I. NEOPROTEROZOIC-PALEOZOIC SUPERCONTINENTAL TECTONICS AND TRUE POLAR WANDER

II. TEMPORAL AND SPATIAL DISTRIBUTIONS OF PROTEROZOIC GLACIATIONS

Thesis by

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The 'Football Earth?' Weren't you the guy who was talking before about the 'Snowball Earth?' Pretty soon they're going to start calling it 'Screwball Earth....'

--my mother, succinctly relating
the two parts of this thesis

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ABSTRACT

The Proterozoic Eon, occupying nearly half of Earth history from 2.5 to 0.5 billion years ago, is marked at its beginning and end by dramatic events in the tectonic, paleoclimatic, chemical, and biological evolution of the planet. The onset of the Proterozoic Eon witnessed the emergence of continents and perhaps the introduction of plate tectonics, Earth's first extensive ice ages, oxygenation of the hydro-atmosphere, and development of eukaryotes. The end of the Proterozoic Eon is characterized by supercontinental turnover and very rapid continental drift rates, a series of glaciations which left their marks on every continent, the rise of atmospheric oxygen to sustain multicelled organisms, and an evolutionary "explosion" of animal life.

Establishment of coherent paleogeographies of these important intervals is a crucial prerequisite for describing the events and understanding the underlying processes. Paleomagnetism is the most direct quantitative method for charting continental drift through time. The purpose of this dissertation is to use paleomagnetism to constrain tectonic and paleoclimatic processes at the beginning and end of the Proterozoic Eon.

A paleomagnetic study of Early Cambrian rocks in western Mongolia finds somewhat ambiguous results and addresses tectonic models of the Paleo-Asian Ocean. Review of the most reliable studies among the Proterozoic-Cambrian global paleomagnetic database permits the hypothesis that an episode of inertial interchange true polar wander (TPW) occurred in Early Cambrian time. The Cambrian TPW event and a previously hypothesized Ordovician-Devonian TPW migration share a common axis and suggest the existence of long-lived mantle mass anomalies inherited from the previous supercontinent, Rodinia. The breakup of Rodinia and subsequent amalgamation of Gondwanaland appear analogous in several ways to the Gondwanaland-Super-Asia supercontinental transition, suggesting a 500-600-Myr cyclicity.

An exhaustive review of paleomagnetic and geochronological constraints upon Neoproterozoic glaciogenic deposits fails to find a convincing high-paleolatitude occurrence. Detailed study of one of these deposits in South China reveals a reliable paleomagnetic pole implying a paleolatitude of 34±2°, with both paleoclimatic and paleogeographic implications. A reliable estimation of 11±5° depositional paleolatitude for 2.2-billion-year-old lavas directly overlying a glaciogenic formation in South Africa, extends the occurrence of low-latitude continental ice sheets further back into the Precambrian.

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PART I. NEOPROTEROZOIC-PALEOZOIC SUPERCONTINENTAL TECTONICS AND TRUE POLAR WANDER

Chapter 1: Introduction to Part I

Deciphering the history of ancient supercontinents is a daunting task, requiring synthesis of geological knowledge from all corners of the world. The pioneering works of Suess (1892), Wegener (1912), and du Toit (1937) paved the way toward recognition of larger landmasses from the geological past, such as Gondwanaland and Pangea, once coherent entities which subsequently fragmented into today's continents. In the 1950s and 1960s, marriage of continental drift theory with the geodynamic mechanism of plate tectonics spawned fresh ideas on these matters. Coeval developments in the technique of paleomagnetism permitted quantitative estimates of continental configurations, verifying Pangea's existence (Irving, 1956; Runcorn, 1956).

But reconstruction of Pangea only takes us back about 200 million years (Myr). Although some geologists recognized the possibility of older supercontinental assemblages, created and destroyed by the repeated opening and closing of ocean basins (Wilson, 1966), geological evidence supporting these more ancient landmasses was scanty. The major hurdles obstructing efforts to reconstruct the pre-Pangean world were and still are: (1) lack of seafloor older than 200 Ma, (2) difficulties in dating Precambrian (older than about 550 Ma) rocks due to the lack of a well ordered macroscopic fossil record, and (3) tectonic, thermal, and paleomagnetic overprinting of Precambrian rocks by subsequent geological events. Despite these shortcomings, several intrepid workers set out to determine the history of supercontinents preceding Pangea (e.g., Sutton, 1963; Burke and Dewey, 1973; Piper et al., 1973; Bond et al., 1984; Hoffman, 1988; Dalziel, 1992; Rogers, 1996).

Whereas the early-recognized peaks in isotopic ages, representing orogenic maxima and thus periods of possible supercontinental assembly, have proven to be correct to first order (Hoffman, 1989), many of the earlier paleomagnetic results must be discounted because of their failure to separate magnetic overprints from primary directions (Van der Voo, 1993). Thus pre-Pangean paleogeography remains speculative. In particular, a flurry of recently proposed continental reconstructions for Neoproterozoic time (Moores, 1991; Hoffman, 1991; Dalziel, 1991, 1994, 1997; Kirschvink, 1992; McKerrow et al., 1992; Powell et al., 1993; Pelechaty, 1996; Li et al., 1996; Torsvik et al., 1996; Veevers et al., 1997) have yet to be tested quantitatively by reliable paleomagnetic studies.

Some models (e.g., Hoffman, 1991; Dalziel, 1991) rely heavily on preexisting tectonic units, such as the 1.3-1.0-Ga "Grenvillian" orogenic belts, as "piercing points" between reconstructed continental margins. In addition, two passive margins of the same age and similar sedimentary record (in this case Vendian-Cambrian) can be rejoined to a supposed pre-rift state (e.g., Bond et al., 1984; Li et al., 1996). Unfortunately, the solutions to this problem are non-unique; if a Neoproterozoic supercontinent (Rodinia) existed as many now suspect, we should find many penecontemporaneous orogenic belts contributing to its formation and many similarly-aged passive margins resulting from its demise. Because rifted continental margins commonly split along previous sutures, many Neoproterozoic continental margins should have contained fragments of Grenvillian-aged orogenic belts (Fig.1), and there are many ways to put the pieces together coherently. The most common reconstructions of Rodinia incorporate a single, continuous belt of Grenvillian-aged continental sutures, although alternative reconstructions of discontinuous sutures, as was the case for Pangea, are also possible.

Other models use faunal realms and lithological climate indicators to reconstruct the continents (e.g., McKerrow et al., 1992). The main problems with this method are (1) precise faunal realms are determined only for Cambrian and younger time, leaving almost the entire Neoproterozoic Era unconstrained, (2) faunal realms are only qualitative measures of continental separation or combination, and (3) lithologic units such as evaporites, carbonates, and tillites are also qualitative and may not be applicable to extreme fluctuations or non-uniformitarian behavior of Earth's paleoclimate (see Part II).

Models based primarily on paleomagnetism (Kirschvink, 1992; Powell et al., 1993) can quantitatively constrain the paleolatitudes (with respect to Earth's time-averaged

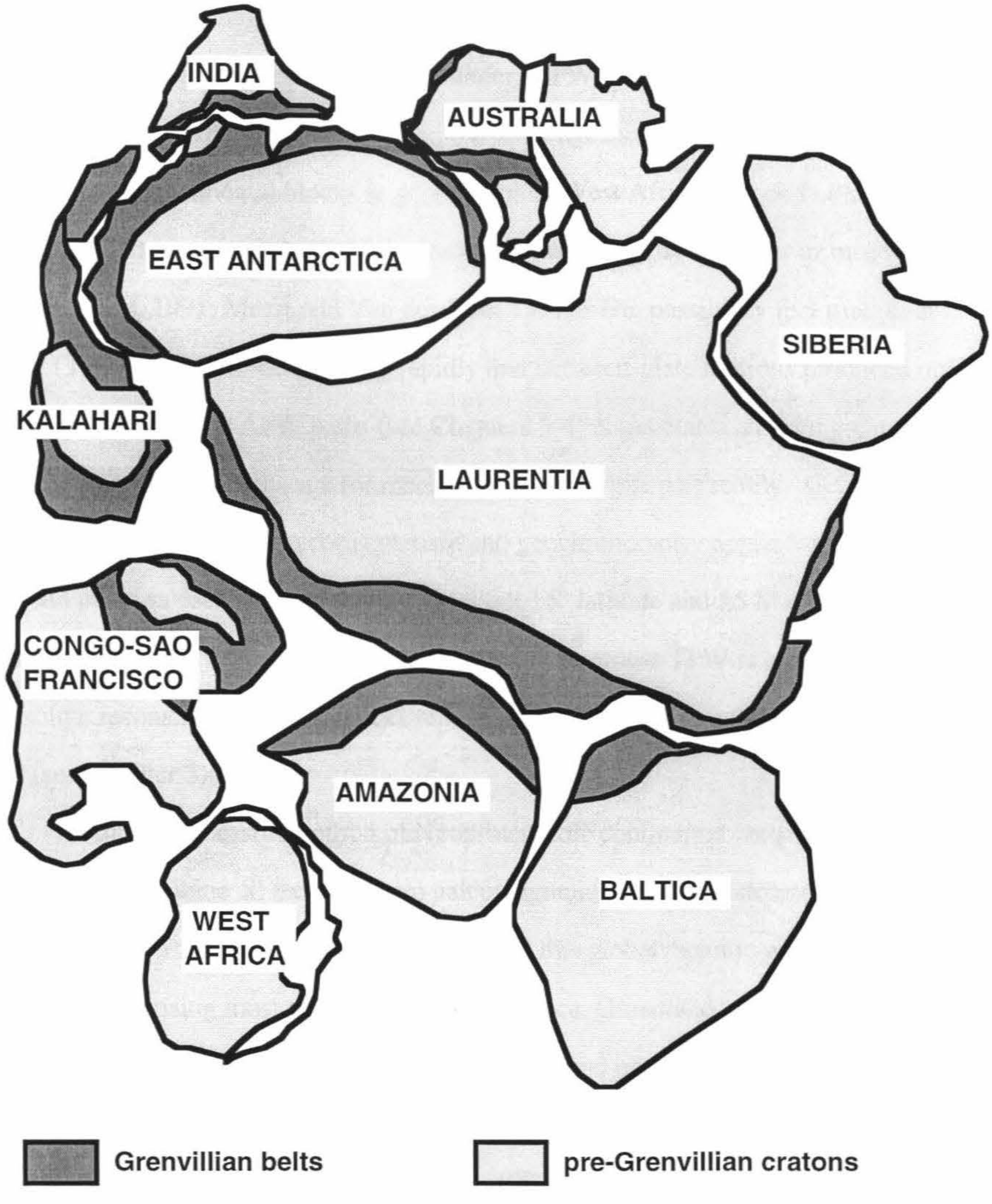

Figure 1. Reconstruction of Neoproterozoic Rodinia, after Hoffman (1991). "Grenvillian" belts are interpreted as the loci of 1.3-1.0-Ga continent-continent collisions.

magnetic dipole) of various continental blocks, given that the data from each block are adequately constrained. These reconstructions are hampered by a shockingly poor Neoproterozoic paleomagnetic database for almost all of the continental blocks. Most studies contributing to an apparent polar wander (APW) path of a given craton have not proven that magnetization is primary, i.e., the same age as the rock; and many Neoproterozoic continental blocks (e.g., Amazonia, West Africa, Rio de la Plata craton) are constrained by at most a few reliable poles, separated in age by 100 Myr or more (Van der Voo and Meert, 1991; Meert and Van der Voo, 1997). The possibility that true polar wander (TPW) may have occurred so rapidly that between-plate motions produced only a small component of the APW paths (see Chapters 3-4) necessitates an even greater understanding of all the datasets for merely a first-order grasp of reality. Given enough data, however, the tools of paleomagnetism and geochronometry applied to Neoproterozoic rocks can position each continent to within about ±5° latitude and ±5 Ma. Longitude is generally unconstrained by paleomagnetic techniques, unless TPW is invoked to produce an absolute reconstruction of all blocks relative to the common rotational pole of the TPW event (see Chapter 3).

The most successful method of Neoproterozoic continental reconstructions, of course, would combine all the data from paleomagnetism, faunal distributions, and tectonic correlations into a coherent picture. The pieces of this global tectonic puzzle were: (1) Laurentia, comprising most of present North America, Greenland, the northern British Isles, and allochthonous terranes in the Andean foreland of Argentina (Hoffman, 1988; Dalla Salda et al., 1992); (2) Baltica, the exposed Fennoscandian shield plus the Russian platform of sedimentary cover (Torsvik et al., 1992); (3) the Siberian craton, the cratonic massif bounded by (CCW from west) the Yenisey Ridge, the Altaid foldbelt, the Verkhoyansk Mountains, and the Taimyr belt (Zonenshain et al., 1990; Sengör et al., 1993); (4) Precambrian blocks of eastern Asia, including Tarim, North China, and South China (Li et al., 1996); (5) East Gondwanaland, comprising present western and central

Australia, East Antarctica, India and Sri Lanka, and Madagascar; (6) the Kalahari craton in southern Africa (Tankard et al., 1982); (7) the Congo-São Francisco craton, straddling the southern Atlantic between central Africa and eastern Brazil (Cahen and Snelling, 1984); (8) West Africa, generally west of 0° longitude but extending into the São Luis block of northern Brazil (ibid.); and (9) Amazonia, occupying a large region in northwestern Brazil and adjacent areas (Trompette, 1994). Other smaller Precambrian continental fragments exist, for example, in Uruguay (Ramos, 1988), Yemen (Windley et al., 1996) and Svalbard (Harland and Wright, 1979).

Part I of this dissertation seeks to address global Neoproterozoic-Paleozoic kinematics using paleomagnetism. New results are obtained from Cambrian rocks in Mongolia (Chapter 2), and the pre-existing global paleomagnetic database is re-examined to reveal previously undiscovered patterns of continental motion. These patterns suggest the existence of substantial and rapid true polar wander (TPW) during and around Early Cambrian time (Chapter 3). The TPW appears to be governed by a geophysical legacy of the vanished Rodinia supercontinent, permitting the possibility that many other TPW events could have occurred during Neoproterozoic-Paleozoic time (Chapter 4). The Neoproterozoic tectonic transition between Rodinia and Gondwanaland is kinematically similar to the Phanerozoic and ongoing transition from Gondwanaland to Super-Asia. A new paradigm for supercontinental cyclicity is proposed, based on continuous transfer of blocks from a waning, old supercontinent, to the waxing, new supercontinent (Chapter 5).

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Chapter 2: Paleomagnetism of the Bayan Gol Formation, western Mongolia

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Abstract

Oriented samples of the Lower Cambrian Bayan Gol Formation from Salaany Gol, Mongolia, were collected at roughly 5m stratigraphic intervals for palaeomagnetic analysis. Progressive alternating-field and thermal demagnetization isolated two magnetic components: a present-field overprint, typically removed by 10 mT fields and ~200 °C heating; and a high-coercivity, high-unblocking-temperature (550-600 °C), predominantly single-polarity component that was imparted to the rocks prior to early or middle Palaeozoic deformation.

Single-polarity magnetization at Salaany Gol contrasts with results from Lower Cambrian rocks on the Siberian platform, previously considered correlative with the Bayan Gol Formation, which show a prominent change in polarity bias near the top of the Tommotian Stage. Two hypotheses can explain this discrepancy. First, the entire Bayan Gol Formation may correlate with the predominantly reversely polarized, lower half of the Tommotian Stage in Siberia. This model is consistent with plausible interpretations of $\delta^{13}C$ profiles for the Zavkhan basin and the Siberian platform. Alternatively, the characteristic magnetic direction from our samples may be a pre-fold overprint. If post-accretionary, then comparison with Siberian palaeomagnetic results suggest a Silurian-Devonian remagnetization age, and existing bio- and chemo-stratigraphic correlations provide the most reliable spatial and temporal links between the Zavkhan basin and the Siberian platform.

If the observed magnetic directions are primary or an immediate overprint then they may be used to constrain the early Cambrian palaeogeography of the Zavkhan basin and the Palaeo-Pacific Ocean. Mean inclination of 62±4° corresponds to a palaeolatitude of 44±5°, several thousand km from the equatorial Siberian craton.

1. Introduction

The Zavkhan basin contains a ~1-2 km thick Vendian-Cambrian sedimentary sequence deposited on Riphean volcanic rocks (Voronin *et al.*, 1982). The Bayan Gol Formation, primarily siliciclastic with calcareous interbeds, contains small shelly fossils that are thought to correlate with Nemakit-Daldynian (ND; Brasier, Kuleshov & Zhegallo, 1996) or ND to lower Atdabanian biozones in Siberia (Voronin *et al.*, 1982). The sequence is well exposed at Salaany Gol, in the Khasagt-Khayr Khan Mountains ~50 km NW of Altay (Fig.1).

The thickness and excellent fossil content of the Salaany Gol locality make it potentially useful for magnetostratigraphic comparisons with other Vendian-Cambrian sections throughout the world. Magnetostratigraphy, in conjunction with carbon isotope stratigraphy, has helped refine palaeontological correlations among basins in Siberia, Morocco, and China (Kirschvink *et al.*, 1991). This study is the first palaeomagnetic analysis of Vendian-Cambrian rocks in the Zavkhan basin.

2. Sampling and analysis

More than 200 oriented core and block samples were collected throughout the Bayan Gol Formation, from three adjacent profiles with slightly overlapping stratigraphy (Figs. 1, 2). The lowest sample, near the base of the Bayan Gol Formation, is coincident with the base of the *Tiksitheca licis-Maikhanella multa* Zone (here and below, *sensu* Voronin *et al.*, 1982). Aside from a 120-m interval of poor exposure near the top of this zone, sample spacing is of the order 5 m throughout the higher zones *Ilsanella compressa*, *Anabarella plana*, *Tannuella gracilis*, and *Stenothecoides*. Two of the sample profiles (A and B, Fig. 1) cross moderately dipping strata, whereas the third (C) transects subvertical layers. The ~40° difference in bedding dip permits a fold test, whereby the age of magnetization can be compared with the age of folding.

Samples were oriented in the field by both magnetic and solar compasses. Once

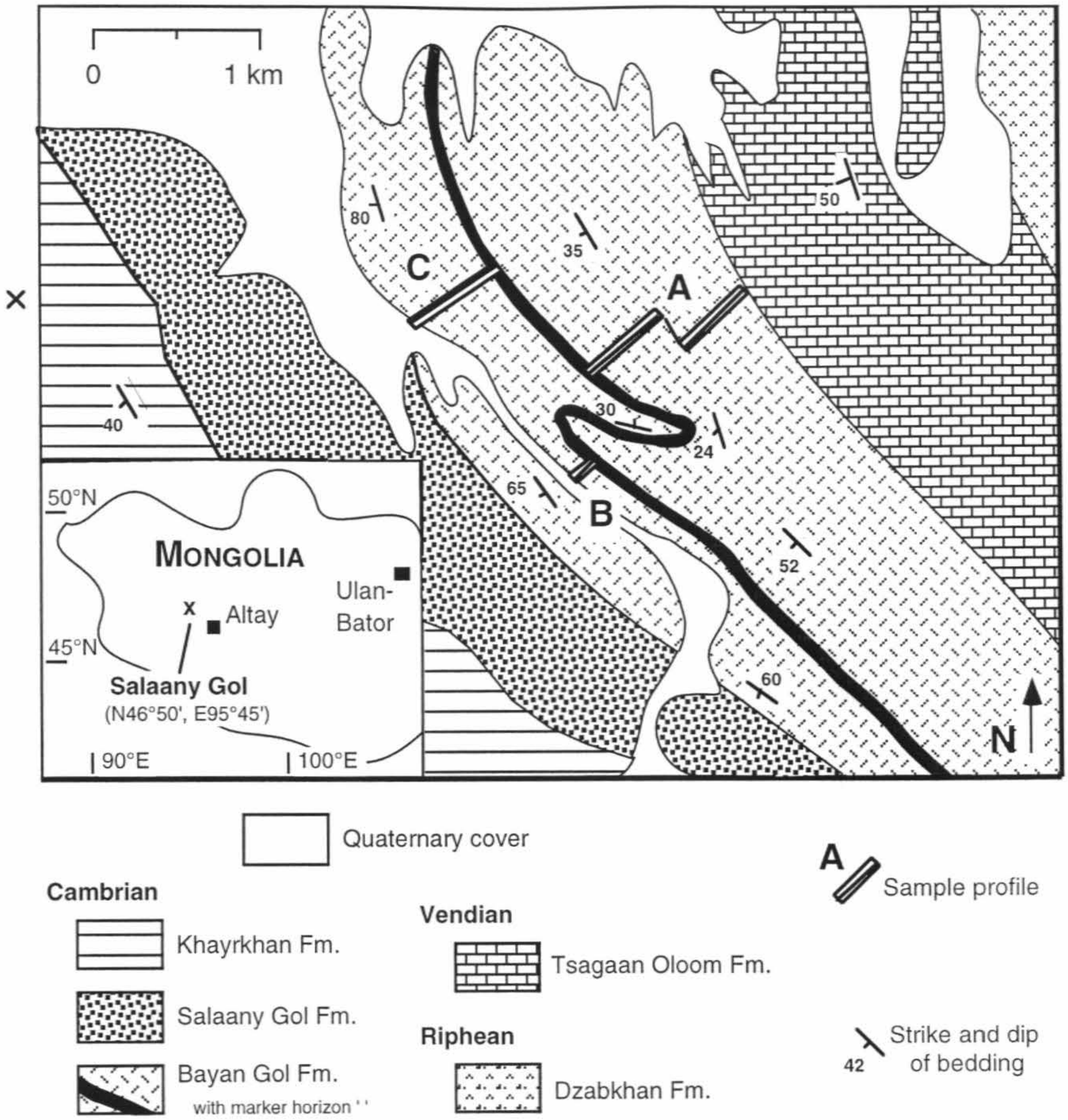
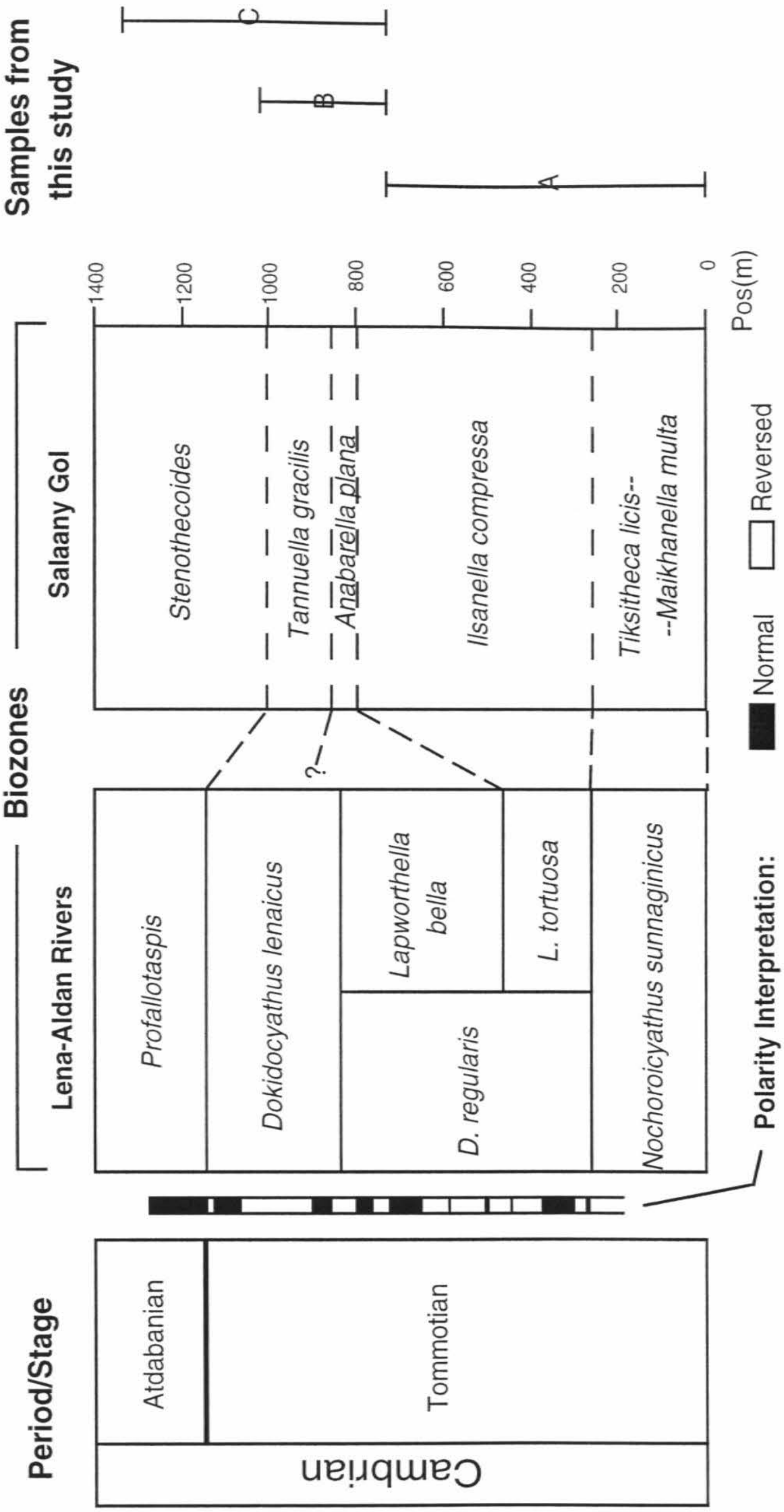


Figure 1. Geologic map of Salaany Gol, after Voronin *et al.* (1982, p. 8).

Figure 2 (next page). Biostratigraphic correlation between Mongolia and Siberia, after Voronin *et al.* (1982, p.13). Magnetic polarity interpretation from Siberian reference sections along the Lena River, according to Kirschvink *et al.* (1991). Stratigraphic thicknesses of Mongolian and Siberian sections are not to scale.



Period/Stage
Cambrian
Atdabanian
Tommotian
Biozones
Lena-Aldan Rivers
Profallotaspis
Dokidocyathus lenaicus
D. regularis
Lapworthella bella
L. tortuosa
Nochoroicyathus sunnaginicus
Salaany Gol
Stenothecoides
Tannuella gracilis
Anabarella plana
Ilsanella compressa
Tiksitheca licis--
--Maikhanella multa
?
1400
1200
1000
800
600
400
200
0
Pos(m)
Samples from this study
A
B
C
Polarity Interpretation:
Normal
Reversed

collected, the 2.5 cm-diameter cylindrical cores were trimmed to 2.3 cm length. From the block samples, cores were drilled, oriented, and trimmed in the laboratory. Samples were measured in an SCT-2G® superconducting moment magnetometer inside a magnetically shielded room. Specimens were subjected to 2.5 mT steps of alternating-field (AF) demagnetization up to 10 mT, followed by thermal demagnetization (heating and cooling in zero field) at 50 °C steps from 200 °C to 650 °C, or until sample intensity dropped to the instrument noise level. Measurements were made after each partial demagnetization step. Magnetic components were separated by linear principal-component-analysis (Kirschvink, 1980).

2.a. Sample data

Examples of sample demagnetization paths are shown in Fig. 3. Natural remanent magnetization (NRM) moments varied from 10^{-9} to 10^{-6} Am^2 (or 10^{-6} to 10^{-3} emu), with typical values of 10^{-8} Am^2 (10^{-5} emu). Almost all of the samples yielded coherent demagnetization paths with distinct components. NRM vectors generally comprise two components. The first of these to be removed, primarily in the AF steps or the 200 °C thermal steps, is aligned with the present geomagnetic field. The (1) low coercivity and (2) low unblocking temperature of this component suggest that it is held (1) partly by multidomain magnetite, and (2) partly by goethite. The second magnetic component is generally removed between temperatures of 350 °C and 550-600 °C. The unblocking temperature of this component suggests that magnetite is the dominant remanence carrier. For each sample, we call the last distinguishable component to be removed the characteristic remanent magnetization (ChRM). In many samples the demagnetization paths do not decay to the origin. Although this may suggest an additional component for each sample, least-squares-plane analysis demonstrates that such components are randomly oriented throughout the suites of samples (Fig. 4c). Thus the chosen ChRM is the most

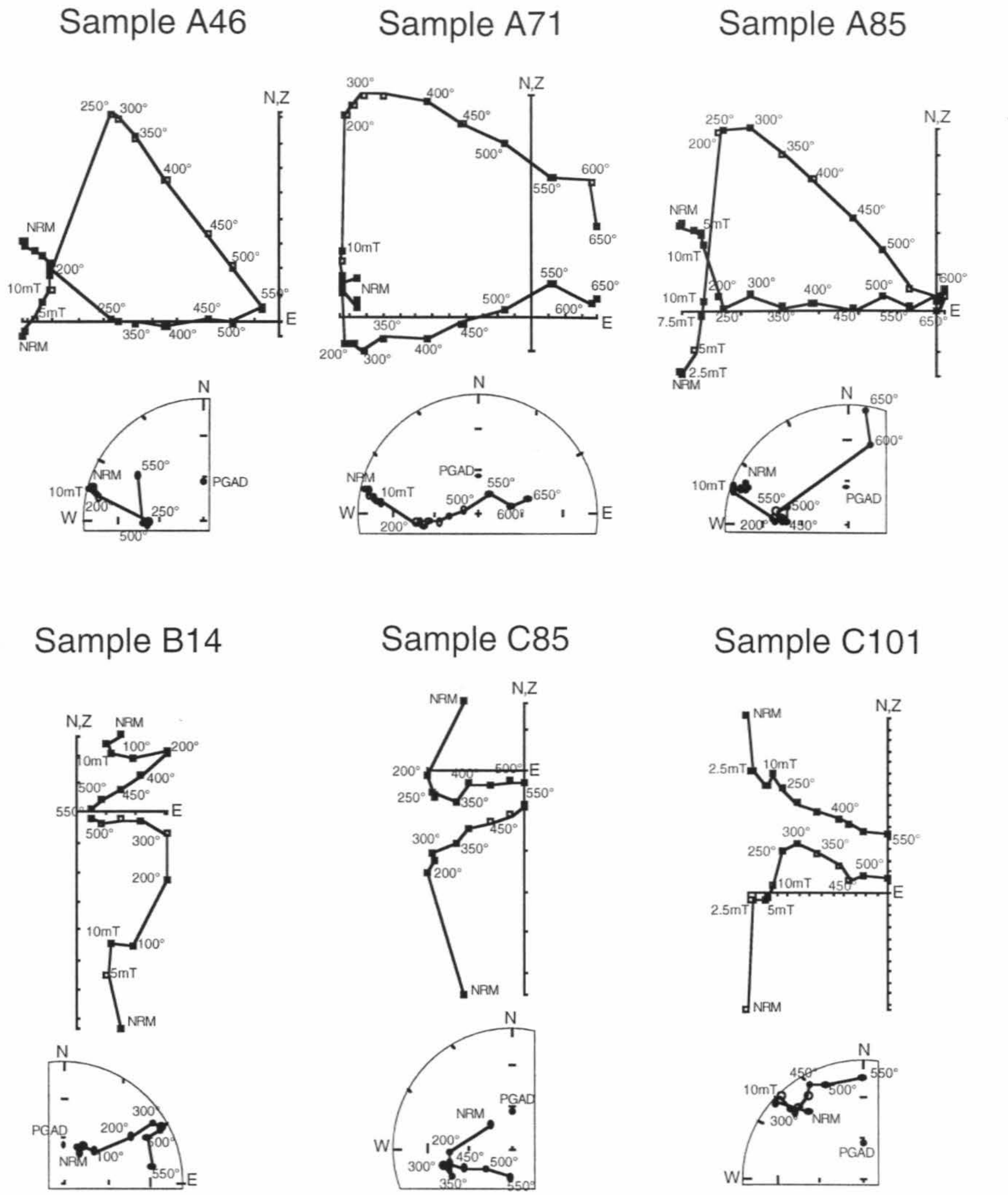


Figure 3. Representative sample demagnetization paths, in present coordinates. For each sample, upper diagram presents superimposed orthogonal projections of sample moment, onto the E-W vertical (open boxes) and horizontal (filled boxes) planes. Temperature steps in °C. NRM = natural remanent magnetization. Tick marks at 10^{-9} Am^2 (10^{-6} emu). Lower diagrams show equal-area projections of the same paths, symbols as in Fig. 4. PGAD = present geocentric axial dipole direction for Salaany Gol.

stable and geologically meaningful component for each sample. All subsequent evaluation of the data concerns ChRMs only.

3. Interpretation of magnetic directions

Twenty-one specimens from a total of 220 were thoroughly remagnetized by the present Earth field; because this direction differs substantially from the other population of ChRM directions, these samples are easily identifiable and will be ignored in the following discussion. ChRM directions from all remaining samples are shown in Fig. 4, in tilt-corrected coordinates. Most show a steep, negatively-inclined (upward), NNW direction in restored coordinates. When viewed in present (*in situ*) coordinates, however, ChRMs from the moderately dipping (profiles A and B) rocks cluster in a significantly distinct direction from ChRMs of the subvertical (profile C) samples (Fig. 5). Thus the component qualitatively passes a fold test, indicating acquisition prior to deformation. Bootstrap statistical analysis (McFadden, 1990) confirms this result (Table 1): parallelism between the two groups in present coordinates can be rejected with >99.9% confidence, whereas the same hypothesis for the tilt-corrected data can only be rejected with <50% confidence. In this case, note that the statistics only reinforce visually obvious results.

Tight folding at Salaany Gol probably occurred during Palaeozoic accretions of the Altaid orogenic collage onto the Siberian platform (Mossakovsky *et al.*, 1992, 1994; Sengör, Natal'lin & Burtman, 1993). The tight, NW-SE fold across which our samples were collected may be cogenetic with a thrust in the eastern part of the range, that is truncated by Palaeozoic granite (Fig. 3 of Khomentovksy & Gibsher, 1996). This granite has been dated as Devonian (Khasin, Borzakovskiy & Zonenshayn, 1973), consistent with subsequent recognition of a regional Devonian magmatic episode (Yashina, Matrenitskiy & Garam, 1979). Therefore, our pre-fold characteristic magnetization is pre-Devonian if our structural correlations and existing geochronological data are correct.

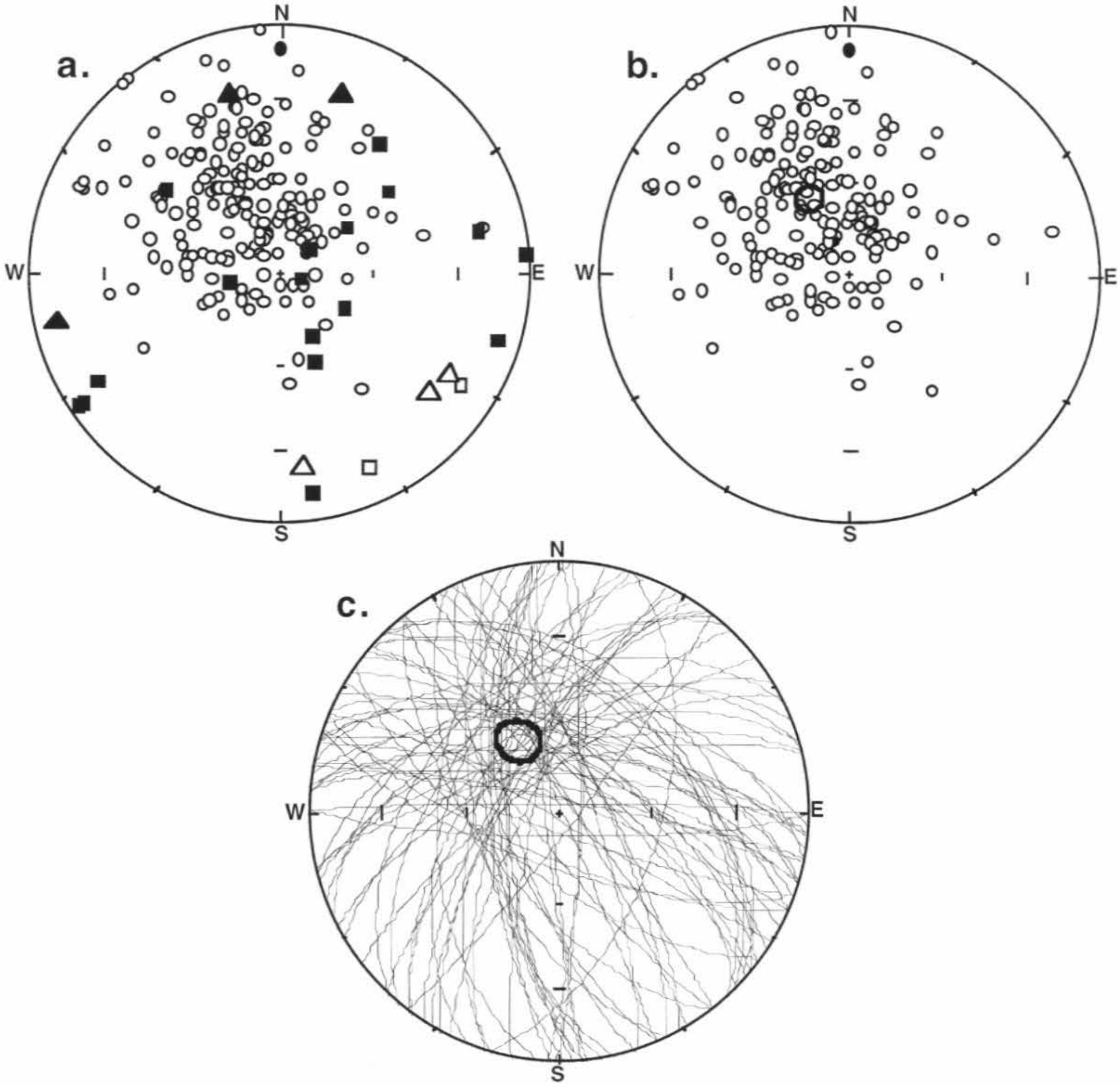


Figure 4. Equal-area projections of ChRM directions in tilt-corrected coordinates. Solid (open) symbols represent vectors projected from the lower (upper) hemisphere of the unit sphere. (a) All non-present-field ChRM directions from Salaany Gol. Squares depict samples with opposite polarity to the modal population. Triangles depict outliers from either polarity. (b) Non-present-field ChRM directions as in (a), but excluding all outliers from the modal population. Dark ellipse is the projection of the 95% confidence cone around the Fisher mean. (c) Least-squares planes from profile C, projected onto the upper hemisphere. Each plane contains a sample's ChRM vector and the origin. Dark ellipse is the projection of the 95% confidence cone around the best-fit common vector among the planes.

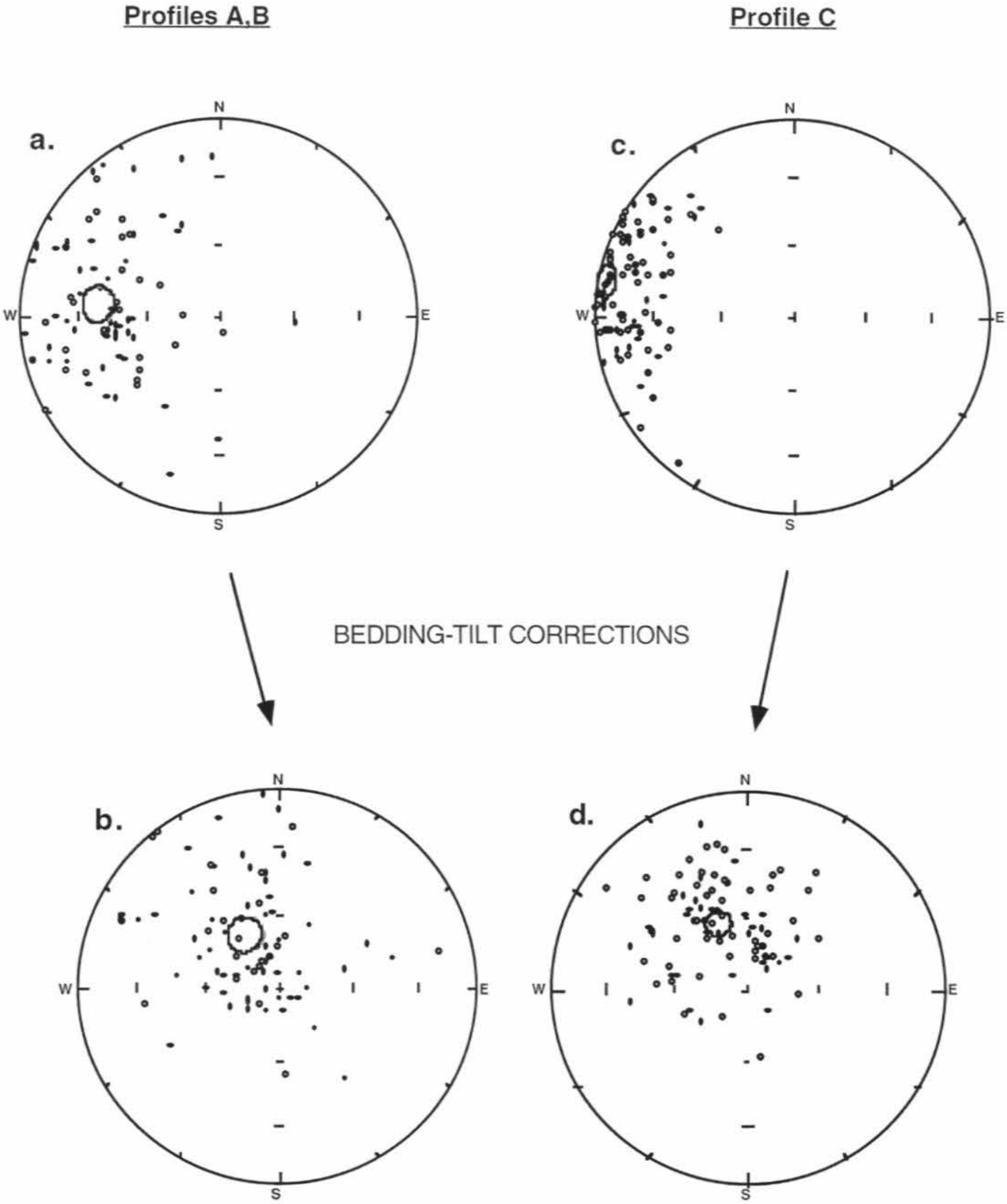


Figure 5. Graphical illustration of the Salaany Gol fold test. Subset of ChRM directions excludes present-field values and distant outliers. Equal-area projections with symbols as in Fig. 4. For statistical data, see Table 1. (a) Samples from profiles A and B in present coordinates. (b) Profile A and B samples in tilt-corrected coordinates. (c) Samples from profile C in present coordinates. (d) Profile C samples in tilt-corrected coordinates. Note close alignment of the two subset means after tilt correction (panels (b) and (d)).

Table 1. Fold test of modal ChRM directions

	Sample profiles A,B				Sample profile C				Test results		
Coordinates	D	I	N	k	D	I	N	k	V_0	n (%)	$V_{(Conf\ \%)}$
Present	275.7	-38.7	84	6.0	280.6	5.0	90	9.3	156	99.9	41
										99	25
Restored	327.4	-64.2	84	5.7	337.3	-60.5	90	9.5	2.8	90	12
										50	3.3
										40	2.4

Bootstrap test from McFadden (1990). For each coordinate system, the two subset distributions are tested for parallelism. V_0 indicates the degree of separation between the groups. $V_{(n\ \%)}$ gives the nth-percentile separation between two groups randomly selected from a common Fisher distribution. If V_0 exceeds $V_{(n\ \%)}$, then parallelism can be rejected with (n %) confidence. Bootstraps were run with 3000 iterations.

Abbreviations: D, I = Fisher mean declination and inclination, k = Fisher precision parameter, N = numper of samples, n = confidence percentile.

Although most specimen ChRM directions lie in the steep negative NNW octant of the unit sphere, several outliers demand attention. Using the discordancy test outlined by Fisher, Lewis & Embleton (1987, p.126) with a 10% probability cutoff, 25 samples are excluded from the modal population. By this filter, 174 samples within ~84° arc length of the mean direction are included in the modal subset. Spurious ChRMs may be explained by unusual occurrences such as close proximity to lightning strikes, local chemical remagnetization, or acquisition in the transitional geomagnetic field during a reversal. Because our samples show predominantly single polarity, however, the latter explanation is less likely than the others. For some outlying samples, discrepancy between magnetic and solar compass measurements on the outcrop suggests they have been remagnetized by lightning strikes.

Extremely outlying samples may record opposite geomagnetic polarity to that of the mode. Consistency of each magnetic polarity among stratigraphically adjacent samples is an important test of the reversal hypothesis (*e.g.*, Kirschvink, 1978). Nineteen samples lie within 84° arc length of the antipole to the mean direction and may be of opposite polarity (Figs. 4, 6). Of these, one (at 258 m) is adjacent to the only wide sample gap, so opposite polarity may continue into that interval. Two are immediately adjacent to each other (~758 m). Others occur in stratigraphic isolation, although several clusters occur within relative proximity (2-8 m; 589-622 m; 755-766 m; 1095-1109 m). These concentrations may represent geomagnetic polarity bias, but the general stratigraphic sporadicity suggests that antipolar samples may be, in fact, members of a superimposed uniform population. Indeed, treating these single-hemispherically distributed outliers as axial data (Fisher, Lewis & Embleton, p.160) shows that we cannot reject uniformity with >90% confidence.

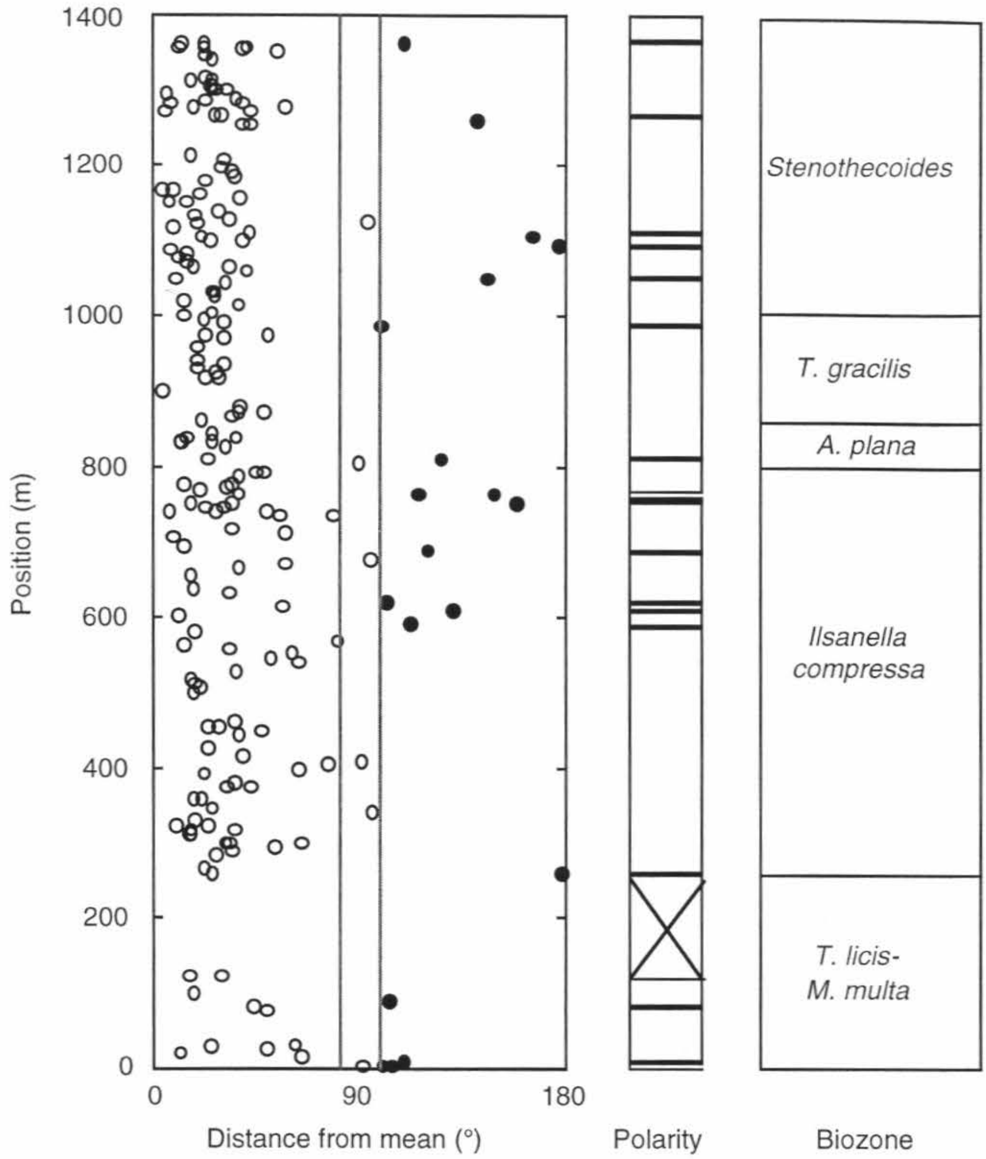


Figure 6. Stratigraphic variation in tilt-corrected ChRM directions, with geomagnetic polarity interpretations. Polarity symbols as in Fig. 2. Arbitrary reversed polarity assigned to majority of samples. Filled circles correspond to samples denoted by squares in Fig. 4.

