Paleomagnetic polarity reversals in Marinoan (ca. 600 Ma) glacial deposits of Australia: Implications for the duration of low-latitude glaciation in Neoproterozoic time

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ABSTRACT

A paleomagnetic investigation of Marinoan glacial and preglacial deposits in Australia was conducted to reevaluate Australia's paleogeographic position at the time of glaciation (ca. 610–575 Ma). The paleomagnetic results from the Elatina Formation of the central Flinders Ranges yield the first positive regional-scale fold test (significant at the 99% level), as well as at least three magnetic polarity intervals. Stratigraphic discontinuities typical of glacial successions prevent the application of a magnetic polarity stratigraphy to regional correlation, but the positive fold test and multiple reversals confirm the previous low paleolatitude interpretation of these rocks (mean D = 214.9°, I = –14.7°, α_{95} = 12.7°, paleolatitude = 7.5°). The underlying preglacial Yaltipena Formation also carries low magnetic inclinations (mean D = 204.0°, I = –16.4°, α_{95} = 11.0°, paleolatitude = 8.4°), suggesting that Australia was located at low paleolatitude at the onset of glaciation. The number of magnetic polarity intervals present within the Elatina Formation and the Elatina's lithostratigraphic relationship to other Marinoan glacial deposits suggest that glaciation persisted at low latitudes in Australia for a minimum of several hundreds of thousands to millions of years.

INTRODUCTION

The Neoproterozoic Era (1000–543 Ma) marks a time of unusual and profound changes on the surface of the Earth. Supercontinents assembled, then broke apart with apparent great rapidity (Moores, 1991; Dalziel, 1991, 1995; Hoffman, 1991; Powell et al., 1993). The first complex metazoan organisms appeared (e.g., Cowie and Brasier, 1989; Lipps and Signor, 1992; Bengtson, 1994; Grotzinger et al., 1995; Xiao et al., 1998), perhaps spurred by enhanced carbon burial that yielded significant amounts of oxygen (Knoll, 1992, 1994). There were two broad intervals of widespread and extended (and possibly repetitive) glaciation in the Neoproterozoic Era: the Sturtian glacial interval (ca. 780–700 Ma) and the Varanger glacial interval (which includes the Marinoan glaciation in Australia and Ice Brook glaciation in Canada, ca. 610–575 Ma; see Knoll and Walter, 1992; Bowring and Erwin, 1998). These glacial intervals are unlike any in Phanerozoic Earth history because of their great severity. Paleomagnetic evidence from glacial strata suggests that continental-scale ice sheets advanced over land at or near sea level and equatorward of 20° latitude (Sturtian of North America; Park, 1994, 1997; Marinoan of Australia; Schmidt et al., 1991; Schmidt and Williams, 1995). On only one other occasion in Earth history, the Paleoproterozoic Huronian glaciation (ca. 2.2 Ga), may glaciation have been equally severe (Evans et al., 1997; Williams and Schmidt, 1997).

Paleoclimatologists accustomed to Pleistocene glacial conditions tend to greet word of low-latitude glaciation with polite disbelief, because the Pleistocene climate models that have guided thinking about climatic processes have never needed to address such an extreme case of global cooling. Most climate models for glacial intervals since the days of the CLIMAP Project (CLIMAP Project Members, 1976) have assumed steep thermal gradients and general cooling of oceanic surface water at middle to high latitudes, with little or no change in tropical ocean surface temperatures. Some researchers have even proposed that the average equatorial sea-surface temperature has remained within 1 °C of its present level for billions of years (Fairbridge, 1991). Only in the past few years has evidence of cooling of the tropics come to light, and the calculated temperature changes are only a few degrees in magnitude (Beck et al., 1992; Stute et al., 1995; Broecker, 1996; Patrick and Thunell, 1997).

In light of Pleistocene climate modeling, the occurrence of low-latitude glaciation in Neoproterozoic time seems bizarre; however, it is the rarity of low-latitude glaciation in the Precambrian that is strange. After all, models of stellar evolution suggest that 600 Ma, the sun's luminosity should have been 3%–6% lower than present levels (e.g., Gough, 1981). Under those conditions, producing an extremely frigid climate should have been simple. However, the existence of water-lain sedimentary deposits dating back to 3.9 Ga indicates that temperatures were warm enough for running water, a problem known as the "faint young sun" paradox (Sagan and Mullen, 1972; Newman and Rood, 1977). To compensate for reduced solar luminosity, researchers have postulated a supergreenhouse effect, caused by elevated levels of greenhouse gases such as methane and CO_2, that declined with the passage of time until the atmosphere achieved a composition similar to that of the present day (Sagan and Mullen, 1972; Newman and Rood, 1977; Owen et al., 1979; Kasting, 1992). (Note that an "effective" level of CO_2 is often used as a replacement for other

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greenhouse gases in climate modeling [Washington, 1992].) Estimates of normal atmospheric CO_2 levels ca. 600 Ma are around 840 ppm (Carver and Vardavas, 1994), compared to the preindustrial value of ~280 ppm. As Kasting (1992) pointed out, having to account for low-latitude glaciation at sea level in a supergreenhouse world presents a rather difficult problem.

