other Gondwanaland continents. The Silurian age misassignment of the Urucum deposits, along with this paleopole, actually formed the basis of an alleged Silurian magnetic overprint affecting a broad range of Gondwanaland rocks cited by Perrin and Prevot (1988).

Rio de la Plata craton (78).–Exposed crystalline basement in Uruguay and eastern Paraguay denotes the Rio Plata craton, which may underlie a large portion of the Paraná basin (Ramos, 1988). During Neoproterozoic time, it may have been distinct from other cratons in Gondwanaland (Hoffman, 1991) or contiguous with Amazonia (Trompette, 1997). Within the Camaquã late- to post-tectonic basin in southernmost Brazil (Gresse and others, 1996), a glaciogenic origin has been interpreted for some of the deposits (Eeroli, 1995); however, this interpretation has been disputed (P. Paim, personal communication). The postulated glaciogenic units occur within the Camaquã Group (*sensu* Gresse and others, 1996), whose ichnofauna suggest a Cambrian or Ordovician age. Paleomagnetism of the Camaquã basin revealed a wide array of magnetic components, of unknown acquisitional ages, from units lying with slight angular unconformity below the postulated glaciogenic deposits (D'Agrella-Filho and Pacca, 1988). Lack of a precise age for the Camaquã Group prohibits application of Vendian-Cambrian paleopoles from the other Gondwanaland continents.

Farther south, in the Tandilia region of eastern Argentina, the Balcarce Formation contains a basal diamictite that is considered to be possibly glaciogenic (Dalla Salda, Bossi, and Cingolani, 1988). Although de Alvarenga and Trompette (1992) mention the possibility that the formation is latest Proterozoic in age, it is more likely to be early Paleozoic because of its purported *Cruziana* and other trace fossils (Dalla Salda, Bossi, and Cingolani, 1988).

IMPLICATIONS FOR PALEOCLIMATIC MODELS

Spatial Distribution of Neoproterozoic Glacial Deposits

Paleomagnetic data bearing on the Neoproterozoic glaciations vary widely in quality and should be culled before any conclusions are generated. First, a caveat: any selection procedure of paleomagnetic results is likely to contain some amount of subjectivity, and the reader may disagree with my conclusions. I use Van der Voo's reliability scale (1990) in table 1, because of its easy-to-use checklist format. The scale identifies seven criteria useful for assessing a reliable paleomagnetic study and produces "Q", the number of satisfied criteria. Whereas the Q scale is designed for constructing continuous apparent polar wander (APW) paths for continents or cratons, the purposes here are purely for estimating paleolatitudes of specific glaciogenic formations within sedimentary successions. For that reason, I do not use a high-pass filter at Q≥3 as was advocated by Van der Voo (1990). Instead, I assign unequal weight to the seven criteria and choose paleopoles based on the points raised in the above discussion.

Of the 80 or so allegedly glaciogenic deposits discussed above, only 16 have somewhat reliable paleolatitudinal constraints (fig. 10). With paleolatitudes shown in bold in table 1, they are: Rapitan Group (deposit #*1*), Toby Formation (*6*), Johnnie Rainstorm Member (*9*), Florida Mountains diamictite (*10*), units in the southern and central Appalachians (*11*), Vestertana Group (*20*), Tarim basin (*33*), basal Sinian of South China (*35*), Elatina Formation (*40*), Walsh Tillite (*42*), Egan Formation (*43*), Schwarzrand Subgroup (*51*), Chuos Formation (*63*), Bthaat Ergil/Fersiga Groups (*65*), Squantum tilloid (*74*), and Gaskiers Formation (*75*). These are the units for which I conclude that the paleomagnetically studied formation is stratigraphically near enough to the alleged glaciogenic unit to be useful, and for which I consider the magnetic remanence to be possibly primary. Some entries in table 1 are queried as possibly primary but are omitted from the above subset. This is because of too much stratigraphic distance separating the glaciogenic unit and the paleomagnetically studied formation (Ice Brook Formation, #*2*; Sêrie Verte, #*71*), uncertain age of the glaciogenic unit