Because our evidence for geomagnetic polarity reversals is not conclusive, the following discussion entertains not only that possibility, but also the chance that the antipodal ChRMs result from some other sources of error and lie coincidentally opposite the modal ChRM directions.

3.a. Bimodal primary magnetization hypothesis

If the consistent magnetization at Salaany Gol is primary and bipolar, then comparison with magnetostratigraphy of Siberia brings a surprise: there, previous work has shown a major magnetic polarity shift from mainly reversed to predominantly normal bias in the uppermost Tommotian Stage (Kirschvink *et al.* 1991; Fig. 2). The Mongolian data presented here are predominantly single polarity, at odds with the previous correlation by Voronin *et al.* (1982). Small polarity zones in the Siberian section are common enough to allow ready correlation of our widely scattered opposite-polarity intervals in a variety of non-unique alternatives, but these must be validated by bio- and chemo-stratigraphic data.

Figure 7 presents one such model. Brasier, Kuleshov & Zhegallo (1996) describe several cycles of $\delta^{13}C$ within the Bayan Gol Formation, which they claim are removed by the sub-Tommotian unconformity on the southern Siberian platform. Their correlation thus bears no prediction for magnetostratigraphy at Salaany Gol; nonetheless, it demands a rather substantial revision to the original biostratigraphic study (Voronin *et al.*, 1982). On the other hand, the palaeontological correlation by Voronin *et al.* (1982) is fairly consistent with the $\delta^{13}C$ data from the two regions (Siberian cycle IV coeval with Mongolian cycle F), but the magnetostratigraphic records are inconsistent with this interpretation. Therefore, we present a compromise that generally satisfies all three stratigraphies, whereby the entire Bayan Gol Formation spans only early Tommotian time, from the *N. sunnaginicus* through the middle *L. bella* zones. In our correlation, Mongolian cycles E and F (Brasier, Kuleshov & Zhegallo, 1996) correspond to Siberian cycles II and III, respectively.

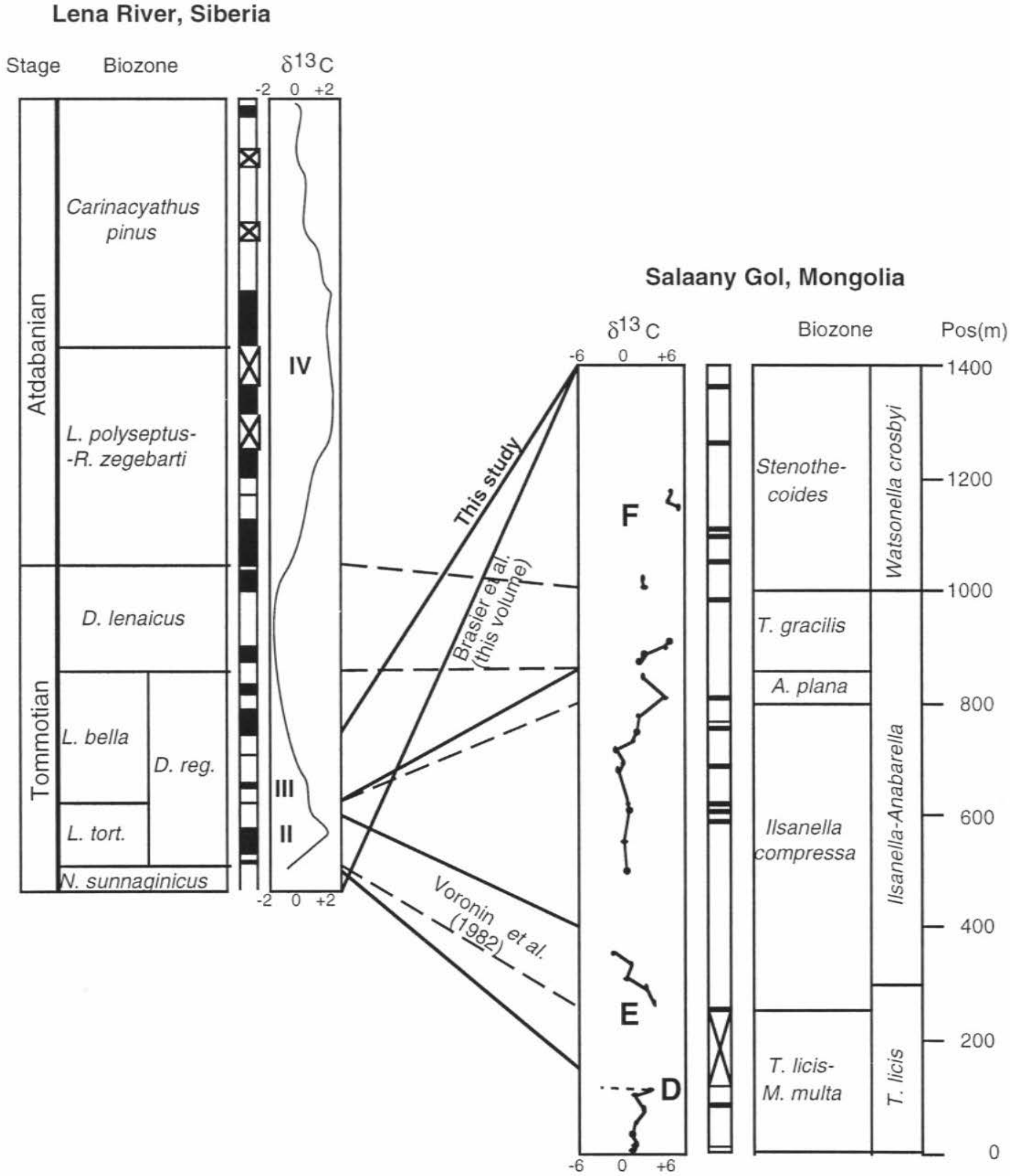


Figure 7. Various correlations of the Salaany Gol sequence with the Vendian-Cambrian of Siberia. Integrated magnetic polarity interpretation (symbols as in Figure 2) and carbon-isotope data for Siberia, from Kirschvink *et al.* (1991). Carbon-isotope data and named cycles for Salaany Gol from Brasier, Kuleshov & Zhegallo (1996); where dashed, from the Bayan Gol locality. Our preferred correlation, based on the assumption of primary detrital remanence with meaningful antipolar directions, is shown by the dark solid lines.

Primary magnetic remanence implies a depositional palaeolatitude of 44 ±6°N (Table 2, first row). By comparison, palaeomagnetic results (Khramov, 1982, p. 302; Kirschvink and Rozanov, 1984; Van der Voo, 1993, p. 100) and archeocyathid bioherms (Kirschvink and Rozanov, 1984) restrict the Siberian platform to tropical palaeolatitudes in Early Cambrian time. Whereas our results reconstruct Mongolia to higher latitudes, thousands of kilometers between a craton and its adjacent back-arc basin is not unreasonable, given, for example, the present width of the Philippine Sea. The palaeotectonic model developed by Mossakovsky *et al.* (1992, 1994) suggests a "Palaeo-Pacific" ocean between the Siberian craton and East Gondwanaland, to which Mongolian terranes primarily belonged. These terranes were rifted from East Gondwanaland, transported, and subsequently accreted to the Siberian craton during Vendian and early Palaeozoic time. Provinciality of Early Cambrian fauna from the Zavkhan basin provides a clue to palaeogeography. The compilation of data from Voronin *et al.* (1982); Voronova *et al.* (1986); Korobov (1989); Wood, Zhuravlev & Chimed Tseren (1993); Esakova and Zhegallo (1996); and Ushatinskaya (1995), shows a distinct trend from Nemakit-Daldynian to Tommotian fauna of affinity with the Yangtze platform, and latest Atdabanian to earliest Botomian fauna of Siberian affinity (A. Zhuravlev, unpub. compilation), supporting the model by Mossakovsky *et al.* (1992, 1994). An alternative model, placing the Tuva-Mongol region, including rocks of the Zavkhan basin, adjacent to Siberia during Vendian-Cambrian time (Sengör, Natal'lin & Burtman, 1993), is inconsistent with this palaeontological compilation.

The Yangtze craton (South China block) was reconstructed to ~5° latitude by palaeomagnetic results from the Precambrian-Cambrian boundary section at Meishucun (Wu, Van der Voo & Liang, 1988/9), but subsequent workers suggested a ~40°S latitude for Sinian-Cambrian placement of South China (Wang, Van der Voo & Wang, 1994). Unfortunately, the paleopole derived in the latter study, which lacks any field stability tests, is similar to an Ordovician pole from South China (Fang, Van der Voo & Liang, 1990) and

Table 2. Fisher statistics of ChRM directions excluding present-field and distant outliers

		Mean direction				Palaeomagnetic pole*		Reliability	
Subset	N	D(°)	I(°)	k	a_{95}	Lat.(°N)	Long.(°E)	1 2 3 4 5 6 7	Q
Including oppositely directed ChRMs†	193	331.9	-62.6	5.8	4.6	4.0	115.6	1 0 1 1 0 1 0	4
Excluding oppositely directed ChRMs	174	333.1	-62.3	7.1	4.3	3.5	114.9	1 0 1 1 0 0 0	3

Abbreviations as in Table 2.1. Reliability criteria from Van der Voo (1990): (1) Age of rock well known, (2) N>24, k 10, and a95 16°, (3) Demagnetization and vector subtraction, (4) Field tests, (5) Tectonic coherence with craton, (6) Reversals, and (7) No resemblance to younger APW path. Q = sum of satisfied criteria 1-7.

* assuming reversed geomagnetic polarity

† reflected through origin

may be a remagnetization of that age. Clearly, the Eocambrian drift history of the Yangtze craton needs to be better understood in order to assess our Mongolian paleomagnetic result in context of the tectonic models by Mossakovsky *et al.* (1992, 1994).

3.b. Coincidental antipolarity, remagnetization hypothesis

Two variations of this conjecture are that magnetization may be either primary or secondary, but in either case it was acquired in a period of single geomagnetic polarity. ChRMs that are directed opposite to the mean direction are due to purely coincidental scatter or error. Final statistical treatment of the data excludes these along with the other widely aberrant directions (Table 2, second row). Fisher analysis gives a palaeomagnetic pole of 4°N,115°E, corresponding to a palaeolatitude of 43±5°N given a reversed field polarity during remanence acquisition.

A single-polarity, primary magnetization is less likely than one of dual polarity, for this requires that <u>all</u> of the normally magnetized Siberian intervals must be excluded from the stratigraphic record at Salaany Gol. More likely, single polarity would be due to a magnetic overprint, whereby no stratigraphic correlations with other Vendian-Cambrian reference sections are permitted. The wide scatter of our ChRMs imply averaging of geomagnetic secular variation, suggesting a stable but heterogeneous chemical remagnetization. Well defined, linear ChRM components (Fig. 3) indicate that the randomly directed component remaining in many samples after 600 °C, does not contribute to the scatter of ChRMs.

Because rocks of the Zavkhan basin are thought to have accreted to the Siberian craton during early Palaeozoic time (Mossakovsky *et al.*, 1992, 1994) and lie north of the projected trace of the Mongol-Okhotsk suture (Gorzhevskiy & Shabalovskiy, 1972), west Mongolian and Siberian APW paths should be coincident from that time to the present, unless post-accretionary vertical axis rotation occurred between the two blocks. Thus, the

Salaany Gol magnetic result should conform with Phanerozoic directions from Siberia, unless our magnetization was attained prior to accretion.

The mean pole position lies nearest to the Silurian portion of Siberia's APW path, although ~60° of post-Palaeozoic clockwise rotation of the Khasagt-Khayr Khan Mountain (KKM) region, relative to Siberia, could have brought Devonian overprints to the observed palaeopole location (Fig. 8). This is in the opposite sense from Permian to Cretaceous anticlockwise rotations of the Tarim block relative to Siberia (Li, 1990; Van der Voo, 1993, p. 207); if the Salaany Gol magnetization is post-Silurian, then subsequent rotations of the KKM region probably occurred independently from Tarim. Pre-Jurassic palaeomagnetic results from eastern Mongolia (Pruner, 1992), across the Mongol-Okhotsk suture, are not applicable to the Zavkhan region.

Alternatively, remagnetization could have occurred prior to Altaid accretion of the Zavkhan basin onto the Siberian craton, possibly during early diagenesis, in which case discrepancies between Siberia's Cambrian APW path and our computed pole would be due to wide separation between the Siberian craton and the Zavkhan basin. In this case similarity between our pole and the middle Palaeozoic Siberian APW path is coincidental. Palaeogeographic implications are similar to those discussed above in section (**3.a**). Rapid closure of Paleo-Pacifica, such that by latest Atdabanian time the Zavkhan basin had entered the Siberian faunal province, was followed by Cambrian-Ordovician accretion of the Zavkhan basin to Siberia during the Salairian orogeny (Mossakovsky *et al.*, 1994). If the folding at Salaany Gol occurred during this accretion, then our positive fold test justifies these well constrained Cambrian palaeogeographic comparisons with the Siberian platform.

4. Tests and summary

The above arguments follow the implications of two alternative hypotheses to explain our palaeomagnetic results from Salaany Gol. These possibilities are testable by further

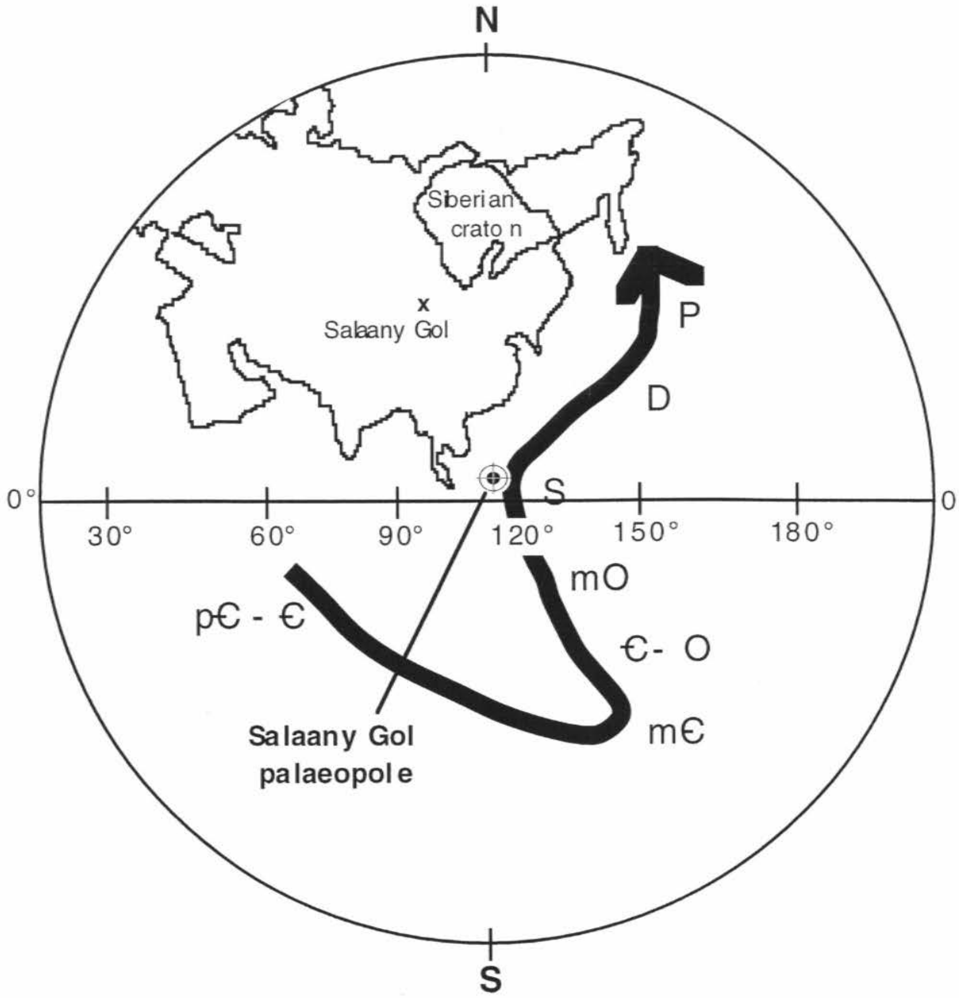


Figure 8. Relation between paleomagnetic pole computed from this study to the Paleozoic apparent polar wander path for the Siberian craton, compiled by Van der Voo (1993, p. 100), except for the Middle Cambrian position according to Khramov (1982, p. 302) and the Precambrian-Cambrian boundary pole from Kirschvink *et al.* (1991). Abbreviations: pC-C = Precambrian-Cambrian boundary, mC = Middle Cambrian, C-O = Cambrian-Ordovician boundary, mO = Middle Ordovician, S = Silurian, D = Devonian, P = Permian.

palaeomagnetic and stratigraphic work. Magnetic conglomerate tests of volcanic clasts in the lower Tsagaan Oloom diamictites (Lindsay *et al.*, 1996) could demonstrate primary remanence of the underlying Dzabkhan Formation and thus low probabilities of regional chemical or thermal remagnetization, or they could demonstrate regional magnetic overprinting. At present sample spacing of ~5 m, the Salaany Gol palaeomagnetic transect may pass over many short-lived magnetic polarity zones; future palaeomagnetic studies of the Bayan Gol and adjacent formations, at <1 m sample spacing in several overlapping stratigraphic sections, should also help test whether our oppositely directed samples truly indicate geomagnetic reversals.

In summary, if the palaeomagnetism recorded at Salaany Gol is primary, then we can interpret not only Cambrian palaeogeography but also magnetostratigraphic comparisons with the Early Cambrian of Siberia. In that case the simplest model that is consistent with the magnetic and carbon-isotope stratigraphy is one whereby the Bayan Gol Formation is equivalent to the lower part of the Tommotian Stage. If the magnetization is an overprint, then it was probably acquired either (1) prior to accretion of the Zavkhan basin with the Siberian craton, implying wide separation between the two regions, or (2) following accretion, during Silurian-Devonian time. In any case the magnetization was acquired at about 44° latitude.

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Chapter 3: Evidence for a large-scale reorganization of Early Cambrian continental masses by inertial interchange true polar wander

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Analysis of Vendian to Cambrian paleomagnetic data shows anomalously fast rotations and latitudinal drift for all of the major continents. These motions are consistent with an Early to Middle Cambrian inertial interchange true polar wander event, during which Earth's lithosphere and mantle rotated about 90 degrees in response to an unstable distribution of the planet's moment of inertia. The proposed event produces a longitudinally constrained Cambrian paleogeography and accounts for rapid rates of continental motion during that time.

The Vendian-Cambrian transition [~600 to 500 million years ago (Ma)] is one of the most intriguing periods in Earth history. Geological evidence points to the breakup of one supercontinent, Rodinia, and the almost simultaneous assembly of another, Gondwanaland (*1*). The sudden appearance of virtually all the animal phyla (*2*) and their exponential diversification are coeval with abrupt shifts in oceanic geochemistry (*3, 4*); recent calibration of this time interval with U-Pb isotopic ages (*5,6*) indicates that these events occurred within a span of 30 million years (My), and the major diversification happened in only 10 to 15 My (Fig. 1). The new ages, along with paleomagnetic data, indicate that continents moved at rapid rates that are difficult to reconcile with our present understanding of mantle dynamics (*7*). We propose that rapid continental motions during the Cambrian Period were driven by an interchange event in Earth's moment of inertia tensor. The age constraints on the geophysical data indicate that the rapid continental motions occurred during the same time interval as the Cambrian evolutionary diversification and therefore the two events may be related.

The most reliable paleomagnetic data for this interval are from magnetostratigraphic studies across well exposed stratigraphic sections that include the Precambrian-Cambrian and Cambrian-Ordovician boundaries (*4, 8-10*). Magnetic polarity patterns from these sections are correlated in different geographic regions, indicating that the rocks have not

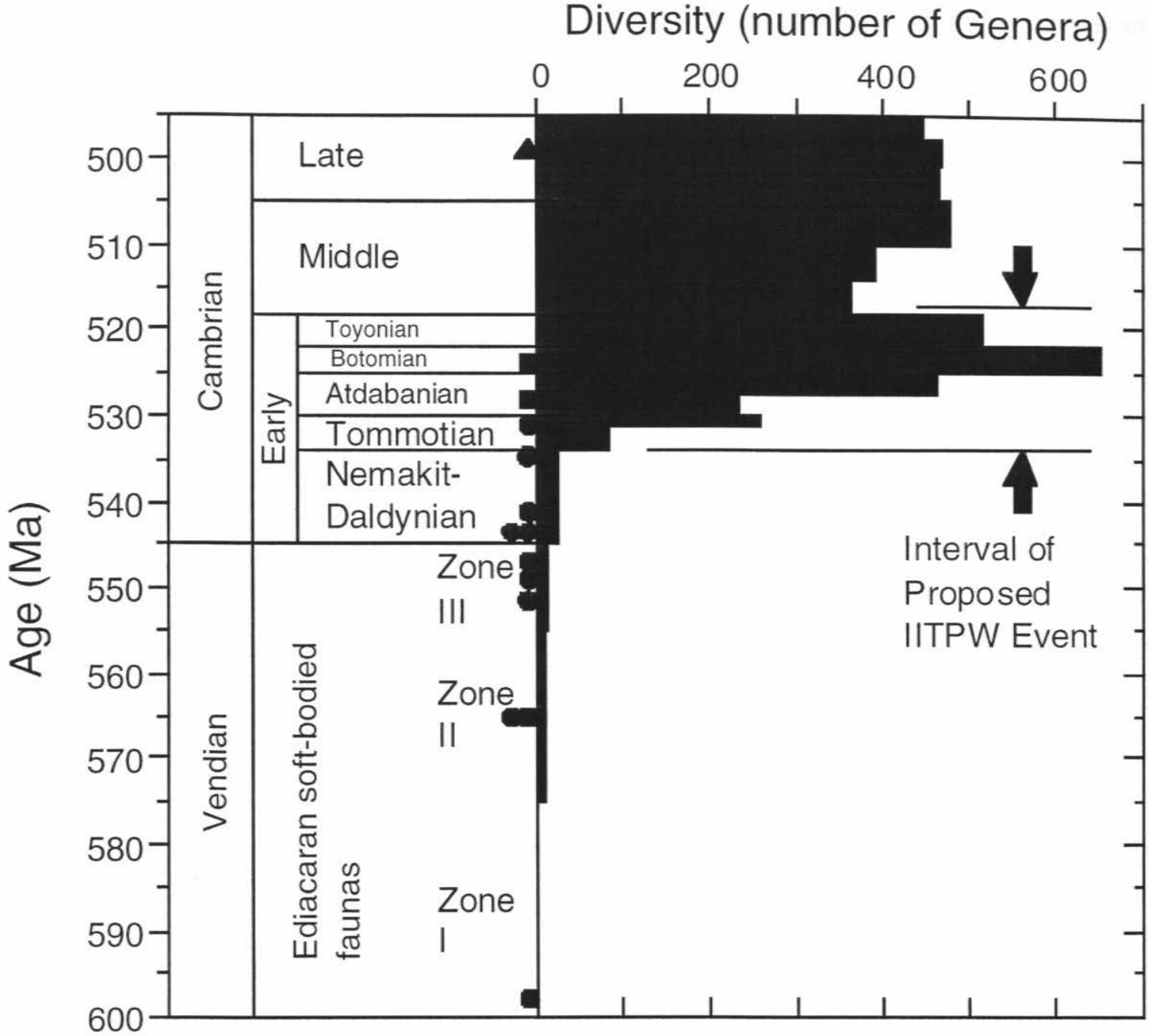

Fig. 1. Isotopic age constraints and generic diversity for Vendian and Cambrian time. The time scale is from Tucker & McKerrow (*5*), with location of the radiometric constraints indicated by solid circles (U-Pb from zircon thermal-ionization mass spectrometry, TIMS), squares (U-Pb zircon measured by sensitive high resolution ion microprobe, SHRIMP), and triangles (Rb-Sr shale isochron). SHRIMP ages have been corrected for a 1 to 2% discrepancy between the TIMS and SHRIMP data (*5*). Data for the generic diversity (including archaeocyathids) (*50*) are plotted simply as the number of taxa reported in each reference time interval. The black arrows indicate the approximate time of the proposed IITPW event.

suffered remagnetization. The strata can be correlated by biostratigraphy, chemostratigraphy, and sequence stratigraphy, usually with a temporal precision better than 1 My. Magnetostratigraphic studies from sedimentary rocks typically span time intervals long enough to average out secular variations in the geomagnetic field. Through these studies, the paleohorizontal is defined by the bedding and sedimentary structures, and these data allow us to determine Earth's paleorotation axis (*11*). We used biostratigraphically dated units with correlatable magnetic polarity patterns to derive our apparent polar wander (APW) paths, and supplement these with the most reliable poles from other sources (*12*). Averaging pole positions over time intervals of 15 to 20 My as done by some authors (*13*) can mask rapid shifts in apparent pole positions, even among sediment-based APW paths. We therefore did not smooth the data.

Australia has the best constrained APW path for Vendian and Cambrian time (Fig. 2). Three key paleomagnetic poles are ranked with the highest reliability (Q = 7 out of a possible 7) on the paleomagnetic quality index (*12*). The oldest of these is from the basal Vendian Elatina Formation of South Australia. It was deposited during the upper Marinoan glaciation at about 600 Ma. The two-polarity remanent magnetization is carried by detrital hematite with Curie temperatures of 680°C, whereas the maximum burial temperature was less than 160°C. The magnetic directions also pass slump-fold and reversal tests, and reversals are stratigraphically bounded and consistent between parallel sections (*14*). Close to this pole is a group of three others from the Pertatataka, Arumbera, and Todd River formations in central Australia *(8, 15)*. The Pertatataka and Arumbera formations contain a late-stage Ediacaran fauna, implying an age of about 560 to 545 Ma. Unconformably overlying sedimentary rocks contain some of the earliest Cambrian fossil tracks and trails. The Todd River formation, one of the top units, contains archaeocyathids and small skeletal fossils of Atdabanian affinity. Dense paleomagnetic sampling ends in sedimentary rocks correlated with the earliest Tommotian *(16)*. In the calibrated time scale (*5, 6*)(Fig. 1), these three poles span the interval of about 560 to 534 Ma. The characteristic

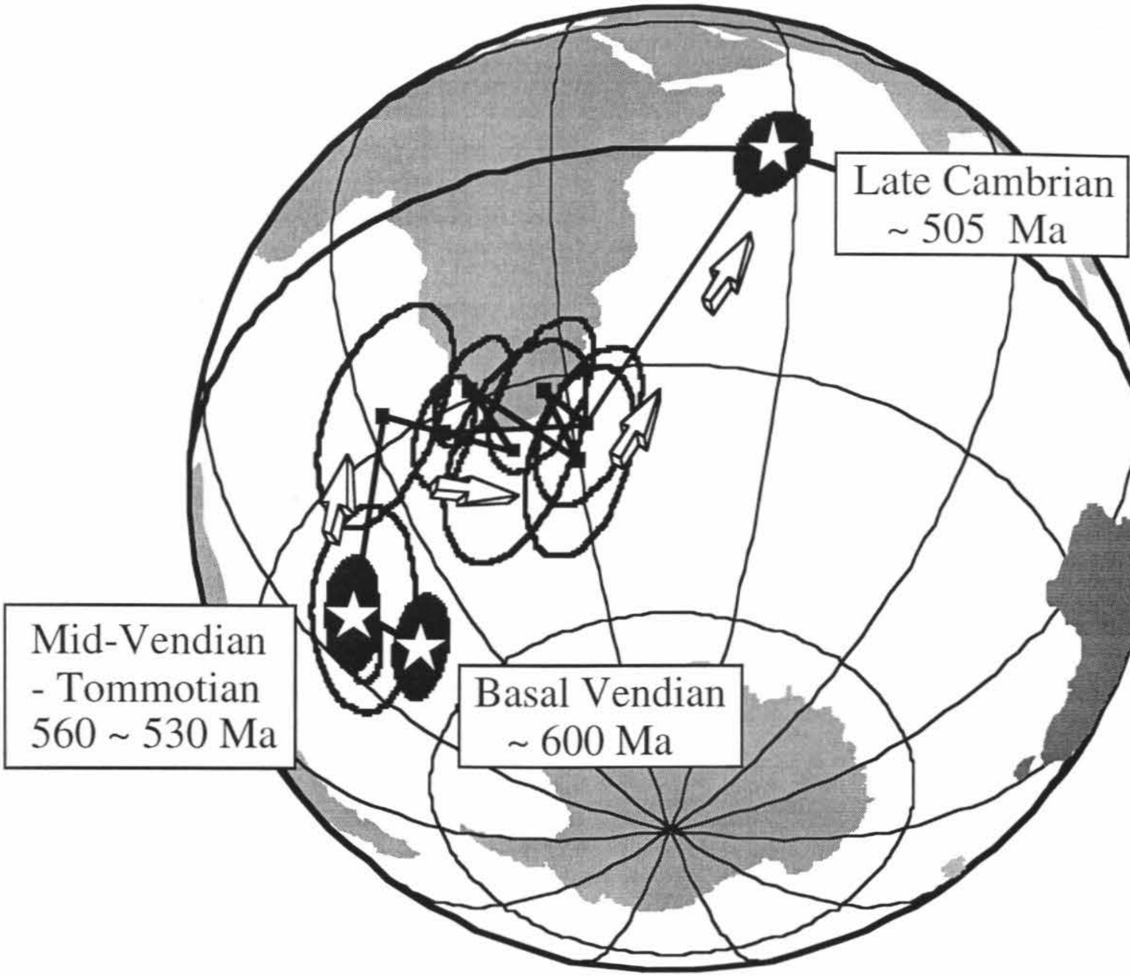

Fig. 2. Paleomagnetic apparent South polar wander curve from the cratonic areas of Australia for Vendian and Cambrian time. Ovals show the boundary of the cones of 95% confidence for the true mean directions; those containing stars are particularly well constrained both in timing and paleomagnetic reliability. Latitude and longitude grid lines are spaced 30° apart, from which it can be seen that the angular distance from the Late Cambrian direction to the cluster of Vendian to Tommotian poles is, within the error limits, 90°. Arrows show the direction of inferred pole motions, based on (*18*). These directions are interpreted to be South poles, based on the Phanerozoic APW path for Australia.

magnetic directions for these poles pass the fold and unconformity tests, and the polarity patterns can be correlated at several other sections within the Amadeus Basin and in the neighboring Ngalia and Georgina basins *(8, 15)*. These data show no detectable motion for Australia in the 65 My before 535 Ma.

A magnetostratigraphic study of the Australian Cambrian-Ordovician boundary interval (~500 to 495 Ma) from Black Mountain in western Queensland (*10*) produced a detailed magnetic polarity profile. The polarity zones were tied to the conodont and carbon-isotope ($\delta^{13}C$) stratigraphies and match the polarity, biostratigraphic, and carbon-isotopic variations in rocks from North China (*17*), indicating a two-polarity primary magnetization. The pole for this sequence is >80° away from the Vendian-Early Cambrian poles, implying that Australia underwent a large rotation while remaining near the equator sometime between Tommotian and Late Cambrian time (Fig. 2). On the basis of the recent age determinations (Fig. 1), this motion started no earlier than 535 Ma and was completed sometime before ~505 Ma. Although not based on magnetostratigraphy, other studies of Lower Cambrian rocks in central and southern Australia (*18*) demonstrate that Australia rotated counterclockwise during this time.

Other parts of the Gondwanaland supercontinent must have been involved in this tectonic rotation; deformation within the Pan-African and Brasiliano orogenic belts occurred mainly from 700 to 550 Ma (*19*), suggesting that Antarctica, India, Africa, South America, and perhaps parts of East Asia also rotated with Australia; this is supported by paleomagnetic data from these blocks (*12, 20*). In addition, lithofacies in the West African and nearby cratons changed during Cambrian time as the continent moved from lower to higher latitudes (*21*). The Australian and Gondwanaland paleomagnetic data imply that at least half of the Earth's continental lithosphere rotated nearly 90° between 534 and 505 Ma.

The paleocontinent of Laurentia (which included most of North America and Greenland) has the next most reliable set of paleomagnetic data in the Vendian-to-Cambrian interval. There are more than 30 published pole positions for this interval that vary in quality and age control. Several authors have reviewed these results (*12*, *13*) (Fig. 3). Three of the poles, all from igneous intrusive complexes, fall within an overlapping group labeled Vendian to earliest Cambrian (575 to 540 Ma). The Catoctin A pole is constrained in age at 564 ± 9 (U-Pb) (*22*); the Callander Complex yields Pb/Pb and K-Ar mineral isochron ages of 575 ± 5 Ma; and the tilt-corrected Sept-Iles B pole has its best constrained 8-point Rb-Sr isochron at 540 ±7 Ma (*23*). This group of Vendian- to earliest Cambrian-aged poles is separated by nearly 70° from a cluster of biostratigraphically dated Middle to Late Cambrian Laurentian poles, all with $Q \geq 5$ (Fig. 3). In order of decreasing age, these include the early Middle Cambrian Tapeats Sandstone, the Dresbachian-aged Taum Sauks Limestone, the Franconian-aged Moore Hollow and Morgan Creek poles, the Franconian-Early Trempealeauan Point Peak Shale and Royer Dolomite, and the earliest Ordovician Oneota Dolomite (*13*). We also show in Fig. 3 the pole from the Puerto Blanco volcanic rocks, for which we now have reproduced a positive conglomerate test *(24)*. Biostratigraphic constraints indicate that the Puerto Blanco volcanic rocks are of Atdabanian or early Botomian age (*25*), which agrees with the location of this pole between the Vendian and Middle-Late Cambrian pole clusters (Fig. 3). Laurentia's APW path suggests that the continent drifted rapidly from polar to tropical latitudes during Early and Middle Cambrian time, with the proto-Cordilleran margin leading and the proto-Appalachian margin trailing. This motion is consistent with the first appearance of abundant Early Cambrian carbonate rocks, where Atdabanian archaeocyathids first appeared in the Cordilleran miogeocline, followed by younger Botomian and Toyonian forms on the Iapetus shelf (*26*).

Comparison of the Australian and Laurentian APW paths demonstrates that nearly two-thirds of Earth's continental lithosphere experienced similar periods of stasis and motion between 600 and 500 Ma; both had little or no motion during Vendian time, and both shifted abruptly during a short time in the Cambrian. Although two fundamentally different paleogeographies exist for the width of the Cambrian Iapetus ocean between Laurentia and Gondwanaland (*27, 28*), even the narrow-Iapetus hypothesis includes

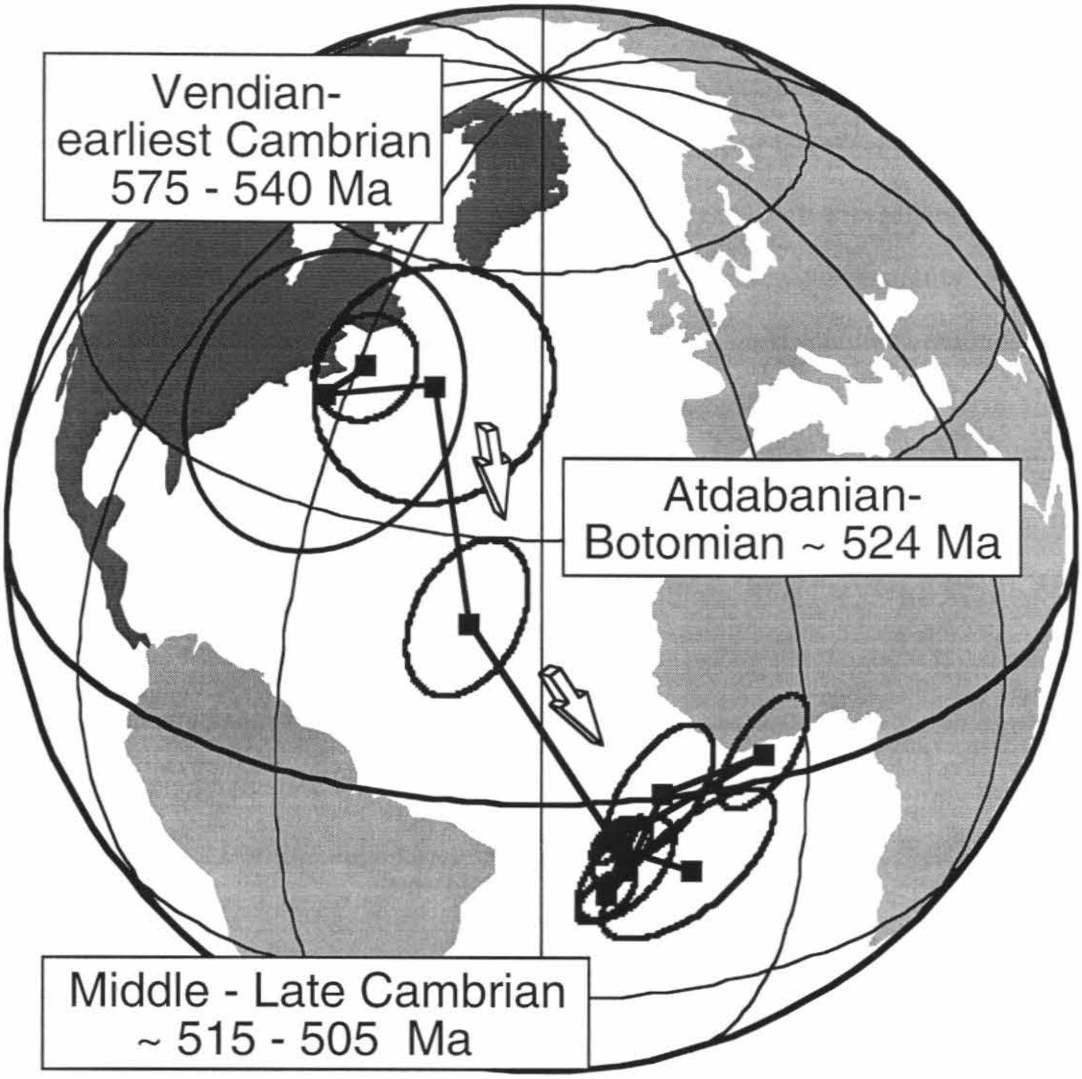

Fig. 3. Apparent South polar wander path for North America. Symbols are the same as in Fig. 2. Data were compiled from (*13*) with results of Q > 4. Vendian poles include the Callander complex, Catoctin, and tilt-corrected Sept-Iles [Laurentian poles 54, 55, and 56 of (*13*)], the Atdabanian-Botomian pole is from the Puerto Blanco Volcanics (*24*), and the Middle to Late Cambrian cluster includes the Tapeats, Taum Sauk, Moore Hollow, Morgan Creek, Royer, Point Peak, and Oneota poles [Laurentian poles 49,44,45,41,43,47, and 40 of (*13*)]. Only the pole from the Buckingham Latites [pole 50 of (*13*), rated a 'Q' of 5] is not used; the contact test only constrains the age to be older than Early Ordovician, and the pole falls in the well defined post-IITPW group.

separation of the two continents onto different plates based on Vendian to Cambrian rift-to-drift sedimentation on the proto-Appalachian margin of Laurentia (*29*). Thus, at least two tectonic plates, involving over two-thirds of the Earth's continental lithosphere, were involved in a rapid rotation of ~90° relative to the spin axis. We speculate that the entire lithosphere may have been included in this rotation. The postulated event began sometime in the first half of the Early Cambrian and ended by earliest Middle Cambrian time, for a total duration of ~15 My. These time constraints are as follows: (*i*) The pole from the upper Arumbera-Todd River formation, which dates from the latest Nemakit-Daldynian or earliest Tommotian stages, plots with the Vendian group, whereas the slightly younger Early Cambrian poles of Klootwijk (*18*) show a shift toward the Black Mountain direction. (*ii*) The pole from the Atdabanian-Botomian Puerto Blanco Formation (*24*) is midway between the Vendian and Middle Cambrian groups. And (*iii*), the pole from the earliest Middle Cambrian Tapeats Sandstone (*30*) lies adjacent to the Late Cambrian poles from Laurentia. Consistent magnetic polarity zones from two sections of the lowermost Tapeats Sandstone suggest that it has a primary magnetic remanence, implying that the polar shift was finished by the early Middle Cambrian *Glossopleura-Alokistocare* trilobite zone.

True polar wander (TPW) is the process through which quasi-rigid spheroids align their maximum moments of inertia with the spin axis, pushing positive mass anomalies toward the rotational equator (*31*). Normal TPW on Mars, the moon, and Earth results when small redistributions of mass lead to excursions in the location of the axis of maximum moment of inertia. The Tharsis volcanic province on Mars, with the largest positive gravity anomaly known from any planet, appears to have reoriented the martian lithosphere to place Tharsis on the equator; similarly, the lunar mascons all lie facing Earth symmetrically about its equator (*32*). Studies of hotspot tracks, paleomagnetic data, and post-mid Cretaceous plate motions on Earth suggest that TPW velocities of up to 0.5°/My were due to small changes in crustal mass anomalies, most likely associated with growth and variation in the size of subduction zones (*33*) and glacial ice volume (*34*). A double burst of TPW has also been suggested for Silurian time, based on matching similar bends on the Gondwanaland and Laurussian APW paths (*35*). Hence, the theory of TPW has observational support from at least three terrestrial planets, including Earth.

A variant of this mechanism, inertial interchange true polar wander (IITPW), involves discrete bursts of TPW of up to 90° in geologically short intervals of time if the magnitudes of the intermediate (I_{int}) and maximum (I_{max}) moments of inertia cross (*36*). This would result in a rapid movement away from the spin axis by the geographic location of the former pole, with rotation of the entire solid Earth centered about the minimum moment of inertia (I_{min}) located on the equator. Because I_{max}, I_{int}, and I_{min} are orthogonal, the simple interchange case (with no independent plate motions) yields a 90° shift. Although such an event has not yet been recognized in the geologic record, the geodynamic consequences of an inertial interchange event have been considered in qualitative terms (*36*). These analyses predict that the 90° rotation of an IITPW event should happen over an interval of ~10 to 15 My.

One way to test the hypothesized occurrence of an IITPW event is to see if it predicts an internally consistent paleogeography, with a rotation about an equatorial axis. Conventional paleomagnetic analyses provide an estimate of paleolatitude of a continent and its orientation with respect to the geomagnetic pole, but the paleolongitude cannot be constrained. During a large and rapid TPW event, however, motion of the geomagnetic dipole relative to the lithosphere should appear to have a common swath from all continents. It can therefore be used to predict an absolute reconstruction, including latitude and relative longitude. Variations due to the much slower normal plate motions would tend to produce only small changes in relative paleogeography at either end of the IITPW event. The IITPW reconstructions (Fig. 4) force the Appalachian margin of Laurentia to lie adjacent to the Andean margin of South America, supporting the narrow-Iapetus model of a consanguineous Appalachian-Andean mountain system during Ordovician time (*28, 29*).

Baltica has a well defined APW path back through Ordovician time (*37*), implying a confidence in the polarity interpretation of Ordovician poles. The Baltica pole closest in age to the termination of the IITPW event is from the lower Ordovician (Arenig-Llanvirn) Swedish Limestones *(37)*. The best pole at the older end of the IITPW event is from the Fen complex, which is dated at 578 ± 24 Ma by Rb-Sr on phlogopite (*38*). The angular gap between these directions is either 82° or 98°, depending upon the polarity of the Fen pole. Ordinarily, either polarity interpretation would be possible without reliable intermediate directions; however, the IITPW paleogeographic analysis indicates that one choice is impossible: the polarity interpretation of Torsvik *et al.* (*37*), with the South pole at 50°N, 144°E, places Baltica on top of the African and Indian portions of Gondwanaland. Reversing this polarity interpretation constrains Baltica's Vendian position to lie near Siberia and Arabia (Fig. 4). This location is similar with past paleogeographic models for the Ordovician that used faunal distributions for approximate paleolongitude control (*13, 27, 39*).

Paleomagnetic data for Siberia are more problematic. Most Late Cambrian and earliest Ordovician results are not documented in enough detail to assess their reliability. However, two magnetostratigraphic studies of Upper Cambrian and Ordovician sedimentary rocks on the Siberian Platform provide independent constraints on the APW path (*40*). The Ordovician magnetic reversal patterns agree with results compiled from other cratons, providing a constraint on the polarity interpretation. Although there are no reliable paleomagnetic results of late Vendian age, the magnetostratigraphic data from the Tommotian Pestrotsvet Formation of the Lena River *(9)* are only slightly younger than the start of the inferred IITPW event. The magnetic polarity interpretation for these directions is constrained by paleontological and stable carbon-isotopic correlations with the Série Lie de Vin in the Moroccan Anti-Atlas Mountains (*4*), which is in turn referenced to the Gondwanaland and Australian APW paths. These two poles, with their stated polarity interpretation, are separated by about 68°; together they yield a plausible tropical position for

Siberia, nearly on the opposite side of the globe from Australia, in agreement with some paleogeographic reconstructions (*13, 27*). On the other hand, if the Ordovician result were of opposite polarity, Siberia would be positioned adjacent to the Pacific coast of Australia, as suggested by other reconstructions (*39*).

Our reconstructions (Fig. 4) are a first-order analysis of a possible IITPW event, limited in several ways by the paleomagnetic and tectonic database. First, the paleomagnetic directions are each only accurate to within about 5°, depending upon the pole. Because each Euler pole (Table 1) depends upon two paleomagnetic pole determinations, positional differences of less than 10° (~1100 km) cannot be resolved. Second, no continent has reliable, biostratigraphically dated poles at both the precise beginning and end of the inferred IITPW event. Normal plate motions before and after the IITPW event will therefore modify the apparent positional change. Third, there are still ambiguities in the reconstructions used for Gondwanaland. These lead to angular differences between East and West Gondwanaland of as much as 20° (*39*), particularly when extrapolating from Australian data to the positions of northern Africa and South America. Although we maintain that between-plate motions should have been small relative to the rapid rotation associated with an IITPW event, we have allowed for these between-plate motions because of our acceptance of APW paths at face value. We view these plate motions as approximate, and emphasize only the >1000-km scale features of these reconstructions. We attribute the apparent conjunction and overlap between Laurentia and Gondwanaland by Cambrian-Ordovician time (Fig. 4B) as an artifact of the several sources of error described above.

Relative to the geomagnetic reference frame, continental drift velocities of ≥30 cm/year are suggested by the well constrained paleomagnetic data from Laurentia and Gondwanaland. This is greater than any plate velocity, continental or oceanic, observed for the past 200 My. Such high velocities of continents during the Vendian-Cambrian interval led some to question the applicability of a maximum rate of plate tectonic motions; still, conventional geodynamic mechanisms must be stretched to account for the fast

Fig. 4 (next page). Paleogeographic reconstructions before (**A**) and after (**B**) the proposed IITPW event, according to the Euler poles in Table 1. The pre-IITPW reconstruction in (A) was made by first rotating the four major continents (Gondwanaland, Laurentia, Baltica, and Siberia) such that their pre-IITPW poles (Vendian-earliest Cambrian) coincide at the South pole, and then rotating each of them laterally about the rotation axis such that their post-IITPW poles (latest Cambrian to early Ordovician) lie on the same longitude. The Late Cambrian reconstruction in (B) was done in a similar fashion, but by reversing the pole groups. For each reconstruction, one group of poles constrains the latitude and orientation with respect to the paleomagnetic pole, while the other constrains relative paleolongitude. Tick marks from the present latitude and longitude grid on the continental fragments are shown at 5° intervals, with 30° intervals for the Cambrian reference latitude and longitude grid (Galls Mercator and equal-area South polar projections). The large plus (+) symbols in Antarctica and Siberia on the diagrams indicate the axis of the minimum moment of inertia, I_{min}, about which Earth's entire lithosphere rotated during the proposed IITPW event. Large arrows in (A) indicate the direction of motion around I_{min} during this event. Color-coding is based on the location of Cambrian terranes within the present continents. As discussed in the text, uncertainties in these reconstructions are at least ~10°; hence the apparent collision shown between Laurentia and Gondwanaland, contrary to geological evidence supporting Cambrian divergence between those two continents *(29)*, is viewed as an artifact of an incomplete dataset.