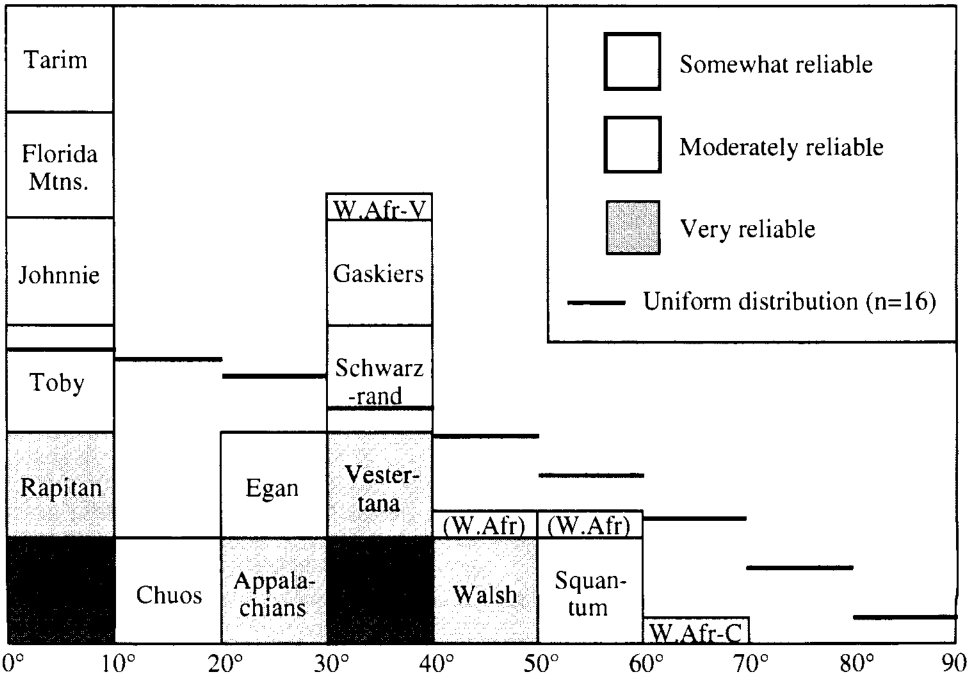


Fig. 10. Histogram of somewhat (clear), moderately (light shade), and very (dark) reliable paleolatitudinal estimates of Neoproterozoic-Cambrian glaciogenic deposits. Note that "reliability" takes into account not only paleomagnetic reliability but also the confidence that the deposits represent regionally significant, low-elevation ice sheets. Unit area is assigned to each deposit, except for the West African "triad" of the Taoudeni basin, whose uncertain age permits a range of paleolatitudes in the context of Gondwanaland's rapid Early Cambrian motion (see fig. 9). The discontinuous steps show the expected density function of a uniform distribution over the sphere.

(Moelv "tillite", #*23*; East European platform deposits, #*24*), or a paleopole of low reliability similar to much younger paleopoles (Sturtian pole MT1, #*39*).

Of the 16 "semi-finalists" whose paleolatitude may be determined accurately by the present database, I consider only five to be reliable at a greater level of confidence. The eleven others are excluded because of doubtful association with a continental ice sheet (Toby, Johnnie Rainstorm, Florida Mountains, southern Appalachians, Schwarzrand, Squantum, and Gaskiers), uncertainty of the paleomagnetic reliability as discussed in detail above (Tarim), uncertain age of the glaciogenic unit (Egan, Jbeliat/Fersiga), or paleomagnetically studied formation (Mbozi complex applied to Chuos). The remaining five are the Rapitan Group, Vestertana Group, South China glaciogenic units, Walsh Tillite, and Elatina Formation including its Marinoan correlatives. The Rapitan paleomagnetic studies by Morris (1977) and Park (1997) are reliable mainly because they independently observed the same high-stability, hematitic components implying low paleolatitude. Regardless of whether the X/R3 or the Z/R2 component represents a primary versus diagenetic remanence, both were acquired when that part of Laurentia was near the Equator. The Vestertana Group pole is pre-folding (Torsvik, Lohmann, and Sturt, 1995); even though it is similar to Ordovician poles from Baltica, special circumstances would be required for it to be an overprint. The Liantuo Formation pole (Evans, 1998b) and the Chengjiang/Nantuo pole (Zhang and Piper, 1997) pass several field stability tests, are determined independently by three laboratories, are broadly consistent, and support the notion that the Chang'an and Nantuo glacial epochs are similar in age (Evans and others, 2000). The Walsh Tillite pole, distinct from all previously determined Neoproterozoic-Paleozoic results from Australia, is pre-folding, and that folding occurred prior to Cambrian time (Li, 2000). Lastly, paleomagnetic results from the Elatina Formation and neighboring stratigraphic units (Schmidt and Williams, 1995; Sohl, Christie-Blick, and Kent, 1999) are quite robust, having passed every test performed on them and having been determined independently by three laboratories. If the lower Marinoan correlations into central and northern Australia are correct, then this result implies extensive equatorial glaciation on that continent. Of the five "finalists," the two judged as "very reliable" are the Liantuo-Nantuo (South China) and Elatina (Australia) results, because of numerous field stability tests and independent verification by several laboratories.