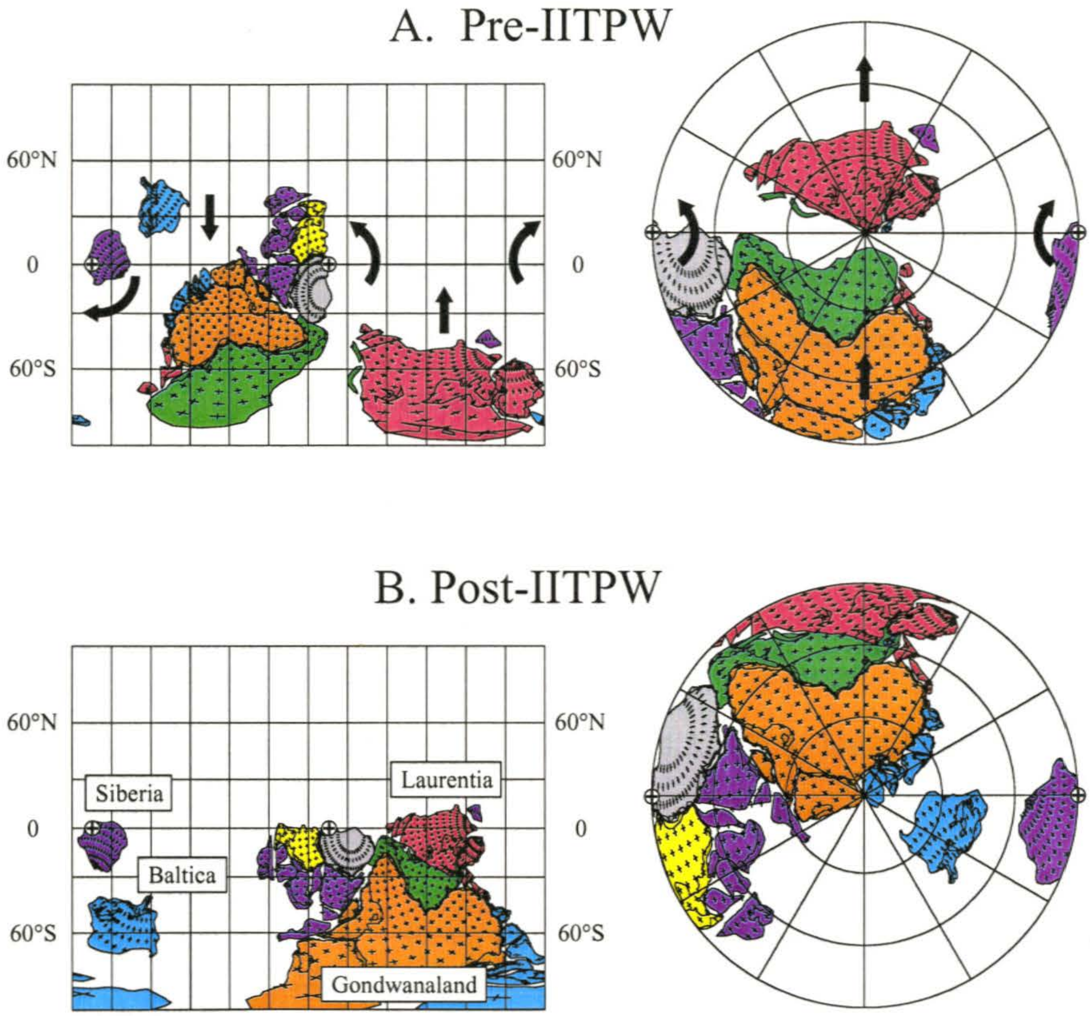
A. Pre-IITPW
60°N
0
60°S
60°N
0
60°S
B. Post-IITPW
60°N
0
60°S
60°N
0
60°S
Siberia
Baltica
Laurentia
Gondwanaland

Table 1. Euler Poles used for producing the absolute IITPW paleogeographic reconstructions shown in Fig. 4. Positions for the paleomagnetic South poles (references in parentheses) are given in order of (north latitude, east longitude), and the Euler poles for moving the continents from their present coordinates to the reconstructed positions are (north latitude, east longitude, CCW angle). Euler poles used for the continental reconstructions of Gondwanaland and small pieces of Laurentia are as given in Kirschvink (*39*), except for the fit between Great Britain and northern Africa; we use a GB-AF pole of (34.2, 15.9, 51.6) to position this further west along the north African margin as indicated by Torsvik *et al.(13)*. Note that the post-IITPW poles provide the relative paleolongitude constraints for the pre-IITPW reconstructions, and vice versa. SS, sandstone; LS, limestone; Fm, formation.

Continent	Pre-IITPW Pole	Post-IITPW Pole	Pre-IITPW Euler Pole	Post-IITPW Euler Pole
Gondwana-land	Arumbera SS(*15*) (-46.6, 337.4)	Black Mountain(*10*) (03.3, 054.0)	49.4, 219.7, -69.3	32.2, 185.9, 118.5
Laurentia	Callander complex, (*48*) (46.3, 301.4)	Tapeats SS,(*30*) (-04.9, 337.7)	14.3, 172.0,-146.5	24.9, 092.9, 96.4
Baltica	Fen complex, (*49*) (-50.2, 324.4)	Swedish LS(*37*) (18.0, 045.8)	40.7, 216.2, -53.4	20.1, 183.0, 132.9
Siberia	Pestrotsvet Fm(*9*) (16.6, 244.5)	Moyero River(*40*) (36.6, 318.6)	20.1, 003.9, 117.3	24.2, 112.0, 156.8

motions *(7)*. If the velocities are due to TPW, however, such geodynamical considerations are obviated because the entire mantle would have rotated along with the lithosphere. Our TPW hypothesis also accounts for the otherwise problematic absence of an orogenic belt on the leading (proto-Cordilleran) margin of Laurentia during its rapid northward motion (*29*).

A ~90° lithospheric and mantle rotation associated with the inferred IITPW event would have rearranged the continents relative to global physiochemical systems, and especially oceanic circulation. Even small changes in circulation can have major effects upon the dynamics of the normal equator-to-pole heat transport; for example, the Pliocene emergence of Panama caused the northward deflection of the Gulf Stream, a second-order global feature but a first-order regional feature of North Atlantic climate *(41)*. Wholesale rotation of the lithosphere is likely to have effected major changes in oceanic circulation and hence, regional climates.

Evidence from carbon isotopes supports the hypothesis that marine circulation was reorganized repeatedly during Early Cambrian time. Pre-Tommotian oceans probably had fairly stable circulation patterns, as evidenced by a steady rise in marine $\delta^{13}C$ (*3, 4*), which imply the sequestration of isotopically light organic carbon either in the deep ocean or in sedimentary reservoirs. Peak $\delta^{13}C$ values were reached immediately before the base of the Tommotian, heralding an interval of about 10 large oscillations of up to 6 parts per mil in $\delta^{13}C$ starting in the late Nemakit-Daldynian, with marine $\delta^{13}C$ values finally stabilizing in the late Botomian. As the positive swings of these oscillations are associated with the first appearance and radiation of new faunal groups, they were first thought to be due to enlargement of the biosphere and the corresponding depletion of ^{12}C from the oceanic reservoir. However, large and rapid swings in $\delta^{13}C$ also could be produced by deep ocean ventilation produced by circulation jumping successively from one quasi-stable pattern to another, coupled by the locally variable sea-level swings predicted for a large TPW event.

Pronounced shifts in seawater Sr isotopic ratios during the Early Cambrian (*42*) provide additional evidence of large-scale changes in the global oceans and related physiochemical systems, including continental weathering and mid-ocean ridge activity.

The repeated reorganizations in global climatic patterns during an IITPW event should have fragmented any large-scale ecosystems that were established, generating smaller, more isolated populations and leading to a higher evolutionary branching rate among existing groups (*43,44*). Comparison of evolutionary rates between Cambrian and late Paleozoic time (*45*), corrected for the new time scale(*5, 6*), indicates that the Cambrian rate was a factor of ~20 times higher. A period of accelerated speciation during the 10- to 15-My interval of the proposed IITPW event could possibly resolve disagreements between estimates of the divergence time for the animal phyla based on molecular (*46*) versus paleontological (*47)* evidence. We suggest, therefore, that the IITPW hypothesis can explain many of the anomalous tectonic, geochemical, and biological trends that occurred during the Precambrian-Cambrian transition.

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Appendix to Chapter 3: Polar wander and the Cambrian; Response to technical comment by Torsvik et al.

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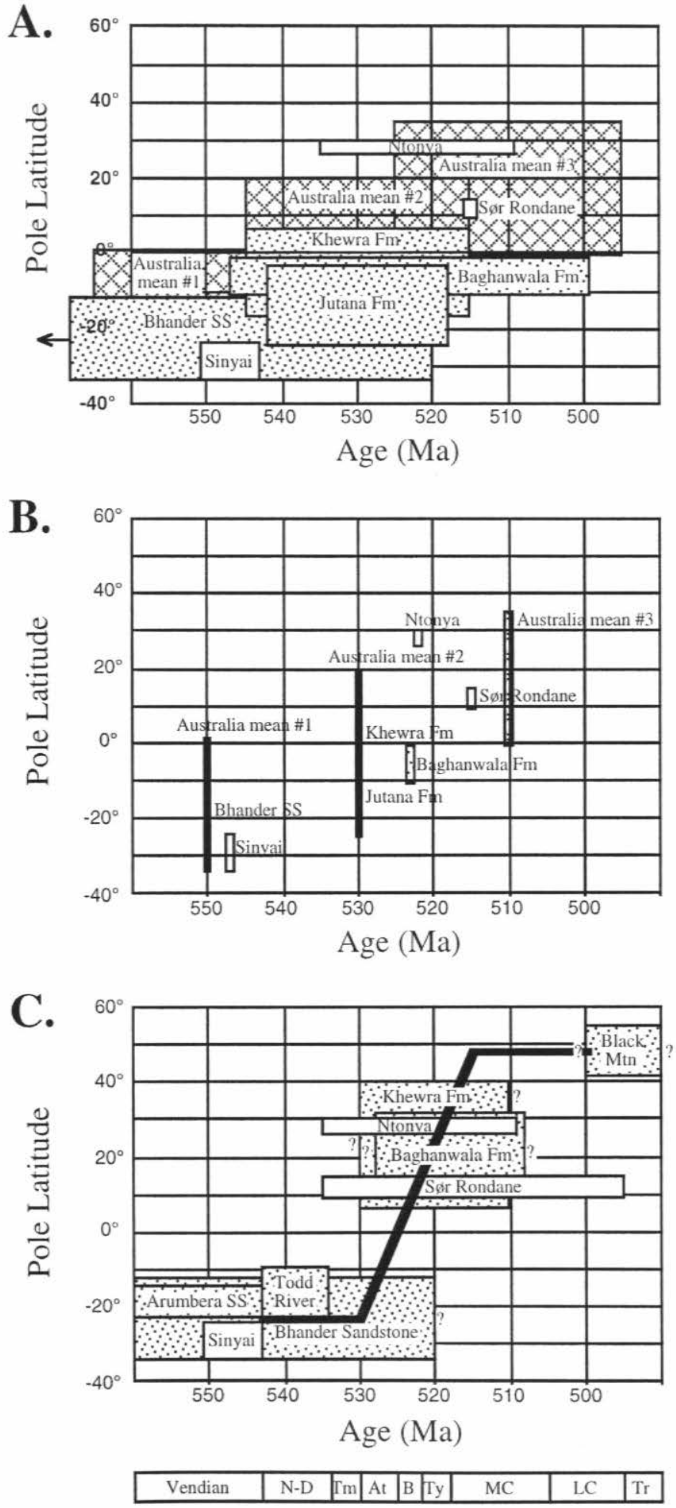
Torsvik et al. question the reliability of the paleomagnetic data used in our report (1). In discussing the Sept Iles B pole (2), which they and other workers have previously treated as one of the most reliable Vendian pole readings from Laurentia (3, 4), they disagree with our recapitulation of its 540-Ma Rb-Sr age (1). The variance in Rb-Sr ages for the Sept Iles complex was discussed thoroughly by Higgins and Doig (5), who developed a geochronologic-petrologic model of the complex that showed emplacement occurring at about 540 Ma. Their well-defined isochron (5) seems perferable to an assignment of 575 Ma that is based on comparison of a paleomagnetic pole with the better-dated Callander complex (3,4); this latter approach seems circular. Contrary to the statement by Torsvik et al. that our "interpretation of Laurentian TPW ... relies exclusively on the contentious Sept Iles Complex result," we note that the angular separation of ~70° between Vendian and Middle-Late Cambrian poles remains the same, whether or not the Sept Iles pole is included [figure 3 of (1)]. Its only relevance to the TPW hypothesis is that it could provide an earliest Cambrian maximum age limit for Laurentia's motion from polar latitudes to the tropics. An older age for the Sept Iles intrusion could relax the constraint on the timing of TPW initiation, but results from earliest Cambrian strata in the Mackenzie Mountains (6) also support the Laurentian APW path that we presented (1). Torsvik et al. also state incorrectly that "a polarity switch of a Vendian pole [for Baltica] ... would have no bearing on Cambrian TPW." As pointed out by Kirschvink et al., the new polarity interpretation of Baltica's Vendian paleopole [permissible, given the lack of Cambrian data from Baltica (1, 3)] generates strict paleogeographic implications and tests for the TPW hypothesis. Regarding the Siberian database, the result of Kirschvink and Rozanov (7) is currently the most reliable paleomagnetic study of the Cambrian Siberian craton: magnetic polarity patterns from that study were found to correlate over wide distances of the Siberian platform, and independent tests of this correlation on an intercontinental scale with the use of marine carbon isotopes strongly support the primary nature of the remanent magnetization (8).

Torsvik et al. present histograms of APW rates for the four Cambrian continents, and use the heterogeneity of their interpolations as a key argument against a Cambrian IITPW event, but the existing Cambrian global paleomagnetic database is far from comprehensive. Torsvik et al. have excluded data they state are problematic [the "anomalous" Lena River pole by Kirschvink and Rozanov (7)], have misassigned ages to the Laurentian poles as described above and below (9), have relied on a 100-Myr interpolation with no data to generate conclusions about Baltica's incremental motion during Cambrian time, and have included incorrect results from Gondwanaland (Fig. 1). We respect that Torsvik et al. have chosen the spline-smoothing technique as the kinematic basis for their geodynamic models (3, 10); however, given the present database, this technique is not yet applicable as a test for Cambrian TPW. Initially developing the spline-smoothing technique, Jupp and Kent explicitly stated (11, p. 45), "fitted spline paths are reasonably stable under *moderate* errors in the data times" [italics ours]. By assigning exact geochronological precision to imprecisely dated and undated paleopoles of low reliability, and by including already smoothed data in the form of means of poles from Australia, Torsvik et al. have constructed specious APW paths that present merely one of several possible interpretations of the data (Fig. 1).

In developing the hypothesis that a single burst of Early Cambrian TPW joined the disparate poles bracketing that interval from all of the major continents, we found that it neatly explained many of the enigmatic features of the Eocambrian geologic record. The model is not a "stretch" relative to geophysical considerations of TPW (12). Until contradicted by solid, reliable paleomagnetic data, the TPW hypothesis remains a viable explanation for the dramatic changes in paleogeography and evolution that occurred during the Vendian-Cambrian interval.

Fig. 1 (next page). Comparison of methods used to analyze the Vendian-Cambrian paleomagnetic database from Gondwanaland. For the sake of facility in presentation, diagrams show an oblique projection of the spherically bounded APW curve onto a purely latitudinal ordinate varying with time. Data are presented as rectangles with dimensions from quoted or employed uncertainties in both paleopole latitude (A95 or $\alpha 95$) and age. Stippled boxes represent results from sedimentary rocks whose ages depend on use of a calibrated time scale of the fossil record [all boundaries as in the study by Tucker and McKerrow (13) except the Vendian-ND boundary at 543 Ma (14)]. Mean poles are cross-hatched. N-D, Nemakit-Daldynian; Tm, Tommotian; At, Atdabanian; B, Botomian; Ty, Toyonian; MC, Middle Cambrian; LC, Late Cambrian; and Tr, Tremadocian. (**A**) Data selected according to a study by Meert and Van der Voo; (15) (which are presumably the basis for the histogram in figure 1E of the comment by Torsvik et al.) are shown at face value. (**B**) Age uncertainties have been eliminated to reflect the nature of the spline-smoothing technique used by Torsvik et al. (**C**) Subset of the most reliable data is taken at face value as used in our report (1). Queried age limits for some results indicate uncertainties in the paleontological ages of the sampled units, as well as potential revisions to the numerically calibrated Cambrian time scale. Arumbera, Todd River, and Black Mountain results are preferred over the mean poles for Australia derived from a smooth interpolation between those data and others of lesser reliability (15). Khewra and Baghanwala poles have been restored to account for ~30° of Neogene CCW vertical-axis rotations in the sampled area [(16, 17), which was not done in other studies (15), and 5° of uncertainty has been added to each pole's latitude. The Jutana pole is not included because rotations of its sampled region were estimated only by comparisons with a poorly defined APW path (17). Age estimates for the undated but probably Vendian-Cambrian Bhander Sandstone are tentatively accepted from (15). The 515 ± 20 Ma age for the Sør Rondane pole is shown as a reasonable 2σ estimate from a compilation of ages (18). Thick solid curve indiates the hypothesis of Kirschvink et al. (1), which appropriately passes through all the uncertainty fields of the data. We (1) did not rely on the more problematic or imprecisely dated poles shown in this panel; they are merely shown here to illustrate their compatibility with our earlier conclusions (1).


A.
60°
40°
20°
0°
-20°
-40°
Pole Latitude
Ntonya
Australia mean #3
Australia mean #2
Sør Rondane
Khewra Fm
Australia mean #1
Baghanwala Fm
Jutana Fm
Bhander SS
Sinyai
550
540
530
520
510
500
Age (Ma)
B.
60°
40°
20°
0°
-20°
-40°
Pole Latitude
Ntonya
Australia mean #3
Australia mean #2
Sør Rondane
Australia mean #1
Khewra Fm
Baghanwala Fm
Jutana Fm
Bhander SS
Sinyai
550
540
530
520
510
500
Age (Ma)
C.
60°
40°
20°
0°
-20°
-40°
Pole Latitude
Black Mtn
Khewra Fm
Ntonya
Baghanwala Fm
Sør Rondane
Todd River
Arumbera SS
Sinyai
Bhander Sandstone
550
540
530
520
510
500
Age (Ma)
Vendian
N-D
Tm
At
B
Ty
MC
LC
Tr

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Chapter 4: True polar wander, a supercontinental legacy

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Abstract

Paleomagnetically determined apparent polar wander (APW) paths should contain components of individual plate motions as well as true polar wander (TPW), the uniform motion of the mantle relative to the spin axis. Quantifying the TPW component for pre-Mesozoic time is hampered by the lack of a representative from the mantle reference frame; nevertheless, TPW can be estimated if all or most of the continents show similar APW paths for a given period of time. Two such estimates have been proposed recently for the early Paleozoic. Plotted relative to Gondwanaland, which may have been drifting over the mantle slowly enough to constitute an approximate mantle reference frame, these TPW swaths are oscillatory and nearly coaxial. A long-lived and stable prolateness of Earth's nonhydrostatic geoid between Cambrian and Devonian time could produce such a pattern of TPW.

Similarly oscillatory APW paths exist for the late Neoproterozoic cratons, perhaps indicating TPW about the same long-lived axis during that earlier interval as well. This long-lived prolate axis of Earth's nonhydrostatic figure, representing a stable configuration of mantle mass anomalies, may have been inherited from the vanished Rodinia supercontinent in the same way that the present nonhydrostatic geoid's prolate axis may be a legacy of Mesozoic Pangea and its peripheral subduction zones. These geoidal legacies may endure for several hundred Myr after supercontinental fragmentation, perhaps indicating a characteristic lag time between two distinct geological phenomena: global-scale surface tectonics recorded by the growth and disassembly of supercontinents, and deep mantle convective structure indicated by the long-term record of TPW.

1. Introduction

For a dynamic planet, true polar wander (TPW) is the migration of the geographic reference system relative to the dynamically conserved angular momentum vector or spin axis; such motion arises from secular trends of changing mass distributions within the planet's interior [1, 2]. In this discussion, I refer to TPW only on timescales longer than relaxation of the hydrostatic equatorial bulge, secular variation in Earth's geomagnetic field, and Milankovitch orbital variations (all < 10^6 yr). Whereas the existence of such long-period Mesozoic-Cenozoic ("Neozoic") TPW on Earth has been verified by several studies involving use of hotspots as a mantle reference frame (earlier studies reviewed by [3]; updated by [4]), any effort to identify pre-Mesozoic TPW suffers from the absence of an extant mantle representative, such as well defined hotspot tracks, for periods older than ~150 Ma.

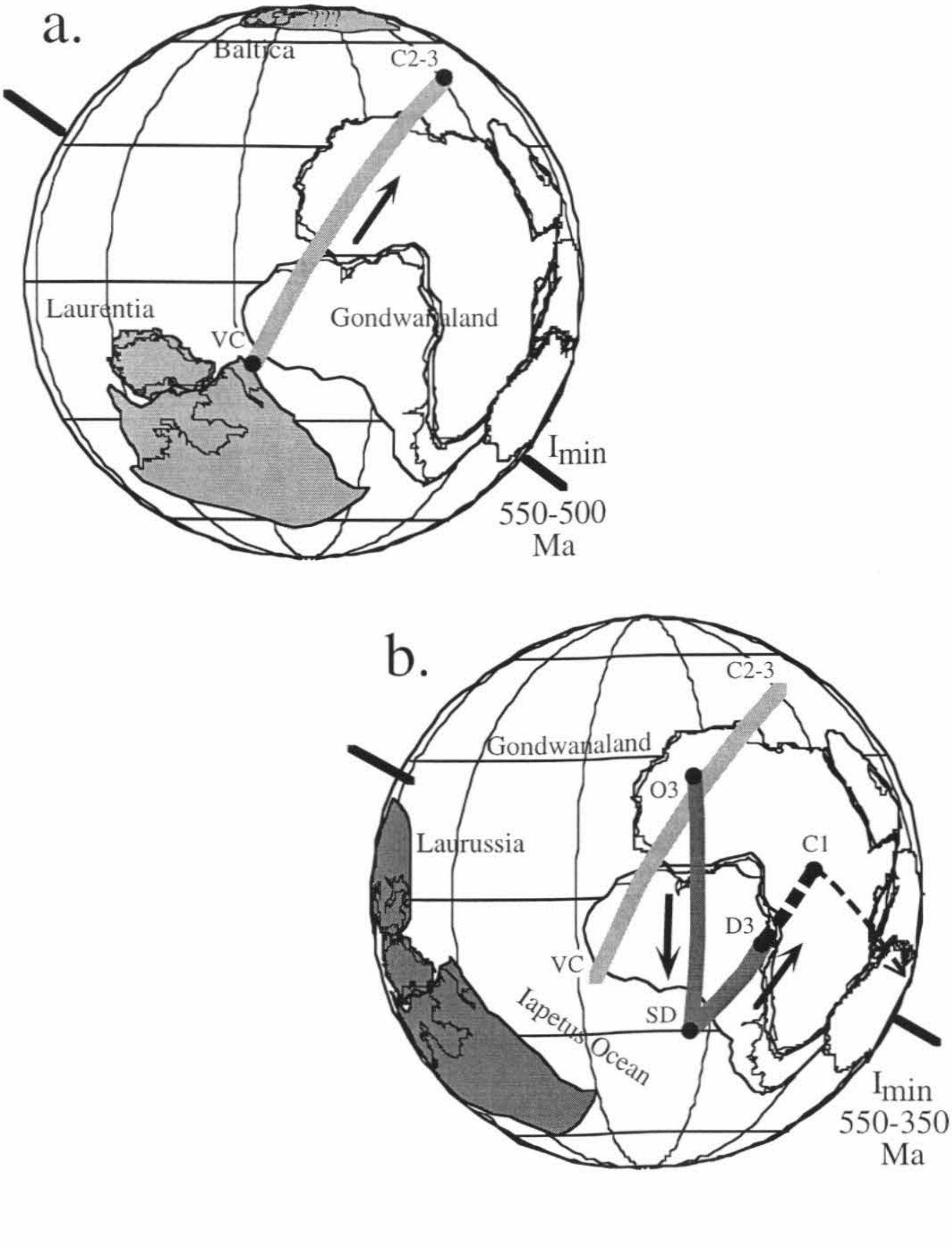
Pre-Mesozoic TPW may be estimated in coarse fashion by comparison of paleomagnetically determined apparent polar wander (APW) paths of continents or cratons during a given interval of time, despite the lack of any extant oceanic lithosphere from these times (predominantly oceanic plates cover nearly half of the present Earth surface). Each APW path depicts the motion of the paleomagnetic dipole--and hence the planet's rotational axis by the axial geocentric dipole hypothesis--relative to a continent or plate. An APW path for a given continent represents the summed motion of two components: the rotation of that continent's plate over the mantle, and the TPW component of whole-mantle migration relative to the spin axis. Thus TPW and APW are not mutually exclusive; rather, TPW (if accepted as a geophysical phenomenon at these timescales) should constitute a non-zero component of *every* APW path segment, common to all continents [5].

Only when TPW occurs at rates much faster than individual plate motions will it compose a significant enough amount of the continents' APW paths to be recognized by this method alone. Two such intervals have been postulated for pre-Mesozoic time: Ordovician-Devonian [6] and Early-Middle Cambrian [7]. The proposed TPW events (Fig. 1) can account for large, rapid rotations of the enormous Gondwanaland continent, previously puzzling to plate kinematicists [8, 9]. Herein I propose a simple geodynamic mechanism for these sweeping rotations, in the context of TPW and supercontinental assembly and fragmentation.

2. TPW on a prolate spheroid

The shape of the Earth is, to good approximation, an oblate spheroid with flattening ~ 1/300. The oblateness is due to the equatorial bulge, a hydrodynamic effect which does not contribute to long-period ($> 10^6$ yr) TPW [2, 10]. The remaining variations of the planet's surface gravity equipotential, the nonhydrostatic geoid, are an expression of the internal mass anomalies which control long-period TPW [3, 11, 12]. Two spheroidal figures of the nonhydrostatic geoid will lead to polar instability, both with the maximum (I_{max}) and intermediate (I_{int}) inertial moments approximately equal: a quasi-sphere and a prolate spheroid. For these figures, small adjustments in internal mass anomalies can generate large TPW much more readily than for an oblate nonhydrostatic geoid. The quasi-spherical case would result in chaotic TPW trajectories, whereas the prolate figure could be recognized by TPW confined generally to a single

Figure 1 (next page). Orthographic projections of portions of the Early Paleozoic South apparent polar wander (APW) path for Gondwanaland, in present African coordinates (reconstruction parameters from [39]). In each panel, I_{min} is determined as the minimum eigenvector of the Bingham distribution [40] of all shown APW vertices. **a.** The Cambrian TPW hypothesis of [7], using both the polar wander path and the longitudinally constrained reconstruction of Laurentia, Baltica, and Gondwanaland. Baltica's position is queried because of its lack of data for that time [7]. (I_{min} at 33°S, 076°E.) **b.** The Late Ordovician-Late Devonian TPW hypothesis and resulting Laurussia-Gondwanaland reconstruction from [6], followed by the Carboniferous path of [39]. The Cambrian APW swath (light gray shading) is retained from (**a**) for comparison. (I_{min} at 25°S, 094°E.) VC, Vendian-Cambrian; C2-3, Middle-Late Cambrian; O3, Late Ordovician; SD, Silurian-Devonian; D3, Late Devonian; C1, Early Carboniferous.


a.
Baltica
C2-3
Laurentia
VC
Gondwanaland
I_{min}
550-500
Ma
b.
C2-3
Gondwanaland
O3
Laurussia
C1
D3
VC
SD
Iapetus Ocean
I_{min}
550-350
Ma

great circle orthogonal to the nonhydrostatic geoid's prolate axis (I_{min}; Fig. 2).

If this axis were long-lived enough to host several oscillations of TPW, and if differential motions between plates were relatively small or slow compared to TPW, then we should observe the oscillations in all of the continental APW paths with simultaneous "turnaround" times and equivalent rates of intervening APW. Such features, in theory, could be used to test hypotheses of pre-Mesozoic TPW [13], but several limitations of the existing pre-Mesozoic paleomagnetic database preclude explicit use of those criteria [14]. First, with rare exceptions, each continent's pre-Mesozoic APW path is constrained only by a few reliable poles, typically separated in age by 10 Myr or more [8, 15]. Second, some dynamical models of TPW allow rotations of ~90° in as little as 5-10 Myr [12, 16], necessitating a much more complete paleomagnetic database to assess the full range of geodynamically possible APW paths. Third, polarity ambiguity [17] and poor age constraints for individual paleomagnetic results, as well as the apparent lack of a plausible geodynamic mechanism to generate rapid rates ($\geq$ 30 cm/yr) of continental motion, both have traditionally spurred paleomagnetists to minimize path length in their APW constructions [18]. For these reasons, the presently favored Neoproterozoic-Paleozoic APW paths may dramatically underrepresent the magnitude of actual APW motions during that interval, and the general distribution of poles is a more reliable test of TPW hypotheses than the comparison of presently favored "turnaround" times of APW swaths from different continents (Fig. 3).

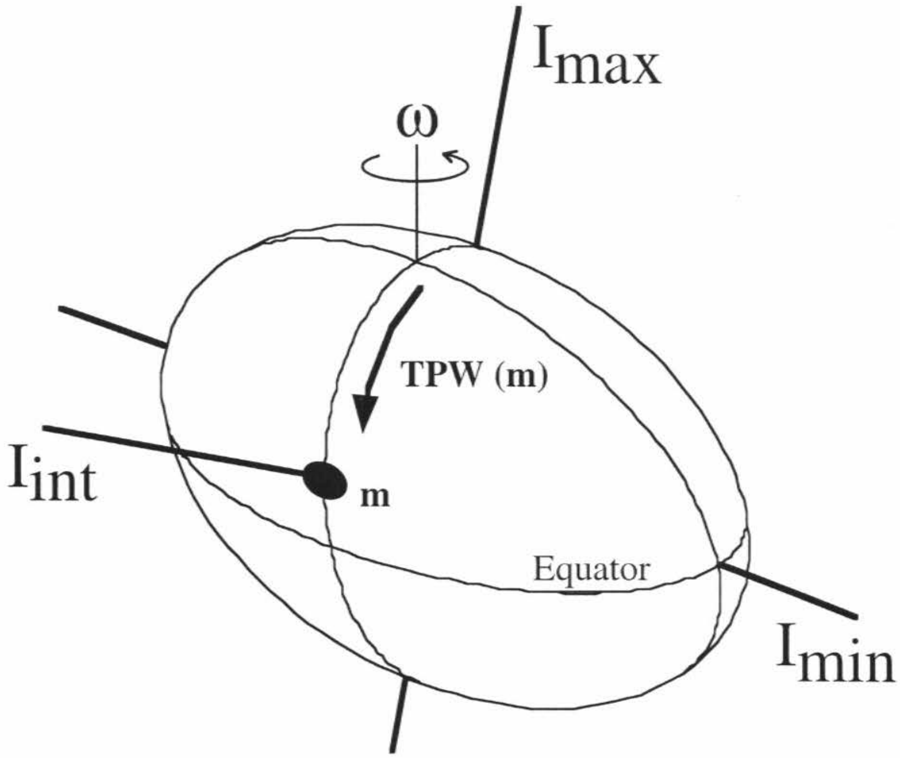


Figure 2. Expected pattern of TPW arising from addition of small mass excess 'm' to an otherwise ideal prolate spheroid. Maximum, intermediate, and minimum moments of inertia are respectively labelled I_{max}, I_{int}, and I_{min}. TPW(m) is the process by which the entire solid planet, including its geographic inertial axes, migrates relative to the angular momentum vector (ω) as a result of the centrifugal torque on 'm'.

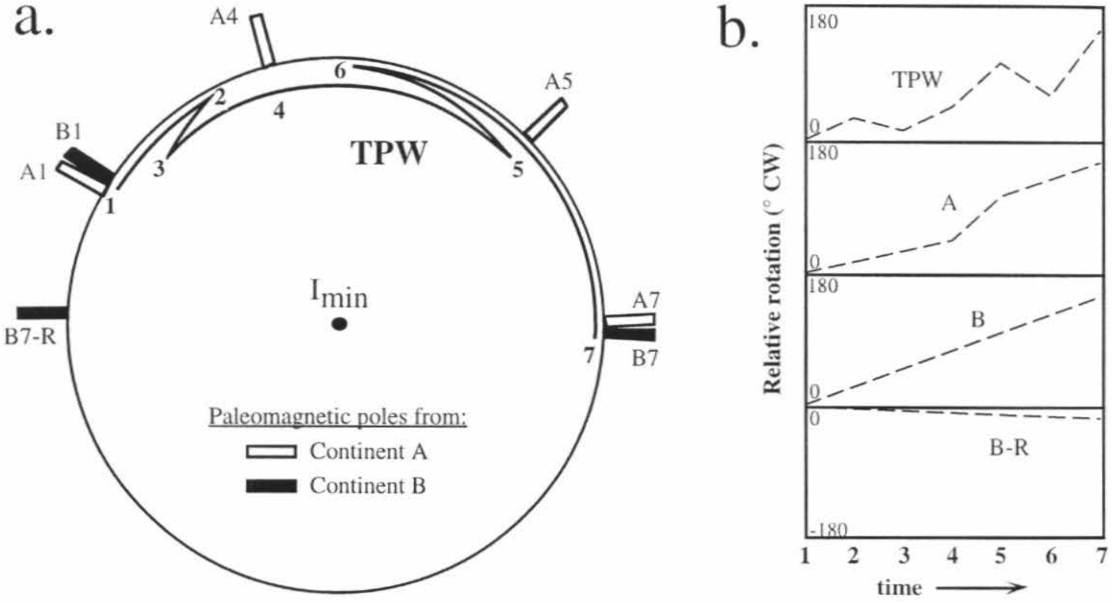


Figure 3. Perils of parsimonious interpolation among sparse paleomagnetic data when testing hypotheses of rapid TPW. **a.** cross-section of the planet in the plane of the short axes I_{max} and I_{int}, with hypothetical TPW swings as shown between times 1 and 7. Two continents with comparatively little motion between them, A and B, have sparse paleomagnetic constraints (tick marks) throughout this interval, with ages of paleopoles exactly known (labels on the tick marks). Pole B7-R is the opposite-polarity interpretation of the B7 result. **b.** Interpolated APW motion during this interval according to the hypothetical TPW path and the constraints from continents A and B. The path for continent A overlooks the 'turnarounds' at times 2 and 5, and most paleomagnetists would erroneously choose path B-R (using the reverse polarity of pole B7) to minimize the length of APW for continent B.

3. Neoproterozoic-Paleozoic TPW

Although the Paleozoic APW path for Gondwanaland has been disputed [18], the most reliable poles fall nearly within the same great circle, at times implying extremely fast velocity of as much as 30 cm/yr [9]. Models of TPW [12, 16] are more easily reconciled with such fast motion of large continents than hypotheses of conventional plate motion requiring special conditions of continental roots, mantle viscosity structure, and plate-driving forces [19]. Figure 1 shows an early Paleozoic polar wander path constructed almost entirely by the two postulated episodes of TPW [6, 7]; if Gondwanaland was drifting slowly over the asthenosphere throughout that interval, then the large continent would provide an effective mantle reference frame for at least qualitative estimation of TPW. The large and rapid early Paleozoic oscillations would then suggest TPW about a long-lived prolate geoid axis intersecting the surface near the continent's Australian-Antarctic side. Subsidiary southeastward (present African coordinates) migration of the pole may reflect slow drift of Gondwanaland or an additional, transverse TPW component.

The latitudinally and longitudinally constrained paleogeographic reconstructions generated by the two postulated TPW events show Laurentia rifting away from the Amazonian margin of Gondwanaland and then joining with Baltica, all between Late Cambrian and Late Ordovician time (Fig. 1). As it did so, it rotated ~30° CCW while remaining near the Equator. This explains why Laurentia's early Paleozoic APW path [15], unlike that of Gondwanaland, does not fall within a single great circle (Fig. 4). Nonetheless, a progression of poorly dated but probably Riphean and Vendian paleomagnetic poles from Laurentia, at times indicating very rapid continental motion, is in fact co-circular with the

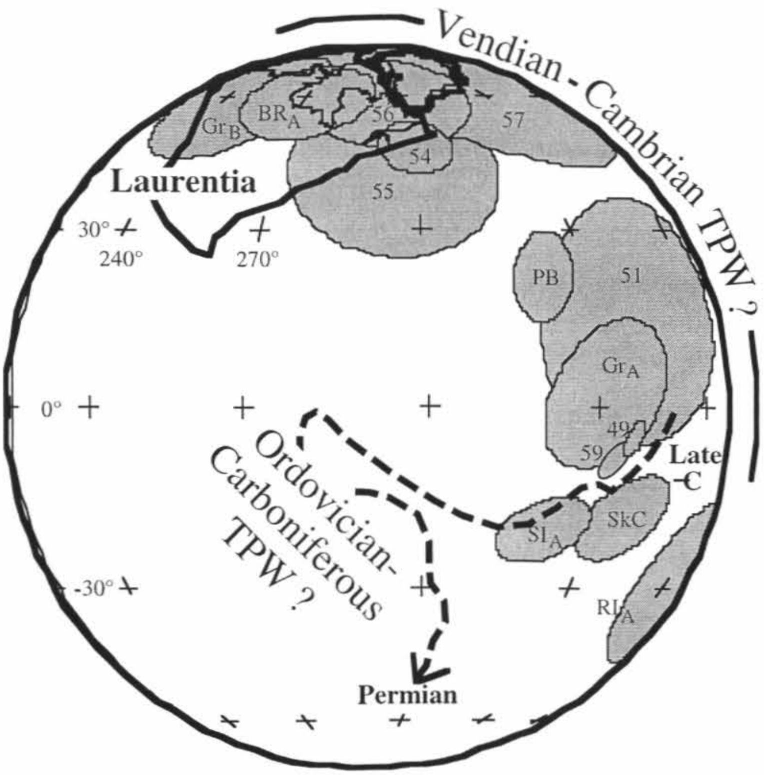


Figure 4. An example of Neoproterozoic-Paleozoic paleomagnetic poles distributed along a great circle, from Laurentia in late Neoproterozoic to Cambrian time (~720-500 Ma). Gray ellipses delimit 95% confidence intervals; poles are numbered according to Torsvik et al. [15] or lettered according to Park [20], except those from the Grenville dikes (GrA, GrB; [41]), Sept Iles 'A' (SIA; [42]), and Skinner Cove Volcanics (SkC; [43]). Many of the ages of magnetization are not well known [15], thus the common model of a single pole-to-Equator migration for the Laurentian continent during Vendian-Cambrian time [15, 20] is not a unique interpretation of the data. Polarity ambiguity for most of the Vendian-Cambrian poles permits their possible locations on the other hemisphere. A smoothed Late Cambrian-Permian South polar wander path [15], including the Late Ordovician to Late Devonian oscillation attributed to TPW [6], is shown for comparison (dashed curve). A complete discussion of the Vendian-Cambrian poles can be found in Chapter 5. Orthographic projection.

Vendian-Cambrian poles [20]; when treated as bidirectional data because of geomagnetic polarity ambiguity, they are evenly distributed around the great circle (Fig. 4). This is consistent with inadequate sampling of a high-frequency TPW oscillation about a prolate nonhydrostatic figure as described above.

The late Neoproterozoic paleopoles from Laurentia, although sparse, suggest the possibility that the prolate axis may have existed as early as ~700 Ma. A similar situation exists for Baltica, whose Riphean-Ordovician APW path swings back and forth along the same great circle [15]. Additionally, recent results from the terminal Proterozoic Bunyeroo Formation in southern Australia [21] support a Vendian APW oscillation which is co-circular with the Cambrian arc attributed to TPW [7]. A test of this sort using the other unassembled Gondwanaland cratons is not presently possible because the Riphean-Vendian APW paths from individual blocks are too poorly constrained to show any coherent patterns [22].

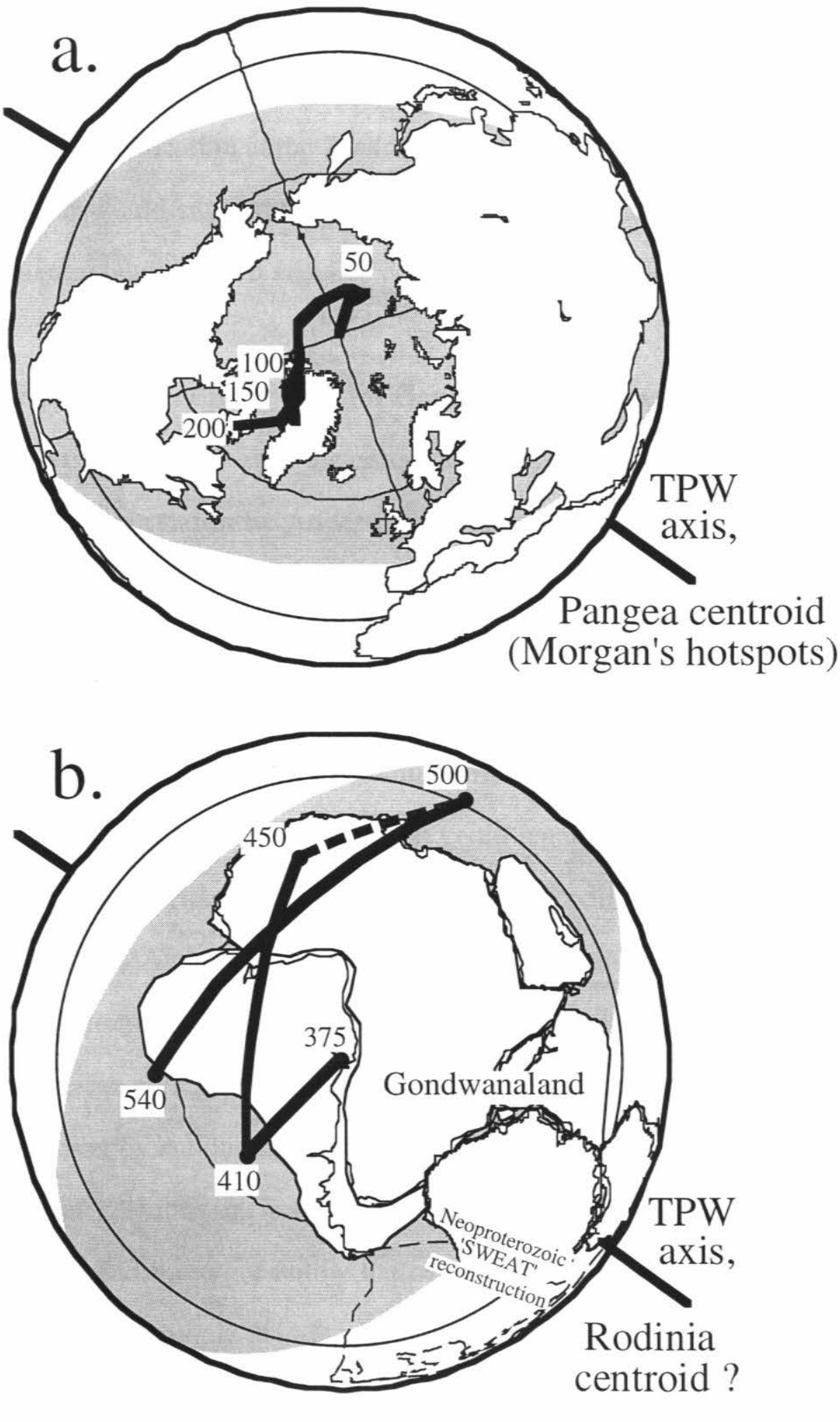
Overall, the limited database permits the hypothesis of predominant TPW during Neoproterozoic-Paleozoic time. If large TPW did *not* occur during that interval, then one must explain with conventional plate mechanisms the high-amplitude, oscillatory nature of APW paths, with implied continental velocities far exceeding today's typical values [7]; also, as these continents sped around the globe their relative positions changed only slightly, as appears to be the case for Laurentia and Gondwanaland (Fig. 1). The TPW hypothesis is attractive because it explains these conundrums of the previous models with a simple and well understood dynamic mechanism. In addition, it successfully reproduces Laurentia's separation from Gondwanaland and convergence with Baltica between Cambrian and Silurian time (Fig. 1), as determined independently by geological arguments [23].

4. A supercontinental legacy

Neozoic TPW has occurred in oscillatory fashion broadly along the great circle marked by 130° and 310° E longitude [4, 24, 25], which lies entirely within the sectoral girdle defined by low values of the long-wavelength nonhydrostatic geoid (Fig. 5a). The prolate contribution of the long-wavelength (harmonic degree 2) nonhydrostatic geoid appears to have remained stable for most or all of that interval of time [11, 26-28]. Enigmatically, the TPW and geoid axes are offset by about 40° longitude, similar to the offset between the present geoid and Pangea's centroid noted and discussed by Le Pichon and Huchon [29]. As both their "absolute" Pangea reconstruction and the Neozoic TPW path [4] are based on Morgan's early hotspot compilation [24], perhaps these discrepancies could be reduced by an updated hotspot model (recent models are uncertain even for 85-130 Ma and completely unconstrained for earlier times [30]). Regardless, the first-order conclusion is that the present nonhydrostatic long-wavelength geoid, as well as Neozoic TPW if verified by further study with an updated hotspot model, both seem to be geophysical legacies of the Pangean world [26, 27].

I propose that the long-lived, early Paleozoic and possibly late Neoproterozoic TPW axis was inherited as a relict prolate figure from the previous supercontinent, early Neoproterozoic Rodinia [31]. Near the center of the Rodinia reconstruction is the "SWEAT" juxtaposition of ~750-Ma rifted margins preserved in Australia-East Antarctica and western North America [32]. If Rodinia had developed an enduring prolate geoid anomaly in the same sense as

Figure 5 (next page). Orthographic polar projections of the present (**a**) and Late Devonian (**b**) world. **a.** Mesozoic-Cenozoic TPW [4] in the Morgan's hotspot reference frame [24], and its relation to the present negative (shaded) nonhydrostatic geoid anomalies (harmonic degree 2 only; [44]). Ages in Ma. The TPW axis at 02°N, 039°E is orthogonal to the Bingham distribution of equally weighted paleopoles and nearly coaxial with Pangea's centroid (from [29] also using Morgan's hotspot reference frame [24]). **b.** Speculative rendering of Late Devonian paleogeography and its accompanying degree-2 geoid. Gondwanaland is reconstructed to its Late Devonian position according to [6], and the APW swaths and postulated TPW axis are the same as shown in Fig. 1. To show coaxiality between the early Paleozoic TPW axis and the antecedent Rodinia supercontinent, Laurentia (dashed outline) is reconstructed to East Gondwanaland along the "SWEAT" juxtaposition [32] according to recent reconstruction parameters [23]. For the Rodinia-Paleozoic TPW connection to be valid, East Gondwanaland must not have drifted appreciably over the mantle between 700 and 550 Ma.


a.
50
100
150
200
TPW
axis,
Pangea centroid
(Morgan's hotspots)
b.
500
450
375
540
410
Gondwanaland
Neoproterozoic
'SWEAT'
reconstruction
TPW
axis,
Rodinia
centroid ?

Pangea, then it is plausible that such an anomaly could persist into the Paleozoic Era, even 300 Myr after the supercontinent began to break apart (Fig. 5b). The prolate TPW axis would approximate the center of the vanished supercontinent. Its early Paleozoic location at the Australian-Antarctic side of the SWEAT reconstruction would suggest that East Gondwanaland did not drift appreciably over the mantle between 750 and 550 Ma, as permitted by limited paleomagnetic data [33].

A rapid apparent polar shift for Gondwanaland between Carboniferous and Permian time abruptly ends the early Paleozoic oscillatory pattern (Fig. 1b). Extending the observation by Anderson [26] that the Late Carboniferous to Late Triassic APW path for Pangea indicates migration of that supercontinent's centroid toward the Equator, Van der Voo [6] suggested that this motion could be a TPW response to the development of a new dynamic mantle upwelling under Pangea. If true, then this places a maximum time constraint of 400 Myr on the duration of the Rodinian supercontinental geoid legacy, between the middle Carboniferous and earliest estimates (~720 Ma) of Rodinia's breakup [23, 33].