Figure 10 summarizes these interpretations. Each paleolatitude determination is given unit area except for the Bthaat Ergil/Fersiga Groups which are split into equal quarters, with the estimated paleolatitudes depending on their age. For this first-order analysis, uncertainty in paleolatitude (see table 1) is not displayed graphically. A uniform spherical distribution with the same integrated area under the curve is shown for reference. The subset of 16 results shows two peaks, at 0° to10° and 30° to 40° paleolatitude, and more low-latitude data exist than would be expected if the suite were chosen randomly from a uniform distribution on the sphere. The most polar determinations are from the Bthaat Ergil/Fersiga Groups and correlative rocks on the West African craton, assuming that their age is middle Early Cambrian rather than Vendian or earliest Cambrian, and the Squantum tilloid from the Boston basin. The latter deposit is likely to be secondarily derived from alpine glaciers in an active orogenic setting, hence not representative of global climate. The bimodal data distribution from the subset of 16 entries persists among the more stringently chosen subset of five, and even the most reliable two, although with so few data it is difficult to believe that the separate modes are meaningful. Still, none of the 16 somewhat reliable results indicate glaciation at polar latitudes.

My compilation carries a fundamentally different conclusion from that of the most recently preceding attempt at this exercise (Meert and Van der Voo, 1994), for several reasons. First, my method is distinct from theirs; the imprecision of age constraints for the majority of glaciogenic deposits, I believe, renders the chronometric-APW approach ineffective for determining their spatial distribution. Second, my compilation considers a more comprehensive list of deposits than theirs. Third, new paleomagnetic results have arisen since their compilation, including four of the five most reliable determinations.

The culled dataset suggests that the Phanerozoic archetype or Pleistocene-analog model fails in describing the spatial distribution of Neoproterozoic glaciations. Although there might appear to be an inordinate abundance of near-equatorial glacial occurrences, the small number of entries prohibits comparisons between the Snowball Earth and high-obliquity models. For example, a bias toward low paleolatitudes resulting from inclination-shallowing due to sediment compaction (all five of the most reliable studies deal with fine-grained siliciclastic rocks) might have deflected the distribution significantly. Yet the paleolatitudes determined from the Rapitan and deposits are in close agreement with those from nearly coeval crystalline rocks (the ~720-Ma Franklin diabase and Natkusiak basalts; Palmer and Others, 1983; Christie and Fahrig, 1983; see fig. 3).

One could argue that the preponderance of low apparent paleolatitudes results from large non-dipole geomagnetic field components in Neoproterozoic time, for which there is compelling observational evidence (Kent and Smethurst, 1998) and a proposed geodynamic mechanism (Bloxham, 2000). However, a 25 percent octupole and 10 percent quadrupole contribution to the dipolar field as postulated by Kent and Smethurst has maximal effect at moderate apparent paleolatitudes, with minimal effect (and even an uncertainty in sign of the required adjustment, due to the hemispheric ambiguity of the geomagnetic field) at near-equatorial apparent paleolatitudes (fig. 11). If a substantial octupole component of the geomagnetic field did exist in Neoproterozoic time, then mid-paleolatitude results from the lower Sinian glaciogenic units, the Walsh Tillite, and the Vestertana Group would need to be corrected poleward by as much as ~15°. This in turn could favor the Snowball Earth model over the high-obliquity hypothesis.

At face value, however, both the Snowball Earth and high-obliquity models are acceptable according to the paleomagnetic data. Whereas the high-obliquity hypothesis cannot be rejected according to paleomagnetic data *per se*, a single, reliable high-paleolatitude determination on a Neoproterozoic glaciogenic deposit would severely undermine its logical foundation. For the interval 780 to 720 Ma, the waning Rodinia supercontinent was likely centered near the equator (Hoffman, 1991); identification of high-latitude glacial deposits from that time may be difficult simply due to a lack of continental occupation of polar regions (see Chumakov and Elston, 1989; Li, 2000). Future paleomagnetic work should therefore focus attempts to identify primary magnetizations from glaciogenic rocks on continents that may have occupied high latitudes toward the end of the Neoproterozoic (for example, Laurentia and Amazonia).

Temporal Distribution of Neoproterozoic Glacial Deposits

Is a general threefold division of ~900, ~800, and ~650 Ma glacial eras (Hambrey and Harland, 1985) justified by recent U-Pb isotopic data? None of those ages finds support with this compilation, particularly because I have rejected whole-rock Rb-Sr analyses on shales and glauconites. Using primarily concordant or only slightly discordant U-Pb results, the following clusters of age emerge: ~720 to 740 Ma (all continents), ~600 to 620 Ma (most continents), ~570 to 580 Ma (Avalonia-Cadomia only?), and ~545 Ma (several continents). These groupings are tentative because so many of the glacial units are poorly dated (fig. 12). If neither the Avalonian-Cadomian nor the terminal Proterozoic alleged glaciogenic deposits are considered indicative of widespread ice ages, then correlation of the entire Neoproterozoic glacial record into only two intervals (Kennedy and others, 1998) is permitted by the most reliable geochronological data. As stated above, carbon- and strontium-isotope stratigraphies have great potential as Neoproterozoic chronometers (for example, Kaufman and Knoll, 1995;