Contrary to Anderson's [26] "continental insulation" mechanism, however, mantle upwelling may be induced under a supercontinent simply because it is large enough to protect the interior region of its underlying mantle from penetration by subducting slabs. The subduction girdle around a mature supercontinent [27] thus encircles two antipodal upwellings, one beneath the supercontinent and the other below the center of the exterior ocean (the present mid-Pacific geoid anomaly is in fact of higher amplitude than the antipodal Pangean geoid high). The upwellings are lower mantle mass deficiencies, but they produce positive geoid anomalies due to dynamic topography [34]; they will therefore be drawn toward the Equator via TPW. By this model, old supercontinents are predicted to lie on the Equator, which seems true for both Jurassic Pangea and 700-Ma Rodinia [33]. After a supercontinent fragments, the girdling subduction zones may recede laterally but the general pattern of mantle convection, with a prolate axis of upwelling, remains stable [11, 35]. In this way the geoid, and hence TPW, becomes a supercontinental legacy, which can persist even after the next supercontinent has begun to form.

The duration of slab preclusion required to generate an upwelling under a mature supercontinent may be as short as ~100 Myr, the approximate length of Pangea's existence between the Carboniferous-Permian Variscan orogeny and the Early Jurassic onset of spreading in the central Atlantic Ocean. Alternatively, the requisite timescale may be much longer; if Gondwanaland's Paleozoic APW path largely reflects TPW, then that large landmass shielded the same region of the underlying mantle from subduction for ~400 Myr, between terminal Pan-African/Brasiliano orogenesis [31, 36, 37] and Cretaceous disaggregation. In either case, the timescale for development of a supercontinent-induced prolate axis is commensurate with the endurance of its post-breakup legacy. Hence the characteristic time for a fundamental turnover in mantle convection, from a downwelling over which a supercontinent amalgamates to a subsequent upwelling above which the supercontinent disaggregates, is suggested by the TPW record to be on the order of ~300 Myr, broadly consistent with previous kinematic [27] and dynamic [38] estimates for that phenomenon.

If old supercontinents generate a mantle structure that is inherently prone to polar instability, then why has there been relatively little TPW during and following the breakup of Pangea (200-0 Ma)? One explanation is provided by modelling the history of Neozoic subducted slabs [11], which suggests that first-order mantle convective structure has changed little since Pangea disaggregated. In addition, Earth's rotation is presently stabilized by the strong oblate component of the moment of inertia, where $(I_{max} - I_{int}) \approx (I_{int} - I_{min})$ [2]. Mature supercontinents and their peripheral subduction zones merely develop a prolate axis (decreasing the eigenvalue magnitude of I_{min}) with little constraint on the other two axes, whose relative magnitudes may be controlled by the distribution of individual slabs within the subduction girdle. Perhaps the late Neoproterozoic to early Paleozoic post-Rodinian subduction system was more evenly distributed than its Neozoic post-Pangean counterpart, inducing substantial TPW during the former but not the latter time.

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Chapter 5: A tethyan-asian paradigm for supercontinental cyclicity

To be submitted to Geology, 1998.

ABSTRACT

Contrary to reigning supercontinental models based on the assembly and fragmentation of Pangea, Phanerozoic tectonics cannot be idealized into separate intervals of continental stasis, divergence, and contraction. The atlantic-pangean supercontinental paradigm fails to explain the geology of Asia, which records continuous growth throughout the Phanerozoic Eon. A new definition of "supercontinent," *i.e.* the continental mass formed by the terminal stage of successive orogenic episodes upon a common cratonic nucleus, and containing many but not necessarily all continental blocks, forms the basis of a supercontinental cycle containing Rodinia, Gondwanaland, and "Super-Asia," the present Asian assemblage plus Arabia and Australia. Forming by two stages of first juvenile, then continental crustal accretion, each supercontinent incorporates cratonic elements derived from its predecessor via transfer through a series of tethyan-type oceans. The tethyan-asian cycle lasts ~500 Myr and is consistent with present knowledge of the Neoproterozoic growth of Gondwanaland at the expense of Rodinia.

INTRODUCTION: THE ATLANTIC-PANGEAN PARADIGM

Eocambrian development of a system of globally concurrent passive margins (Bond et al., 1984) is traditionally thought to represent the early stages of a supercontinental cycle that eventually converged upon the late Paleozoic assembly of Pangea, which itself fragmented during Mesozoic time, starting the cycle anew (Nance et al., 1986; Veevers, 1994). In these models, Atlantic-type or "interior" oceans grow as the old supercontinent splits apart; simultaneously, the Pacific- or Panthalassan-type or "exterior" ocean shrinks to accommodate the continental divergence. This stage lasts about 200 Myr, until the interior oceans attain their greatest width; subduction within these oceans begins the next ~200-Myr stage of global contraction and supercontinental assembly. An important feature of the atlantic-pangean cycle (hereafter uncapitalized to distinguish the general cycle from specific paleogeographic entities) is the temporal distinction between episodes of stasis, fragmentation, and assembly (Nance et al., 1986; Veevers, 1989). The atlantic-pangean supercontinental cycle has been proposed to explain Phanerozoic sea levels and glaciation (Worsley et al., 1984; Nance et al., 1986; Veevers, 1990), and has been linked to geomagnetic superchrons through "superplume" events (Larson, 1991) and/or supercontinental heat anomalies (Veevers, 1994).

Upon critical examination, however, the atlantic-pangean supercontinental cycle fails in many kinematic respects. Most recently, the Mesozoic and Cenozoic dispersal of Pangea is concurrent with the Alpine-Himalayan collision in assembling Asia, itself an aggregation of continents (Fig. 1a). Models which allege a Cenozoic stage of global divergence must ignore this enormous orogenic belt. Secondly, even during the Carboniferous-Permian culmination of Pangean assembly, the supercontinent underwent rifting, known as Veevers' (1994) problematic "Extension I," whereby Pangea was fragmenting even before it had finally assembled. In fact, Permian extension involved separation of the Cimmerian continent from the Tethyan margin of Gondwanaland (Audley-Charles, 1983; Sengör, 1984; Metcalfe, 1996). In Eocambrian time, development of rift-related passive margins, which could be attributed to the divergent stage in an atlantic-pangean cycle, coincided with final Gondwanaland assembly within the Pan-African and Brasiliano orogenic systems (Hoffman, 1991). From these examples it is readily seen that Phanerozoic tectonics cannot be simplified into distinct periods of assembly and disaggregation of continents.

Other geologic records are equally troublesome for the atlantic-pangean cycle. For example, the Cenozoic rise of oceanic $^{87}Sr/^{86}Sr$ has been attributed to the Alpine-Himalayan orogen (Richter et al., 1992) and a similar rise during Vendian time to the Pan-African orogeny (Asmerom et al., 1991; Edmond, 1992), but the supposed paroxysm of Pangean amalgamation leaves no such Sr isotopic record (Fig. 2). Instead, late Paleozoic to early Mesozoic marine sedimentary rocks show the lowest values of $^{87}Sr/^{86}Sr$ for the entire Phanerozoic Eon (Burke et al., 1982). Glaciations and geomagnetic superchrons seem to

Figure 1 (next page). Tectonic comparison of eastern Asia and Gondwanaland. Because of the enormous map areas depicted in these projections, the scale bar is only approximate. Stage II accretionary zones (light stipple) may contain smaller blocks not delineated. **a.** Paleotectonic elements of Asia, modified from Sengör and Natal'in (1996). Cratons: An=Annamia (Indochina), H=Helmand (Afghanistan), I=India, L=Lhasa, NC=north China, Q=Qiangtang, SA=Saudi Arabia, SC=south China. **b.** Cambrian reconstruction showing Pan-African tectonic elements of Gondwanaland, from Hoffman (1991), Trompette (1994), and Unrug (1997). The continental edge is idealized, and East Gondwanaland includes a hypothetical area which would have contained various blocks now found in eastern Asia. Cratons: A=Angola, Am=Amazonia, C=Congo, K=Kalahari, RP=Rio de la Plata, SF=São Francisco, SPP=Sahara polycyclic province, WA=West Africa. Orogens: Ar=Araguaias, AW=Aracuai/West Congo, B=Brasilia, D=Damara/Dom Feliciano/Kaoko, EAO=East African Orogen, G=Gariep/Saldania, M=Mozambique, P=Paraguay, R=Rokelide, SO=Sergipe/Oubanguide, TS=Trans-Sahara.

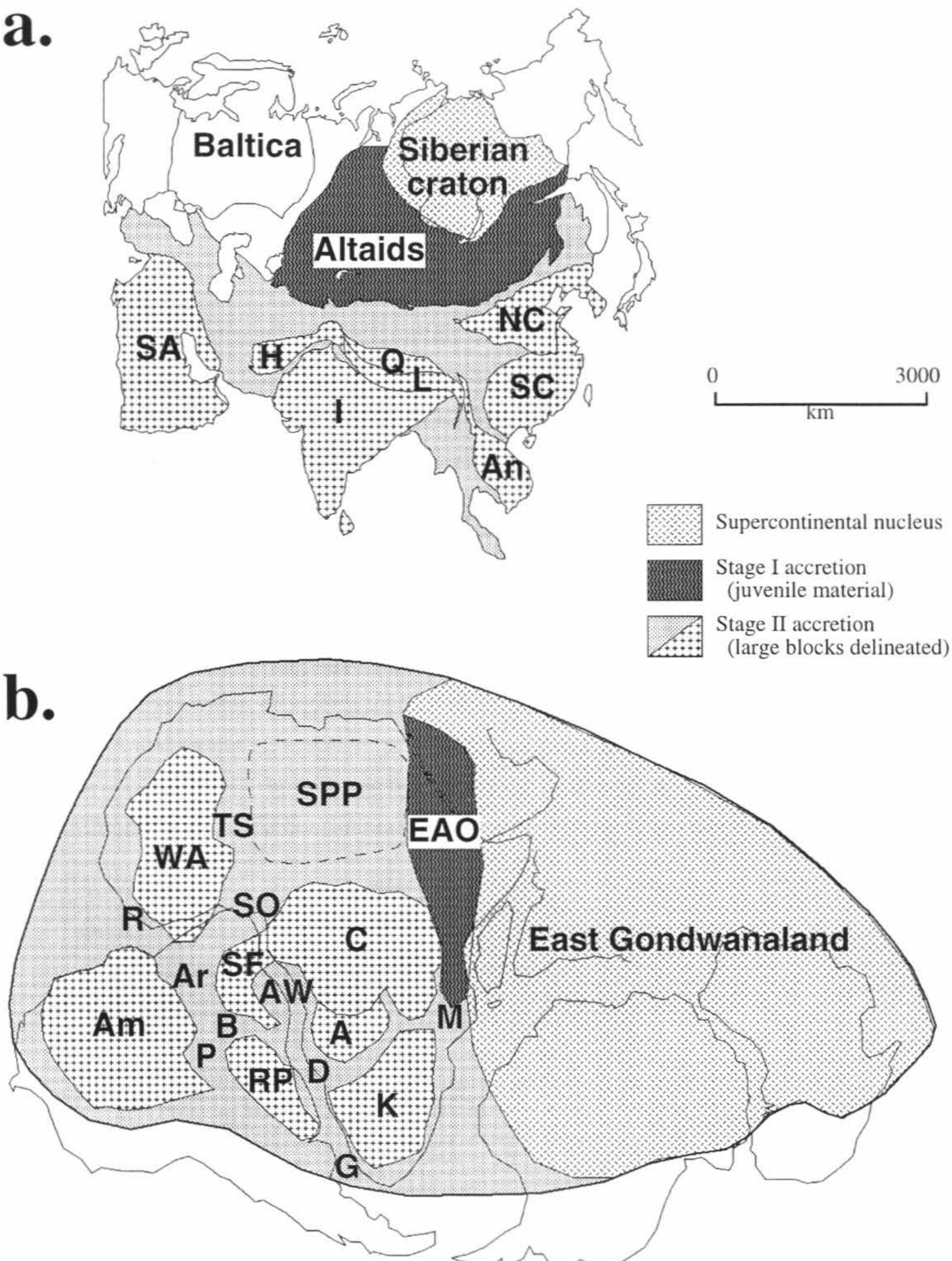
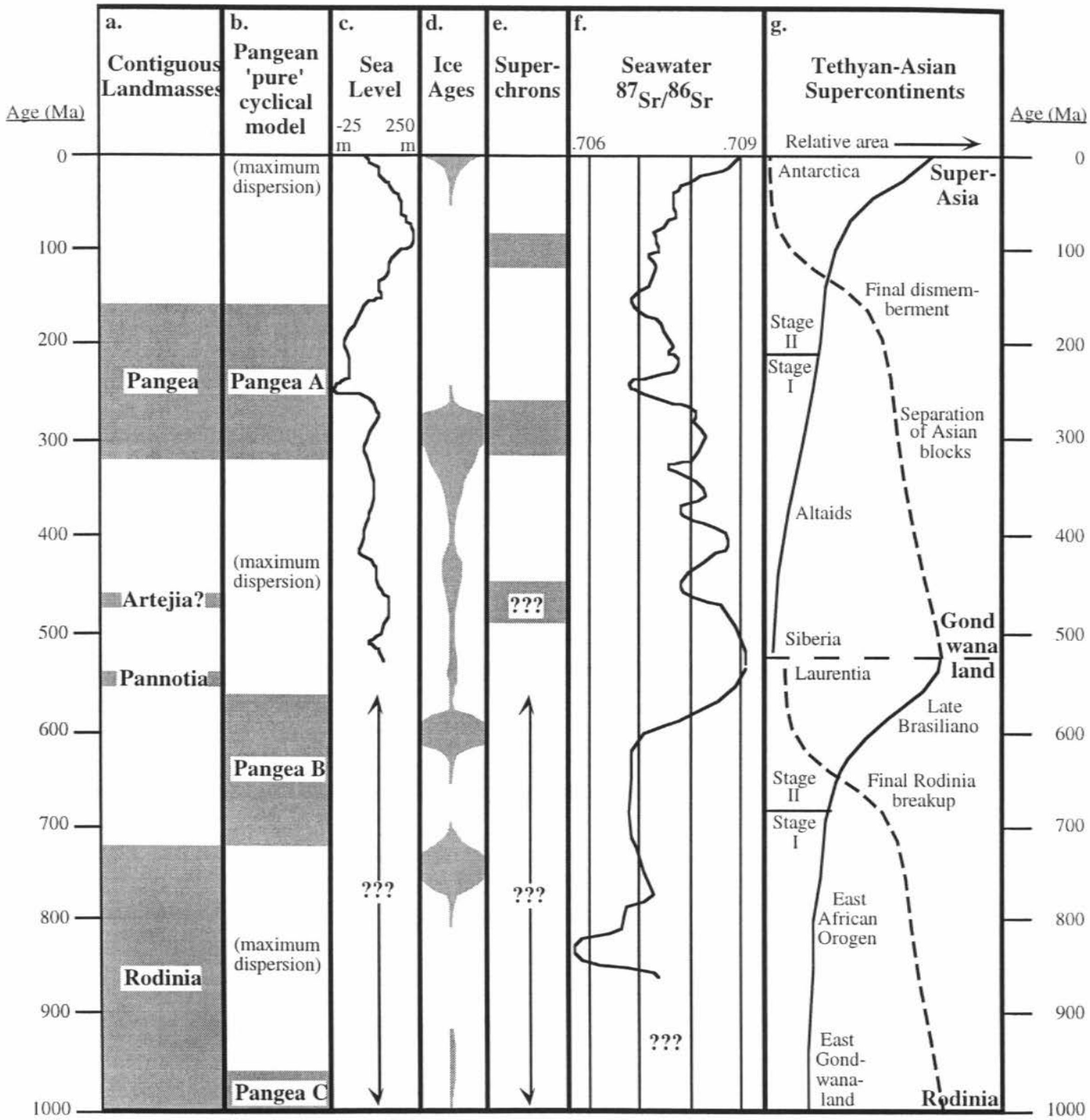

a.
Baltica
Siberian craton
Altaids
SA
H
Q
L
NC
SC
I
An
0
3000
km
Supercontinental nucleus
Stage I accretion (juvenile material)
Stage II accretion (large blocks delineated)
b.
SPP
TS
WA
EAO
R
SO
C
East Gondwanaland
Ar
SF
AW
Am
B
A
M
P
RP
D
K
G

Figure 2 (next page). Supercontinental models and first-order geological events of the Phanerozoic. **a.** Contiguous "supercontinental" landmasses, from Dalziel (1997). **b.** Idealized atlantic-pangean cyclicity, after Veevers (1994) and Veevers et al. (1997). Note that the major differences between panels **a** and **b** for the interval 1000-600 Ma are due to differing interpretations of the ages of Cordilleran rifting (terminating Rodinia or Pangea B) and closure of the Mozambique belt (forming Pannotia or Pangea B). **c.** Sea level, from the "average" model of Algeo and Seslavinsky (1995). **d.** Ice ages, schematically drawn to illustrate the abundance of deposits compiled by Hambrey and Harland (1981). **e.** Geomagnetic superchrons, from Algeo (1996). **f.** Isotopic composition of seawater strontium measured from carbonate rocks, from Burke et al. (1982) and Asmerom et al. (1991). **g.** Tethyan-Asian supercontinental cycle proposed in this paper, schematically represented by the growth (solid curve) and decline (dashed) of each supercontinent. Supercontinental growth stages I and II are discussed in the text.



a.
Contiguous Landmasses
b.
Pangean 'pure' cyclical model
c.
Sea Level
-25 m
250 m
d.
Ice Ages
e.
Super-chrons
f.
Seawater $^{87}Sr/^{86}Sr$
.706
.709
g.
Tethyan-Asian Supercontinents
Relative area
Age (Ma)
0
100
200
300
400
500
600
700
800
900
1000
Pangea
Artejia?
Pannotia
Rodinia
(maximum dispersion)
Pangea A
Pangea B
Pangea C
???
Antarctica
Super-Asia
Final dismem-berment
Stage II
Stage I
Separation of Asian blocks
Altaids
Siberia
Gond wana land
Laurentia
Late Brasiliano
Final Rodinia breakup
East African Orogen
East Gond-wana-land
Rodinia

occur regardless of atlantic-pangean cyclicity and independently of each other (Fig. 2). For example, the two well documented geomagnetic superchrons, Permian and Cretaceous in age, occurred during both alleged Pangean-icehouse and breakup-greenhouse states, respectively (Anderson, 1994). Glaciation has occurred not only on Carboniferous-Permian Pangea, but also the Quaternary continents during maximum extent of the "interior" oceans. If Earth's ice ages are tectonically driven, then they are probably related more to second-order changes in oceanic circulation patterns (Berggren and Hollister, 1977) than atlantic-pangean cyclicity. Long-term sea level trends are not well correlated with Pangean supercontinental stages, for low Neogene levels are inconsistent with predictions (Fig. 2).

In sum, the Pangea-based supercontinent model, with its distinct episodes of accretion and separation, is inconsistent with many first-order features of Phanerozoic tectonics, geochemistry, and global climate. In light of several other ephemeral "supercontinents" proposed for the Paleozoic Era, *e.g.*, Pannotia and Artejia (Powell, 1995; Dalziel, 1997), one may wonder not only whether the atlantic-pangean model is valid, but whether global tectonics conform to a regular supercontinental cycle at all (Gurnis, 1988). A recent study attempted to match the available Neoproterozoic tectonic record into a regular 400-Myr cycle (Veevers et al., 1997), but that interpretation differs markedly from other global syntheses (e.g., Dalziel, 1997) of multi-continental amalgamations (Fig. 2).

THE TETHYAN-ASIAN PARADIGM

A new type of supercontinental cycle, with a Tethyan- or Indian-Ocean perspective, is presented here. The cycle involves simultaneous fragmentation of one supercontinent and growth of the next via the transfer of continental blocks from the former to the latter. Based on the growth of Asia by way of such blocks crossing the various Tethyan Oceans, this model is called the tethyan-asian paradigm (also uncapitalized to distinguish the cycle from specific paleogeographic and geographic entities). Whereas the tethyan-asian model, like the atlantic-pangean cycle, may not completely describe every aspect of global tectonics, it is more consistent with current geological and tectonic syntheses of Asia, Gondwanaland, and Rodinia, described below.

A supercontinent, in the traditional sense inherited from Wegener (1912), is commonly defined by a contiguous landmass of all Earth's continental lithosphere, or a minimum number of independent continental masses (Gurnis, 1988; Veevers, 1994). Alternatively, a supercontinent can be defined in terms of orogenic episodes, representing the terminal stage of continental amalgamation within a super-orogenic cycle involving a common cratonic nucleus, whether or not the assembled landmass contains every block of continental lithosphere. The new definition of supercontinent is more likely to be of use for Precambrian tectonics, for it is unlikely to be demonstrated conclusively that all of the world's continental lithosphere was ever connected as a single, contiguous landmass. By the new definition, Gondwanaland and Super-Asia qualify as supercontinents. Pangea was not built along a single, continuous succession of orogenic episodes against a common cratonic nucleus: it was instead the mere conjunction of a Laurasian sector, which arose from Paleozoic collisions and accretions of arcs, and a Gondwanaland sector, which had assembled by earliest Paleozoic time and was already dispersing as Laurasia grew.

FROM GONDWANALAND TO SUPER-ASIA (PHANEROZOIC)

Eastern Asia is a tectonic collage comprising a Siberian cratonic nucleus flanked by Gondwanaland-derived cratons and intervening juvenile lithosphere from the vanished Tethyan (*s.l.*) oceans (Fig. 1a); it was assembled in two broad stages, approximately separated by the Paleozoic-Mesozoic boundary and by a tectonic "divide" within the continent (Sengör and Natal'in, 1996). During the first stage, a vast amount of juvenile (volcanic arc and oceanic) material with microcontinental fragments became accreted to the growing Siberian nucleus in continuous succession (Sengör et al., 1993). This juvenile terrain, joined from Cambrian to Carboniferous time, has been hypothesized to originate from Siberia in a backarc setting (Sengör et al., 1993), but another model poses a Gondwanaland origin for these areas (Mossakovsky et al., 1994). In any case, the first 300 Myr of Asian growth, from Cambrian to Permian time, consisted of addition of largely juvenile material.

From Triassic to Recent time, further growth of the Asian supercontinent arose predominantly via continental collisions with blocks derived from Gondwanaland (Audley-Charles, 1983; Sengör, 1984; Metcalfe, 1996). Arriving in several episodes, these continental fragments, including North and South China, Qiangtang, Lhasa, India, Sibumasu, and others, successively crossed various generations of Tethys (*sensu lato*). Separation of these blocks began in Devonian (Metcalfe, 1996), or perhaps even Cambrian (C.McA. Powell, pers. comm.) time, coeval with Altaid evolution in Asia, and continued as Pangea assembled and dispersed. Cretaceous rifting completely dismembered Gondwanaland, sending large blocks northward to collide with Asia. Accretion of India, Arabia, and Australia would thus complete the tethyan-asian cycle of Gondwanaland's demise.

Growth of Asia and dismemberment of Gondwanaland were parts of a continuous process encompassing the entire Phanerozoic Eon, but still marked by an orogenic peak--beginning in the Cretaceous as the climactic fragmentation of the southern continents and continuing to the present and the near geological future as the aggregation of some of those continents to form Super-Asia. This orogenic maximum is recorded by a general rise in Mesozoic-Cenozoic seawater $^{87}Sr/^{86}Sr$ from its lowest values attained during the end of the juvenile-accretionary stage (Fig. 2). Dynamically, it is plausible that the Cretaceous evisceration of Gondwanaland may have been due to thermal insulation of the mantle (Anderson, 1982), for that supercontinent had capped the southern hemisphere for a period of 400 Myr since the Early Cambrian (Van der Voo, 1993). Indeed, the virtual absence of Paleozoic continental flood basalts, and their preponderance in the Mesozoic-Cenozoic record, may reflect first-order changes in Earth's thermal budget driven by the supercontinental cycle. Whereas these changes were previously linked to the breakup of ephemeral Pangea (Veevers, 1994), I propose that they are better viewed in terms of the breakup of long-lived Gondwanaland.

FROM RODINIA TO GONDWANALAND (NEOPROTEROZOIC)

The tethyan-asian supercontinental cycle is consistent with present understanding of the Pan-African and Brasiliano orogens that welded Gondwanaland during Neoproterozoic to Cambrian time. East Gondwanaland, comprising present Australia, East Antarctica, India, Sri Lanka, and Madagascar, as well as the restored East Asian blocks, served as the nucleus to the Eocambrian supercontinent (Fig. 1b). The juvenile accretionary stage was manifested by terranes now constituting the Arabian-Nubian Shield of the East African orogen (Stern, 1994; Sengör and Natal'in, 1996). Age control is relatively poor, but sedimentation of arc-related sequences commenced in early Neoproterozoic time (~900 Ma), shortly after "Grenvillian" assembly of Rodinia. The region became consolidated by ~750-650 Ma (Stern, 1994), ending the first stage of Gondwanaland's growth. At about the same time, large continental blocks from West Gondwanaland began to arrive.

"West Gondwanaland," made of the Kalahari, Congo-Sao Francisco, West African, Amazonian, and several smaller cratonic blocks, probably did not ever exist as a distinct, coherent entity. This view contrasts with earlier ideas (*e.g.*, McWilliams and Kröner, 1981) of a climactic collision between two halves of Gondwanaland along the Mozambique belt, but is more consistent with recent geological syntheses of the Pan-African and especially the Brasiliano orogens (Trompette, 1994; Unrug, 1997). In general, peak metamorphic ages decrease from present east (~700-600 Ma) to west (~600-500 Ma) across these belts, culminating in the Paraguay, Rokelide, and perhaps the poorly known Araguaia orogens, all bordering the Amazon craton (Trompette, 1994). Deforming Cambrian rocks, these belts may represent the final collisional event of Gondwanaland's formation. In this view, Cambrian ages of granulites in the Mozambique belt, Madagascar, southern India, Sri Lanka, and East Antarctica (Shiriashi et al., 1994) are interpreted to result from "hinterland" deformation from the coeval collisions located farther to the west.

The cratons of "West Gondwanaland" are commonly considered to be derived from the proto-Appalachian margin of Laurentia as part of the ~1000-Ma Rodinian supercontinent (Hoffman, 1991; Dalziel, 1992). In this way they may be akin to the Gondwanaland-derived cratonic blocks now forming east Asia, having crossed tethyan-type Neoproterozoic oceans en route to the East Gondwanaland nucleus. This large-block accretionary stage of Gondwanaland's assembly lasted approximately 150-200 Myr, until final closure during Cambrian time.

GLOBAL SUPERCONTINENTAL CYCLICITY?

Like the Phanerozoic construction of Asia, the assembly of Gondwanaland was a relatively continuous process spanning several hundred million years. Like Asia, Gondwanaland grew in two temporally distinct stages, first by juvenile accretion, then by addition of large continental blocks. An orogenic maximum, signified by a dramatic rise in $^{87}Sr/^{86}Sr$ (Edmond, 1992), occurred during the second stage of the cycle (Fig. 2). The West Gondwanaland blocks are commonly depicted as crossing a wedge-shaped oceanic realm with poles of rotation near the oceanic vertex (Hoffman, 1991), a geometry mirroring the Tethyan crossings that led to Asian growth. The Neoproterozoic continental-flood-basalt record does not match its Mesozoic-Cenozoic counterpart, but giant radiating dike swarms within Neoproterozoic shield areas are common deeper-level analogues (Ernst et al., 1995) probably related to Rodinia's dispersal (Park et al., 1995).

Success of the tethyan-asian *vs.* atlantic-pangean paradigms in describing supercontinental cyclicity lies in accurately describing Precambrian cycles. By present understanding, the Eocambrian tectonic record favors the tethyan-asian model; Pan-African/Brasiliano orogenesis was coeval with the development of widespread passive margins elsewhere (Bond et al., 1984; Hoffman, 1991). Both models invoke similar periodicity of about 500 Myr, roughly consistent with an episodicity of isotopic-age maxima in the Precambrian and allowing for possible secular trends in global heat production (Hoffman, 1989; Rogers, 1996). The first tests of the tethyan-asian model will be more precise determinations of rifting and closure ages within the Pan-African and Brasiliano belts, and paleomagnetic estimations of the Neoproterozoic drift histories of the West Gondwanaland blocks. Subsequent tests will require accurate paleogeographic reconstructions of the supercontinental transition that begat Rodinia.

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PART II. TEMPORAL AND SPATIAL DISTRIBUTIONS OF PROTEROZOIC GLACIATIONS

Introductory Remarks

The Proterozoic Eon (2.5-1.6 Ga) is punctuated by two broad episodes of widespread glacial deposits (Hambrey and Harland, 1981). The older, Paleoproterozoic glaciation is represented by deposits in the shield regions of North America (Huronian succession and equivalents in Wyoming and Northwest Territories), Fennoscandia (Sariolian), South Africa (Transvaal Supergroup), Western Australia (Turee Creek Group), and perhaps eastern Brazil (Minas Supergroup). The deposits are typically constrained in age between about 2.4 and 2.2 Ga, permitting but certainly not requiring simultaneity (within the same ~10-50 Myr) among the glacial units. The younger, Neoproterozoic glacial era contains widespread glacial or allegedly glaciogenic deposits distributed on all seven present continents. As some of them are associated with warm-climate indicators and others were apparently deposited near the paleo-Equator, they are the subject of intense debate (reviewed in Chapter 8). The Proterozoic glacial deposits, widespread and commonly occupying low latitudes, are thus an enigma of long-term paleoclimate. Identifying the spatial and temporal distributions of glacial formations laid down during the two Proterozoic ice eras, can help point the way toward better understanding of the processes governing their duration and severity.

Chapter 6 is a review paper of the stratigraphic, geochronological, and paleomagnetic constraints on the Neoproterozoic glaciogenic deposits, to be submitted to the SEPM special volume on terminal Proterozoic stratigraphy in conjunction with International Global Correlation Program (IGCP) Project 320. Therein I review the arguments for and against low-latitude glaciation, summarize the stratigraphic, geochronological, and paleomagnetic constraints on each of the 78 deposits, and cull the dataset for the most reliable results. I conclude, first and foremost, that not enough deposits have good enough constraints to identify the spatial distribution to which they belong. Only 15 deposits are somewhat reliably constrained in paleolatitude, and only four are very reliably constrained. Nevertheless, those four deposits are entirely located within 35° of the paleo-Equator, at lower latitude than the farthest advance of the Pleistocene Laurentian ice sheet. Of the three models commonly considered for the Neoproterozoic glaciations, the Phanerozoic archetype of purely moderate- to polar-latitudes for ice sheets can be ruled out. Both the global refrigeration model ("Snowball Earth") and the high-obliquity hypothesis are permitted by the most reliable subset of the database.

Chapter 7 is a detailed paleoamagnetic study of Neoproterozoic sediments interstratified with glaciogenic rocks in South China. Reconnaissance paleomagnetic work by J.L. Kirschvink in 1986 led to the first collection of 58 samples from the upper Liantuo Formation (ULF). Laboratory analyses of these specimens in 1995 and early 1996 revealed a change in magnetic polarity at one level within the stratigraphic section; the second sampling trip to the Three Gorges in 1996 (103 samples from the upper Liantuo Formation) was devoted primarily toward an attempt to identify the same polarity zonation in an independent section of the same lithostratigraphic interval. The sections are exposed in the hills above Liantuo village at 30° 51' N, 111° 09' E. Due to ten years' worth of overgrowth by vegetation and perhaps anthropogenic alteration of the countryside, I could not locate precisely the holes from Kirschvink's 1986 sampling; however, his conspicuous avoidance of yellowish sandstone marker beds (Fig. 3a) allowed me to estimate an approximate location (±30 cm) of the change in polarity. My drilling of that interval was concentrated to 2-3 cm of stratigraphic spacing between samples, near the limit of resolution using standard portable drilling techniques, as well as the uncertainties of lithostratigraphic correlation of individual beds over lateral distances of meters.

Of the three distinct magnetic components observed, one is of two polarities, with polarity zones consistent to within 5cm among the three outcrops. This result, which I interpret to be a primary magnetic remanence, is similar to but better defined than previous results from correlative rocks in South China. My result implies a depositional paleolatitude of 34 ±2° for the overlying and underlying glaciogenic formations, a value neither high enough nor low enough to discredit any of the proposed models for the Neoproterozoic ice ages. However, it is a reliable result (one of the four in the "very reliable" subset identified in Chapter 6) against which future results may be compared for establishing the global distribution of Precambrian glaciations. My result also carries implications for paleogeography, apparently negating two recently proposed reconstructions of South China and Laurentia for ~750 Ma. Finally, one of the overprint components indicates that South China occupied high paleolatitudes at some time between 750 and ~550 Ma. This is the first such indication of polar paleogeography for South China, and it invites further study to determine when the poleward drift occurred.

Chapter 8 delves further back into the geological past, to the Paleoproterozoic Era and its glaciations. The Transvaal Supergroup in South Africa is least deformed on its western side, where 2.2-billion-year-old Ongeluk lavas rest directly atop a widespread glaciogenic deposit called the Makganyene Formation. Extending a comprehensive paleomagnetic survey of the Transvaal Supergroup begun in 1986 by Joe Kirschvink and our South African collaborator Nic Beukes, I performed a detailed paleomagnetic study of the Ongeluk lavas, including two additional sampling trips.

The lavas exhibit extremely stable magnetic behavior for their old age, with only two identifiable and easily distinguishable components. The first to be removed in the laboratory is most likely a present-field overprint. The second and most stable component, residing in magnetite, is tightly clustered among all of the *in situ* flow units studied. Upon detailed sampling of pillow clasts in a hyaloclastic breccia, however, it was found that the observed high-stability, magnetite-borne direction is significantly less well grouped. This suggests that the clasts acquired their magnetization prior to brecciation, and hence, that the observed magnetic remanence is not an overprint. The primary direction thus demonstrated implies a paleolatitude of 11 ±5° for the Ongeluk lavas.

The stratigraphic contact between the Ongeluk lavas and the underlying, glaciogenic Makganyene Formation is interpreted to be conformable, based on regional stratigraphic relations, lack of a weathering horizon between the two, and appearance of volcanic shards in the uppermost strata of the Makganyene Formation. Thus the paleolatitude determined for the lavas is also inferred for the underlying glaciogenic deposit. This result was the first reliable determination of a tropical paleolatitude for a Paleoproterozoic glaciogenic deposit, and it extends the Neoproterozoic climatic paradox (Chapter 6) to apply throughout the Proterozoic Eon.

Chapter 6: Paleomagnetic and geochronological constraints upon the Neoproterozoic climatic paradox

Submitted to SEPM special publication on terminal Proterozoic stratigraphy.

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 the 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 indicate slow, 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 years 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, such deposits 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). 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 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 are of higher quality. Concurrently, the deposits have become better constrained in age. After a superficial step backwards 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 of those earlier results suspect, many of the relevant successions are now being dated with apparently more reliable chronometers.

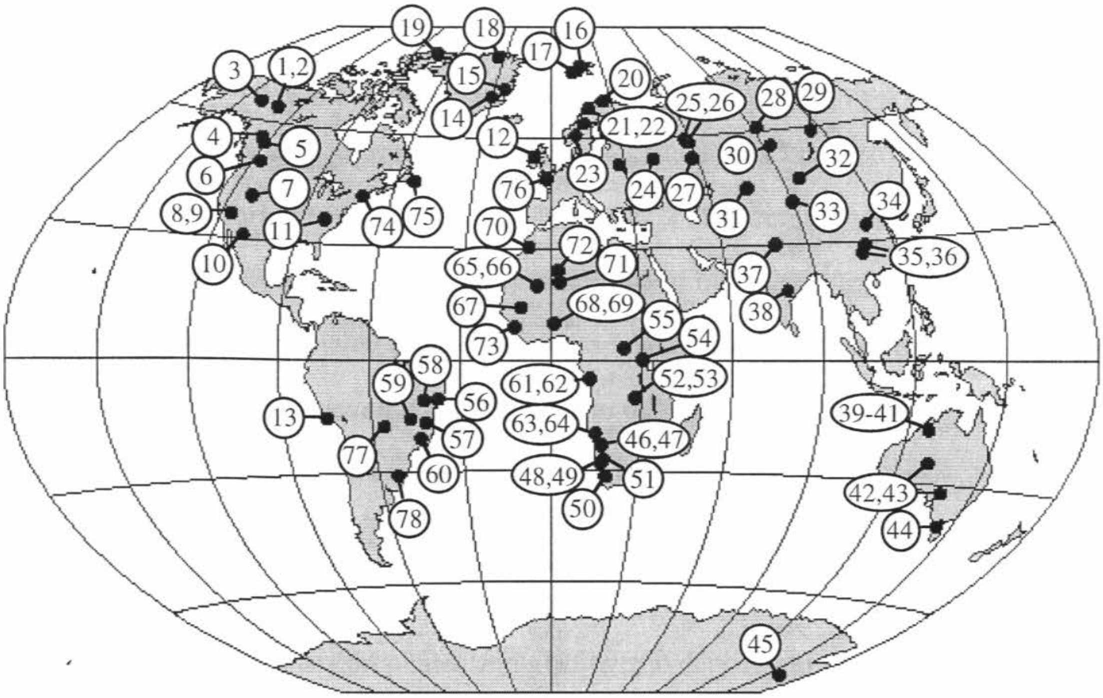


Figure 1. Present distribution of alleged Neoproterozoic glaciogenic deposits. The numbering scheme is used throughout the text and tables.

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 infer 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; 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 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 et al., 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 regarding 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 et al. (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 of the major continental blocks during Neoproterozoic time. As new paleomagnetic results were obtained during the early part of this decade, however, new apparent polar wander loops emerged that suggested poleward excursions during Neoproterozoic time (reviewed by Torsvik et al., 1996). These data were used by Meert and Van der Voo (1994) to conclude that within uncertainty, all of the Neoproterozoic ice sheets were located outside of the tropics. Their view was contested (Williams et al., 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 4800m elevation (Haeberli et al., 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 would effect 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. Finally, for interpreting deposits with apparently ice-rafted debris but no evidence for direct or adjacent upslope ice contact, Crowell (1983) notes 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 mid-latitude glaciation.

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 et al., 1995) rocks seem to indicate a self-reversing geomagnetic field with characteristics very 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 (i.e., low paleolatitudes) seems to characterize the Paleozoic and Precambrian paleomagnetic database (Kent and Smethurst, in preparation). This important observation 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 which 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 ~10^4 yr, then we can consider models to explain sea-level 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) allows a slightly more severe climate to generate continental ice sheets to 25° latitude, a value generated by a computer circulation model whose parameters included the expected 6% 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).

The second model is that of global refrigeration (Harland, 1964) recently dubbed the "Snowball Earth" (Kirschvink, 1992). Contrary to the hyperbole of its new moniker, this model does not suggest a freezing over of the entire world including all of the ocean basins; rather, it merely suggests that the latitudinal range of continental ice sheets advanced from the polar regions into the equatorial belt. Obviously, surface temperatures above freezing must have existed to allow deposition of the primarily subaqueous Neoproterozoic glaciogenic sediments. The main opposition to this model is that first-order energy-balance models (Budyko, 1969; North, 1975) and computer-generated circulation models (Appendix A of Wetherald and Manabe, 1975) suggest that if continental ice sheets were to

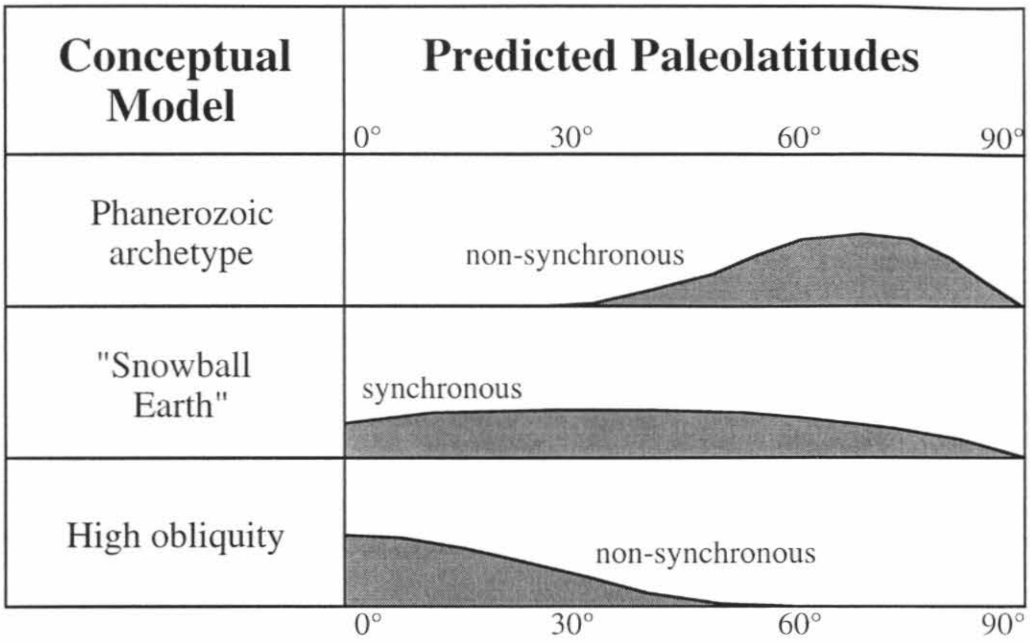


Figure 2. Approximate paleolatitude distributions of glacial deposits predicted by three conceptual models. The Phanerozoic archetype predicts purely polar glaciation; the Snowball Earth model predicts deposits at all paleolatitudes, with an increased abundance at ~20-30° because of greater proportional surface area in equatorial regions; and the high-obliquity model predicts the greatest concentration of glaciers at the Equator.

exist on the Equator, then climatic feedback mechanisms would prevent subsequent recovery from the "ice catastrophe." Given the uncertainties inherent in these simple climate models (c.f. Walsh and Sellers, 1993), it is better to take the paleolatitudinal distribution of Neoproterozoic glacial deposits at face value and develop conceptual models accordingly.

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. The model can be visualized easily for the case of 90° obliquity, where each pole would experience a six-month-long "day" of sunlight followed by a six-month-long "night." The high-obliquity hypothesis can be compatible with either diachronous glaciation as continents moved through tropical latitudes (Kröner, 1977), or a combination with the Snowball Earth model. It should be noted, however, that high obliquity would create an inherently mild climatic regime, even for the tropics; in the conceptually simple end-member case of 90° obliquity, equatorial regions would receive zenithal mid-day sunlight at the vernal and autumnal equinoxes, gradually changing to low-angle 24-hour sunlight during the solstices. For analogy on the present Earth, imagine travelling completely around an opposite pair of meridians, throughout one year, so that each pole is attained at the 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. Whether or not glaciers could grow in that environment probably depends greatly on the annual patterns of precipitation, which themselves likely depend on specific continental arrangements and are thus difficult to predict from a first-order climate model.

Alternative, more imaginative, models 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 end-member models; presumably, they are 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 strictly low-latitude (Piper, 1973) or strictly high-latitude (McElhinny et al., 1974) deposition. Most of the paleomagnetic data cited by those reviews are now obsolete. After a period of 15-20 years, 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 data since the Hambrey and Harland (1981) compendium and its companion manuscript (Hambrey and Harland, 1985), and a wealth of stratigraphic and geochronological information has appeared since. The paleomagnetic data are subjected to a high level of scrutiny, without concern for the low number of studies which pass the quality filter.

METHODS

In the most recent review of paleomagnetic constraints on the Neoproterozoic glaciogenic deposits, Meert and Van der Voo (1994) opted for the general approach of constructing apparent polar wander (APW) paths for the various cratons, then assigning numerical ages to the glacial deposits and estimating paleolatitudes by interpolation from the APW paths. In many instances, however, the glacial rocks are not precisely dated, as is discussed case-by-case below. Thus, for their approach Meert and Van der Voo (1994) needed to assume general ages of ~750 or ~600 Ma for many of the deposits. Such an exercise would actually tend to favor the Snowball Earth model, for it would impart a degree of global synchroneity to the glacial units that might not actually exist. Given (a) the uncertainty in APW paths for most of the Neoproterozoic cratons, (b) the dearth of precise numerical ages for many of the glacial deposits, and (c) 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 the more concise synopsis by Hambrey and Harland (1985). Besides, that is not my primary task: as the interpretation of some of these units has vascillated between glacial and nonglacial, 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. For views which discount glaciogenesis (at least in terms of continent-scale ice sheets) of many of these deposits, see Schermerhorn (1974) or Eyles (1993).

The deposits are numbered and grouped by paleocontinent or craton, and cases are discussed 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 grouping 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 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-isotopic correlation among separate basins (Magaritz et al., 1986; Knoll et al., 1986; Ripperdan, 1994; Kaufman and Knoll, 1995; Kaufman et al., 1997), I acknowledge that the database is continuously growing and encourage readers to draw their own correlations of isotopic excursions. Kaufman et al. (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. An emerging sequence stratigraphy of the Neoproterozoic (e.g., Christie-Blick et al., 1995) is likewise omitted here.

These omissions are justified for two reasons. First, as 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 were 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 (e.g., Hoffman, 1991; Dalziel, 1991; Powell et al., 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 (e.g., Gower and Owen, 1984; Torsvik et al., 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 (e.g., Compston et al., 1992; Heaman et al., 1992; Bowring et al., 1993; Grotzinger et al., 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 et al., 1993; Grotzinger et al., 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, 10-20 Myr at the maximum (Grotzinger et al., 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 from the limits set by the α_{95} error field of the dataset, or by the A_{95} of a paleopole where appropriate. Paleolatitude confidence limits derived from the former 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. These methods all overestimate the true paleolatitude uncertainties, which require cumbersome numerical methods of computation (Demarest, 1983). Those deposits lacking paleomagnetic constraints are presented with applicable geochronological data in Table 2.