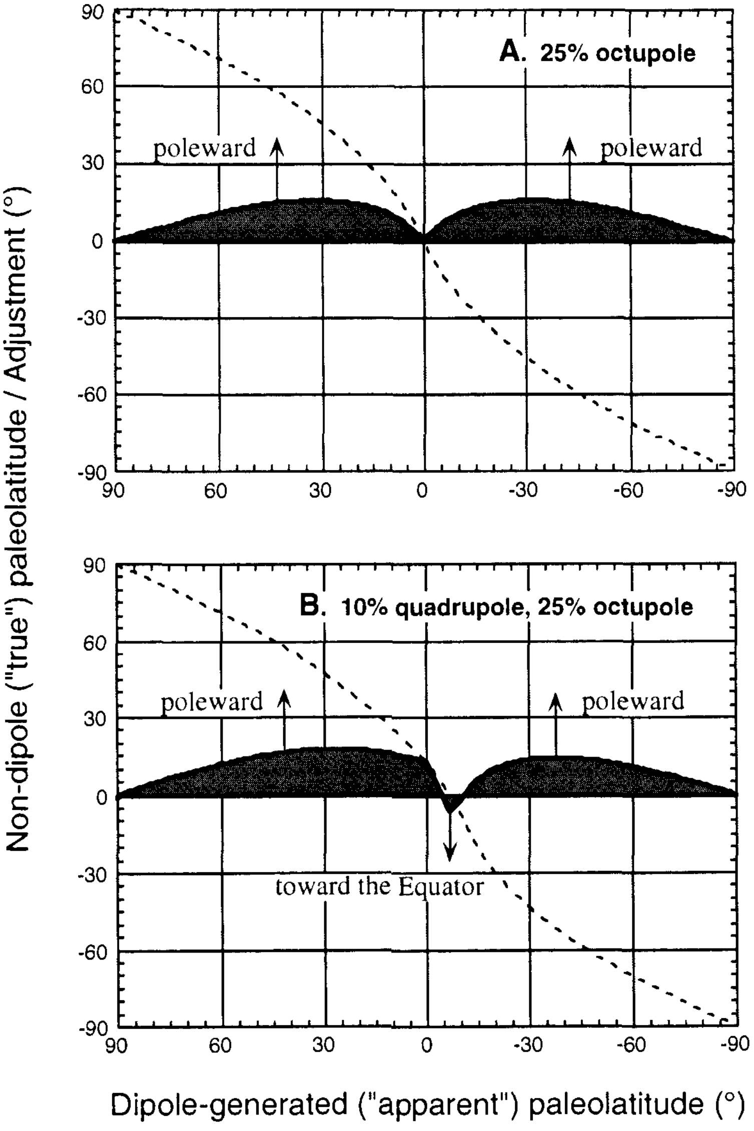
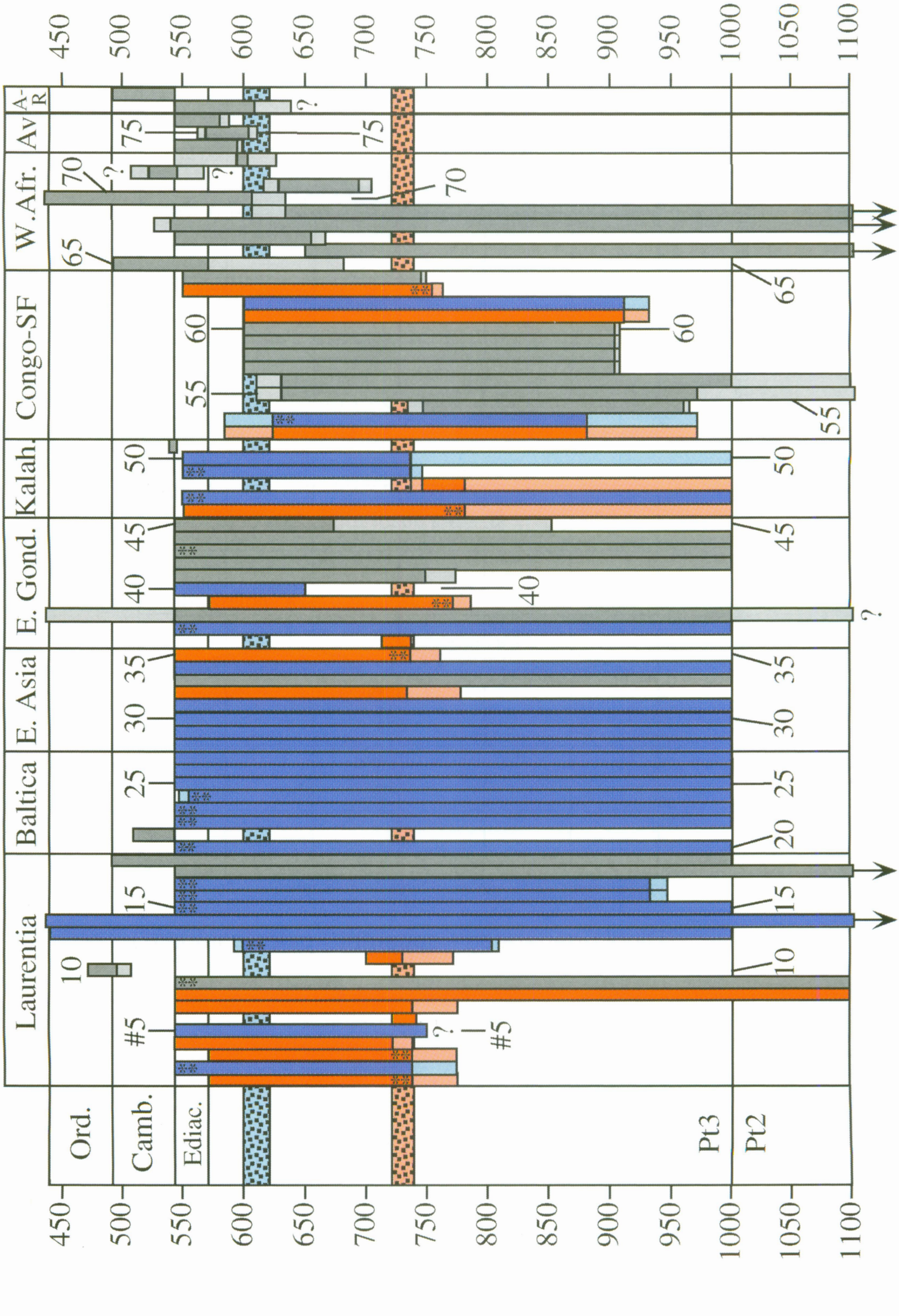


Fig. 11. Effect of non-dipolar components of the geomagnetic field upon interpretation of paleolatitudes based on the axial geocentric dipole hypothesis, calculated from Kent and Smethurst (1998). Assuming that various non-dipolar components truly existed in the Proterozoic, the dashed curves show "true" paleolatitudes as a function of "apparent" paleolatitudes; solid curves (outlining the shaded regions) indicate magnitudes of the adjustments from "apparent" to "true" values. (A) Adjustments that would be required if an octupolar component of the same sign and 25 percent of the dipolar magnitude were added to the geomagnetic model. The adjustments are symmetric about the Equator and indicate a poleward correction by as much as 17° for moderate apparent latitudes, but no correction necessary for precisely equatorial or polar results. (B) As in (A), but adding a 10 percent magnitude quadrupolar component to the geomagnetic model. This is the preferred, although admittedly non-unique, model of Kent and Smethurst (1998), as one explanation for the inordinate abundance of low-latitude results within the Paleozoic and Precambrian global paleomagnetic database. The antisymmetric quadrupolar component introduces an uncertainty in the necessary adjustment, due to the hemispheric ambiguity of paleomagnetic results from ages that pre-date the continuous portions of continental APW paths. For example, depending on the hemisphere, a 10° "apparent" paleolatitude could be adjusted either poleward by 16° or toward the equator by 2°.


Ord.
Camb.
Ediac.
Pt3
Pt2
Laurentia
Baltica
E. Asia
E. Gond.
Kalah.
Congo-SF
W. Afr.
Av
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Kaufman, Knoll, and Narbonne, 1997; Kennedy and others, 1998; Jacobsen and Kaufman, 1999); however, I have omitted most of those data, particularly the high-frequency carbon-isotopic record, from the present analysis in order to present the temporal constraints with as little interpretation as possible.

Of the five moderately reliable paleolatitude determinations discussed above, two (Rapitan, Chang'an/Nantuo) are commonly grouped with the ~720 to 740 Ma cluster of glaciogenic deposits, and the other two (Vestertana, Elatina) are commonly grouped with a ~600 to 620 Ma age. Those correlations may be correct, and if so, they would satisfy the synchroneity requirement of the Snowball Earth model. Nonetheless, it is entirely possible that those two pairs, and many other of the commonly grouped glaciogenic deposits, will be found to be diachronous through forthcoming geochronological research.

Whether or not glaciogenic units can be correlated regionally or globally as chronostratigraphic markers (the "mega-events" of Ojakangas, 1988) carries important consequences for tectonic syntheses as well as paleoclimatic issues. For example, Trompette's (1994, 1997) syntheses of the relative timing for continent-continent collisions represented by the Pan-African and Brasiliano super-orogenic systems is in some ways defined by assumption of regionally synchronous glaciogenic deposits. The isotopic data reviewed here suggest diachroneity of previously correlated glaciogenic units in West Africa. In addition, consideration of a glaciogenic unit or its "cap" carbonate as a global stratotype boundary level for terminal Proterozoic time is rooted in the presumption that several such horizons around the world were deposited roughly synchronously. The present geochronological database contradicts the notion of, for example, coeval deposition of the Avalonian diamictites and the classic "Vendian" Port Askaig Tillite (see the discussion of age constraints on the "Varanger" glaciation in Grotzinger and others, 1995).

Association of glaciogenic deposits with iron- or manganese-formation cannot be used as a reliable chronostratigraphic tool. True, many of these deposits could fall into the ~720 to 740 Ma group, but others, particularly in South America and Africa, are likely to be younger. In many examples (for example, around the São Francisco and West African cratons), appearance of iron-formation is related to a facies transition to a more distal, deep-water environment. This evidence contradicts the claim by Kennedy and others (1998) that glacial association with sedimentary iron-formation can be used as a lithostratigraphic characteristic of definitively "Sturtian" age.

DISCUSSION AND CONCLUSIONS

According to this review of the paleomagnetic constraints on Neoproterozoic glaciogenic deposits, a small subset of reliable results indicates purely low- to mid-latitude occurrences. Thus the Phanerozoic archetype seems to be inapplicable toward Neoproterozoic (and earlier) times, but the Snowball Earth and the high obliquity models are permissible. A single, reliable, high-latitude deposit could undermine the