Table 1. Paleomagnetically determined paleolatitudes of Neoproterozoic glacial deposits

Paleocontinent Glacial deposit	Age*	Related paleomagnetic study Formation; age**	Code	1234567	Q	λ' (°)†	Magn. age‡	Source
Laurentia								
1. Rapitan Gp. (Fe)	E / 755±18	d	Y	0110010	3	74 +12/-11[a]	Cretaceous	Morris, 1977
		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**[c]	primary	Christie and Fahrig, 1983; Palmer et al., 1983
9. Johnnie Rainstorm	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**[c]	primary	Geissman et al., 1991
11. S. Appalachians	C / 742±2	Franklin; 723±3	comb.	1111110	6	**20 ± 4**[c]	primary	Christie and Fahrig, 1983; Palmer et al., 1983
	[if Vendian]	Unicoi; V-C	rotated	1010010	3	09 +7/-8[a]	Camb-Ord?	Brown and Van der Voo, 1982
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 et al., 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)[b]	primary?	Torsvik et al., 1995
24. East European	551±4 / Pt3	Nyborg; V		0011100	3	34-46 (±17)[b]	primary?	Torsvik et al., 1995
27. South Urals (Fe)	E / Pt3	d	comb	0001000	1	26 ± 7	pre-Perm	Danukalov et al., 1982
East 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 et al., 1996
33. Tarim (3 levels)	C / Pt3	d		0111111	6	**08 ± 8**[c]	primary?	Li et al., 1991
34. Luoquan	C / Pt3	d	undoc	0???1?0	?	50-65	???	Mu, 1981
	C / Pt3	d		0110110	4	32 ± 3[a]	Mesozoic?	Piper and Zhang, 1997
35. Chang'an (Fe)	C / 900	Liantuo; 748	prelim	1010101	4	38 +8/-7[a]	primary?	Li et al., 1996
		Liantuo; 748	A	1111111	7	**34 ± 2**[a]	primary	Chapter 8, this thesis
36. Nantuo	C / 748	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
East Gondwanaland								
37. Blaini	E / Pt2	d		0000010	1	57 +10/-9[a]	Carb-Perm	Jain et al., 1981
41. Egan	E	Pertatataka		1011111	6	**21 ± 8**[b]	primary	Kirschvink, 1978b
42. Sturtian (Fe)	E / 802±10	d	MT1	0100110	3	06-21 (±6)[b]	primary?	McWilliams and McElhinny, 1980
		d	MT2	0100110	3	53-76 (±17)[b]	Mesozoic	McWilliams and McElhinny, 1980
		d	MT3	0000110	2	13-33 (±8)[b]	Camb-Ord	McWilliams and McElhinny, 1980
43. Marinoan	E / 802	d	Y1	0000100	1	47-58 (±26)[b]	Cenozoic?	McWilliams and McElhinny, 1980

		d	Y2	0000100	1	63-76 (±14)[b]	Mesozoic?	McWilliams and McElhinny, 1980
		d	VGP	0110100	3	05 ± 2[c]	primary	Embleton and Williams, 1986
		d	VGP	0111100	4	03 ± 1[c]	primary	Schmidt et al., 1991
		d		0011110	4	**03 +3/-4**[a]	primary	Schmidt and Williams, 1995
Kalahari								
46. Blaubeker/Court	(Kaigas)	d	NBX	0000100	1	13 +11/-10[a]	Camb-Ord	Kröner et al., 1980
51. Schwarzrand	539 / 543	d	many	N.A.		N.A.	N.A.	Kröner et al., 1980; Meert et al., 1997
		Sinyai; 547±4		1110111	6	**38 ± 3**[b]	cooling	Meert and Van der Voo, 1996
Congo-São Francisco								
63. Chuos/Varianto (Fe)	589±40 / <u>747</u>	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**[b]	primary	Meert et al., 1995
64. Ghaub Fm. (±Fe)	589±40 / 747	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								
65. Jbeliat Gp. / "triad"	Ord. / E	d	LPO	0110100	3	06 ± 2[a]	Carb-Perm	Perrin et al., 1988
		overlying strata	CO7-8	0010111	4	07 +6/-5[a]	???	Perrin et al., 1988
		C: Ntonya (522±13)		1100101	4	**65-81 (±2)**[c]	primary	Briden et al., 1993
		E: Sinyai (547±4)		1110111	6	**23-38 (±5)**[b]	cooling	Meert and Van der Voo, 1996
66. Basal Atar Gp.	<u>Pt3</u> / Pt1	overlying strata	I9	0010110	3	03 +5/-4[a]	Carb-Perm	Perrin et al., 1988; Perrin and Prevot, 1988

		underlying strata	I2	0110100	3	12 +2/-1[a]	Carb-Perm?	Perrin et al., 1988
67. Mali Gp/Bakoye (Fe)	<u>C</u> / 650	(see item #65)						
71. Série Verte	~615 / 696	Adma diorite: ~615		1110000	3	70 ± 16[c]	primary?	Morel, 1981
72. Série Pourprée	519 / 556	(see item #65)						
Avalonia-Cadomia								
74. Squantum 'tillite'	C / <u>596±2</u>	d		0111010	4	**55 +8/-7**[a]	primary?	Wu et al., 1986
75. Gaskiers Fm.	565 / <u>607</u>	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 et al., 1997
Amazonia-Rio Plata								
77. Puga et al. (Fe)	E / 623	d		0000000	0	24 ± 7[a]	Cambrian?	Creer, 1965

* 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 ± I converted to pole space; ([b]) using dp or dm, whichever more appropriate, of a pole derived from afar; ([c]) using A95 of a mean of virtual geomagnetic poles (VGPs). All values rounded to the nearest whole number.

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

<u>Abbreviations:</u> 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.

Table 2. Age constraints for paleomagnetically lacking Neoproterozoic glacial units

Paleocontinent		*Paleocontinent*	
Glacial deposit	Age	Glacial deposit	Age
Laurentia		*Kalahari*	
3. upper Tindir Gp. (Fe)	(Rapitan)	47. Blasskrans / Naos (Fe)	(Numees)
4. Deserters Range	C / 728+9/-7	48. Kaigas Fm.	741±6 / 780?
5. Vreeland Fm.	E / ~750?	49. Numees (Fe)	E / 741±6
7. Idaho / Utah	(Rapitan)	50. Aties (Fe)	(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)	620±20 / 948
16. Polarisbreen Gp.	C / 939±8	53. Petit Conglomérat	620±20 / 948
17. W. Spitsbergen	(Polarisbreen)	54. Tshibangu	739±7 / 962±2
18. Moræneso Fm.	C / ~1230	55. Akwokwo / Bandja	~700? / ~970?
19. Pearya	Ord. / Pt3	56. Sergipe diamictites	(Bebedouro)
		57. Jequitaí / Macaubas / (±Fe)	600 / 906±2
Baltica		58. Bebedouro / Rio Preto (±Fe)	(Jequitaí)
21. Sito / Vakkejokk	(Varanger)	59. Ibia / Cristalina (Fe)	(Jequitaí)
22. Långmarkberg	(Varanger)	60. Carandái	(Jequitaí)
25. North Urals	(South Urals)	61. W.Congo lower	600 / 1027
26. Central Urals	(South Urals)	62. W.Congo upper	600 / 1027
East Asia		*West Africa*	
29. Patom	Pt3	68. Tamale Gp.	Pz / ~620
30. Sayan rift	Pt3	69. Kodjari Fm. (Fe)	~620 / Pt1
31. West Altaids (Fe)	C / Pt3	70. Tiddiline Gp.	~570 / 615±12
		73. Rokel River Gp.	O / Ar3
East Gondwanaland			
38. Penganga	Pt3?	*Avalonia-Cadomia*	
39. Landrigan / Fargoo	(Sturtian?)	76. Brioverian	C / 584±4
40. Walsh / Moonlight V.	(Marinoan)		
44. Cottons Breccia	Pt3?	*Amazonia*	
45. Goldie Fm.	C / 762±90	78. Camaquã Gp.	Cambrian

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

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, western Spitsbergen, and probably suspect basement inliers within the Andean orogen (Hoffman, 1988; Dalla Salda et al., 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 counterclockwise 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 consanguineous with intrusive bodies (Morris and Park, 1981; Narbonne and Aitken, 1995) dated at 779 Ma (unpublished data cited in Heaman et al., 1992) and 778 ±2 Ma (Jefferson and Parrish, 1989); alternatively, they may be coeval with the Franklin-

Natkusiak episode (Ross et al., 1995) dated at 723 +4/-2 Ma (Heaman et al., 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 et al., 1995), a minimum (Link et al., 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, 1991), occurring above simple discoid megafossils but below the more diverse Ediacaran fauna (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 et al., 1995) ages. As listed in Table 1, however, I prefer an age somewhat closer to Ediacaran for the Ice Brook Formation, and an age 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 et al. (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 et al., 1987) 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 were 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 component (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% 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 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 diamictite-bearing sedimentary sequences occur locally throughout the Canadian and U.S. Cordillera. Described as the "diamictite and volcanic succession" (Link et al., 1993), the glacial rocks are indeed commonly (though not universally) associated with extrusive basalts interpreted to have been deposited in a continental-rift environment. Some diamictites are better dated than others, but most are generally considered as correlative with the Rapitan Group in the Mackenzie Mountains (Link et al., 1993; Ross et al., 1995). Of course, substantially diachronous rifting across the Cordillera could negate these correlations (noted by Knoll et al., 1981). To compound matters, thermal subsidence of the passive margin did not occur until near the Precambrian-Cambrian boundary (Bond et al., 1985), thus allowing a large degree of freedom in assigning absolute ages to each of the glacial deposits within perhaps greatly diachronous rift sequences.

From north to south, the Neoproterozoic rift-related diamictite-bearing units of the Canadian and U.S. Cordillera are: the upper Tindir Group (Allison et al., 1981); Mount Lloyd George Diamictites (Eisbacher, 1981b); Misinchinka Group (Evenchick et al., 1984; Evenchick, 1988); Toby Formation (Aalto, 1981); Pocatello Formation (Link, 1981); Formation of Perry Canyon, Mineral Fork Formation, Sheeprock Group, and Horse Canyon Formation (Blick, 1981); and the Kingston Peak Formation (Miller et al., 1981). Note that Oberbeck et al. (1994) found a single shocked-quartz-bearing clast out of 62 analyzed specimens collected from Sheeprock-Group-derived colluvium. Although this was cited as critical evidence for an impact-origin of the alleged glaciogenic Dutch Peak diamictite within the Sheeprock Group, detailed stratigraphic studies of this formation and its nearby correlatives establish a more likely glaciogenic origin (Link et al., 1993; Link et al., 1994).

The upper part of the Tindir Group is a diamictite-dropstone-bearing succession similar to the Rapitan Group, even including local iron-formation (Young, 1982), yet an absence of reliable numerical ages or tightly constrained biostratigraphy precludes a surefire correlation. Nonetheless, $^{87}Sr/^{86}Sr$ isotopic data for carbonates above the glaciogenic units show low values consistent with a pre-"Varanger" age for the glacial deposits (Kaufman et al., 1992). This contrasts with K-Ar determinations ranging from 644 ±18 to 532 ±11 Ma (reported in Van Kooten et al., 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-Myr 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 et al., 1996), the partly glaciogenic upper Tindir Group is also likely to be Neoproterozoic in age. No paleomagnetic results have come directly from the Tindir Group, so its paleolatitude must be estimated indirectly.

The same dearth of direct paleomagnetic constraints exists for the other glaciogenic deposits of the Canadian and U.S. Cordillera. Age constraints are generally lacking or one-sided, for example in the Deserters Range, where several diamictite horizons (which may or may not be glaciogenic) occur within the Misinchinka Group nonconformably overlying 728 +9/-7 Ma basement gneiss (Evenchick et al., 1984) but with only a Lower Cambrian younger age limit (Evenchick, 1988). Two-sided age control is only provided for the Toby Formation (southeastern British Columbia and northeastern Washington State), which lies unconformably atop gneiss with a late-stage leucocratic phase dated at 736 +23/-17 Ma (discordant U-Pb zircon age; McDonough and Parrish, 1991) and directly below volcanic rocks dated at 762 ±44 Ma (preliminary Sm-Nd mineral and whole-rock isochron; Devlin et al., 1988). These ages overlap within uncertainty and suggest, assuming that all the "Toby" diamictites are indeed correlative with each other, that this glacial episode occurred at ~720-740 Ma. A similar stratigraphic relation of volcanic rocks immediately overlying a possible metatillite apparently occurs in Idaho, where the volcanic rocks have been dated at 699 ±3 Ma (U-Pb SHRIMP; K.V. Evans et al., 1997). Smith et al. (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. Volcanic rocks and iron-formation occur in the Kingston Peak Formation (Miller, 1985), allowing an additional lithostratigraphic correlation with the Rapitan and upper Tindir Groups. The Kingston Peak Formation, in the upper part of the Pahrump Group, occurs several disconformities below Early Cambrian fossils, and it is most likely younger than 1.07-1.09-Ga sills which may immediately postdate sedimentation of the basal Pahrump Group (Heaman and Grotzinger, 1992).

Link et al. (1993) and Ross et al. (1995) propose a coeval episode of ~720-Ma rift-related glaciation throughout the Canadian and U.S. Cordillera. If this is correct, then the 723-Ma Franklin Dikes pole (Palmer et al., 1983; Christie and Fahrig, 1983) could be used to determine a range of paleolatitudes from 18° in the Yukon (upper Tindir and Rapitan Groups) to 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 using this pole with absolute confidence, the paleomagnetic results provide indirect paleolatitudes with reasonable certainty for the relatively well dated Toby Formation, as shown in Table 1. For the sake of completeness, I note here that Runcorn (1964) made an early attempt at direct paleomagnetic study of the Big Cottonwood Formation, at a stratigraphic level "below the supposed tillite." His study, not reliable by present standards, yielded a paleopole in the vicinity of that from the Franklin dikes, but also similar to a cluster of Late Cambrian poles from Laurentia (Figure 2; Torsvik et al., 1996; Kirschvink et al., 1997).

Examples of younger, possibly Ice Brook-correlative glaciogenic deposits also occur in the Canadian and U.S. Cordillera. The locally exposed, 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 et al., 1993; Ross et al., 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 closely underlie conspicuous carbonate units that may correlate with Ediacaran-bearing strata from the upper Miette Group (Hofmann et al., 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 render 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 (*i.e.*,

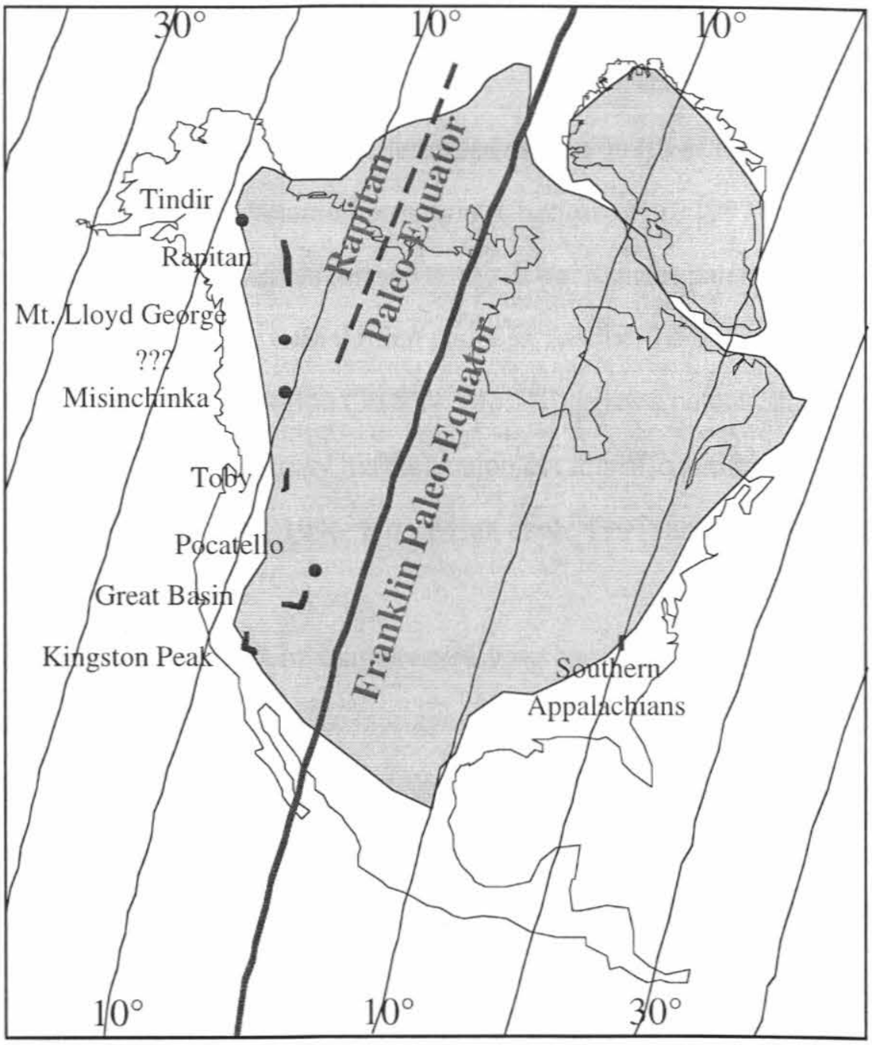


Figure 3. Paleolatitudes of early syn-rift glaciogenic strata along the Laurentian margins, according to a combined pole for the Franklin diabase-Natkusiak basalts (723 Ma) at 08°S, 336°E (Palmer et al., 1983; Christie and Fahrig, 1983). Note that the Rapitan Group has its own paleolatitude determination 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 et al. (1965). Deposits are queried where ages or correlations with the Rapitan Group are especially uncertain.

older) stratigraphic range for those fossils (Knoll et al., 1981). Indeed, an Ice Brook-equivalent sequence boundary has been postulated for terminal Neoproterozoic successions in Utah and Idaho (Link et al., 1993). Finally, in the Death Valley region of eastern California, a diamictite has recently been described as canyon fill of a major incision into the Rainstorm Member of the Johnnie Formation (Charlton et al., 1997). This unit may or may not be glaciogenic and deserves further study. It has a direct paleolatitude determination from 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 Cambrian-Ordovician poles (Torsvik et al., 1996; Kirschvink et al., 1997) 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 sedimentary rocks and thus was originally considered to be Neoproterozoic-Cambrian in age, but 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 et al., 1991) yield a pole similar to other Middle-Late Cambrian paleopoles for Laurentia (Torsvik et al., 1996; Kirschvink et al., 1997), implying a low paleolatitude for the diamictite, whether glaciogenic or not.

Southern Appalachians (11).-

In the southern Appalachians, probable glaciogenic strata occur in rift-related volcanic-clastic successions atop crystalline rocks in two parauthochthonous 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 et al., 1995) and the Vendian-Lower Cambrian Chilhowee Group (Walker and Driese, 1991), including the paleomagnetically studied Unicoi basalts (Brown and Van der Voo, 1982). 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). These dates permit correlation between the Konnarock and Grandfather Mountain glacial units, but there is a ~200-Myr freedom in assigning numerical ages to each of the deposits. Nonetheless, a tentative estimate of 20 ±4° paleolatitude by using the Franklin-Natkusiak pole (assuming a ca.740-Ma age for the Appalachian glacial deposits) is shown in Figure 3. If an alternative age closer to Cambrian were chosen, then paleomagnetic results from the overlying Unicoi basalts could be used to constrain Konnarock paleolatitudes. Unfortunately, the two-polarity direction found by Brown and Van der Voo (1982) lacks any field stability tests and, when rotated 20° clockwise to restore a post-Taconic overprint from the same thrust sheet to match expected directions from autochthonous North America, produces a pole similar to Cambrian-Ordovician directions from Laurentia (Van der Voo, 1993; Torsvik et al., 1996).

Note that Neton and Driese (1993) rejected a glacial association for the upper Grandfather Mountain diamictites, and 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, the latter author admitted that those features are "somewhat atypical" for shocked deposits.

Northern British Isles (12).-

The Port Askaig Tillite, actually a sequence reflecting 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 et al., 1994). Lying 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 (c.f. Rogers et al., 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). Indeed, Dempster and Bluck (1995) seem to have recanted their earlier conclusions, deciding instead for Dalradian deposition within a fragmenting supercontinent (compare their reconstruction with those of Soper, 1994a, and Dalziel, 1994).

The Port Askaig Tillite, at the base of the Argyll Group (Harris et al., 1994) pre-dates the 595 ±4 Ma Tayvallich volcanics (Halliday et al., 1989). A maximum age for the glacial horizon has been suggested at ~750 Ma (Rb-Sr on syntectonic muscovite; Piasecki and van Breemen, 1983) or 806 ±3 Ma (U-Pb on syntectonic monazite; Noble et al., 1996) based on pegmatite dikes associated with shearing that is not observed in the Appin and Argyll Groups; however, direct constraints are lacking. I tentatively accept the U-Pb age as a maximum limit for the Port Askaig Tillite, which in any case should be much closer in age to the overlying Tayvallich volcanics.

The Kinlochlaggan Boulder bed, previously considered as an older glaciogenic deposit within the lower part of the Dalradian or Grampian Group (Treagus, 1981), has now been correlated with the Port Askaig Tillite based on detailed mapping near its type locality (Evans and Tanner, 1996). All of the stratigraphic data from the Dalradian block thus point to a single interval of glaciation, somewhat older than ~590 Ma but otherwise poorly constrained in age. Although the Dalradian sequence has been correlated with

Neoproterozoic successions in north Greenland, east Greenland, and northeast Svalbard (Hambrey, 1983; Soper, 1994b; see below), those correlations are based strongly upon the notion of a single "Varanger" glacial episode, which for the purposes of this paper is not a useful assumption.

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 et al. (1982) showed that diamictite clasts carried the same direction, indicating remagnetization which they considered to be Early Ordovician. In light of subsequent clarification of Laurussia's Paleozoic apparent polar wander path (Torsvik et al., 1996), however, the data are also consistent with a post-tilting remagnetization of Middle-Late Ordovician age, a conclusion favored by Trench et al. (1989).

New geological and isotopic data have revealed another possible glacial 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 et al., 1996). The new data, along with similarly aged Rb-Sr determinations on boulders within the Torridonian succession (~1.2 Ga; Moorbath et al., 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). That study showed a pre-Torridon Group (pre-1.0 Ga using the new data from Turnbull et al., 1996) paleolatitude of 9 ±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 the Bullard et al. (1965) fit is similar to 1.1-Ga poles from the Lake Superior and Grand Canyon regions (summarized by Link et al., 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).

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 et al., 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 et al., 1995) and is intruded by granites with K-Ar hornblende cooling ages of ~440 Ma (unpublished data cited in Shackleton et al., 1979). Further study is now being undertaken to determine better the age and cratonic association of this allegedly glacial unit (D. Carpenter and I.W.D. Dalziel, pers. comm.). 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 (i.e., Laurentian) trilobites. A loose upper 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).

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 700-600 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 (i.e., pre-Devonian) direction similar to Cambrian-Ordovician directions from autochthonous North America (Van der Voo, 1993; Torsvik et al., 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 et al., 1993, 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 et al., 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 et al., 1995). A Varanger- or lower Vendian-equivalent age is generally assumed, as well as correlation with the Tillite Group in East Greenland (Hambrey, 1983; Harland et al., 1993; Fairchild and Hambrey, 1995). 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

Svalbard (or East Greenland) diamictites and the type-Varanger glacial units on the Baltic shield is not substantiated precisely.

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 (e.g., three pre-Carboniferous tectonic domains of Svalbard; Harland and Wright, 1979), precludes indirect assignment of paleolatitudes from extrabasinal Vendian rocks of Laurentia or Baltica.

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 a single glacial episode containing several cycles, correlated with the better preserved exposures of "Vendian" glaciogenic strata to the northeast (Harland et al., 1993). 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.

North Greenland (18).-

A glacially derived diamictite-bearing sequence occurs near Independence Fjord as the Moræneso Formation (Clemmensen, 1981; Higgins, 1986; Collinson et al., 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 glacial 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).

Baltica

Finnmark (20).-

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 nonglacial debris-flow mechanisms (Jensen and Wulff-Pedersen, 1996), leading the latter workers to question the existence of the entire Varanger glacial episode. The 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 1-2μm 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 Ma 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. More robust age constraints are paleontological. Ediacaran fauna occur ~200m above the Mortensnes Formation (Farmer et al., 1992), and microfossils from the unconformably underlying Tanafjord Group suggest an "Early Vendian" age (Vidal, 1981). 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 et al., 1995a). Although the resulting pole falls near Early Ordovician results from Baltica, and the fold test probably only provides a pre-Silurian (Caledonian) age, the authors considered the magnetization to be primary.

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). Except for the Lillfjället Formation within the unfossiliferous Särv nappe, all of these units closely underlie Cambrian strata, permitting a general correlation with the Varanger glacial succession (*e.g.*, Kumpulainen and Nystuen, 1985; Vidal and Nystuen, 1990; Vidal and Moczydlowska, 1995). The Vakkejokk breccia occurs directly above shale containing *Kullingia concentrica*, whose genus ranges into the lowermost Cambrian (Narbonne et al., 1991), and directly below shale containing Early Cambrian hyolithid fauna; this suggests that the Vakkejokk breccia may actually be younger than the glaciogenic deposits in Finnmark (Føyn and Glaessner, 1979). 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 overlying the Moelv Formation, 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 612 ±18 Ma (Vidal and Nystuen, 1990; Vidal and Moczydlowska,1995) which I consider to be unreliable until verified by other isotopic data. If all of the Caledonide glacial deposits are roughly synchronous with the Varanger succession (see caveats for the Vakkejokk breccia above) then paleolatitudes can be applied indirectly from the Nyborg paleomagnetic data (Torsvik et al., 1995a), given that those pre-fold magnetizations are indeed primary (Fig. 4).

East European platform (24).-

Vendian diamictites exist in numerous paleo-depressions across the East European platform, known only from boreholes throughout the region and a few outcrops along the Dniester River in Podolia-Moldavia (Brochwicz-Lewinski, 1981; Aren, 1981; Chumakov, 1981c-g). Some of these units (e.g., the Vilchitsy and Blon Formations) may correlate with each other representing the "Laplandian Horizon," whereas others (e.g., 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); this is 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 et al., 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

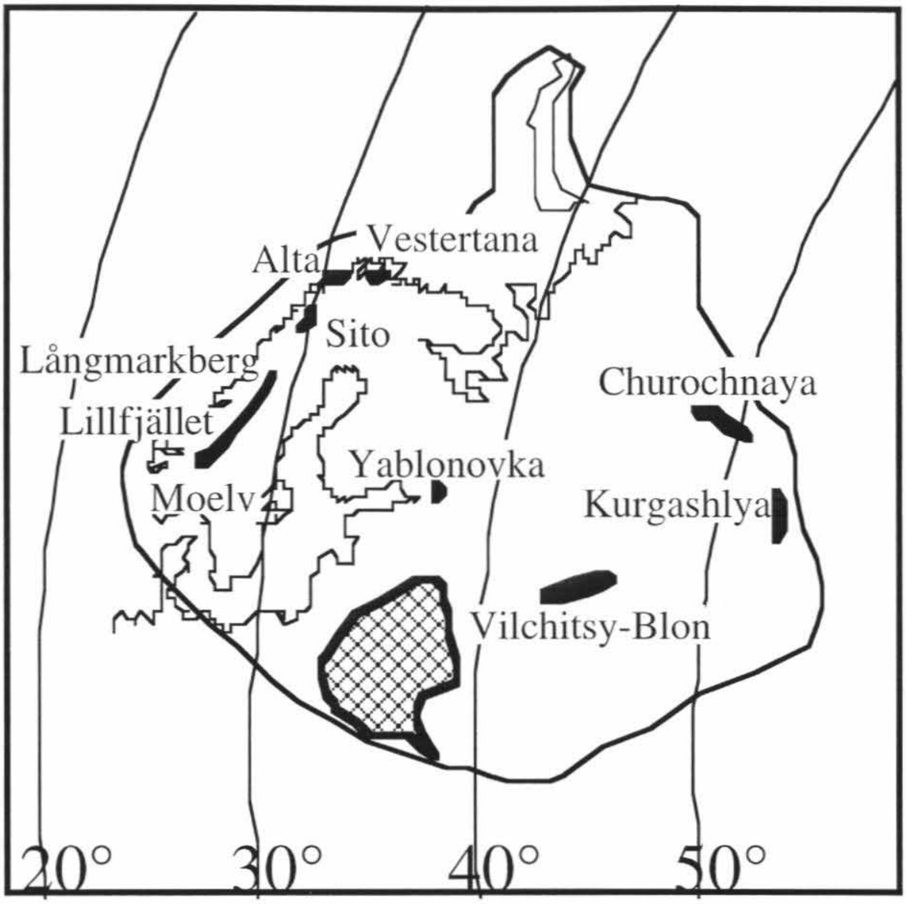


Figure 4. Paleolatitudes across Baltica determined by the Nyborg pole (Torsvik et al., 1995a). Approximate areal extents of Neoproterozoic glacial deposits are shown in black shading or hatchures; they may or may not be truly coeval (see text).

tillites on Baltica, then the Varanger ice age occurred shortly before 550 Ma. This conclusion is 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. The Lublin slope age agrees well with the close stratigraphic proximity between the Finnmark and Vakkejokk diamictites and overlying strata with Ediacaran and Cambrian fauna (see above). Of course, the various tillites and tilloids on the East European platform may be diachronous, representing several glacial intervals, but testing synchroneity vs. diachroneity will require further high-resolution stratigraphic or geochronological work. In the absence of precise numerical ages, only direct paleomagnetic estimates of the glaciogenic strata or conformable units are considered here; these unfortunately do not exist for the East Europan platform except in brief mention with inadequate documentation (e.g., 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) Tana 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 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 (Tana 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, i.e., 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 et al., 1982) that fall generally near the middle Paleozoic apparent polar wander path for Baltica. Positive fold tests for these and other Vendian formations in the vicinity (P. Mikhailov, pers. comm.) nevertheless only constrain the magnetization ages to pre-Permian, based on the age of deformation across the Bashkirian foreland (Brown et al., 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 et al., 1995a) poles are primary, then the substantial apparent polar wander (APW) separating the two results suggests 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 faster between-plate rates (Meert et al., 1993; Torsvik et al., 1996) or true polar wander (Kirschvink et al., 1997; Evans, 1998).

East Asia

Siberian craton (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 terriginous 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. 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" (i.e., "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 poles for Siberia (Torsvik et al., 1995b), and may be either primary or an overprint of possibly early Paleozoic age.

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 690-630 Ma, but these are of unknown reliability. Within the Sayan foldbelt, the late Neoproterozoic Darkhat-Khubsugul rift contains a glacial level comprising tillite and dropstone-bearing shale (Ilyin, 1990).

Kazakhstan (31).-

Diamictites of probably nonglacial origin exist throughout Kazakhstan and northern Kyrgyzstan, deposited in active tectonic settings. Known in central Kazakhstan as the Baykonur (Kheraskova, 1981a,b) and 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 correlated with the Baykonur Formation or named Dzhetym tilloid (Korolev et al., 1981), and to the east in the Tian Shan they occur at two levels within the Tyshkan Group (Korolev, 1981). Many of these deposits are associated with iron-formation, usually in the form of hematitic shales. An active tectonic setting 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 contain clasts of fossiliferous Riphean carbonates or volcanic rocks dated at 800-850 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 et al., 1981). Several levels of tilloids are perhaps distinguished in central Kazakhstan (Kheraskova, 1981a,c).

All the deposits occur within the southwesternmost imbrications of the 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 occur within Neoproterozoic strata of the Zavkhan basin in western Mongolia, within the basal Maikhan Uul Member of the Tsagaan Oloom Formation (Lindsay et al., 1996). This deposit is bracketed in age between unconformably underlying volcanic deposits of the Dzabkhan Formation (isotopic ages of 732-777 Ma, cited by Lindsay et al., 1996) and paraconformably overlying deposits of lowermost Cambrian age. From the Lower Cambrian part of the succession, a pre-fold magnetic remanence suggested a paleolatitude of ~44°, but that result may be a Silurian-Devonian overprint (Evans et al., 1996). The Tsagaan Oloom glacial deposits are therefore still unconstrained in paleolatitude.

Tarim (33).-

Sinian (terminal Proterozoic) glacial deposits occur within the Tarim block; most of these deposits are exposed along the margins of the Tian Shan which bounds the block to the north (localities 1-3 of Wang et al., 1981), although Gao and Qian (1985) report an occurrence in the western Kunlun Mountains. 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 at the best exposed area near Quruqtagh. Any of these levels may be missing from other localities (Wang et al., 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 et al., 1981). The lower two glacial levels are commonly correlated with those of the South China craton (e.g., Li et al., 1996; see below), but direct dating of the Tarim deposits is lacking. A possibly fourth glacial level lies among the pre-Sinian Palgang Group at Quruqtagh, but further study is needed to verify its alleged glacial origin (Gao and Qian, 1985).

A magnetostratigraphic study of the Tarim Sinian revealed a two-polarity remanence with apparently stratabound polarity zones broadly consistent among several stratigraphic sections (Li et al., 1991). The section spans two "tillite" horizons with little change in the measured paleolatitude of about 8°. The invariance of this result through a great stratigraphic thickness and presumably through a long span of time, implies that either the Tarim block experienced little latitudinal or rotational motion, or the paleopole represents a magnetic overprint, perhaps during Late Cretaceous time based on similarity to poles of that age (Zhao et al., 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 et al., 1981; Mu, 1981; Zheng et al., 1994). The age of this level is debatable, ranging from ~750 Ma (i.e., correlated with the Nantuo or Gucheng tillite; see below) to Early Cambrian (references in Xiao et al., 1997). The unit disconformably overlies probably Mesoproterozoic sedimentary rocks (Xiao et al., 1997) and is disconformably overlain by Early Cambrian strata (Piper and Zhang, 1997)

Mu (1981) reports paleomagnetic data of unknown quality for the Luoquan Formation, which would indicate paleolatitudes of ~50-65° 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 et al., 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.

South China block (35, 36).-

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 straightforward (Fig. 5). In the type locality of the lower Yangtze Gorge, only one diamictite exists: the Nantuo Formation, disconformably overlying the Liantuo Formation,

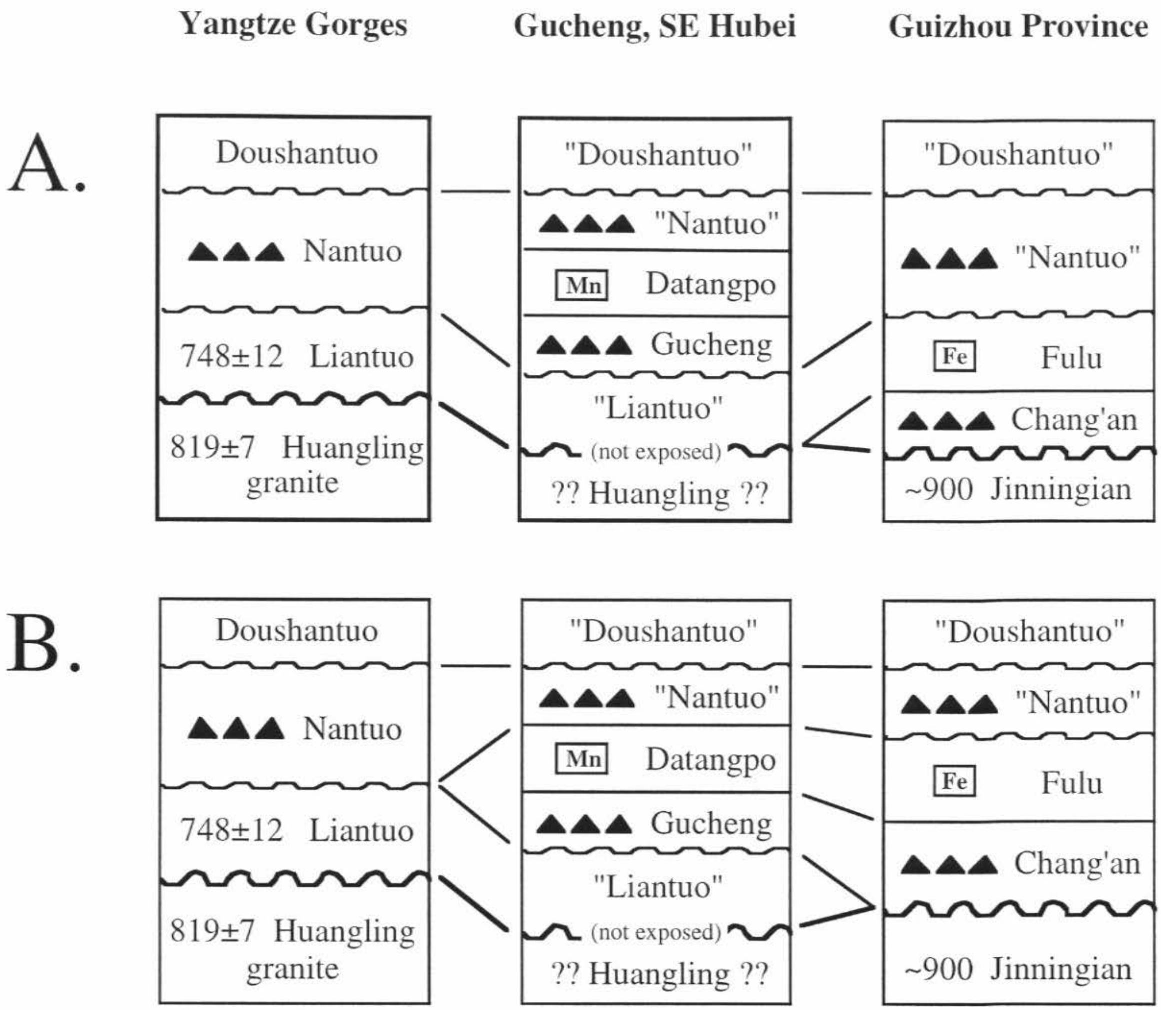


Figure 5. Alternative correlation schemes for Neoproterozoic (Sinian) volcanic-sedimentary successions on the South China block. Dark contacts, unde, indicate profound stratigraphic breaks such as angular unconformities or nonconformities; light wavy curves show regional disconformities; straight contacts 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. (a) Correlation scheme of Liao (1981) and Wang (1986). (b) Alternative correlation by Lu et al. (1985), adopted by Li et al. (1996).

a fining-upwards siliciclastic-volcanic unit which has been dated at 748 ±12 Ma and rests nonconformably upon the 819 ±7 Ma Huangling granite (U-Pb SHRIMP; Ma et al., 1984). Slightly farther southeast at Gucheng, Hubei, two diamictite levels are separated by the Mn-bearing Datangpo Formation (Wang et al., 1981). Because a siliciclastic unit underlies the lower of the two diamictites (called Gucheng), that unit is mapped as "Liantuo" and correlated with the type Liantuo Formation (Wang et al., 1981). In southeastern Guizhou Province, two diamictite horizons occur but are assigned to different ice ages; the lower Chang'an unit is separated from the upper diamictite (called "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. 5a).

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 et al. (1985) considered the Chang'an diamictite as an equivalent of the Gucheng Formation (Fig. 5b). This correlation is problematic because it rejects the Liantuo-Fulu equivalence long favored by stratigraphic syntheses of South China (Wang et al., 1981; Liao, 1981; Wang, 1986). Nonetheless, it is favored by Li et al. (1996).

The alternative correlations carry different implications for the absolute ages of the Sinian glacial levels. In the scheme of Lu et al. (1985), the 748-Ma Liantuo age provides a maximum limit for all Sinian glaciations in South China (Fig. 5b). According to the correlations by Wang et al. (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. 5a). In the latter case, a maximum age for the Chang'an glaciation is provided by the Jinningian movement (Wang, 1986), the folded strata of which unconformably underlie relatively undeformed Sinian. The pre-Sinian orogen contains ~900-Ma ophiolites, granites, and volcanic rocks (Li et al., 1997). 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-Ma date is a maximum for the Nantuo level.

An early attempt at paleomagnetic study of the Gucheng and Nantuo Formations (Zhang and Zhang, 1985) produced results of generally high scatter, low reliability, and similarity to Mesozoic results from the South China block (Van der Voo, 1993; Zhao et al., 1996). Other studies are difficult to assess (reviewed by Zhang and Piper, 1997). Recent work, however, is converging upon a consistent paleolatitude for the Sinian glacial deposits. Preliminary data reported by Li et al. (1996) are consistent with an unpublished Sinian paleolatitude of 40 ±7° quoted by Meert and Van der Voo (1994). A similar result has been fully documented by Rui and Piper (1997); their most stable, hematitic, two-polarity direction yields a paleolatitude of 37 +6/-5° for strata immediately underlying and overlying the "Nantuo" diamictite in Yunnan Province (probably a correlative of the Nantuo *s.s.* as defined in the Yangtze gorges). They consider their result to be primary based on a soft-sediment fold test which I consider inconclusive: the precision of directions *in situ* is the same as that of the modal dataset (k=15), and an increase to k=30 upon 20% unfolding may be fortuitous. Nevertheless, their result is similar to a stratabound two-polarity magnetization (paleolatitude 34 ±2°) from equivalent stratigraphic sections of the type Liantuo Formation in the lower Yangtze Gorge (see Chapter 8). This combined dataset supports moderate depositional paleolatitudes for the Sinian glaciations.

East Gondwanaland

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, and subsequent workers; Brookfield, 1987). Limits listed in Table 1 are determined by Middle Riphean stromatolites within unconformably underlying strata (Valdiya, 1969), and Ediacaran fauna within overlying Krol carbonates (Mathur and Shanker, 1990). A glaciogenic origin for the Blaini diamictites has been debated, but the unit has many features that would suggest at least indirect glacial influence (summarized by Brookfield, 1987). A possibly correlative unit, the Tanakki diamictite with an overlying "cap" dolostone, occurs in northern Pakistan (Brookfield, 1994).

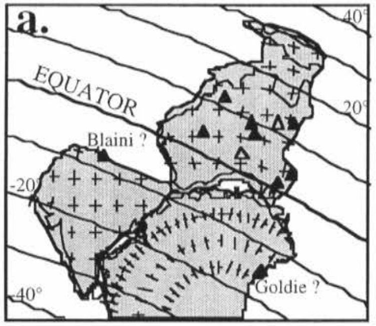
A paleomagnetic study of the Blaini and Krol succession was hampered by the assumption of a Carboniferous-Permian age for the glaciogenic strata (Jain et al., 1981). Given such limitations, that study assigned magnetization ages (in the absence of a fold test) through liberal use of vertical-axis rotations and comparisons with the Carboniferous-Recent apparent polar wander path for India. Obviously, if a Neoproterozoic age had been considered, more freedom would exist in choosing viable magnetization ages; unfortunately, the data are not thoroughly presented in that paper, prohibiting 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 glacial deposits (e.g., Rapitan and Urucum deposits as described above and below; N.J. Beukes, pers. comm.). Second, 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 true diamictites (purely descriptive terminology *sensu* Hambrey and Harland, 1981) and invite further study. The Chanda Limestone is probably Neoproterozoic in age, as determined by microfossil assemblages (Bandopadhyay, 1989).

Australia (39-44).-

Neoproterozoic glacial rocks are widespread throughout Australia (Dunn et al., 1971): on or near the Kimberley block, within the "Centralian Superbasin" (Savory, Officer, Amadeus, Ngalia, and Georgina structural basins; Walter et al., 1995), and in the Adelaide "geosyncline" (Fig. 6). Most of these areas contain two glacial levels, commonly considered to represent an older Sturtian and younger Marinoan episode (c.f. Preiss and Forbes, 1981). Locally, each level may contain glacial-interglacial stratigraphic repetition (c.f. Brookfield, 1994). From present north to south they are described below, followed by a discussion of possible correlations and relevant paleomagnetic data.

In the Kimberley block and along its margins, several diamictites and overlying "cap" carbonates occur in discontinuous exposure (Fig. 6a,b). Each of the glacial formations except one (Fargoo) rests locally upon a striated pavement. Initial mapping of the Kimberley region led to the correlation described by Plumb (1981a), with the Marinoan episode represented by the Egan Formation, and the Sturtian episode represented by the Walsh, Landrigan, and Moonlight Valley/Fargoo Formations. The latter pair would record several glacial cycles within the Sturtian interval. An alternative correlation by Coats and Preiss (1980) and preferred by Kennedy (1996), assigned the Landrigan and Fargoo diamictites to the Sturtian episode, and the Walsh, Egan, and Moonlight Valley diamictites to the Marinoan interval. Recently, Plumb (1996) and Corkeron et al. (1996) have favored the first set of correlations but have shifted the entire stratigraphic package younger relative to the Sturtian and Marinoan intervals of central and southern Australia (Fig. 6b). Thus, there may be no Sturtian glacial deposit in the Kimberley region (except perhaps the Fargoo a.



b. Kimberley region

Coats and Preiss, 1980				Plumb, 1981			Plumb, 1996; Corkeron et al., 1996
Walsh	Egan	Moonlight Valley	M		Egan		p€/€
	Landrigan	Fargoo	St	Walsh	Landrigan	Moonlight Valley/Fargoo	M

c. central Australia

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

d. southern Australia

	Adelaide region	Broken Hill
M	Yerelina Subgroup	Teamsters Creek
St	Yudnamutana Subgroup	Yancowinna Subgroup

Figure 6. Correlation of Neoproterozoic glaciogenic deposits of Australia. (a) Reference map with Marinoan paleolatitudes according to Schmidt and Williams (1995), including possibly correlative deposits in other sectors of East Gondwanaland (restored according to Powell and Li, 1994). Solid (open) triangles represent deposits that are confidently (tenuously) correlated with the Marinoan. (b) Correlations within the Kimberley region of Western Australia. (c) Correlations within the "Centralian Superbasin" of central Australia (Walter et al., 1995). (d) Correlations within southern Australia (Coats and Preiss, 1987). M = Marinoan, St = Sturtian glacial epochs.

Formation) and the Egan Formation would solely represent a younger, only locally developed, glaciation occurring near the Precambrian-Cambrian boundary.

Farther to the southeast, a vast region of central Australia contains three major structural basins (Officer, Amadeus, and Georgina) and subsidiary outliers (e.g., the Ngalia basin) separated by E-W trending horsts which expose earlier Proterozoic crystalline basement. Neoproterozoic sedimentary deposits of the basins have been correlated into a "Centralian Superbasin" (Walter et al., 1995), following earlier correlations of lithology (Preiss and Forbes, 1981) and tectonic subsidence (Lindsay et al., 1987). As in the Kimberley region, glaciogenic deposits can be correlated into two episodes coinciding with the Sturtian and Marinoan ages. Both of these occur in the Amadeus (and Ngalia, cited in parentheses) basins, the lower represented by the Areyonga Formation and its equivalents (Naburula Formation), and the upper comprising the Olympic (Mount Doreen) Formation (Wells, 1981). Within any other individual basin, only one interval is preserved by true glaciogenic deposits. A correlative of the Areyonga Formation in the southern Georgina basin is the Yardida Tillite and its equivalents (Walter, 1981). Farther to the east, the Little Burke Tillite of the Mount Birnie beds are believed to correlate with the Marinoan interval based on their distinctive "cap" carbonate (Plumb, 1981b). On the westernmost side of the Centralian Superbasin, the Savory basin, the Marinoan interval is manifested by the glaciogenic Boondawari Formation (Walter et al., 1994). To the southeast, however, the Officer basin contains only a Sturtian representative, the Chambers Bluff Tillite (Preiss, 1993); this correlation is tenuous, however, due to 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).

The Sturtian and 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). The older, Sturtian episode is associated with iron-formation and commonly comprises two levels of diamictites (note that the stratigraphic names may include the term "Tillite" even though the rocks may be non-tillitic, glacially derived deposits). 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, Marinoan 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 (c.f. Williams and Schmdt, 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 equivalents.

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 (basement) and the Cambrian (overlying sedimentary rocks). Following Coats and Preiss (1987), Young (1992b) correlated the Yancowinna and Teamsters Creek deposits with the Sturtian and Marinoan, respectively, of the Adelaide geosyncline.

Farther south, on King Island and in Tasmania, sporadic outcrops of possibly Neoproterozoic diamictites occur (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 *Bavlinella faveolata*, widely considered diagnostic of the Vendian; however, this acritarch was also found in the Mineral Fork Formation of the western United States, possibly ~700-750 Ma (see above), as well as the Cambridge argillite from

Avalonia, which may be ~640 Ma (see below). The Cottons Breccia is overlain conformably or disconformably by thin, laminated dolostone which may correlate with the Nuccaleena "cap" of the Marinoan in the Adelaide geosyncline. Alternatively, thick basaltic and pyroclastic rocks which occur higher in the section suggest a correlation with the Chambers Bluff tillite in the Officer basin, a possible Sturtian equivalent (Coats and Preiss, 1987).

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 of the glaciogenic deposits can be ascribed to either a Sturtian or a Marinoan glacial epoch. The Rook Tuff, in the Callana Group two unconformities below Sturtian deposits, has yielded a U-Pb date of 802 ±10 Ma (Fanning et al., 1986). Ediacaran fauna occur several km above the Marinoan succession. Several whole-rock Rb-Sr ages on shale, of 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 Marinoan succession (reported by Coats, 1981; and Preiss, 1993). In addition, the Bunyeroo shale, ~1 km higher than the Marinoan deposits but below Ediacaran strata, has yielded a whole-rock Rb-Sr age of 588 ±35 Ma (cited by Preiss, 1993). Three detrital zircons from the Marino Arkose, lying stratigraphically between the Sturtian and Marinoan glacial deposits, yielded ages of ~650 Ma, which could provide a maximum date for the Marinoan if no significant Pb-loss had occurred (Ireland et al., 1998). In sum, the two glacial epochs are constrained in age only between 802 ±10 Ma and ~550 Ma, the onset of diverse Ediacaran assemblages (Grotzinger et al., 1995). Presumably, the Sturtian horizon is closer in age to ~800 Ma, and the Marinoan is closer to ~550 Ma (Table 1).

The Neoproterozoic glaciogenic deposits from Australia have universally yielded low paleomagnetic latitudes. Beginning with the northern deposits, if the Egan Formation is equivalent to the Julie Member of the Pertatataka Formation of the Amadeus basin

(Corkeron et al., 1996), then the paleopole from the latter region (Kirschvink, 1978b) determines a paleolatitude of 21 ±8° for the younger ice age recorded in the Kimberley area. A direct study of the Kimberley glacial deposits and adjacent units with "possibly primary" directions (McWilliams, 1977) yields generally consistent results but has not yet been published.

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 et al. (1980), which may not be useful as determined by Meert et al. (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 et al., 1992). MT1 may be a primary remanence, but it is similar to the poles from the Marinoan Elatina Formation and the Precambrian-Cambrian boundary (see below), possibly indicating remagnetization of the Sturtian rocks. The Marinoan interval, including the Mount Curtis Tillite, was 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. The paleolatitudes derived from these poles (Table 1) are a range of values extrapolated from the sampled region in the northern Flinders Ranges near Adelaide 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 et al. (1987) performed a fold test on allegedly soft-sediment features, indicating that the previously reported remanence was primary. This test was reproduced by Schmidt et al. (1991). The preceding data were discounted by Meert and Van der Voo (1994), who suggested that the Elatina result did not average paleosecular variation of the Earth's geomagnetic field; thus the observed paleolatitude could be as much as 20° in error from the true paleolatitude. That interpretation was met with some contention (Williams et al., 1995). Finally, Williams and Schmidt (1995) reported two polarities from a suite of additional outcrop and borehole samples, and Sohl and Christie-Blick (1995) have demonstrated stratabound polarity reversals through several stratigraphic sections spanning pre-, syn-, and post-glacial strata. These later studies show that the Elatina Formation yields a robust and primary paleomagnetic pole, indicating a depositional paleolatitude of 2.7 ±3.7° (Williams and Schmidt, 1995).

Lastly, McWilliams (1977) obtained more "possibly primary" paleopoles the Mount Birnie beds and the Cottons Breccia. All of the results are generally similar to the Elatina and Pertatataka poles described above, supporting the correlations of those deposits into Marinoan (Elatina-equivalent) and Precambrian-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 et al., 1988). The alleged glacial units are interbedded with mafic basalts dated at 762 ±90 Ma (Sm-Nd, Borg et al., 1990; with more conservative error estimates from Storey et al., 1992). To the contrary, Walker and Goodge (1994) found detrital zircons as young as 650-700 Ma from the Goldie Formation near the Beardmore Glacier, ca.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 Myr, 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 et al., 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 (Fig. 7). 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 et al., 1997). No diamictite or

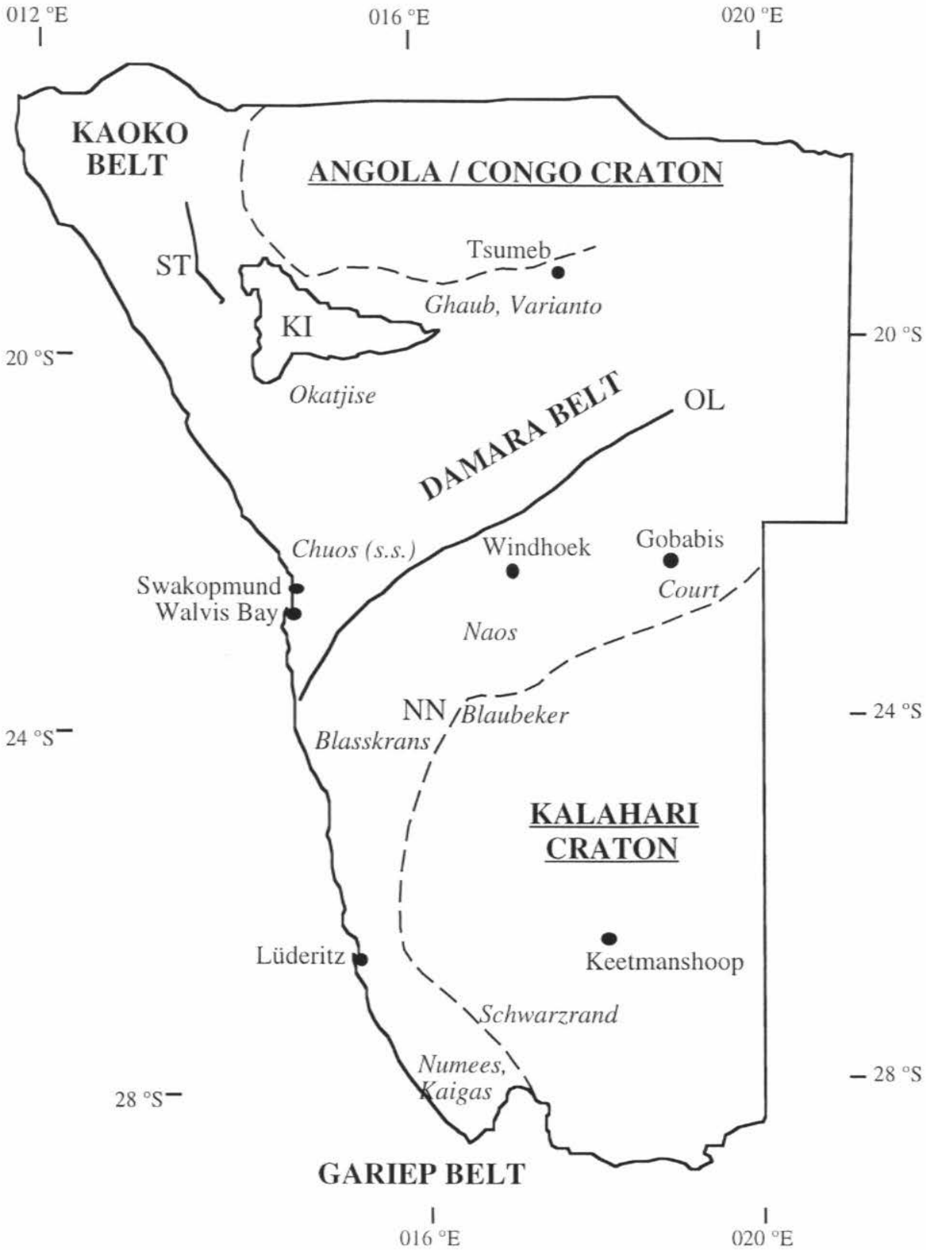


Figure 7. Location map for Neoproterozoic-Cambrian glaciogenic formations (names italicized) in Namibia. Solid curve labelled "OL" = Okahandja lineament; KI = Kamanjab inlier; NN = Naukluft nappes; ST = Sesfontein thrust. Dashed curves demarcate the approximate limits of Pan-African deformation related to the Kaoko, Damara, and Gariep belts.

tuff is present directly beneath the Bushmansklippe Formation, but an older glaciogenic unit occurs in that region, called the Court diamictite (Martin et al., 1985) or the Blaubeker Formation *sensu lato* (Kröner, 1981).

Within the allochthonous southern margin zone of the Damara orogen, the Naos Formation (formerly called the Chuos Pebbly Schist; Kröner, 1981) comprises mainly 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 et al. (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 yielding four single-zircon Pb-evaporation ages regressed to 741 ±6 Ma (Frimmel et al., 1996). This minimum age is slightly more restrictive than 717 ±11 Ma on Gannakouriep dikes (Reid et al., 1991) which locally intrude the Kaigas Formation and adjacent strata. Farther to the south, the 780 ±10 Ma Lekkersing granite (Allsopp et al., 1979) may be intrusive to or nonconformably underlying the Gariep Group (Frimmel et al., 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 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 contains both schistose diamictite and iron-formation, and underlies a "cap" carbonate thought to correlate with the post-Blasskrans carbonate (Hoffmann, 1990; Saylor et al., 1995a). 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 et al., 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 et al., 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 et al., 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).

Summary of correlations and paleomagnetic results.-

For correlation of the glacial deposits on the periphery of the Damara-Gariep belts, 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, and nonconformably underlying Namaqua gneisses (Frimmel et al., 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 similar to Cambrian-Ordovician poles from Gondwanaland (Kröner et al., 1980). The second glacial episode lies unconformably below basal Nama Group sediments dated at ~549 Ma (Grotzinger et al., 1995) and is perhaps significantly younger than the ~741 Ma Rosh Pinah volcanics. No paleomagnetic data exist for the younger Neoproterozoic glaciogenic deposits.

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 Ma and 539.4 ±1 Ma (U-Pb on zircon from interbedded ash layers; Grotzinger et al., 1995). On the other hand, Saylor et al. (1995b) 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 Subgroup were first analyzed paleomagnetically by Kröner et al. (1980), who observed three imprecise clusters of bipolar directions. Subsequently, Meert et al. (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 of 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 palolatitude 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 (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 (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?). It contains 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 preferred 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 with the lower Kundelungu extensional event. A much younger age for the Kundelungu glacial deposits may come from the Zambezi belt, where a metamorphosed supracrustal succession resembling the Roan Supergroup contains basal metarhyolites dated by U-Pb zircon at 879 ±19 Ma (R. Warslaw, unpub., cited by Hanson et al., 1994). With such a correlation, this young age would invalidate the older K-Ar determination. 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 et al., 1961; decay constants updated by Cahen and Snelling, 1984, p.132).

In eastern Kivu, Zaire, the basal diamictite unit of the Tshibangu Group (Cahen et al., 1979a) contains clasts of the ~960-970 Ma pegmatites (Romer and Lehmann, 1995). The diamictite was deformed and then intruded by the 739 ±7 Ma Ruvubu syenite (U-Pb on zircon; Tack et al., 1984). If the Tshibangu diamictite is glaciogenic and correlates with the Grand Conglomérat as Cahen (1982) suggests, then the ~740-Ma minimum age may constrain Neoproterozoic glacial deposits throughout the eastern Congo craton. However, the Tshibangu diamictite may be younger, perhaps correlative with the Petit Conglomérat.

Northern Congo/São Francisco (55, 56).-

Farther north and west, the Bunyoro diamictite in western Uganda (Byørlykke, 1981) has been correlated with the Luma and Loyo Group (Cahen and Snelling, 1984, p. 246) and also the Akwokwo "Tillite" (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 Luma-Loyo successions are apparently involved in deformation coeval with the sub-Tshibangu unconformity (Cahen and Snelling, 1984, p. 205). They may correlate with the Grand Conglomérat, permitting the Tshibangu-Petit Conglomérat correlation suggested above.

In the Lindian Supergroup, the Akwokwo diamictite probably correlates with a diamictite in the Bakouma region and the Bandja "tillitic complex" in the Sembé-Ouesso region (Cahen, 1982; Cahen and Snelling, 1984, p. 246). A higher, allegedly "fluvio-glacial" level, overlain by a pink carbonate rock resembling the "cap" of the Petit Conglomérat, also occurs near Bakouma (Bonhomme and Weber, 1977; Cahen and Snelling, 1984, p. 245). For the last step to the West-Congo foldbelt, Cahen (1982), Cahen and Snelling (1984, p. 175), and Trompette (1994, p. 220) equate the Bandja diamictite-volcanic association with the lower diamictite or both diamictites in the West-Congolese Supergroup.

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, correlated with strata above the Akwokwo and Bandja diamictites, (Bonhomme and Weber, 1977, using $\lambda^{87}Rb = 1.42 \times 10^{-11}\ yr^{-1}$) could provide a minimum age for the glacial rocks, although the Rb-Sr chronometer on shale is suspect without independent verification. A rough 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 E-W trending Oubanguide-Sergipe orogen, possibly correlative diamictites occur within the Sergipe and Rio Preto belts (Trompette, 1994, p. 217 and 244). The latter region contains a transition of facies from the cratonic equivalents of the Bebedouro Formation and neighboring strata (see below) to correlative basinal deposits. Recognizing the potential for stratigraphic revisions along the northern border 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. For all 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 glacial deposits may be reevaluated with further work.

São Francisco craton and southern marginal foldbelts (57-60).-

Glaciogenic rocks occur within and along all of 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 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). The latter succession, commonly associated with iron-formation or manganese concentrations, constitutes an important stratigraphic marker within the Araçuai foldbelt (Trompette, 1994, p. 141) and is most likely correlative with the Jequitaí Formation. Elsewhere within less-deformed regions of the craton, on the Chapada Diamantina to the northeast, 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). Even farther to the 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 Brasilia 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 amounts of tectonic imbrication. These sedimentary rocks, deposited in more distal regions from the craton, may be glacial 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 of 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 et al. (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 (Machado et al., 1989). This contradicts the commonly quoted age of ~950-1000 Ma for the Jequitaí Formation and correlative glacial rocks of the São Francisco craton (Bonhomme et al., 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 (Iyer et al., 1995) suggests correlation with post-Rapitan (younger than ~750 Ma; see above) carbonates in the Mackenzie Mountains of Canada (Kaufman et al., 1995). For a minimum age constraint, the São Francisco Supergroup is involved in Brasiliano orogenesis (Chemale et al., 1993) dated at ~580 Ma from the Ribeira belt (Machado et al., 1996) and at ~600 Ma from the internides of the Brasilia belt (Pimentel et al., 1996).

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 grey, 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 (*e.g.*, 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 Aracuai 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 1027 ±56 Ma on the unconformably underlying Mativa granite (strongly discordant array of U-Pb results; Cahen et al., 1978) 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), ~730-740 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 et al., 1979b), or 604 ±58 Ma (lower intercept of strongly discordant U-Pb results from gneiss within the internal zone; Maurin et al., 1991). Trompette (1994, p. 153) opts for dating the West Congo orogen at 600-620 Ma, primarily using constraints from its Brazilian counterpart, the Aracuai belt.

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 occurring 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; P. Hoffman, pers. comm.). Farther west, across the Sesfontein thrust, the schistose diamictite formerly called "Chuos" but probably correlative with the Ghaub Formation (Hoffman et al., 1994), is associated with iron-formation (Henry et al., 1993; Dingeldey et al., 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 similar level 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), and an iron-rich tilloid occurs at a similar level in a more internal zone immediately west of the Sesfontein thrust (Henry et al., 1993). The lower, ferruginous diamictite, retains the name "Chuos" in the nomenclature of Hoffmann and Prave (1996).

South of the Kamanjab inlier, the Okatjise Formation contains a conglomerate/diamictite which is traditionally placed at the Varianto/Chuos *s.s.* stratigraphic level (Miller, 1974). The formation is sandwiched between volcanic rocks dated at 746 ±2 Ma below and 747 ±2 Ma above (Hoffman et al., 1996); if the correlations are correct, then the older northern Namibian glaciation is very precisely dated. Nonetheless, the unit mapped as "Chuos" in this region, which immediately overlies the upper volcanic deposit, may be correlative with either the Chuos *s.s.* or the Ghaub. Given the ongoing stratigraphic revision in this region, I have provided conservative age estimates in Table 1. A younger age constraint for the two glacial successions is provided by the crosscutting Omangambo pluton dated at 589 ±40 Ma (discordant U-Pb on zircon; Miller and Burger, 1983).

Paleomagnetic constraints.-

Most of the glaciogenic or alleged glacial deposits from the Congo/São Francisco craton are too poorly constrained in age for application of one of the few reliable paleomagnetic results available. For example, in east-central Africa the Bukoban

Supergroup has paleomagnetic constraints (Meert et al., 1995) but is one of the few basins in that region which lack a glacial horizon; in addition, no definitive correlation exists between the Bukoban and the diamictite-bearing successions.

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 et al., 1998), and the other (NQ2) similar to the Cambrian-Ordovician boundary magnetostratigraphic pole from Queensland, Australia (Ripperdan et al., 1992; restored to Africa). 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. Evans, unpublished data). Paleolatitudes from these results are computed in Table 1 for a reference locality at 20°S, 015°E.

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 et al. (1990) are older than 1000 Ma and hence predate the glacial deposits by >100 Myr if the pre-glacial date of 906 ±2 Ma (Machado et al., 1989) is correct. From across the Congo craton, paleomagnetism of the post-Ubendian Mbozi complex (Meert et al., 1995) may provide a paleolatitudinal constraint on the Chuos Formation. A cooling age of 755 ±25 (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. The paleomagnetic results indicate a 10 ±5° paleolatitude for the Chuos Formation (calculated for a reference locality at 20°S, 015°E).

West Africa

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. Within this succession are several levels of glaciogenic strata, including a well dated Upper Ordovician deposit. 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. Correlations from this region 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 et al., 1982), ages which I consider suspect until verified independently by other isotopic systems. Otherwise, aside from a general collection of Neoproterozoic microfossils within the sedimentary succession (Trompette, 1994, p. 50), the only direct age constraints for the Bthaat Ergill and Jbeliat Groups are their nonconformity over Paleoproterozoic basement of the Reguibat shield and their disconformity below indubitably Upper Ordovician strata.

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. 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 Early Proterozoic rocks of the northeast Reguibat shield (Trompette, 1994, p. 23).

Farther to the northeast, a recent discovery of simple Ediacaran discoid fossils from strata unconformably underlying a "triad"-like succession of the Fersiga Group (Bertrand-Sarfati et al., 1995) has apparently provided a new, late Vendian or possibly Cambrian age for the entire Fersiga-Bthaat Ergill glaciogenic succession (Moussine-Pouchkine and Bertrand-Sarfati, 1997). Such an age would conflict with the previously mentioned Rb-Sr ages from the Adrar of Mauritania, but it is more consistent with fossil findings from the southwestern margin of West Africa, as discussed next.

Taoudeni basin, southwest (67).-

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 et al., 1989; Proust and Deynoux, 1994). Similarity between this "triad"-like sequence and those of the northwest Taoudeni basin suggests correlation.

Still farther to the southwest, 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 Atdabanian age (Culver et al., 1988). Some debate regarding these ages has arisen; if one assumed that the glacial rocks, at the base of the Mali Group, were deformed by 550-560-Ma tectonism (Dallmeyer and Villeneuve, 1987), this required more than ~30 Myr of 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). The 550-560-Ma deformation in the Mauritanides is, however, not very prevalent in the area near autochthonous Mali Group deposits (Dallmeyer and Villeneuve, 1987). Moreover, because the entire "triad" is involved in that episode of folding, then any deformation of the Mali Group <u>must</u> be younger than Early Cambrian, in light of the new fossil findings (noted by Trompette, 1996, 1997). In that case, the diamictite's age is not tightly constrained by Mauritanide deformation; the glacial unit may also be of Early Cambrian age, obviating the need for a cryptic unconformity separating it from its post-glacial sequence. This same conclusion was attained on different grounds by Bertrand-Sarfati et al. (1995) as discussed above.

Volta basin and Dahomeyides (68, 69).-

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 634-607 Ma ($^{40}Ar/^{39}Ar$ on muscovites of the northern region; Attoh et al., 1997), providing a minimum age for the Oti-Pendjari equivalents that is supported by identification of *Chuaria* sp. in post-glacial strata (Amard, 1992).

Periglacial and/or glacial deposits are also cited for basal strata in the unconformably overlying Tamale Group, considered as a molassic unit to the Dahomeyide deformation, of probably Neoproterozoic-Cambrian but possibly middle Paleozoic age (Trompette, 1994, p. 57; Villeneuve and Cornée, 1994).

Summary of correlations.-

As shown above, direct age constraints for the Neoproterozoic-Cambrian glacial deposits of West Africa are scarce or one-sided. Combining results from the Hoggar shield and 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 et al., 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 Ergill-Jbeliat, Fersiga, Bakoye, Mali Groups, all equivalent to the Série Pourprée (see below) with an Early Cambrian age (Bertrand-Sarfati et al., 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 Table 1. 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.

Paleomagnetic constraints.-

Direct paleomagnetic study of the Adrar de Mauritania spanned the glacial horizons (Perrin et al., 1988). 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 the Bthaat Ergill glaciogenic strata. Samples from the Bthaat Ergill Group itself yielded a direction that Perrin et al. (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 Kiaman reversed superchron. Results from the middle of the Atar Group, although judged by Perrin et al. (1988) as probably primary, was 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 which conformably overlie the glaciogenic Bthaat Ergill Group and conformably underlie fossiliferous Cambrian-Ordovician strata (Perrin et al., 1988). This result carries two polarities and is unlike any other accepted Phanerozoic paleopole from Africa or Gondwanaland, both factors suggesting a primary magnetization. Its anomalous position relative to other Cambrian paleopoles from Gondwanaland (c.f. Meert et al., 1995; Meert and Van der Voo, 1997; Kirschvink et al., 1997; Evans et al., 1998), however, renders it suspect. I include this direct constraint from the Adrar de Mauritanie 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 (Meert and Van der Voo, 1997; Kirschvink et al., 1997).

Because the polar motion is so rapid, a Nemakit-Daldynian vs. Atdabanian age of the "triad" can make the difference between deposition in moderate vs. high latitudes (Fig. 8). Better constrained ages, or more direct and reliable paleomagnetic studies, of the West African glacial deposits are required to distinguish among these possibilities.

Anti-Atlas Mountains (70).-

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 et al., 1992). The Tiddiline tilloids contain clasts of the Bleida granodiorite dated at 615 ±12 Ma (U-Pb, Ducrot, 1979) or 608 ±11 Ma (Rb-Sr, unpublished, cited by Hefferan et al., 1992). 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 et al., 1983).

Hoggar shield (71, 72).-

Traditional stratigraphic summaries of the Pan-African sedimentary succession within the Pharuside belt along the northeast 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). Its age is bracketed between

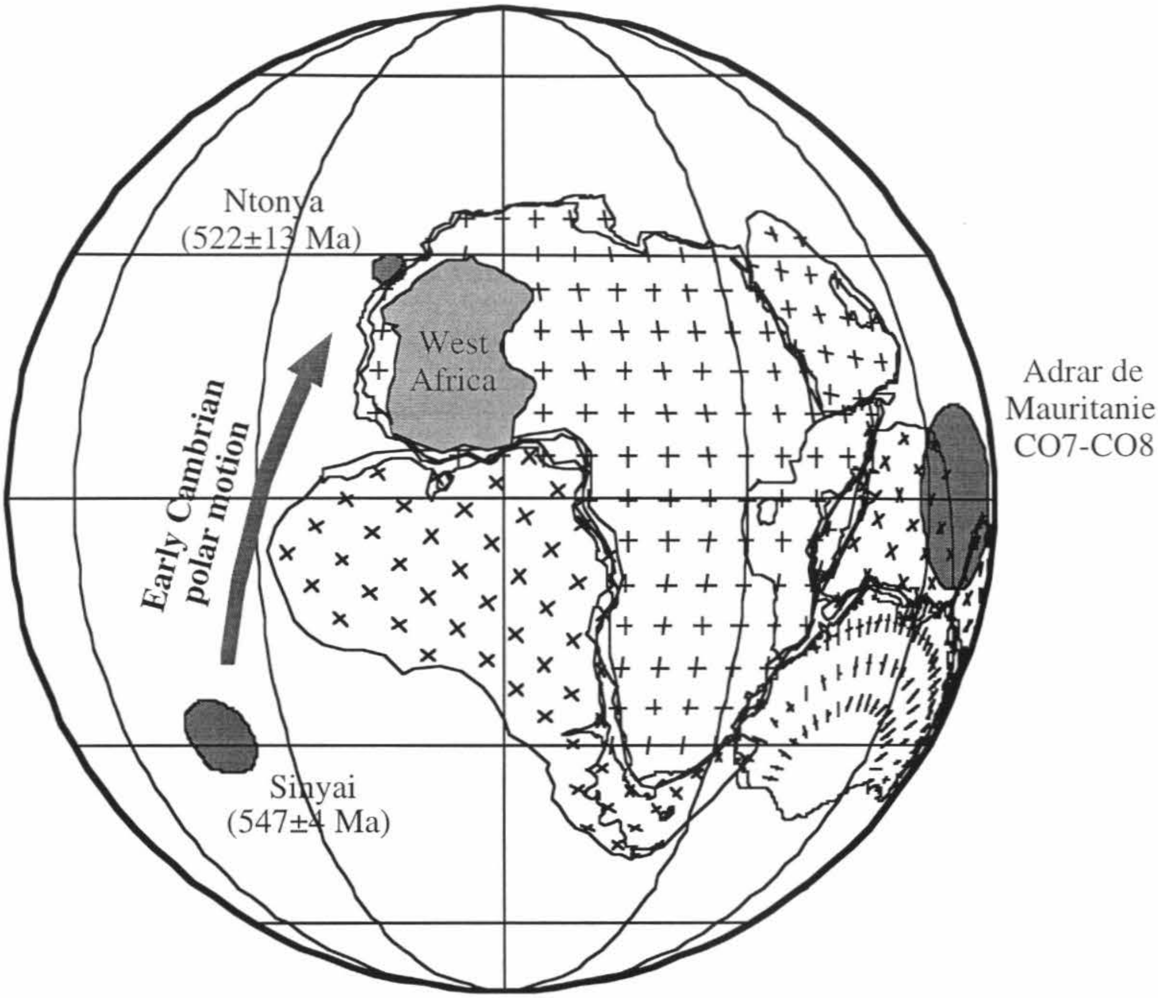


Figure 8. Terminal Proterozoic to Early Cambrian paleomagnetic poles for Gondwanaland, showing rapid poleward motion of the West African craton. The Sinyai metadolerite pole is from Meert and Van der Voo (1996), the Ntonya ring structure pole is from Briden et al. (1993), and the anomalous Adrar de Mauritanie pole, shown only for comparison, is from Perrin et al. (1988). Gondwanaland reconstruction in African coordinates, from Powell and Li (1994). Orthographic projection, grid lines spaced 30° apart.

unconformably underlying quartz diorites at 696 +8/-3 Ma (concordant U-Pb on zircon; Caby and Andreopoulos-Renaud, 1985) and intruding diorite and quartz monzonite respectively at 616 ±11 and 613 ±3 Ma (unpublished U-Pb zircon data cited by Black et al., 1979). A possibly correlative succession in the eastern Hoggar, the Tirririne Formation (Caby and Fabre, 1981a), is older than sills dated at 660 ±5 Ma (U-Pb on zircon; Bertrand et al., 1978). The Tirririne succession contains distinctive conglomerates of disputed glacial origin (Trompette, 1994, p. 253).

The Série Pourprée comprises two stratigraphic entities, the lower Tagengan't and the upper Ouellen-in Semmen units (Deynoux et al., 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 et al. (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 620-580 Ma (Rb-Sr and U-Pb using outdated decay constants, no isotopic data presented; Allègre and Caby, 1972), 560 ±10 Ma (ibid., 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 et al., 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 using updated decay constants) or 519 ±11 Ma (ibid., cited by Allègre and Caby, 1972; also recalculated).

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 et al., 1979).

A paleomagnetic study of the Adma diorite/adamellite intrusion, a post-Série Verte plutonic suite dated at ~615 Ma (unpublished U-Pb ages cited by Black et al., 1979), 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 (Meert et al., 1995; Meert and Van der Voo, 1997; Evans et al., 1998).

Rokelides (73).-

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 cm-diameter outsized clasts in laminated mudstone (Culver et al., 1980). The Tabe Formation is nonconformable atop Neo-Archean ("Liberian") crystalline rocks and is deformed by folding of the Rokelide orogen, marked by a host of ~500-550-Ma whole-rock Rb-Sr and K-Ar mineral ages from nearby basement rocks (Cahen and Snelling, 1984, p. 327). 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.

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, U.S.A., 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). 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 et al., 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 correlated with the Cambridge Argillite, for which a latest Precambrian age is suggested by microfossil assemblages, including *Bavlinella faveolata* (Lenk et al., 1982). A recent U-Pb date (abstract only) of 643 ±6 Ma from the type locality of the Cambridge Argillite, however, suggests that it can no longer be correlated with purple mudstones overlying the glaciogenic Roxbury Conglomerate (Johnston et al., 1995); the older age for the Cambridge Argillite is supported by the possibility that *Bavlinella* acritarchs may have a lower stratigraphic range than purely Vendian (Knoll et al., 1981). Contacts between the Boston Bay Group and the Lower Cambrian North Attleboro and Weymouth Formations are not exposed. Hence, there is effectively no younger biostratigraphic age constraint on the Squantum tilloid; a structural lower age limit is provided by post-Westphalian thrust faults, associated with the latest Carboniferous to early Permian Alleghanian orogenic episode, which also cross-cut the Squantum unit (c.f. Fig. 1 of Smith and Socci, 1990). The Squantum member is cleaved to much greater extent than the 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 et al., 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 et al., 1986) and paleomagnetic results from other Avalonian terranes; however, it is noted that Avalonia occupied similar moderate-high latitudes in Llanvirn time (Mac Niocaill et al., 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 comprise slump deposits from an unstable volcanic-arc setting, whose glacial influence was restricted to higher elevations.

The Gaskiers Formation is well dated between the nearly immediately underlying Harbour Main Group at 606 +3.7/-2.9 Ma (U-Pb on zircon; Krogh et al., 1988) and the more distantly overlying Mistaken Point Formation at 565 ±3 Ma (U-Pb on zircon with no isotopic data given; Benus, 1988). A tighter upper age constraint may be provided by a late-stage rhyolitic dike intruding the Harbour Main Group (585.9 +3.4/-2.4 Ma; Krogh et al., 1988), but correlations between this region and those containing glaciogenic strata are tenuous (A.F. King, pers. comm.).

Direct paleomagnetic study of the Gaskiers Formation has yielded disappointing results. Gravenor et al. (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 et al., 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 et al., 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 et al., 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 et al., 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 et al., 1990). Both of the sedimentological studies by Doré et al. (1985) and Eyles (1990) refuted the existence of any glacial influence for the Granville Formation, and so it will not be discussed further here.

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 (NW) to the more internal zone of the orogen (SE). Many of these deposits are associated with sedimentary iron- or manganese-formation, including the banded manganese ores of Urucum (Jacadigo Group). The glacial deposits are conformably overlain by the Araras Formation, a carbonate-rich succession containing Ediacaran-like fauna (summarized by O' Connor and Walde, 1985). Trompette (1994, p. 75) cites a Rb-Sr age of 623 ±15 Ma on quartz porphyry below the Boqui Group, similar to locally occurring volcanic rocks found directly under glaciogenic deposits in the Paraguay belt. If this age is reliable, and if the Jangada-Puga-Jacadigo-Boqui glacial deposits are coeval, then they are relatively well constrained in age.

A direct paleomagnetic study of the Jacadigo Group was undertaken as early as Creer (1965), who thought the Urucum deposits were Silurian. His 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 Late Ordovician age from other Gondwanaland continents. The Silurian age mis-assignment 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 et al., 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 et al., 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.

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 15 have somewhat reliable paleolatitudinal constraints (Fig. 9). 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 Appalachians (*11*), Vestertana Group (*20*), Tarim basin (*33*), Chang'an Formation (*35*), Egan Formation (*41*), Elatina Formation (*43*), Schwarzrand Subgroup (*51*), Chuos Formation (*63*), Jbeliat/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 probably primary. Some entries in Table 1 are queried as possibly primary but are omitted from the above subset. This is because of too

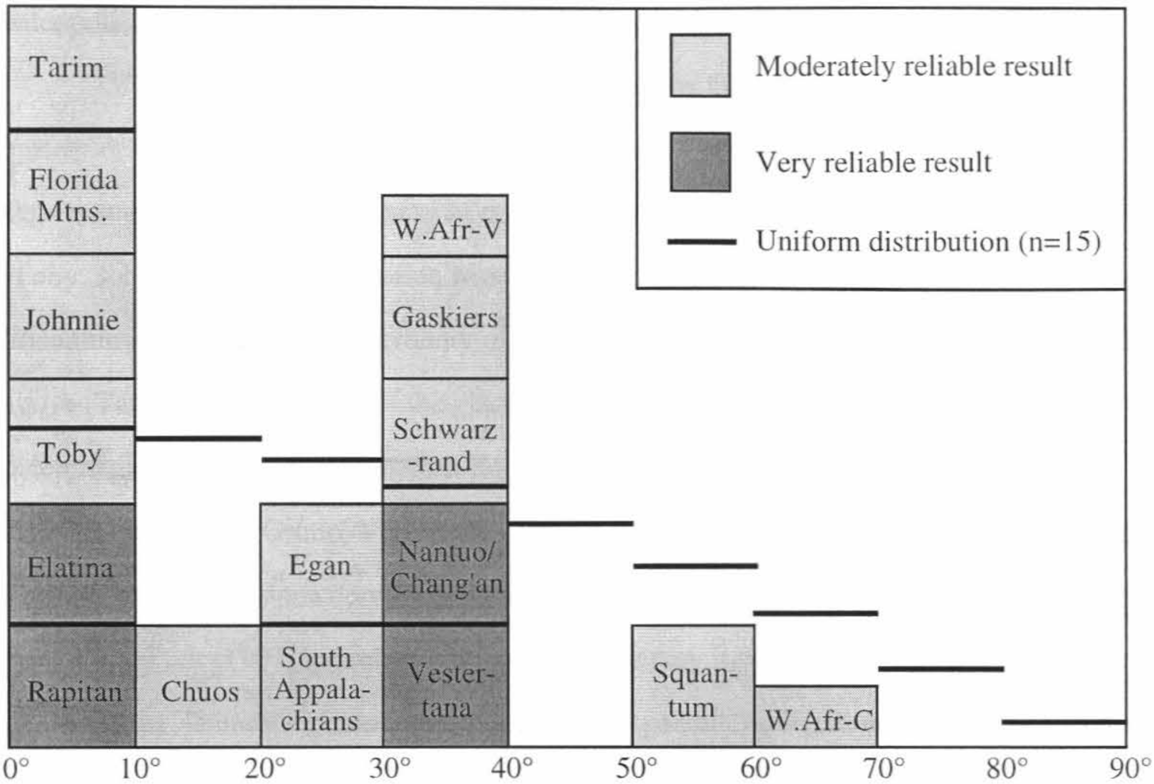


Figure 9. Histogram of moderately (light shade) and very (dark) reliable paleolatitudinal estimates of Neoproterozoic-Cambrian glaciogenic deposits. Unit area is assigned to each deposit. For comparison, the discontinuous steps show the expected (non-integral) density function of a uniform distribution over the sphere.

much stratigraphic distance separating the glaciogenic unit and the paleomagnetically studied formation (Ice Brook Formation, *#2*; Série Verte, *#71*), uncertain age of the glaciogenic unit (Moelv "tillite," *#23*; East European platform deposits, *#24*), or a paleopole of low reliability similar to much younger paleopoles (Sturtian pole MT1, *#42*).

Of the 15 "finalists" whose paleolatitude may be determined accurately by the present database, I consider only four 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 four are the Rapitan Group, Vestertana Group, South China glaciogenic units, and Elatina Formation plus 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 vs. diagenetic remanence, both were acquired when that part of Laurentia was near the Equator. The Vestertana Group pole is pre-folding (Torsvik et al., 1995a), but it is similar to Ordovician poles from Baltica. Regardless, special circumstances would be required for it to be an overprint. The Liantuo Formation pole (Chapter 8) and the Chengjiang/Nantuo pole (Zhang and Piper, 1997) are consistent with each other and support the notion that the Chang'an and Nantuo glacial epochs are similar in age. Lastly, the Elatina pole (Schmidt and Williams, 1995; Sohl and Christie-Blick, 1995) is one of the most robust Precambrian paleomagnetic results, having passed every test performed on it and been determined independently by three laboratories. If the Marinoan correlations into central and northern Australia are correct, then this result implies extensive equatorial glaciation on that continent.

Figure 9 summarizes these interpretations. Each paleolatitude determination is given unit area except for the Jbeliat/Fersiga Groups which are split into equal halves, 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 15 results shows two peaks, at 0-10° and 30-40° paleolatitude; 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 Jbeliat/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, not representative of continental ice sheets. The bimodal data distribution from the subset of 15 entries persists even among the more stringently chosen subset of four, although with so few data it is difficult to believe that the modes are meaningful. Still, none of the four very 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 three of the four most reliable determinations.

The culled dataset suggests that the Phanerozoic archetype 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 non-dipole geomagnetic field components (Kent and Smethurst, in preparation) or inclination-shallowing due to sediment compaction (all four of the most reliable studies deal with fine-grained siliciclastic rocks) might have deflected the distribution significantly. At present, both the Snowball Earth and high-obliquity models are acceptable according to the paleomagnetic data.

Temporal Distribution of Neoproterozoic Glacial Deposits

Is a general threefold division of ~900 Ma, ~800 Ma, 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: ~740 Ma (all continents), ~600-620 Ma (most continents), ~570-580 Ma (Avalonia-Cadomia only?), and ~545 Ma (several continents). These groupings are tentative because so many of the glacial units are poorly dated. As stated above, carbon-isotope stratigraphy has potential as a Neoproterozoic chronometer (e.g., Kaufman and Knoll, 1995); however, this paper has ignored those data in order to present the temporal constraints with as little interpretation as possible.

Of the four most reliable paleolatitude determinations discussed above, two (Rapitan, Chang'an/Nantuo) are commonly grouped with the ~740-Ma cluster of glaciogenic deposits, and the other two (Vestertana, Elatina) are commonly grouped with a ~620 Ma age. Those correlations may be correct, and if so, 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 upcoming 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 requires 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.

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 ~740-Ma group, but others, particularly in South America and southwest Africa, are demonstrably younger. In many examples (e.g., around the São Francisco craton), appearance of iron-formation is related to a facies transition to a more distal, deep-water environment.

SUMMARY 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 negate the 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 (i.e., 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 may exist, including non-dipole field effects biasing the data (Kent and Smethurst, in preparation), or, if the data are unbiased, any combination of processes such as lower solar luminosity (Crowley and Baum, 1993), tectonically induced CO_2 drawdown (Young, 1991), or other factors.

A new item of interest is the possibility of substantial true polar wander (TPW) during late Neoproterozoic time, as a consequence of a long-lived, rotationally unstable pattern of mantle convection induced by the long-lived Rodinian supercontinent (Evans, 1998). The geodynamical theory of TPW permits large shifts as rapid as 90° per 5-10 Myr (Steinberger and O' Connell, 1997). Rapid TPW can explain the abrupt transition into and out of Neoproterozoic glaciations (Fairchild, 1993), but it cannot explain low paleomagnetic latitudes obtained directly on the glacial deposits (e.g., 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 exposures 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 TPW or plate-tectonics alone (Gurnis and Torsvik, 1994)--could create a strong positive feedback toward the growth of ice sheets.

If Precambrian paleoclimate is governed by non-uniformitarian processes or boundary conditions such as presented in the high-obliquity hypothesis, then the abrupt transition to the Phanerozoic archetype is intruiging. 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 et al., 1995) and occupied polar latitudes would extend the Phanerozoic paleoclimatic paradigm to that period as well. If the Egan Formation is latest Proterozoic in age (Corkeron et al., 1996; Plumb, 1996), then its low paleolatitude would mark the end of the Precambrian climatic system. Dynamic models of planetary obliquity may be able to determine whether such a rapid transition is physically plausible.

More than fifteen years 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 five years. We are 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. Additional geochronological studies (using advanced techniques to measure very small mineral samples with great precision), new methods of chronostratigraphy (e.g., secular variations in isotopic ratios of sedimentary rocks) 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 upcoming years.

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Chapter 7: Paleomagnetic results from basal Sinian sediments of the South China block

Abstract

Neoproterozoic (Sinian) sediments are exceptionally well-preserved in the Three Gorges region (western Hubei Province) of the South China block. A paleomagnetic study of 143 samples from the basal Sinian Liantuo Formation at its type locality, using thermal demagnetization and least-squares line analysis, reveals three distinct magnetic components among the suite of samples. The most stable of these is a two-polarity remanence, residing in hematite, with a change in polarity that is stratigraphically consistent among three outcrops. The corresponding tilt-corrected paleomagnetic pole is from 69 directions that are free of contamination by overprints ("A"; 01.5°N, 161.7°E, dp=1.8°, dm=2.6°, Q=7).

Two such overprints can be distinguished from each other by laboratory unblocking temperatures. The first to be removed ("C"), always annihilated below 600 °C, is common throughout the dataset but is amenable to least-squares line-fitting in only 37 samples. It yields a pole which in present coordinates resembles Mesozoic overprints identified from previous studies in the Three Gorges region (73.6°N, 177.1°E, dp=6.0°, dm=8.3°, Q=4). The higher unblocking-temperature overprint ("B"), always subsidiary to the "A" component, is more prevalent than "C" and identified by line-fitting in 67 samples. The "B" direction is very steep and generates a paleopole whose *in situ* coordinates do not resemble the Mesozoic-Cenozoic apparent polar wander path for South China, and whose tilt-corrected coordinates (20.6°N, 105.5°E, dp=7.2°,dm=7.3°, Q=5) bear no resemblance to any reliable Phanerozoic paleopoles from the South China block. The steep "B" direction was probably acquired sometime in the 200-Myr interval between deposition of the Liantuo Formation at ~750 Ma and Cambrian time.

The "A" pole is distinct from the allegedly primary pole from correlative strata in Yunnan, which may be explained by the streaked distribution of the Yunnan dataset hampering the isolation of magnetic components within those samples. The "A" pole from this study is considered to be primary based on its thermal stability and its magnetostratigraphic consistency, and constrains the depositional paleolatitude of the

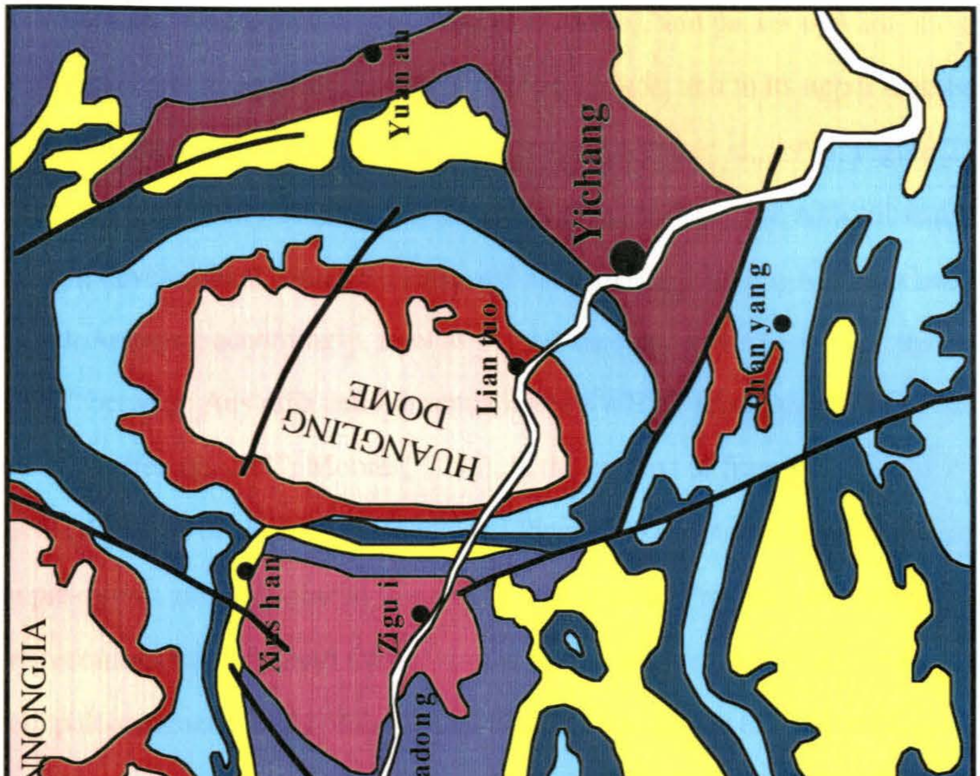
Liantuo Formation to 33.8 ±1.8°; this may also apply to the stratigraphically adjacent Chang'an glacial deposits or the disconformably overlying Gucheng-Nantuo glacial deposits. South China's position in Rodinia has been debated; the present study provides constraints on the various proposed reconstructions. In particular, South China may have lain between Australia and Laurentia only in an orientation different than originally proposed in the "missing link" hypothesis. Likewise, a paleoposition adjacent to northwestern Australia requires a specific relative orientation between the two blocks.

Introduction

Sub-Cambrian, unmetamorphosed, stratified rocks are widespread throughout China (Grabau, 1922). Grouped into the Sinian System, the deposits occur both north and south of the Qinling-Dabie Mountains (Wang, 1986), which contain the Mesozoic suture between the North China and South China blocks (Enkin et al., 1992). In the South China block (SCB), Sinian strata occur among several provinces; the type section is in the lower (Xiling) of the Three Gorges of the Changjiang (Yangtze River).

There, a structural dome affecting all pre-Cretaceous units and therefore related to the so-called Indosinian orogeny during late Mesozoic time, exposes the regional stratigraphy in "onion-skin" fashion around the Huangling dome (Fig. 1). Pre-Sinian basement in the core of this feature consists of the Huangling batholith intruding Archean to Paleoproterozoic cyrstalline rocks (Wang et al., 1996). The phase of the batholith which nonconformably underlies Sinian strata has been dated by the SHRIMP U-Pb method on zircon at 819 ±7 Ma (Ma et al., 1984). The basal Sinian, arkosic to argillitic Liantuo Formation contains interbedded volcanic ash horizons, one of which was dated at 748 ±12 Ma (Ma et al., 1984). Overlying the Liantuo Formation with apparent disconformity, the Nantuo Formation is the classic Sinian "tillite"; it actually spans a range of glacial-sedimentary facies throughout the SCB (Wang et al., 1981; Liao, 1981). The Doushantuo Formation, primarily muddy dolostone with minor shale and bedded or nodular chert, is

Figure 1 (next page). Geology of the Three Gorges of the Changjiang, after Wang et al. (1996). Liantuo is located at (30° 51' N, 111° 09' E); Yichang at (30° 42' N, 111°, 19' E). Z = Sinian, C = Cambrian, O-S = Ordovician and Silurian, D-P = Devonian through Permian, Tr = Triassic, J = Jurassic, K = Cretaceous, Cz = Cenozoic.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apparently disconformable upon the Nantuo diamictites. It contains the Miaohe biota of Ediacaran or perhaps pre-Ediacaran age (Ding et al., 1996), and passes upward into the Dengying Formation bearing Ediacaran fauna, Vendotaenids, and in its upper member, small skeletal fossils marking the lowermost Cambrian (Wang et al., 1996; Figure 2).

The Sinian has been interpreted as a continental-rift succession with subsequent passive-margin development, similar in style and age to those found in Australia and western North America; accordingly, Li et al. (1995) reconstructed the SCB as the "missing link" between Australia and Laurentia in the SWEAT reconstruction (Eisbacher, 1985; Bell and Jefferson, 1987; Moores, 1991), in the context of fragmentation of Rodinia (Powell et al., 1993; Li et al., 1996). Zhang and Piper (1997) obtained a paleomagnetic pole from pre-, syn-, and post-glacial Sinian strata in Yunnan Province and proposed an alternative reconstruction of South China, similar to that suggested by Kirschvink (1992a). The present paleomagnetic study of the Sinian deposits was aimed to resolve these discrepancies, as well as contribute one more datum to the spatial distribution of widespread Neoproterozoic glacial deposits, the subject of a long-standing debate (Harland, 1964; Schermerhorn, 1974; Williams, 1975; Crowell, 1983; Hambrey and Harland, 1985; Embleton and Williams, 1986; Kirschvink, 1992b; Meert and Van der Voo, 1994; Williams et al., 1995; Evans, 1997; Chapter 7).