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Fig. 12. Summary of age constraints (in Ma) for alleged Neoproterozoic glaciogenic deposits. Early Paleozoic timescale from Bowring and Erwin (1998). "Ediac." = informal designation of the approximate age range (no older than 570 Ma) for higher-diversity Ediacaran fossils (Bowring and Erwin, 1998; Grotzinger and others, 1995). Numbering scheme for the deposits matches that of the text, tables, and other figures. Darker shades show the range of possible ages for each of the alleged glaciogenic formations; lighter shades indicate the uncertainty ranges of the maximum or minimum constraints. Asterisks denote more likely "true" ages toward one end or another of the permitted range (see text for case-specific details). Blue = deposits most commonly grouped with a "Varanger" or "Marinoan" glaciation; red = deposits commonly grouped with a "Rapitan" or "Sturtian" glaciation; gray = deposits of disputed or uncertain affiliation. Stippled colored regions show possible age ranges for "Sturtian" and "Varanger" glaciations that satisfy all of the existing age constraints from the various deposits. A = Avalonia-Cadomia belts; R = Amazonia and Rio de la Plata cratons; SF = São Francisco craton.

high-obliquity model; a demonstration of substantial diachroneity among deposits worldwide could refute the Snowball Earth model. A combination of high-obliquity and Snowball Earth models is possible (that is, predominantly low-latitude deposits but allowing for the rare occurrence of polar glaciers), but the high-obliquity model was devised as an alternative rather than a supplement to global refrigeration (Williams, 1975). Other possibilities to explain the low-latitude data, including non-dipole-field biasing effects (Kent and Smethurst, 1998; Bloxham, 2000), cannot fully account for the most reliable, most equatorial paleomagnetic results.

Neither the Snowball Earth nor the high-obliquity hypotheses state directly how the glaciations may have been initiated. Any combination of processes such as lower solar luminosity (Crowley and Baum, 1993) and supercontinental rift-related chemical drawdown of atmospheric CO_2 (Young, 1991, 1995) may have contributed to planetary refrigeration. The chemical and albedo effects of continents are exacerbated if the landmasses are centered on the equator, as geodynamic theory would predict for old supercontinents (Anderson, 1982; Evans, 1999). Hoffman and Schrag (2000) noted additionally that a global paleogeography incorporating a low-latitude supercontinent and polar oceans, such as that suggested for Rodinia at ~750 Ma, could effectively eliminate the negative feedback of ice cover and silicate weathering that helps regulate Quaternary climates. Finally, Hoffman and others (1998a) emphasized the physical implications of Rodinia's breakup, suggesting that newly developed passive continental margins became efficient repositories for the sedimentary burial of organic carbon, causing a long-term atmospheric CO_2 reduction.

A new item of interest is the possibility of substantial true polar wander (TPW) during late Neoproterozoic time, due to a rotationally unstable pattern of mantle convection induced by the long-lived Rodinian supercontinent (Evans, 1998a). The geodynamical theory of TPW permits large shifts as rapid as 90° per 5 to 10 my (Steinberger and O' Connell, 1997). Rapid TPW could explain the abrupt transition into and out of Neoproterozoic polar-centered glaciations (Fairchild, 1993), but it cannot–and was never intended to–explain low paleomagnetic latitudes obtained directly on the glacial deposits (for example, the Elatina Formation). However, if many rapid TPW swings occurred during the Neoproterozoic, then the global climate may have responded in unusual ways. For example, rapid continental drift across polar and equatorial latitudes could subject high-relief areas to alternating intervals of physical weathering (freezing and thawing of ice) and chemical weathering (silicate leaching). The latter process contributes to drawdown of atmospheric CO_2 but cannot operate at high rates unless the surface area of rock exposure is increased by the former process. The two processes acting together in a scenario of rapid continental drift, whether due to multiple episodes of TPW (Evans, 1998a) or enhanced rates of plate-tectonics (Gurnis and Torsvik, 1994), could create a strong positive feedback toward the growth of continental ice sheets.

If Precambrian paleoclimate is governed by different boundary conditions such as presented in the high-obliquity hypothesis, then the abrupt transition to the Phanerozoic archetype is intriguing. By Late Ordovician time, the planet's ice cap appears to have coincided with its geomagnetic and, by assumption, rotational axes (Smith, 1997). Cambrian glaciogenic deposits are relatively rare, but the possibility that the "triad" marker unit found throughout West Africa is of Early Cambrian age (Bertrand-Sarfati and others, 1995) occupying polar latitudes, would extend the Phanerozoic paleoclimatic paradigm to that period as well. If the Egan Formation is latest Proterozoic in age (Corkeron and others, 1996; Plumb, 1996) then its low paleolatitude would mark the end of the non-uniformitarian Precambrian climatic system. In this respect, further sedimentological study of the low-latitude, Cambrian-Ordovician Florida Mountains diamictites would be especially welcome, to assess critically their alleged glaciogenic origin.