Paleomagnetic field and laboratory methods

Reconnaissance paleomagnetic work by J.L. Kirschvink in 1986 led to the first collection of 40 samples from the upper Liantuo Formation (ULF). Laboratory analyses of these specimens in 1995 and early 1996 revealed a change in magnetic polarity at one level within the stratigraphic section; the second sampling trip to the Three Gorges in 1996 (103 samples from the upper Liantuo Formation) was devoted primarily toward an attempt to reproduce the polarity zonation in an independent section of the same lithostratigraphic interval, as well as perhaps to obtain transitional directions from within a Precambrian

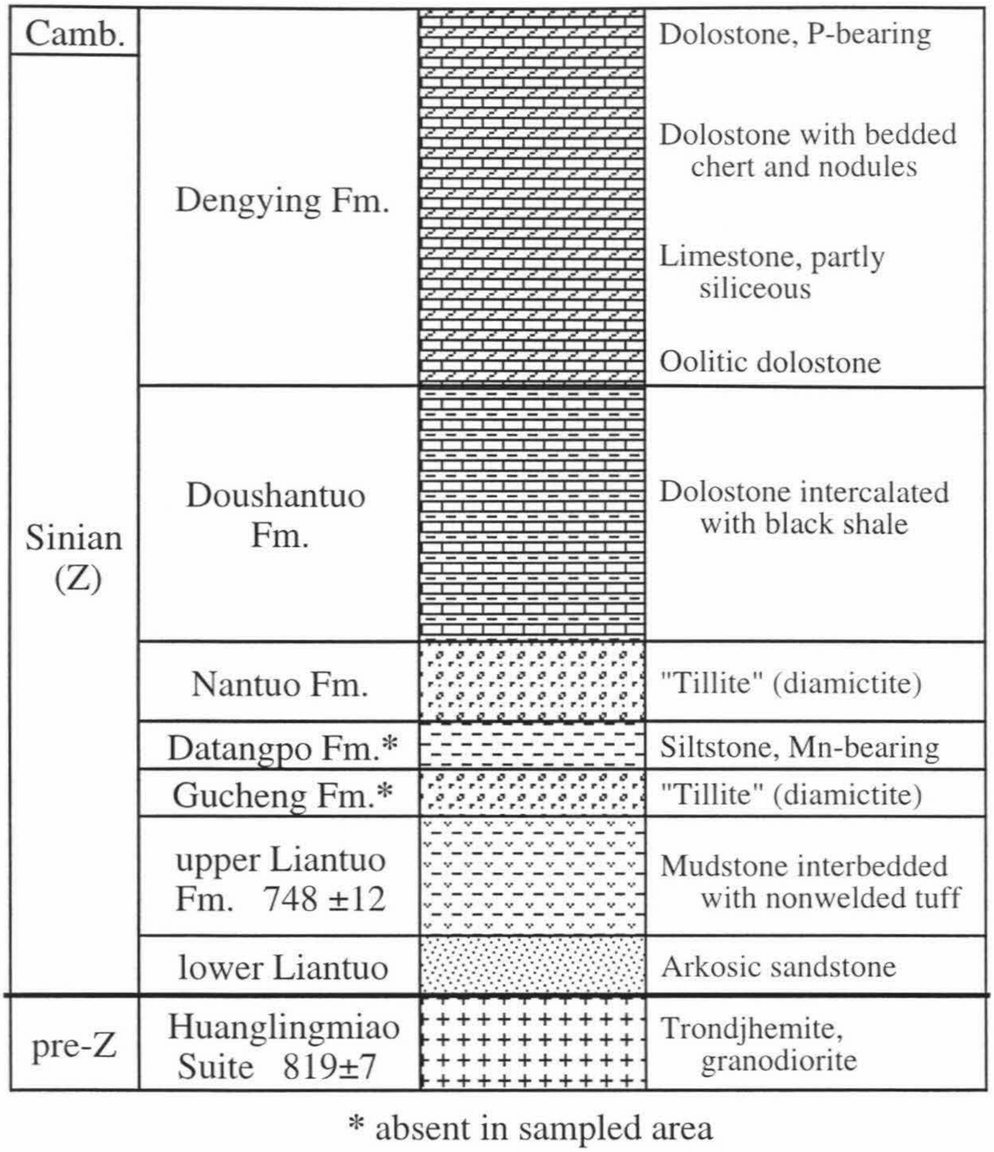


Figure 2. Generalized Sinian stratigraphy in the Three Gorges region, after Wang et al. (1996).

geomagnetic reversal. The sections are exposed in the hills above Liantuo village at (30° 51' N, 111° 09' E). Sampling in the zone of the estimated polarity change was concentrated at 2-3 cm of stratigraphic spacing, near the limit of resolution using standard portable drilling techniques, as well as the limit in uncertainties of lithostratigraphic correlation of individual beds over lateral distances of meters to decameters. The stratigraphic distribution of paleomagnetic samples from this study is shown in Figure 3. Most of the samples are purple-red mudstone, siltstone, or fine sandstone; some specimens include green mottling which generally follows certain layers but also cuts across bedding in many places. This green coloration is interpreted to indicate the flow of chemically reducing fluids through the rocks at some unknown time after deposition.

The samples were brought to Caltech and trimmed to right-cylindrical specimens of 2.5-cm height, with one specimen per sample analyzed thus far. Laboratory analysis involved a SQuID cyrogenic magnetometer and an automatic sample-changing system allowing as many as 400 consecutive, unaided measurements. For the 1986 samples, measurement of natural remanent magnetization (NRM) was followed by alternating-field and thermal demagnetization at 2.5, 5.0, 7.5, and 10.0 mT; and 150, 250, 350, 450, 500, 550, 600, 630, 650, 660, 667, 672, 675, 678, and 682 °C. For the 1996 samples, the steps were NRM; 5.0, 10.0, and 15.0 mT; and 250, 450, 570, 630, 650, 660, 667, 672, 676, 680, and 683 °C. All 143 specimens received the full treatment of partial demagnetization, until their directional behavior became unstable due to acquisition of spurious magnetizations. The latter are probably the result of partial-thermoviscous remanent magnetizations (pTVRM) acquired in the furnace, despite maintenance of ~30 nT or lower field strength throughout the heating chamber and less than 10 nT in the cooling chamber. Because the samples were placed in the oven in opposite orientations during alternating temperature steps, oscillatory demagnetization trajectories observed in some samples at temperatures greater than 650 °C betray the oven-induced pTVRM as the likely cause of this noise. To be certain that no more information remained in the samples, some

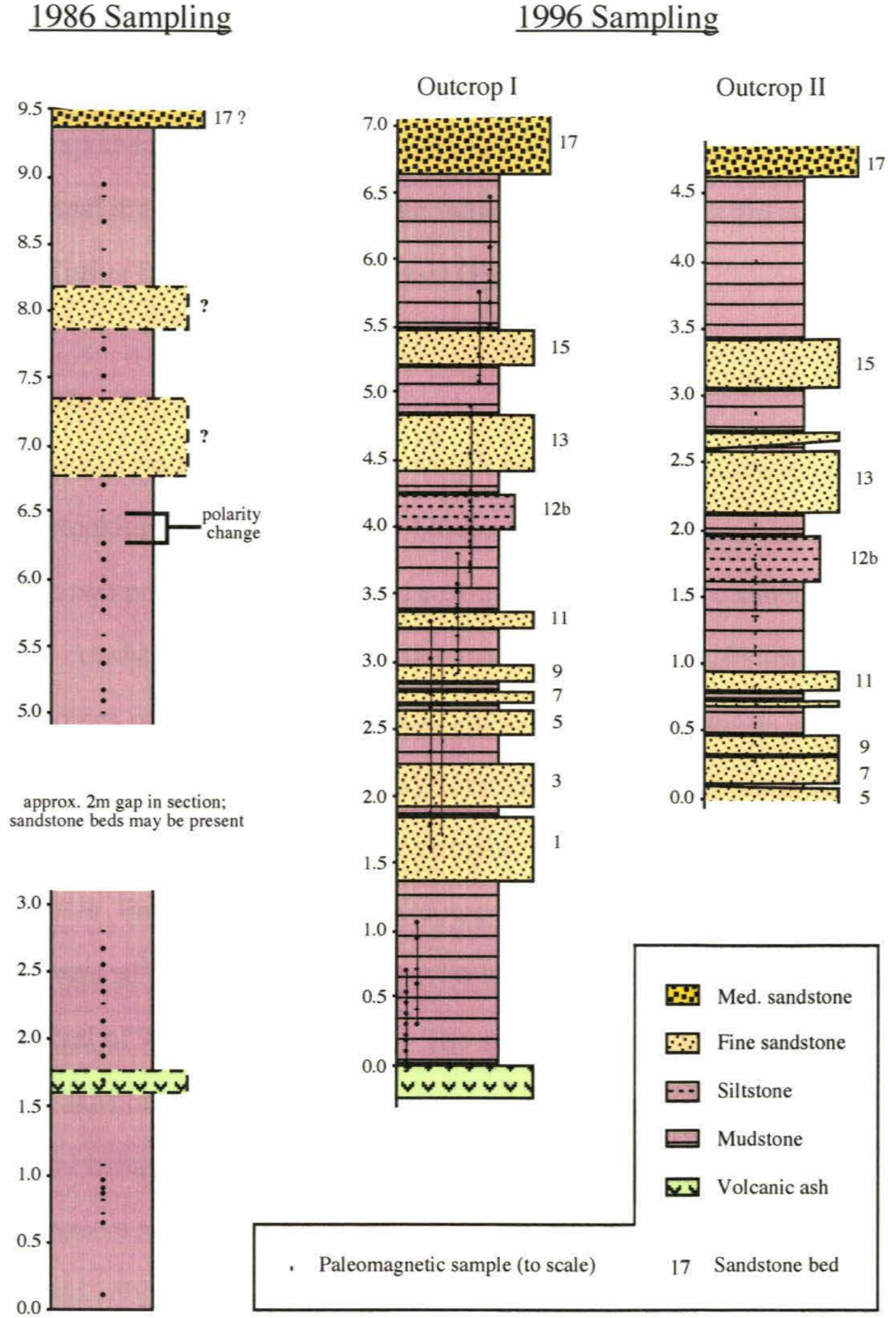


Figure 3. Paleomagnetic sampling of the upper Liantuo Formation. Solid dots are the stratigraphic positions of 2.5-cm-diameter paleomagnetic cores, drawn to scale. Vertical lines connecting samples in Outcrop I represent sampling profiles, separated laterally by approximately 5m.

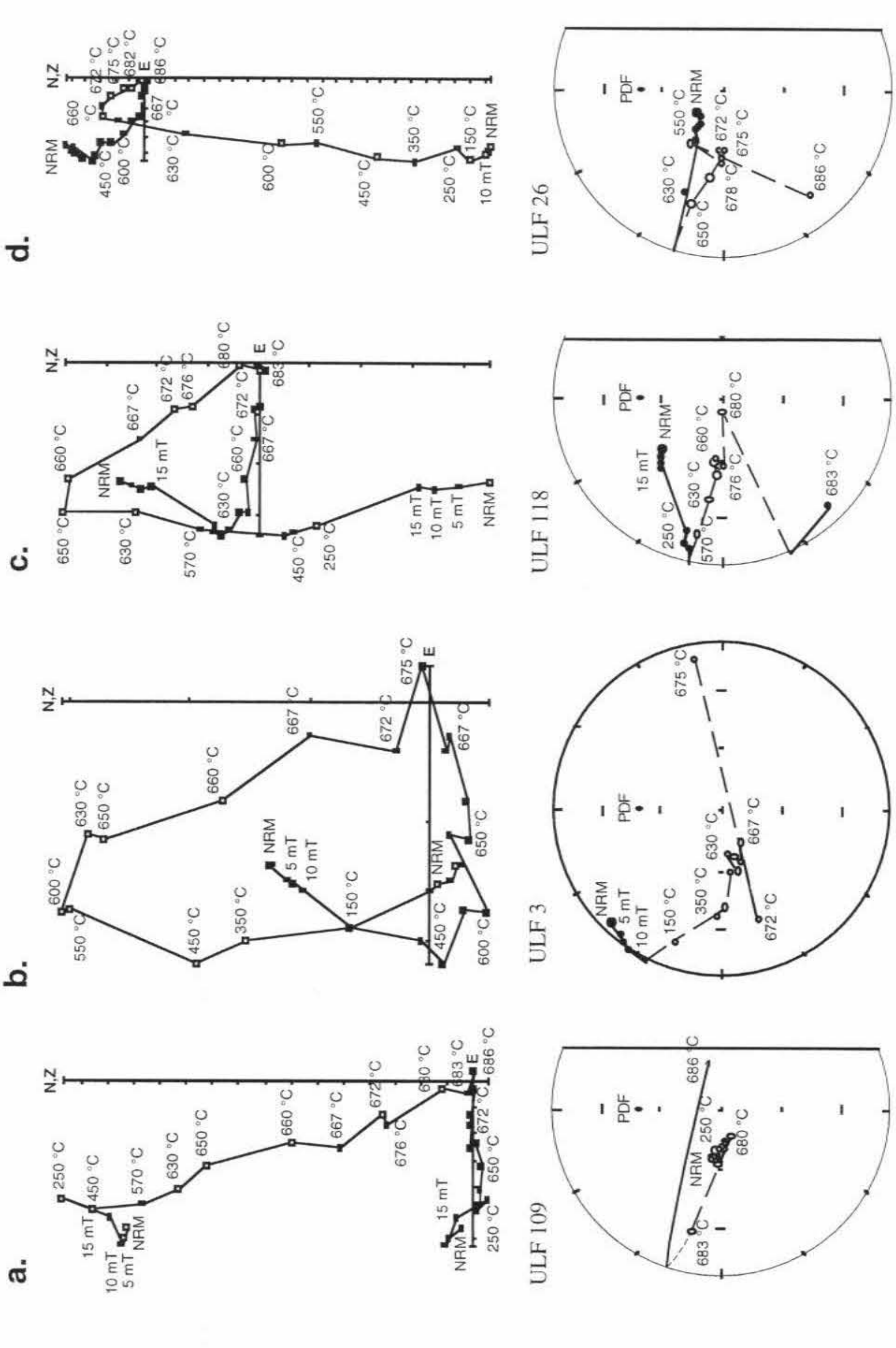
were demagnetized at additional thermal steps of 686, 690, and 695 °C. Surprisingly, some specimens retained stable behavior even to these high temperatures, a pattern also reported in the study of probably correlative strata in Yunnan (Zhang and Piper, 1997). The majority of specimens from this study retained useful information to 675 or 680 °C. In most cases, thermal unblocking spectra of the individual components did not overlap, permitting line-fitting by least-squares analysis (Kirschvink, 1980).

Directional data

Three distinct magnetic components were identified by their directional groupings and thermal unblocking spectra (Figs. 4,5). Of these, the most stable component (labelled "A") yielded the two polarities (west-up and east-down in present coordinates) to be tested for stratigraphic consistency. In order to understand the factors contributing to the directional groupings of the three components, and to facilitate a reversals test on component "A," results from the two polarity zones are discussed separately below.

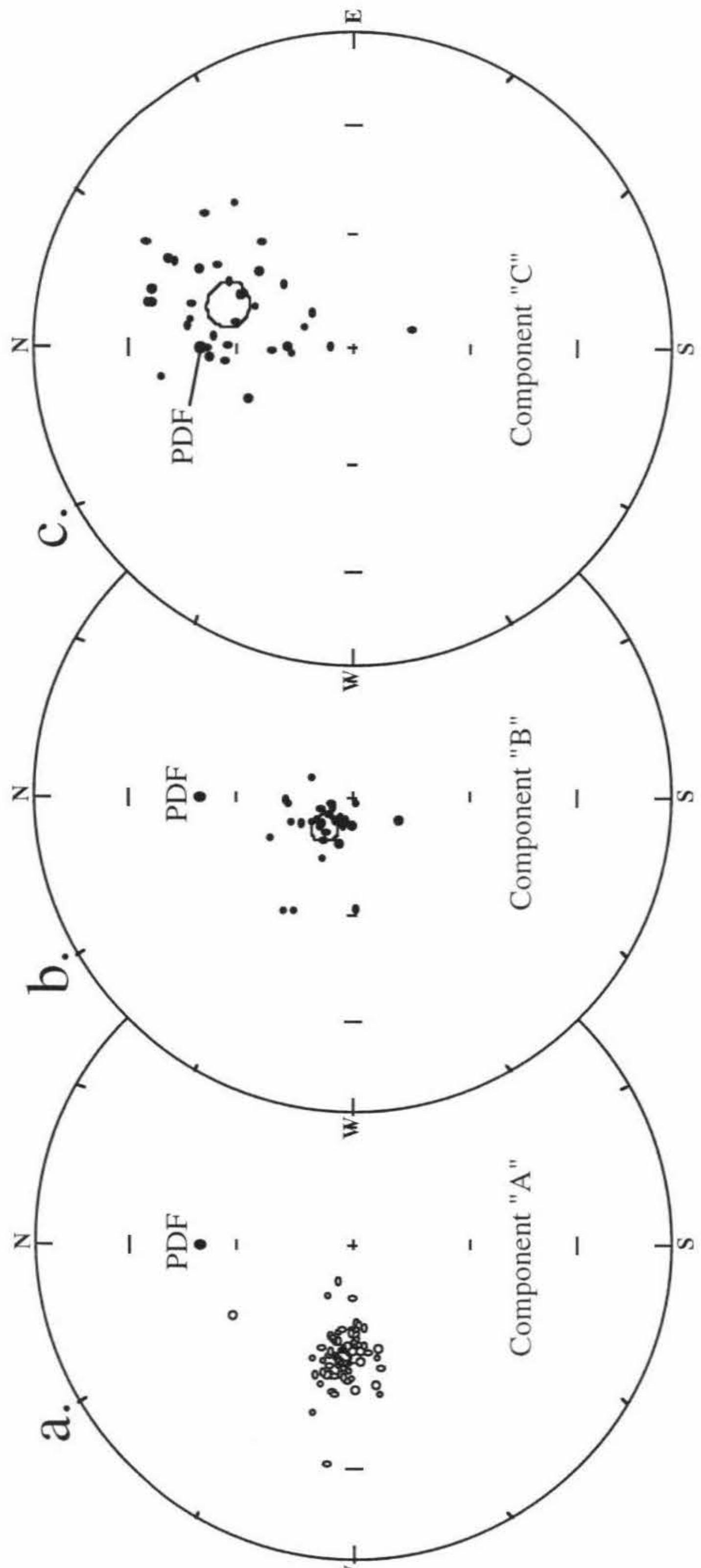
From the lower polarity zone, where component "A" is westward and upward in present coordinates (Fig. 5a), two additional components are observed overprinting "A." Figure 4 shows typical demagnetization behavior among three subgroups of specimens, representing a total of 91 out of 99 samples. The subgroups are distinguished by relative strength of the NRM held by the various components. The first subgroup (Fig. 4a) shows single-component behavior with only "A" present, or perhaps an ill-defined and small northward and downward component removed by the alternating-field and low thermal steps. The second subgroup (Fig. 4b,4c) includes the greatest number of samples and is characterized by subequal magnetizations of "A" and an overprint component (usually "C," rarely "B"). Figure 4c shows one of the few samples containing distinct and linear trajectories of all three components. The third subgroup (Fig. 4d) contains samples whose NRM is dominated by an overprint direction (usually "B," rarely "B" and "C," in two samples only "C"). For most samples within this subgroup, the "A" component is

Figure 4 (next page). Typical demagnetization trajectories, in present coordinates, of Liantuo mudstones from the lower polarity zone. Upper panels are superimposed-orthogonal-projection diagrams of measurements after the natural remanent magnetization (NRM) and subsequent partial demagnetization steps, with solid symbols in the horizontal plane and open symbols in the N-S vertical plane. Tick marks are at intervals of 10^{-8} Am^2. Lower panels are equal-area projections of the same data (symbols as in Figure 5). (a) Single-component ("A") behavior representative of 22 samples. (b-c) Two components with equivalent vectorial lengths of magnetization, representative of 46 samples. In panel (b), component "C" overprints "A"; in panel (c), all three components are identifiable. (d) Overprint-dominated magnetization representing 23 samples. In this case, "B" overprints "A."



a.
b.
c.
d.
N,Z
E
PDF
NRM
ULF 109
ULF 3
ULF 118
ULF 26

Figure 5 (next page). Equal-area projections of least-squares line-fitted component vectors from the lower polarity zone, in present coordinates. Open (solid) symbols are plotted on the upper (lower) hemisphere. The large ellipses are projections of the 95% confidence cones around the Fisher means. PDF = present dipole field direction at the sampling site. (a) Component "A." Nine samples have been excluded from this population because of mean-angular-deviation values greater than 10°. (b) Component "B." (c) Component "C."


a.
b.
c.
PDF
PDF
PDF
Component "A"
Component "B"
Component "C"
N
S
W
E

recognizable only by great-circle trajectories toward west-up on the stereonet; these data are amenable to polarity interpretation but not least-squares line-fitting. In some cases, however, a stable endpoint is reached with linear decay toward the origin (e.g., Fig. 4d). Of the eight samples not included in the three subgroups, the polarity of the "A" component is recognizable in four through great-circle analysis. Figure 5a shows the distribution of "A" among 69 least-squares lines whose mean angular deviation values (Kirschvink, 1980) are less than 10°.

Components "B" and "C" are distinguished by their thermal unblocking spectra: "C" is always completely removed by 600 °C, whereas "B" is not. This criterion separates the overprint directions neatly into two Fisherian or circularly symmetric datasets (Fig. 5b,c). As stated above, "C" generally occurs equal in magnitude to "A," whereas "B," when present, tends to dominate the NRM. Although no trends are obvious when comparing sample coloration to demagnetization behavior, it should be noted that much of the green mottling affecting the section occurs in strata within the lower polarity zone; as will be demonstrated below, "C" is largely absent from the upper polarity zone. This pattern, combined with the entirely sub-600-°C unblocking spectrum for "C," suggests that the magnetic carrier of that component is magnetite formed by chemical reduction via secondary fluid percolation through the section.

Components "B" and "A," with unblocking spectra ranging above 600 °C, are likely carried by hematite. Ferromagnetic remanence in this mineral can arise from a number of depositional and post-depositional processes (Butler, 1992, p.197-203), and stability of thermal unblocking in the laboratory does not necessarily correspond with age of remanence acquisition. Therefore, from the preceding discussion alone, there is no compelling reason to accept "A" as the primary magnetic component, despite its universally higher stability than "B" and narrow thermal unblocking spectrum near the 675-°C Neel temperature of hematite. At most one of the components can be primary, however, and presence of two polarities, if verified by stratigraphic consistency, would strongly suggest that "A" is primary and "B" secondary (see below).

Within the 44 samples from the upper polarity zone, two subgroups are identified (Fig. 6). In all 29 samples of the first subgroup, "B" and the eastward-downward polarity of "A" are identifiable and clearly distinguishable in both superimposed-orthogonal-projection diagrams and the equal-area stereonets (Fig. 6a). In the second subgroup, the two components are not easily identified because a stable endpoint "A" direction is not attained. In half of these samples, component "B" can be least-squares line-fitted, whereas only the polarity of "A" is identifiable by great-circle migration away from the "B" direction on the stereonet (Fig. 6b). A component similar to "C" from the lower polarity zone is largely absent, insignificant, or nonlinear in samples from the upper polarity zone. As discussed above, "C" may be related to localized chemical reduction and such greenish mottling is not prevalent in the upper polarity zone.

For all of the upper-polarity-zone samples, the straightest segments of the demagnetization trajectories were chosen without regard to unblocking temperature. Like the analysis for the lower polarity zone, mean-angular-deviation values greater than 10° were excluded from final statistical compilation. Nevertheless, the distribution of components "A" and "B" from the upper polarity zone is streaked (Fig. 7) and appears to define a mixing plane between two endmembers. The elongate distribution may be attributed to slightly overlapping thermal unblocking spectra between the two components within individual specimens; whereas the overlap could be easily identified and avoided in samples of the lower polarity zone because of the nearly 180° inter-component angle, in the upper polarity zone "A" and "B" are separated by only ~30°, allowing some contamination of the overlapped part of the unblocking spectrum into the least-squares line analysis.

Both means for components "A" and "B" from the upper polarity zone are displaced toward the centroid of the combined dataset (Figure 7). This incomplete separation of the two components renders both a failed common mean test for "B" and a failed reversals test

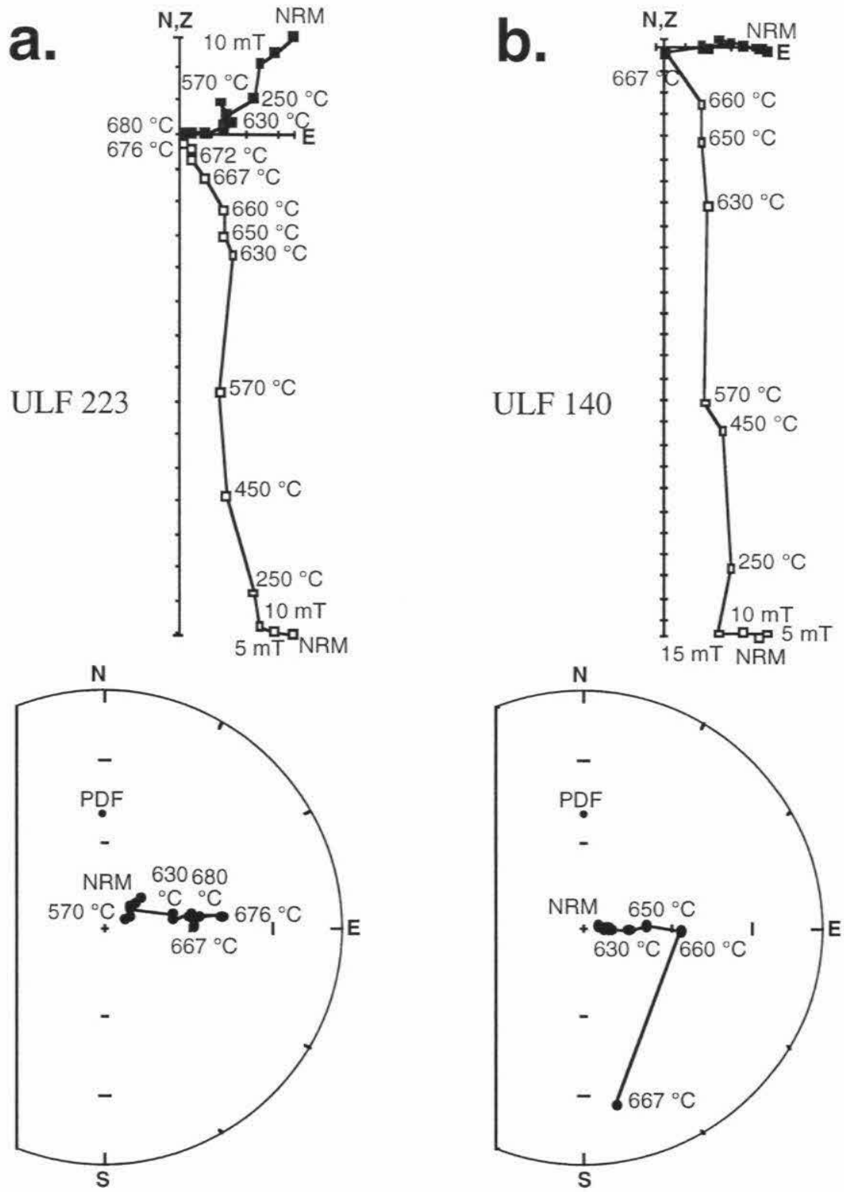


Figure 6. Typical demagnetization trajectories, in present coordinates, of Liantuo mudstones from the upper polarity zone. Symbols as in Figure 4. (a) Components "A" and "B" relatively easy to separate, representative of 29 samples. In this case, component "C" may be present at the early demagnetization steps, and components "A" and "B" are separated by the 630 °C step. (b) Components "A" and "B" difficult to separate, as observed in 14 samples.

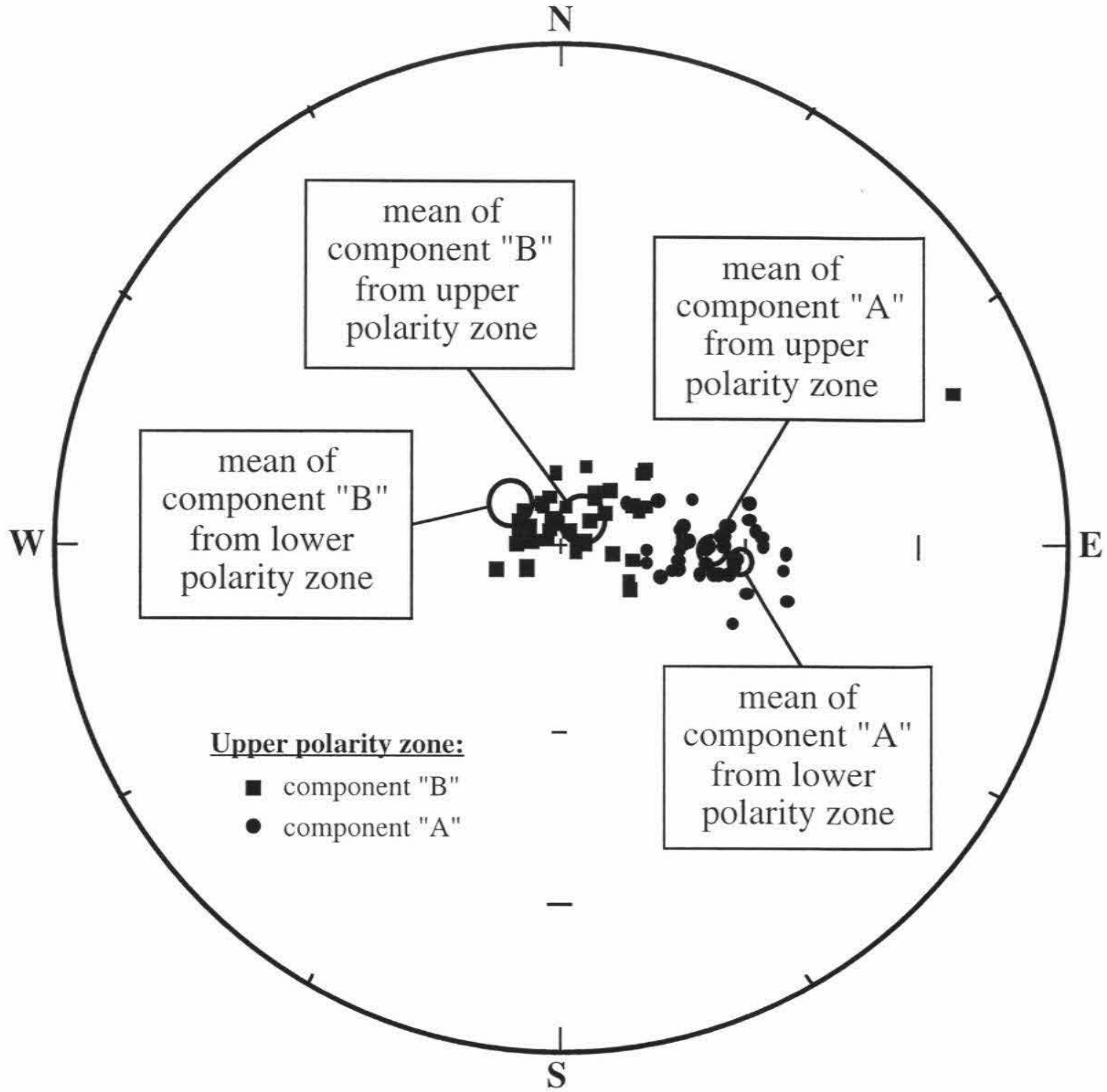


Figure 7. Lower-hemisphere equal-area plot of overprint "B" (squares) and characteristic component "A" (circles) from the upper polarity zone, in present coordinates. Dark ellipses are projections of 95% confidence cones about the means of components "A" and "B"; means from both polarity zones are shown for comparison.

for "A" (bootstrap tests adapted from McFadden and McElhinny, 1990) at the 99% level, between the two polarity zones. For this reason, although the difference in means between upper and lower polarity zones is not great (e.g., less than 5° for component "A"), in subsequent generation of paleomagnetic poles I use only the Fisherian distributions from the lower polarity zone (Table 1).

Magnetostratigraphy

Despite the statistical failure of the reversals test due to minor overlap in thermal unblocking spectra between components, it is clear that the "A" component is a nearly pure, dual-polarity remanence (Fig. 8) exhibiting consistently high magnetic stability. Stratigraphic variation in the "A" remanence is also quite stable and consistent among the three sections studied, where a single polarity switch is observed (Fig. 9). Following the results from the 1986 sample suite which originally identified the polarity change within a stratigraphic interval of ~20cm, the two 1996 outcrops were sampled quite densely in that interval. The density of 2-4 cm between samples approaches the limit of resolution for the 2.5-cm diameter drilling apparatus, as well as the 1-cm precision of stratigraphic measurements where limited outcrop availability may require as much as decimeters of lateral distance separating stratigraphically adjacent samples.

Within the 1996 outcrops I and II, the polarity change occurs respectively within 41-46 cm and 40-46 cm below the base of sandstone bed #13. A more subtle lithological distinction appears approximately at this level in both sections: the base of a siltstone unit (#12b in Fig. 3) similar in color to the mudstones but slightly more indurated (especially noticeable as an increased resistance during drilling). The polarity change is coincident with the base of this unit in outcrop I, but approximately 5cm above the base in outcrop II.

Several conceptual models can be put forward to try to explain these data. First, the "A" component may be a two-polarity overprint acquired long after deposition and hence yielding no information regarding the depositional paleolatitude. If so, then that overprint

Table 1. Least-squares-line data, upper Liantuo Formation, Three Gorges region (30.9°N, 111.2°E), South China

Polarity zone, Component (coordinates)	N	Fisher statistics D (°)	I (°)	k	α_{95}	Paleomagnetic pole λ (°N)	ϕ (°E)	dp	dm	Reliability 1234567	Q
Lower "A" (in situ)	69	274.9	-61.0	82.0	1.9	16.8	161.7	2.2	2.9		
(tilt-corrected) *		291.4	-53.2	82.4	1.9	01.5	161.7	1.8	2.6	1111111	7
"B" (in situ)	29	311.7	79.0	57.5	3.6	43.4	089.1	6.5	6.8		
(tilt-corrected) *		206.6	84.2	53.6	3.7	20.6	105.5	7.2	7.3	1110101	5
"C" (in situ) *	37	018.7	55.8	17.8	5.8	73.6	177.1	6.0	8.3	1110100	4
(tilt-corrected)		037.5	64.3	17.6	5.8	57.5	162.7	7.4	9.3		
Upper "A" (in situ)	37	091.7	65.2	103.2	2.3	21.1	157.7	3.0	3.7		
(tilt-corrected)		111.6	57.4	92.7	2.5	03.9	158.2	2.7	3.7		
"B" (in situ)	38	041.2	84.3	37.6	3.8	39.1	120.6	7.4	7.5		
(tilt-corrected)		131.0	77.8	35.3	4.0	14.5	129.0	7.1	7.5		

* = used for final calculation of paleopoles; N = number of samples; D = mean declination; I = mean inclination; k = best estimate of Fisher precision parameter; α_{95} = semi-radius, in degrees, of confidence cone (standard error of the mean); λ = paleopole latitude; ϕ = paleopole longitude; dp,dm = lengths of minor, major semi-axes (in degrees) of 95% confidence ellipse about paleopole; reliability criteria from Van der Voo (1990), note that some criteria are satisfied by relations outside of a particular column (see text).

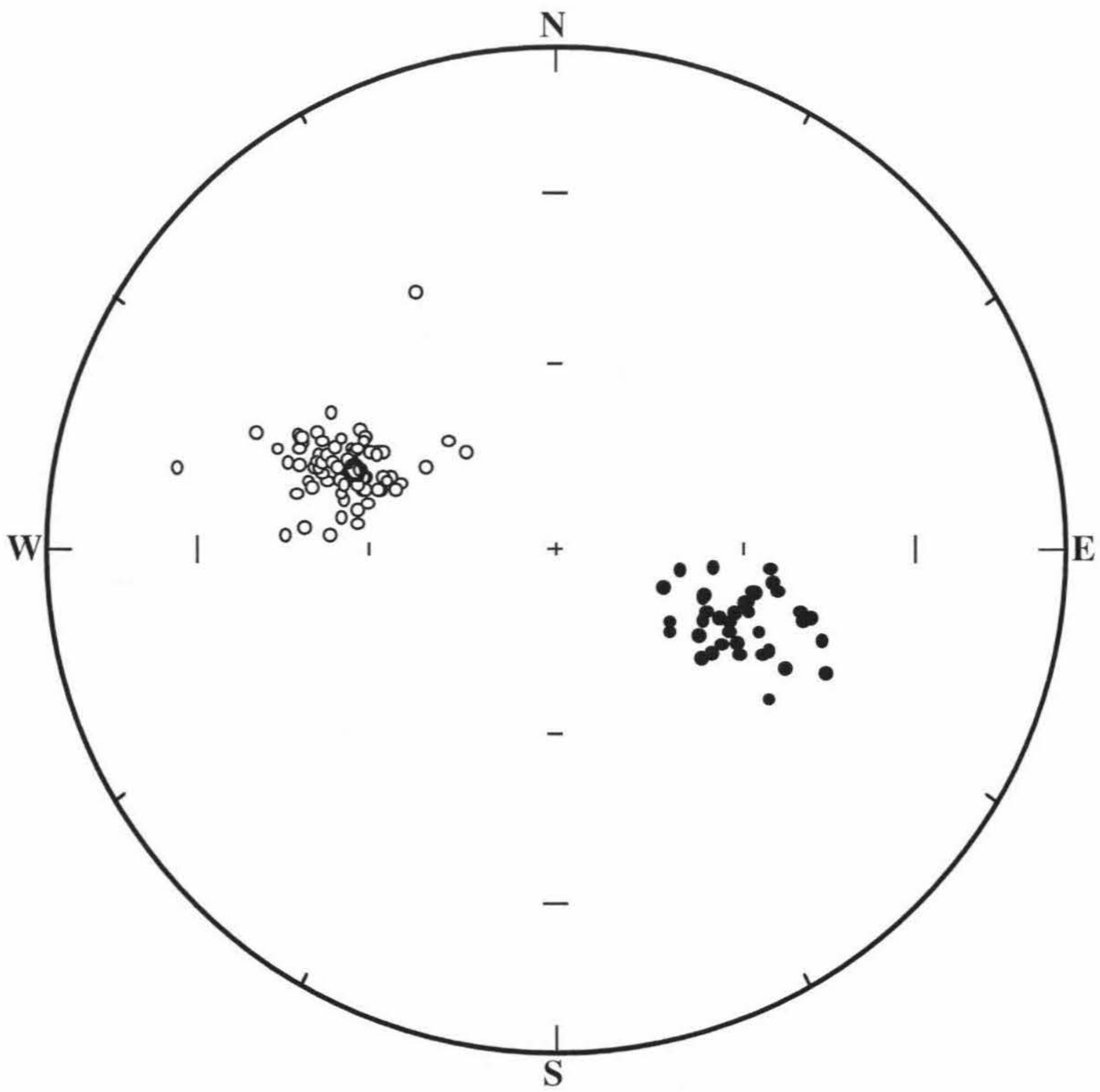
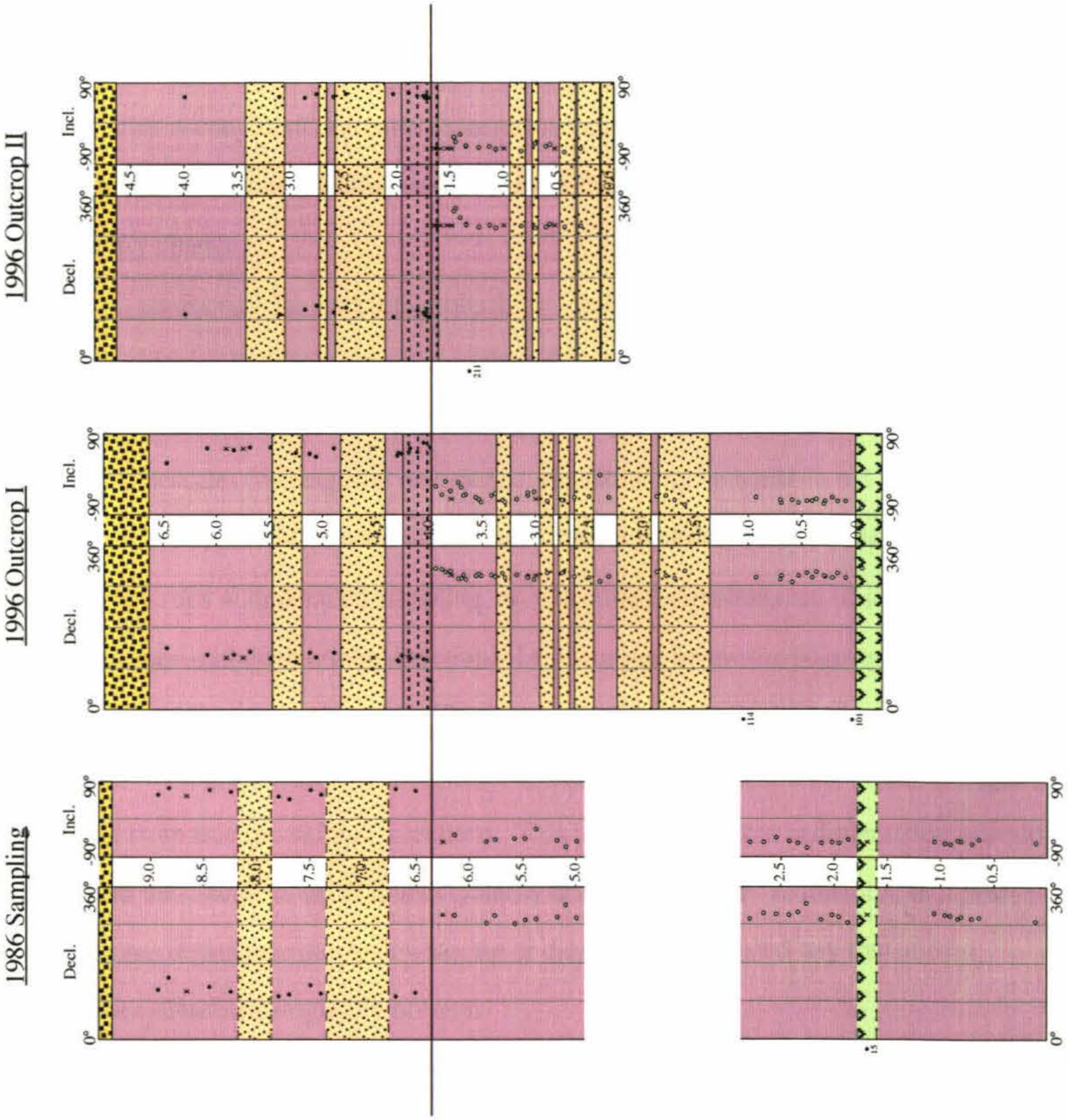


Figure 8. Equal-area projection of tilt-corrected component "A" from both polarity zones. Symbols as in Figure 5. Dark ellipse is the 95% confidence cone about the Fisher mean of all samples, after inversion of the lower-hemisphere directions.

Figure 9 (next page). Stratigraphic variation of component "A" from the three sampled sections, showing declinations (left) and inclinations (right) in tilt-corrected coordinates. Lithological patterns as in Figure 3. Stratigraphic height in meters, in the center of each column. Open (solid) circles depict least-squares line-fitted data from samples, approximately to scale, in the lower (upper) polarity zone. Samples not amenable to least-squares line-fitting, but whose polarity designations are obvious, are demarcated by 'x' at the mean direction. The four samples with no determinable polarity designation are labelled by sample number at the left side of the columns. The horizontal line spanning the three sections denotes the alleged change in geomagnetic polarity.


1996 Outcrop II
1996 Outcrop I
1986 Sampling
Decl.
Incl.
0°
360°
-90°
90°

would have migrated diachronously through the exposures over an interval of a few thousand years, the time considered to span a typical geomagnetic reversal. In doing so, it would have a sharply defined and remarkably consistent remagnetization "front" maintaining a position 40-45 cm below sandstone bed #13. Neither a chemical (i.e., fluid-related) nor a thermal overprinting mechanism would be expected to maintain such stratigraphic regularity, for the flow paths of those sources would be expected to follow lithological boundaries between rocks of different porosity or conductivity. The fact that "A" polarity is consistent among both sandstone and mudstone in all three outcrops suggests that diachronous diffusive flow, either fluid or thermal, was not responsible for the polarity change.

Considering a geomagnetic polarity reversal occurring near the time of deposition, then, it is difficult to determine whether the polarity change records the reversal in "real-time" or whether there is a time lag of some sort, e.g., due to lithification processes. The range of timescales for magnetic remanence acquisition by hematite in red beds is reviewed by Butler (1992, p.197-203) and is not easily determined in the laboratory. Note that from the two outcrops with detailed sampling, no transitional directions are observed between polarity zones--although erratic directional behavior just below the polarity change can be observed in both outcrops I and II (Fig. 9), this is due to less linear demagnetization trajectories in these units rather than well-defined lines in unusual orientations as might be expected from true transitional directions. This implies that either sedimentation was slow relative to the change in geomagnetic polarity during "real-time" acquisition, or a post-depositional remanence due to dewatering or diagenesis obliterated any transitional remanence directions originally present.

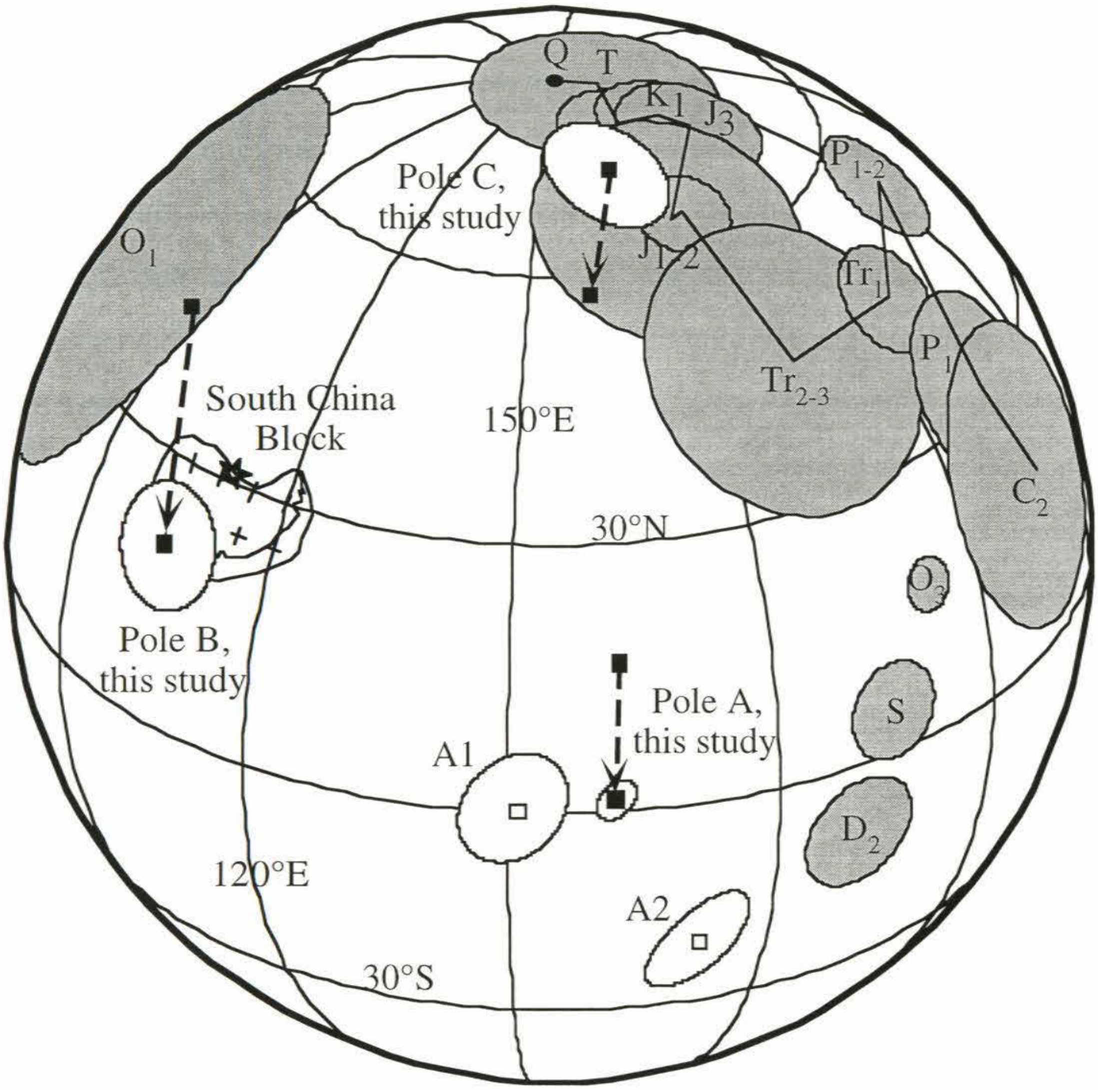
Another possible model of immediate post-depositional remanence is that during the reversal Earth's field strength declined, causing the transitional directions to be of low ferromagnetic stability. As geomagnetic field strength rebounded into the new polarity chron, unlithified sediments deposited during the transition were remagnetized with the new polarity. In these cases, the apparent reversal near the base of unit #12b may actually have occurred during deposition of, say, sandstone bed #13. The 5-cm discordance between outcrops I and II, of the polarity change relative to the base of unit #12b, can be explained in the "real-time" scenario by an onlap relationship of the siltstone from outcrop II to outcrop I, whereby the base of unit #12b in outcrop I would represent a mild disconformity. Such an onlap gradient of 5 cm over a lateral distance of ~100 m is plausible for the sedimentary environment of the upper Liantuo Formation, considering the lenticular dimensions of the lithological units (Fig. 3).

Ages of magnetic components

The stratigraphic consistency of the polarity zones observed in three independently logged lithostratigraphic sections strongly suggests acquisition of magnetic component "A" either concurrently with or shortly after deposition, at the time of a geomagnetic reversal. Therefore, assuming an axial geocentric dipole magnetic field, the unbiased mean "A" direction calculated from the lower polarity zone can provide constraints upon the depositional paleolatitude of the basal Sinian deposits and global paleogeography at 748 ±12 Ma. Given that consistent stratigraphic polarity zonation among several sedimentary sections constitutes a legitimate field stability test, the paleomagnetic pole corresponding to component "A" rates a perfect seven on the 'Q' scale of reliability (Van der Voo, 1990).

Figure 10 shows the paleomagnetic poles derived from this study, compared with previously determined poles from basal Sinian and Phanerozoic rocks of the SCB. Pole "A," regarded to be primary as defended above, lies between the "A1" and "A2" poles from basal Sinian rocks in Yunnan (Zhang and Piper, 1997). In that study, the distribution of least-squares lines appears non-Fisherian, streaked in the same direction as was observed from some samples in this study (compare their Fig. 7 with the individual sample behavior shown in Fig. 3 of this paper). Zhang and Piper (1997) divided their dataset at a local minimum in the density of least-squares directions, at an inclination of -30°. The local

Figure 10 (next page). Orthographic projection of paleomagnetic poles generated from components "A," "B," and "C," relative to existing paleopoles from the South China block. The star indicates the Three Gorges region. Dashed arrows denote the tilt correction upon paleopoles from this study. Poles "A1" and "A2" are from the recent study of basal Sinian strata in Yunnan by Zhang and Piper (1997). Shaded poles are from Phanerozoic rocks and summarized by Zhao et al. (1996). O_1 = Early Ordovician, O_3 = Late Ordovician, S = Silurian, D_2 = Middle Devonian, C_2 = Late Carboniferous, P_1 = Early Permian, P_{1-2} = mid-Permian, Tr_1 = Early Triassic, Tr_{2-3} = Middle-Late Triassic, J_{1-2} = Early and Middle Jurassic, J_3 = Late Jurassic, K_1 = Early Cretaceous, T = Tertiary, and Q = Quaternary (present rotational axis).



Q
T
K1
J3
P1-2
Pole C,
this study
O1
J1-2
Tr1
Tr2-3
P1
South China
Block
150°E
C2
30°N
O3
Pole B,
this study
S
Pole A,
this study
A1
D2
120°E
A2
30°S

minimum is not convincing as a significant feature of the dataset, however, and the -30° cutoff seems otherwise arbitrary. Individual sample behavior in the study by Zhang and Piper (1997) displays the same sensitive demagnetization behavior as observed in this study for the temperature range 630-680 °C. Apparently, even the careful, 10 °C increments used in that study were not able to isolate component "A" completely.

It is unlikely that both "A" and "A1" are unbiased and coeval, for this would imply that the present geographical separation of ~1000 km between Yunnan and the Three Gorges region is entirely the result of post-Sinian extensional displacement between the two areas. Nonetheless, the general consistency between the tilt-corrected "A" pole from the Three Gorges region, and the streaked pair "A1" and "A2" from Yunnan, suggests that to first order the South China block has remained structurally coherent, especially with regard to vertical-axis rotations, since Sinian time.

The "B" pole, derived from the steep, hematite-borne overprint observed in both polarity zones, is far-removed from Cretaceous to Recent paleopoles from the SCB and therefore must be considered in pre-tilt coordinates. Its restored paleopole lies directly atop southern China, documenting here for the first time a polar paleogeography of the SCB. Unfortunately, the age of this pole is not well constrained. It is dissimilar to all previously reported Phanerozoic poles, except in broad proximity to the pre-fold result from Early Ordovician sedimentary rocks in Yunnan (Fang et al., 1990; "O1" in Fig. 10), discounted as anomalous by Zhao et al. (1996), and a two-polarity result of lesser reliability from Precambrian-Cambrian boundary strata in Guizhou (Wang et al., 1994; pole at 17°N, 059°E, not shown in Fig. 10). If those results were real, then component "B" from this study may have been acquired around Cambrian time; however, it should be cautioned that other Cambrian poles of questionable reliability exist for the SCB (reviewed by Zhao et al., 1996), and that the platform-carbonate-dominated succession throughout the early Paleozoic of southern China (Wang et al., 1996) would argue against such a high paleolatitude at that time.

Alternatively, the "B" pole may be applicable to Sinian time, sometime within the 200-Myr interval between Liantuo deposition at 748 ±12 Ma and the Cambrian Period. A uniformitarian view of the latitudinal ranges of carbonate vs. glaciogenic sediment deposition would suggest that the "B" component would have been acquired during the Nantuo ice age, whose deposits disconformably overlie the Liantuo Formation. This is inconsistent, however, with the results of Zhang and Piper (1997), which demonstrate consistency of the A1/A2 remanence in both the Chengjiang (probably Liantuo equivalent) and the overlying Nantuo Formations. The Nantuo diamictites may be greatly diachronous across southern China, deposited during the apparent polar wandering from "A" to "B." On the other hand, strict uniformitarianism may not be applicable to Neoproterozoic paleoclimate, for at least one glacial deposit of that time period appears to have formed in cold conditions at sea-level, near the Equator (Williams et al., 1995). Given these uncertainties, I interpret the "B" pole to be of Sinian age, perhaps or perhaps not related to the late stages of the Nantuo ice age. On the 'Q' scale of reliability (Van der Voo, 1990), the "B" pole rates a five, failing the categories of field stability tests and the presence of reversals.

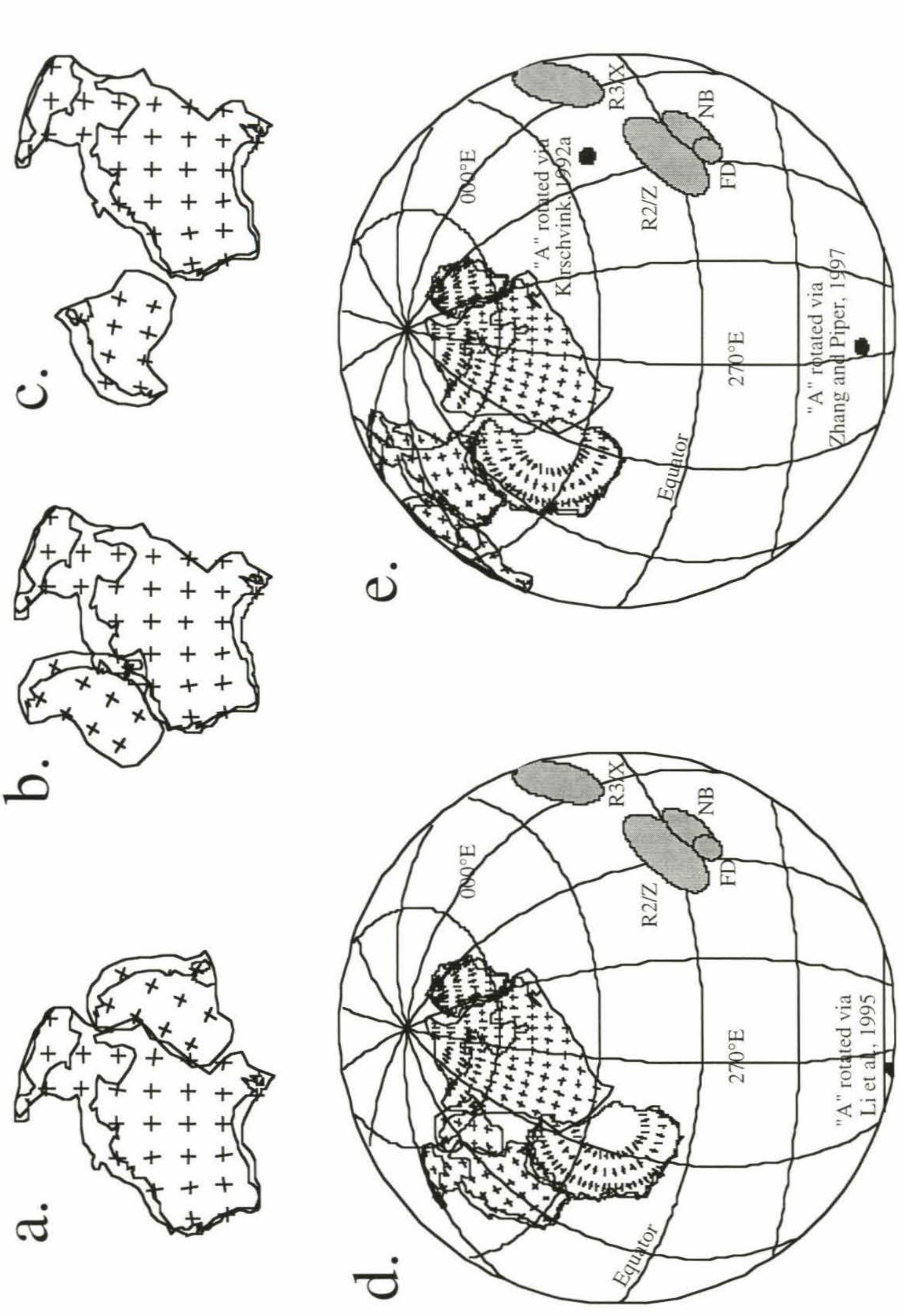
The "C" pole associated with low thermal stability and secondary greenish coloration in the rocks, is located prior to tilt-correction near previously determined Jurassic-Cretaceous poles from the SCB (and in fact all of China; Zhao et al., 1996). The location is near poles derived from a widespread overprinting episode affecting Paleozoic sedimentary rocks in the Three Gorges region (Kent et al., 1987), and distinct from the present rotation axis. Upon correction for the tilt of bedding, the "C" pole becomes distinct from the existing group of late Mesozoic poles, suggesting that "C" was acquired soon after Indosinian tilting, in Jurassic-Cretaceous time. The "C" pole has reliability Q=4, lacking field stability tests and dual polarity, and showing similarity to late Mesozoic poles.

Discussion

A primary age for the "A" component implies a depositional paleolatitude of 33.8 ±1.8° for the Liantuo Formation. Most regional stratigraphic syntheses present a disconformity between the Liantuo Formation and the overlying Nantuo or Gucheng glaciogenic sediments. General consistency between the "A" pole from this study with results from both pre- and syn-glacial sediments in Yunnan (Zhang and Piper, 1997; see above), however, suggests that if the correlations are valid, then the intervening disconformity is not of great duration. Stratigraphic correlations with southeastern China are disputed, thus the Liantuo may correlate either with strata conformably above (Wang, 1986) or disconformably below (Lu et al., 1985; Li et al., 1996) the Chang'an glacial deposits. Even if the ~34 ±2° paleolatitude of the Liantuo Formation applies to one or both of the Chang'an or Nantuo ice ages, the result is neither high nor low enough to negate one of the existing models of Neoproterozoic glaciations (e.g., Harland, 1964; Williams, 1975; Meert and Van der Voo, 1994). If the "B" component could be identified confidently with the late stages of the Nantuo glaciation, however, this would refute the high-obliquity hypothesis of Williams (1975).

A primary paleomagnetic pole from the SCB at 748 ±12 Ma permits evaluation of several paleogeographic models of the Rodinian supercontinent. Whereas most workers prefer to reconstruct South China adjacent to Australia during Neoproterozoic time, the precise location and relative orientation is unclear. Figure 11 shows three proposed reconstructions of the SCB and Australia, from Li et al. (1995, 1996; panel a), Zhang and Piper (1997; panel b), and Kirschvink (1992a; panel c). Unfortunately, the paleomagnetic database for mid-Neoproterozoic Australia, and all of East Gondwanaland, is very poor (Powell et al., 1993); however, if the paleogeographic models are extended to include reconstructions of Australia against Laurentia, then the Liantuo "A" pole may be compared with more reliable poles from that paleocontinent.

Figure 11 (next page). Hypothesized reconstructions of the South China block relative to Australia, East Antarctica, and Laurentia, and comparisons of the rotated "A" pole with 725-750-Ma poles from Laurentia. (a) The "missing link" model approximated from Li et al. (1995, 1996), generated by a right-handed Euler rotation of (21.6°N, 160.0°E, +90.7°) from South China to Australia. (b) Alternative reconstruction by Zhang and Piper (1997), achieved by an Euler rotation of (11.0°N, 146.5°E, +68.5°). (c) Third possibility proposed by Kirschvink (1992a), with an Euler rotation of (05.8°N, 111.0°E, +162.82°). (d) Comparison of the "A" pole from this study, rotated to North American coordinates according to the "missing link" model of Li et al. (1995, 1996; Australia to Laurentia Euler rotation of 34.6°N, 134.6°E, +126.9°), with approximately coeval poles from Laurentia. (e) Same as panel (d), but showing the Zhang and Piper (1997) and Kirschvink (1992a) South China-Australia reconstructions and rotating them to Laurentia according to Dalziel (1997; Euler parameters of 28.9°N, 126.0°E, +132.1°). FD = Franklin dikes pole (Christie and Fahrig, 1983), NB = Natkusiak basalts pole (Palmer et al., 1983), Rapitan Group components "R2" and "R3" (Park, 1997) respectively equated with components "Z" and "X" isolated by Morris (1977; means not shown). Panels (d) and (e) are equal-area projections.



a.
b.
c.
d.
e.
000°E
270°E
Equator
R2/Z
R3/X
NB
FD
"A" rotated via
Li et al., 1995
"A" rotated via
Kirschvink, 1992a
"A" rotated via
Zhang and Piper, 1997

The comparably aged Laurentian poles come from two lithological groups: basic igneous rocks precisely dated at 723 +4/-2 Ma (Heaman et al., 1992) from the Franklin-Natkusiak magmatic episode, and glacial/interglacial rocks in synrift strata of the Rapitan Group of the Mackenzie Mountains in northwest Canada (Eisbacher, 1985). According to unpublished data cited in Ross et al. (1995), a Rapitan diamictite contains a granitic clast with U-Pb zircon age of 755 ±18 Ma. Elsewhere in the Canadian and U.S. Cordillera, Rapitan-correlative deposits are associated with ~700-750-Ma volcanism (Link et al., 1993). The poles from the Rapitan Group (Park, 1997; similar to results obtained by Morris, 1977) are in the vicinity of those from the Franklin-Natkusiak episode, supporting a ~700-750-Ma age for glaciation in northwest Canada and permitting the lithostratigraphic correlation of the basal Sinian and Raptian units (Li et al., 1995). Although the choice of which Rapitan pole is primary is debatable (Park, 1997), either will suffice for the first-order comparisons made below.

The model by Li et al. (1995, 1996), depicted in Fig. 11a, includes not only the South China-Australia connection, but also their reconstruction against Laurentia in the "missing link" hypothesis modified from SWEAT (Moores, 1991). Figure 11(d) shows the "A" pole from this study rotated to Laurentia according to the reconstruction by Li et al. (1995). It falls far from the Laurentian poles, apparently negating the reconstruction. The "missing link" model could be modified to match the poles if the SCB were rotated ~90° about a local vertical axis, but then the continents would overlap substantially.

The South China-Australia reconstruction by Zhang and Piper (1997), shown in Fig. 11b, was recognized by those authors not to conform with ~720-750-Ma poles from Laurentia via the SWEAT hypothesis (Fig. 11e; updated rotation parameters from Dalziel, 1997). Instead, the reconstruction was based on a swathe of previously determined late Neoproterozoic-Cambrian poles. By arguing that this paleogeography should have been valid for earlier times such as 750 Ma, they used the incompatibility of their "A1" pole with coeval Laurentian poles to invalidate the SWEAT connection. However, the Chinese paleomagnetic database upon which their argument is built is rather poor, thus their reconstruction is non-unique.

The global Vendian-Cambrian paleogeography proposed by Kirschvink (1992a) provides an additional reconstruction of the SCB and Australia to be tested (Fig. 11c). That reconstruction was based on Cambrian biogeography (Burrett and Stait, 1986) and magnetostratigraphic results from the Precambrian-Cambrian boundary (Kirschvink, 1978a,b; Fang et al., 1988/9) and Cambrian-Ordovician boundary (Ripperdan, 1990; Ripperdan and Kirschvink, 1992) on both continents. Upon restoring the "A" pole from this study to Australia according to Kirschvink (1992a) and then to Laurentia via SWEAT (Dalziel, 1997), it falls fairly near to the Laurentian dataset. This evidence alone would support the first-order validity of the Kirschvink recontruction for the interval 750-500 Ma.

The first-order lithostratigraphy of the Late Neoproterozoic in southern China and northwest Australia, however, creates problems for the above model. For example, the Nantuo/Gucheng/Chang'an glacial units in southern China are commonly considered to be entirely older than 700 Ma, although this is partly based on a probably unreliable whole-rock Rb-Sr age of 700 ±5 Ma on shale from the upper part of the Doushantuo Formation (Ma et al., 1984). In the Kimberley region of northwest Australia, the ages of several discontinuously exposed glacial deposits have been debated, but recent correlations (Plumb, 1996; Corkeron et al., 1996) suggest that they may be Marinoan (estimated at ~600 Ma) and younger (immediately pre-Ediacaran, perhaps ~560 Ma). If all of the above correlations are correct, then the absence of ~750-Ma glacial deposits in the Kimberley region and of ~600-550-Ma glacial rocks in the SCB would tend to discredit the extrapolation of Kirschvink's (1992a) reconstruction back to 750 Ma.

In conclusion, more reliable paleomagnetic poles are needed from South China, Australia, and Laurentia, before the paleogeographic models can be assessed fully. In particular, the SCB lacks high-quality poles from the interval 700-500 Ma (Zhang and Piper, 1997); Australia lacks reliable poles for the period 800-650 Ma (Powell et al., 1993), and Laurentia lacks good results for the time span of 700-600 Ma (Torsvik et al., 1996). These long intervals with little or no kinematic constraints must be filled with reliable paleomagnetic data before the Rodinian supercontinent's paleogeography can truly take shape.

Acknowledgments

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Chapter 8: Low-latitude glaciation in the Palaeoproterozoic Era

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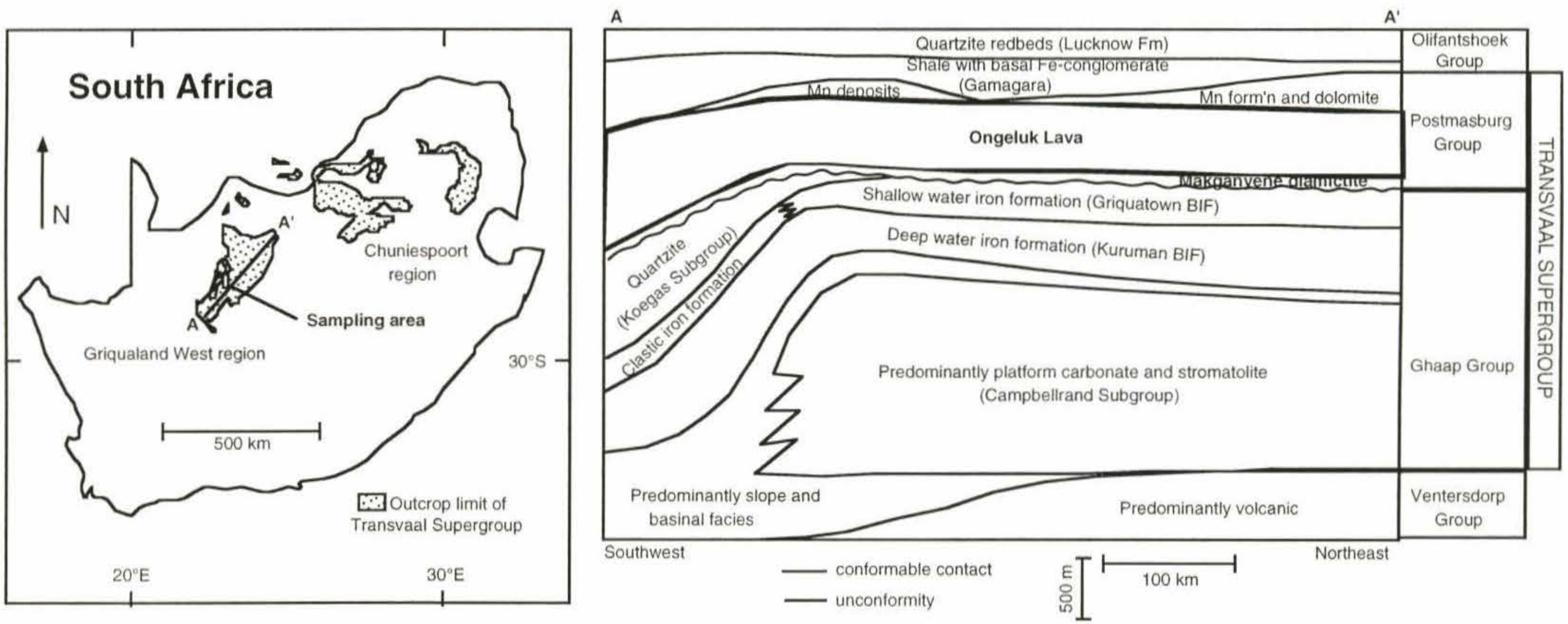
One of the most fundamental enigmas of the Earth's palaeoclimate concerns the temporal and spatial distributions of Precambrian glaciations. Through four billion years of Precambrian history, unequivocally glacial deposits have been found only in the Palaeoproterozoic and Neoproterozoic record[1]. Nonetheless, some of these deposits are closely associated with tropical--rather than just polar--palaeolatitudinal indicators such as carbonate rocks, red beds, and evaporites[1,2]. These observations are quantitatively supported by palaeomagnetic results indicating a ~5° latitude for Neoproterozoic glaciogenic rocks in Australia[3-5]. Similarly reliable palaeolatitudes for the older, Palaeoproterozoic glaciogenic rocks have not yet been obtained, as such deposits commonly suffer from poor preservation and secondary magnetic overprinting. The Archaean-Palaeoproterozoic 'Transvaal Supergroup' on the Kaapvaal craton in South Africa is, however, exceptionally well preserved and is thus amenable to palaeomagnetic determination of depositional palaeolatitudes. Within this supergroup the ~2.2 billion-year-old Ongeluk lavas are a regionally extensive, largely undeformed and unmetamorphosed, extrusive volcanic succession[6], which conformably overlies glaciogenic deposits (the Makganyene diamictite). Here we report a palaeomagnetic estimate of 11±5° depositional latitude for the lavas, and hence for the underlying contemporaneous glacial rocks. The palaeoclimate enigma is thus deepened; a largely ice-free Precambrian world was apparently punctuated by two long ice ages, both yielding glacial deposits well within tropical latitudes.

The Ongeluk Formation comprises ~500-1,000 m of extrusive, basaltic-andesitic lavas deposited in the marine facies succession of the Transvaal Supergroup[7], exposed in the Griqualand West region of the Northern Cape Province (Fig. 1). A correlative volcanic unit,

Figure 1 (next page). Regional map (left) and schematic stratigraphic overview of the Transvaal Supergroup in the Northern Cape Province[7] (right). (BIF, banded iron formation.)


South Africa
N
A'
A
Chuniespoort region
Sampling area
Griqualand West region
30°S
500 km
Outcrop limit of Transvaal Supergroup
20°E
30°E
A
A'
Quartzite redbeds (Lucknow Fm)
Shale with basal Fe-conglomerate (Gamagara)
Mn deposits
Mn form'n and dolomite
Ongeluk Lava
Makganyene diamictite
Shallow water iron formation (Griquatown BIF)
Deep water iron formation (Kuruman BIF)
Quartzite (Koegas Subgroup)
Clastic iron formation
Predominantly platform carbonate and stromatolite (Campbellrand Subgroup)
Predominantly slope and basinal facies
Predominantly volcanic
Olifantshoek Group
Postmasburg Group
Ghaap Group
Ventersdorp Group
TRANSVAAL SUPERGROUP
Southwest
Northeast
conformable contact
unconformity
500 m
100 km

the Hekpoort Formation, occurs in the Chuniespoort region to the northeast (Fig. 1). Isopach trends[8] indicate that the two volcanic units were deposited in a single, contiguous basin across the Kaapvaal craton[9]. The Ongeluk lavas have been Pb/Pb-dated at 2,222 ± 13 Myr old[10], and the Hekpoort Formation has a recalculated Rb-Sr age[11,12] of 2,177 ± 21 Myr. Studies using the U-Pb system in zircon give ages of 2,432 ± 31 Myr for the underlying Griquatown Iron Formation[13] and ~2,050 Myr for the Bushveld complex[14] which intrudes the Transvaal sequence; these results firmly bracket the age of the Ongeluk-Hekpoort Formations between about 2.4 and 2.1 Gyr. The Hekpoort andesites were studied palaeomagnetically by Briden[15], who obtained near-present-field directions by alternating-field demagnetization of samples from only four sites. Of all other penecontemporaneous rocks (2.0 - 2.4 Gyr) on the Kaapvaal craton, only the Bushveld complex has palaeomagnetic constraints, yielding a range of palaeolatitudes from ~25° to ~60° (ref. 16).

Palaeomagnetic investigations of other Palaeoproterozoic successions have not conclusively demonstrated primary magnetic remanence. For example, the partly glaciogenic Huronian succession (2.45-2.22 Gyr) in North America has been the object of several palaeomagnetic studies[17-19], but those results vary greatly, and a more recent study of the 2.22-Gyr Nipissing diabase and other early Palaeoproterozoic intrusive rocks of the Superior craton casts doubt upon the established apparent polar wander path for the Huronian interval[20]. Huronian palaeolatitudes thus remain unresolved, partly because of greenschist- or higher-grade regional metamorphism, multiple magnetic components within and among sampling sites, and a lack of unequivocal tests for determining the ages of those components.

In the Northern Cape Province, autochthonous sedimentary rocks of the Transvaal Supergroup are better preserved. Warped into broad, open folds with bedding dips usually less than 5°, the Ongeluk Formation is exposed in several structural troughs in this area. From the mineralogy of the underlying Kuruman Iron Formation, maximum metamorphic temperatures of <170°C were estimated[21]. All localities discussed herein are road-cut exposures of fresh, fine-grained, massive or pillowed, green-grey lava, except for two sites

("OLL") whose coarse grain size may indicate that these rocks are intrusive, perhaps feeder dykes for subsequent flows. The remarkably low degree of deformation, low metamorphic grade, and fresh exposures of the Ongeluk lavas provide an ideal opportunity for preservation of primary magnetic directions from the Palaeoproterozoic.

The stratigraphic succession in the Griqualand West region was widely misinterpreted for many years[6,22] because a large thrust fault was not recognized. In its place, some chose to assign unconformities between virtually all of the formations to account for their juxtapositions; this view has been rejected by subsequent observations and geological syntheses[23] that incorporate the thrust fault and only one major unconformity within the Transvaal sequence, between the Ghaap and Postmasburg Groups (Fig. 1).

We present the following field evidence in support of the results outlined by Beukes and Smit[23], that no significant hiatus separates the Ongeluk lavas and the underlying, glaciogenic Makganyene Formation. First, although thicknesses of individual members of the Makganyene Formation vary considerably, this is to be expected in a glacial deposit[24], where thickness variations may be due to original facies variability rather than subsequent removal by erosion. Indeed, the lower sections of the Makganyene Formation have widely variable lithology, as seen in boreholes[25]. Second, the Hekpoort andesite is intercalated with the underlying Boshoek Formation, partly glaciogenic and correlative with the Makganyene Formation, at several localities[26]. Third, we and other workers[25,27] have observed, in borehole cores, apparently conformable contacts with no palaeo-weathering between the diamictite and the Ongeluk lava. Fourth, volcanic shards are abundant within the upper strata of the Makganyene Formation[25], indicating that volcanism had already commenced during late-stage glacial times. From the above arguments we contend that no significant plate motion could have occurred between deposition of the diamictite and the lava, and that palaeolatitude estimates of the latter should apply equally to the former.

Visser[28] demonstrated a glacial origin for the Makganyene diamictites, citing abundant striated and pockmarked clasts in the Griqualand West region. In addition, during our field work we found cobbles with multiple polished facets and several directions of striations from Makganyene-derived colluvium. Such textures are unlikely to have formed within nonglacial debris flows, but are common in subglacial or glaciomarine environments[24]. The glacial units are fluvial-deltaic throughout the Chuniespoort region, whereas glaciomarine conditions existed in the Griqualand West region[9]. In the latter region, rafted pebbles within mudstone imply a distal environment with input from floating icebergs, and sandstone beds between diamictite bodies are interpreted as subaqueous debris flows[25]. The sub-Makganyene unconformity involves gentle warping of the lithosphere, less than 1° in the Griqualand West region, which was followed by sheetlike sedimentation over virtually all of the Kaapvaal craton (Fig. 1). Sedimentary indicators all across the craton point to a continental (northeastern or eastern) source for the deposits, ultimately the Limpopo belt or some other subsequently rifted craton[9]; thus the Makganyene Formation cannot be ascribed to local alpine glaciation sourced from either an advancing (offshore to the present west) mountainous terrain or the so-called "Vryburg arch"[8] which is a post-Transvaal anticlinal flexure now separating the once-contiguous Griqualand West and Chuniespoort 'sub-basins' (Fig. 1). The vast regional extent of the Makganyene and equivalent formations in the Chuniespoort region also suggests proximity to a continental ice sheet rather than a local montane source.

Twenty flow-units of the Ongeluk lavas were sampled from twelve localities in the Griqualand West region. Virtually all of the Ongeluk specimens exhibited two magnetic components (Fig. 2). The first to be removed, at the low alternating-field and thermal steps, is directed moderately-to-steeply up and north, coincident with the present local field (PLF) at the sampling sites (Fig. 3a). The coercivity/unblocking spectrum of this component, which is undoubtedly of Recent age, suggests that it is held partly as a viscous-remanent magnetization by multidomain grains, and partly as a chemical-remanent magnetization by goethite, which is visible along fractured surfaces within a few of the samples.

The higher-coercivity, high-temperature component, observed in 114 of 122 samples of pillow lava or massive andesite, is westerly and of shallow negative (up) inclination (Fig. 3a).

Figure 2 (next page). Representative sample demagnetization behavior. **a.** Sample from *in situ* flow-unit. **b.** Sample from breccia clast. Orthogonal projection diagrams (top): open (solid) symbols depict projection onto the vertical (horizontal) plane. Equal-area projections (bottom): open (solid) symbols show projection onto the upper (lower) hemisphere. AF, alternating-field level in mT; PLF, present local field; NRM, natural remanent magnetization. Samples were either drilled in the field as 2.5-cm cores, or collected as oriented blocks and drilled in the laboratory. Whenever possible, samples were oriented by both magnetic and solar compasses. Cores were trimmed to 2.2 cm length. Measurement of natural remanent magnetization was followed by alternating-field (AF) and thermal demagnetization, typically measured at the following steps: 5, 10, 15, and 20 mT; 150, 200, and 250 °C; and 30 - 80 mT with 10 mT intervals. Following the low AF steps, a few samples were thermally demagnetized to 580 or 665 °C; demagnetization was halted when either intensity dropped to zero or spurious magnetizations appeared.

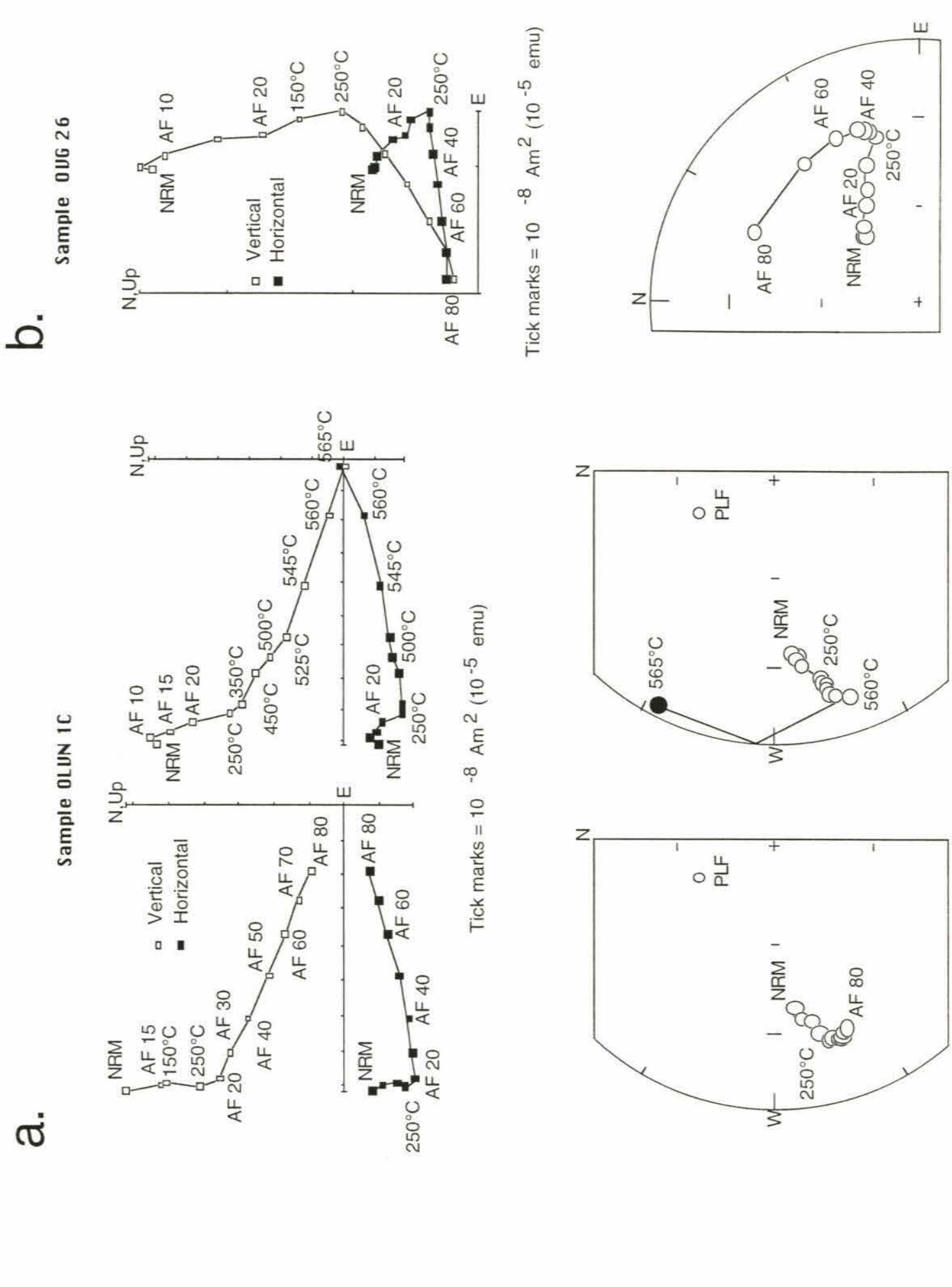
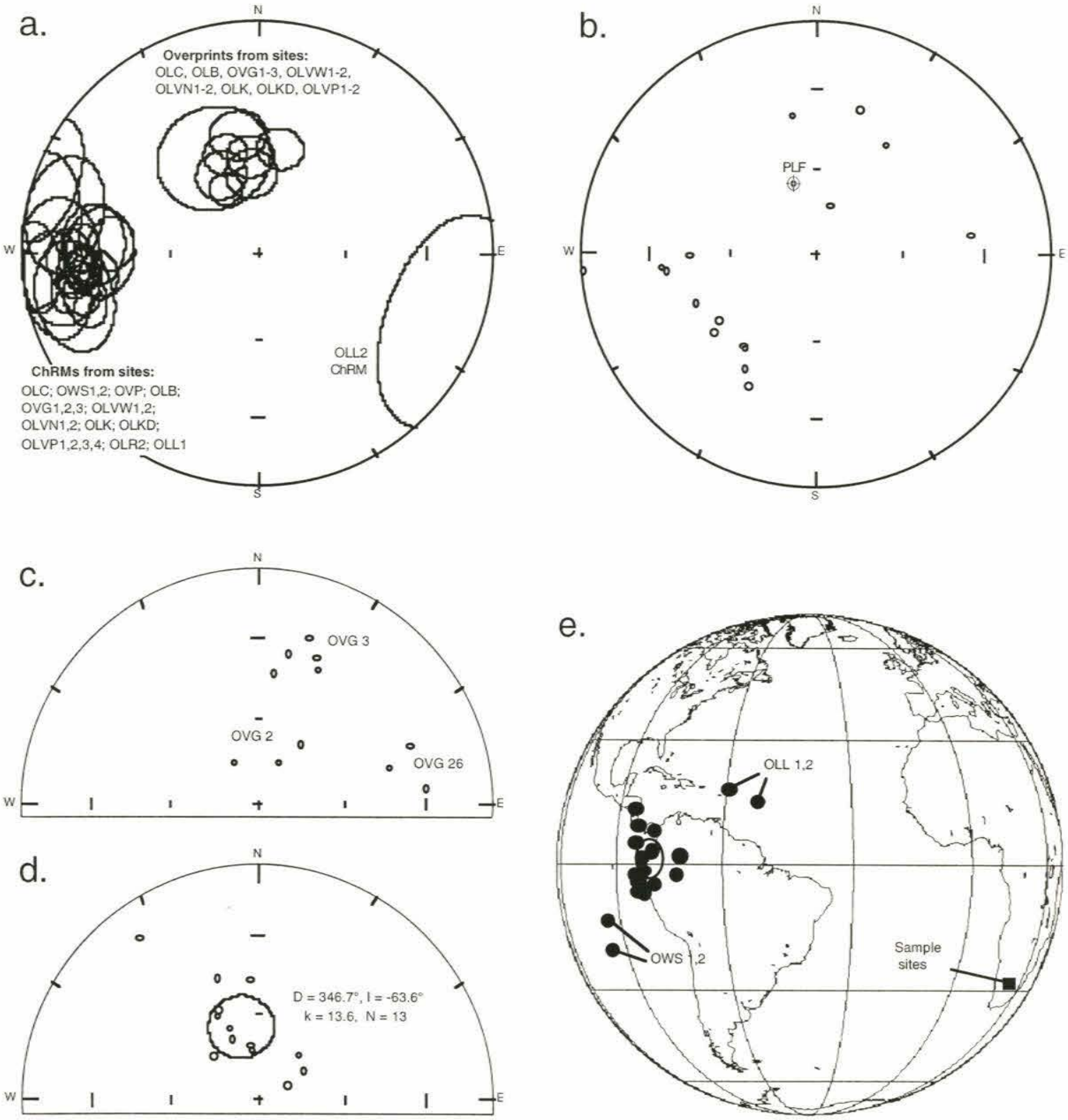

a.
Sample OLUN 1C
Vertical
Horizontal
N,Up
E
NRM
AF 15
150°C
250°C
AF 20
AF 30
AF 40
AF 50
AF 60
AF 70
AF 80
350°C
450°C
500°C
525°C
545°C
560°C
565°C
Tick marks = 10 -8 Am 2 (10 -5 emu)
N
S
W
PLF
b.
Sample OUG 26
AF 10
150°C
250°C
Tick marks = 10 -8 Am 2 (10 -5 emu)

Figure 3 (next page). Summary of directional data. Large ellipses are 95% confidence intervals around Fisher mean directions from individual sites (panels **a-d**), or the mean palaeopole of virtual geomagnetic poles from all sites (panel **e**). **a-d**, Upper- hemisphere equal-area projections. **a**, Site means of characteristic remanent magnetizations (ChRMs) and overprints. Overprints are coincident with the present local field (PLF) at the sampling sites. **b**, Individual sample ChRMs from the hyaloclastic breccia test, each sample from a distinct clast. **c**, Reproducibility of discordant clast ChRMs. Internal consistency within each clast shows that the large dispersion in **b** is not spurious. **d**, Individual sample overprints from the breccia test. Good grouping about the present field direction at the sampling sites indicates the expected failure of the breccia test of the Recent overprint, and it demonstrates that the wide dispersion in **b** is not caused by errors in sample orientation. D, declination; I, inclination; k, Fisher precision parameter; N, number of samples. **e**, Orthographic projection of virtual geomagnetic poles calculated from site-mean ChRMs (pole from site OLL2 inverted).


a.
N
W
E
S
Overprints from sites:
OLC, OLB, OVG1-3, OLVW1-2,
OLVN1-2, OLK, OLKD, OLVP1-2
OLL2
ChRM
ChRMs from sites:
OLC; OWS1,2; OVP; OLB;
OVG1,2,3; OLVW1,2;
OLVN1,2; OLK; OLKD;
OLVP1,2,3,4; OLR2; OLL1
b.
N
W
E
S
PLF
c.
N
W
E
OVG 3
OVG 2
OVG 26
d.
N
W
E
D = 346.7°, I = -63.6°
k = 13.6, N = 13
e.
OLL 1,2
OWS 1,2
Sample
sites

Five samples of the coarse-grained rock at a single site ("OLL2") are oppositely directed; as this site may be intrusive to and slightly younger than the others, its opposite polarity is not surprising. The broad, 30-80 mT coercivity of the westerlycomponent, as well as a narrow ~580 °C unblocking temperature spectrum, suggests that it is carried by single-domain magnetite. Most samples yielded extremely stable, linear demagnetization behaviour (Fig. 2a), conducive to principal-component analysis[29]. For some of the samples which did not demagnetize to the origin, the trajectories tended towards the present field direction, probably indicating incomplete removal of a haematitic (coercivity > 80 mT) component of the Recent chemical-remanent magnetization.

A palaeomagnetic breccia test at one locality ("OVG") places a constraint on the relative ages of the magnetic components. The hyaloclastic breccia at this site is thought to have formed by explosion of the partly solidified lava on entering shallow water[30]. The larger clasts have the same megascopic lithology as the overlying and underlying pillowed and massive lavas exposed in the same road-cutting, and in some instances can be identified as fragments of chilled-rim pillows[30]. Thus, if the large volcanic clasts entrained in the breccia show similar magnetic behaviour to our samples taken from *in situ* flows, then magnetization is primary in the clasts if and only if it is primary in the flows.

As distinct from a conglomerate, whose rounded clasts imply tumbling into truly random orientations, the explosion-induced hyaloclastic breccia is likely to show some differential rotation but not completely random orientations among the clasts. Our breccia test is a statistical modification of the standard palaeomagnetic conglomerate test: instead of testing against a truly uniform distribution, we compare precision of the suite of clasts with the grouping from each of the *in situ* flows. If the precision of directions from a given magnetic component is significantly lower among the clasts than the *in situ* samples, then the test is positive and magnetization of that component probably occurred prior to brecciation. Alternatively, if the precision among the clasts is similar to precision from each other site, then that component post-dates the brecciation and is not reliable for estimating depositional palaeolatitudes.

Sixteen clasts from the breccia at locality OVG show considerable scatter of characteristic remanent magnetization relative to the massive and pillowed andesite at the same and other localities (Fig. 3b, Table 1). Individual clasts carry reproducible directions (Fig. 3c), and the suite carries a well grouped PLF overprint similar to that found at other sites (Fig. 3d). Relative precision parameters are scaled against an F-ratio distribution, to determine the likelihood of indistinguishable precision[31]. Common precision between the distribution of each other site and that of the breccia suite can be rejected with >99% confidence for all flow-units, except for site OLL2 (~80%), whose coarse grain size may effect different bulk magnetic properties. The cluster of some of the clast directions toward the southwest may be due to minimal rotation of these clasts during explosive brecciation, or subsequent heating by later flows, or immediate hydrothermal alteration after brecciation. Notably, these samples are from the same proximity in outcrop, with anomalously low magnetic susceptibilities; this may indicate a chemical or mineralogical change due to immediate and localized hydrothermal alteration after emplacement of the breccia[10]. Considering the suite of clasts as a whole, however, their wide dispersion rules out any pervasive, secondary remagnetization at locality OVG, and the similarity of directions from *in situ* flows at OVG with other sites supports the interpretation that the characteristic remanent magnetization components were all obtained during deposition of the lavas.

This positive breccia test, in addition to a reversely directed site, implies that the high-coercivity, high-unblocking-temperature, characteristic remanent magnetization component from the *in situ* samples, is probably a thermal-remanent magnetization acquired by initial cooling of the lava flows. Within-flow scatter of directions is small, typically smaller than between-site scatter (Table 1), and the axially symmetric (Fisherian) distribution of virtual geomagnetic poles (Fig. 3e) supports the model that site-means are thermal-remanent magnetizations of geomagnetic secular variation about an axially geocentric dipole field. We

Table 1. Mean characteristic remanence directions and virtual geomagnetic poles for the Ongeluk Formation, Griqualand West region, South Africa.

Site/flow	South Lat.	East Long.	n/N	GDec	GInc	k	TDec	TInc	k	Plat(N)	Plong(E)
OWS1	27°07'	022°59'	5/7	239.6	-22.5	100	238.5	-26.9	99.1	-19.8	264.3
OWS2	27°07'	022°59'	6/6	246.2	-27.9	67.3	245.1	-32.4	70.0	-12.7	265.3
OVP	28°17.60'	023°19.68'	5/5	271.6	-26.0	89.8	271.6	-26.0	89.8	07.8	281.9
OLB	28°53.85'	022°48.52'	7/7	257.8	-16.9	118	257.8	-16.9	118	-06.4	279.2
OVG1	28°53.80'	022°49.26'	5/6	266.7	-26.2	54.9	267.5	-21.4	55.8	03.1	281.9
OVG2	28°53.80'	022°49.26'	6/6	265.8	-29.9	103	264.7	-25.4	103	01.9	278.5
OVG3	28°53.80'	022°49.26'	6/6	261.5	-18.8	17.9	260.9	-14.1	17.9	-04.4	282.1
Breccia	28°53.80'	022°49.26'	16/16	256.5	-62.5	3.4	255.5	-57.7	3.4	not used	
OLVW1	28°53.70'	022°50.20'	5/6	266.1	-05.8	34.5	266.1	-05.8	34.5	-02.0	288.4
OLVW2	28°53.55'	022°51.51'	6/7	270.0	-08.0	13.0	270.0	-08.0	13.0	02.0	289.3
OLVN1	28°53.55'	022°51.82'	6/6	259.8	-25.3	251	259.8	-25.3	251	-02.3	276.3
OLVN2	28°53.55'	022°51.82'	6/6	259.4	-21.9	159	259.4	-21.9	159	-03.6	277.8
OLK	28°53.60'	022°53.10'	6/6	261.3	-30.7	60.7	258.7	-24.3	59.4	-03.5	277.3
OLKD	28°53.66'	022°53.30'	8/8	264.2	-25.9	61.0	264.2	-25.9	61.0	01.6	278.2
OLVP1	28°53.62'	022°56.11'	4/4	267.4	-24.9	36.5	270.3	-33.4	36.5	09.0	277.0
OLVP2	28°53.62'	022°56.11'	4/4	269.8	-29.6	89.5	273.6	-37.8	89.8	13.0	275.7
OLVP3	28°53.68'	022°57.02'	6/6	266.8	-30.7	25.0	266.8	-30.7	25.0	05.2	276.9
OLVP4	28°53.68'	022°56.65'	6/6	261.5	-16.3	12.7	262.0	-22.1	12.7	-01.3	279.0
OLR2	28°53.59'	022°57.45'	6/6	256.6	-20.6	14.9	256.6	-20.6	14.9	-06.3	277.1
OLL1	28°53.61'	023°02.01'	6/6	289.0	-03.7	14.9	289.0	-03.7	14.9	17.5	300.8
OLL2	28°53.54'	023°02.77'	5/8	110.0	-11.1	6.0	110.0	-11.1	6.0	-14.6	128.0
Mean of 20 sites (OLL2 inverted):			114/122							00.5	280.7
										($K = 38.3$, $A_{95} = 5.3°$)	

n, number of samples used; N, total number of samples; GDec, mean geographic (*in situ*) declination; GInc, mean geographic inclination; k, estimate of Fisher precision parameter; TDec, mean tilt-corrected declination; TInc, mean tilt-corrected inclination; Plat, pole latitude; Plong, pole longitude; K, precision parameter of site means; A_{95}, semi-radius of 95% cone of confidence.

find the tilt-corrected, normal-polarity palaeomagnetic pole at 0.5°N, 280.7°E, implying a depositional palaeolatitude of 11±5° for the Griqualand West region. On the seven-point 'Q' scale[32] of reliability, this pole rates a perfect seven.

The low palaeolatitude of the Makganyene diamictite compounds the enigma of Precambrian glaciations. Although glaciogenic deposits of generally Palaeoproterozoic age occur on many cratons, our study provides the first direct and reliable determination of any of these units. It is now documented that both of the broad intervals of Precambrian glaciation, near the beginning and end of the Proterozoic Eon, include glaciogenic sediments deposited in equatorial latitudes. This may indicate a global climatic system fundamentally different (due, for example, to changes in Earth's orbital obliquity[33]) from that of the past 500 Myr, when glacial deposits were restricted largely to the polar regions. Alternatively, the low-latitude Precambrian glacial deposits could indicate severe, globally inclusive ice ages (a model called the "Snowball Earth"[34]). In that case, our planet's subsequent recoveries to more mild temperatures would indicate a remarkable resilience to extreme perturbations in climate.

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