The abrupt transition from the Proterozoic to the Phanerozoic glacial paradigm can be explained by both the high-obliquity and Snowball Earth hypotheses, although there are additional implications for each. G.E. Williams (1993) suggested dissipative core-mantle coupling as a means for rapidly reducing planetary obliquity at ~500 Ma, but the phenomenon has been shown to be incapable of the desired effect (Pais and others, 1999). As an alternative mechanism, D.M. Williams, Kasting, and Frakes (1998) suggested the obliquity-oblateness feedback, operating in reverse of its conventional application as a way to increase Earth obliquity (Rubincam, 1993). The change in sign is allowed if and only if a somewhat non-intuitive range of values is adopted for the phase lag, ξ_i, between solar insolation and the decay and growth of ice sheets: the process only works to reduce obliquity if the ice sheets either disappear nearly instantaneously upon increases in luminosity ($\xi_i < 25°$) or continue growing even after the insolation has increased and reached a maximum ($205° < \xi_i < 360°$). Also, reduction of obliquity by the oblateness feedback past the critical value of 54° should be hindered by little or no forcing at that time ($\Delta J_2 \approx 0$; no concentration of ice at any latitude and hence an insignificant effect on oblateness even during intense ice ages). Finally, it appears that there is a delicate range of high-enough obliquity values capable of maintaining of orbital stability; between 60° and 90° is an unstable zone due to the Moon's influence (Laskar, Joutel, and Robutel, 1993; see also fig. 3 of Williams, Kasting, and Frakes, 1998).

The Snowball Earth hypothesis also includes possible reasons for a secular change in the planet's fundamental paleoclimatic system at the Proterozoic-Cambrian transition (Hoffman and others, 1998a), namely the evolutionary development of bioturbation inhibiting the atmospheric CO_2-depleting process of organic-carbon burial and rising oceanic oxygen concentrations inhibiting globally averaged rates of photosynthetic productivity. These boundary conditions would of course be absent during the Paleoproterozoic, also a time of low-latitude glaciation (Evans, Beukes, and Kirschvink, 1997; Williams and Schmidt, 1997).

Almost 20 yrs have passed since Hambrey and Harland (1981) produced their exhaustive compilation of Neoproterozoic (and other pre-Pleistocene) glacial deposits. The paleoclimatic paradox itself, the alternative models, and the means of testing those models, all have changed little during the intervening years, but many new stratigraphic, geochronologic, and paleomagnetic data have arisen, especially in the last 5 yrs. We are only now beginning to constrain some of the Neoproterozoic glaciogenic deposits adequately enough to address this fundamental issue of Earth's long-term paleoclimate. Most of the glacial units, however, are still poorly constrained in time and space. Further work in emerging methods of chronostratigraphy (for example, focussed geological mapping in conjunction with intra- and inter-basinal correlation using a variety of techniques), additional geochronological studies (using advanced methods to measure very small mineral samples with great precision), and more focussed paleomagnetic work (with emphasis on proving primary magnetic remanences and testing the axial geocentric dipole hypothesis) will certainly provide important new constraints during the coming years.

ADDED IN PRESS

Periglacial features have now been reported within the Ikorongo group of the Tanzania Craton (see fig. 8; Pinna, P., Cocmerie, A., Thiéblemont, D., and Jezequel, P., 2000, the Kisii Group of Western Kenya: An end-archean (2.53 Ga) late orogenic volcano sedimentary sequence: Journal of African Earth Sciences, v. 30, p. 79–97). The Ikorongo Group has undergone paleomagnetic study, but it is apparent from the streaked distribution of sample directions that the alternating-field demagnetization methods were incapable of isolating a "pure" characteristic component within the redbeds (Piper, J. D. A., 1975, Palaeomagnetic correlations of precambrian formations of East-Central Africa and their tectonic implications: Tectonophysics, v. 26, p. 135–161). Recent *computer simulations of Snowball Earth conditions, using a general-circulation climate model, provide support for the "mild" version of the hypothesis, maintaining areas of liquid water near the equator during the modeled ice ages (Hyde, W. T., Crowley, T. J., Baum, S. K., and Peltier, W. R., 2000, Neoproterozoic "Snowball Earth" simulations with a coupled climate/ice sheet model: Nature, v. 405, p. 425–429).*

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