The entire Proterozoic climate scene may seem too far removed from present-day conditions to be of any interest to paleoclimatologists working in more recent geologic intervals. However, the physics of climatic processes does not change—only the boundary conditions do. While great strides have been made in improving our understanding of the various forcings and feedbacks that control climate today and in the recent geologic past (e.g., CLIMAP Project Members, 1976; Hays et al., 1976; Hansen and Takahashi, 1984; Broecker and Denton, 1989), the role that these perturbations play in long-term climate trends and the possible range of climate variability is still poorly understood. Understanding the circumstances under which the Earth's climate can be compelled to change from unusual warmth to extreme cold, as well as the time scales over which various processes interact, may provide unique insights that improve our understanding of the current climatic conditions.

In this paper we approach the low-latitude glaciation question by analyzing the most comprehensive evidence yet acquired from the Neoproterozoic of Australia in support of the hypothesis. We then discuss the implications of these data for the duration of low-latitude glaciation during the Varanger glacial interval. In the interests of opening the discussion of the low-latitude glaciation hypothesis to a broader interdisciplinary audience, we have also provided some additional background material on selected topics.

PALEOMAGNETISM AND THE LOW-LATITUDE GLACIATION HYPOTHESIS

The hypothesis of low-latitude glaciation during the Proterozoic dates back nearly 40 years, when researchers first noted some major differences between Neoproterozoic glacial deposits and the much younger (and more familiar) glacial deposits of the Cenozoic (Harland and Bidgood, 1959; Girdler, 1964; Harland, 1964a, 1964b). First, the extensive Neoproterozoic glacial deposits are widely scattered across the globe: they can be found on every continent, with the possible exception of Antarctica (cf. Stump et al., 1988). Such a distribution is unlike the more areally limited distribution of Pleistocene glacial deposits. Second, the glacial deposits appeared to be in close spatial and temporal relationships with rocks normally associated with warm, arid, or tropical settings. Redbeds, carbonates, and/or evaporites are commonly found stratigraphically beneath the glacial rocks, and the end of glacial deposition for both the Sturtian and Varanger intervals is typically marked by the presence of thin, laterally persistent (over hundreds of kilometers), laminated, buff to pink dolomites now called cap carbonates (because they cap the glacial deposits; Roberts, 1976; Williams, 1979; Deynoux, 1985; Young, 1992; Kennedy, 1996). Third, early paleomagnetic studies of rocks associated with (but not from) Varanger-age glacial deposits in Greenland and Norway yielded low paleolatitudes (Harland and Bidgood, 1959; Bidgood and Harland, 1961). The widespread distribution of the glacial deposits, their unusual relationship with carbonates and other arid-climate deposits, and the paleomagnetic data led Harland (1964a) to postulate that somehow, the entire Earth had become glaciated during an interval he referred to as "the great Infra-Cambrian glaciation" (see also Chumakov and Elston, 1989; Kirschvink, 1992). As the field of paleomagnetism developed and methods for measuring and analyzing paleomagnetic remanence improved, additional studies were undertaken in rocks associated with certain glacial units in Africa (Kröner et al., 1980; McWilliams and Kröner, 1981), North America (Morris, 1977), China (Zhang and Zhang, 1985; Li et al., 1991), and Australia (McWilliams and McElhinny, 1980; Embleton and Williams, 1986; Schmidt et al., 1991). All of these studies seemed to confirm that low-latitude glaciation had occurred over much of the existing landmass during the Neoproterozoic.

The low-latitude glaciation hypothesis has met considerable resistance over the years because it doesn't fit a familiar rule of thumb: sedimentary deposits commonly linked with particular climatic conditions (e.g., carbonates, coal, bauxite) are supposed to lie within certain latitudinal ranges, because the climates that influenced those deposits have latitude-dependent mean temperatures in the more modern world (Briden and Irving, 1964). (It is ironic that this rule of thumb was based upon a frequency analysis of the paleolatitudes of such deposits as determined by paleomagnetism.) Objections to the hypothesis on sedimentological grounds have centered on the possibility that widespread diamictites are nonglacial, perhaps related to debris flow in tectonically active basins (Schermerhorn, 1974) or meteorite impact ejecta blankets (e.g., Oberbeck et al., 1993; Rampino, 1994; cf. Young, 1993). For the most part, such objections are easily addressed because there is ample evidence of glaciation. Thick successions (to thousands of meters) of marine diamictite contain glacially striated and faceted stones (e.g., Hambrey and Harland, 1981, 1985; Eyles, 1993). Associated laminated siltstones contain dropstones that are inferred to have been ice rafted (Crowell, 1964; Ovenshine, 1970; Vorren et al., 1983) rather than transported by some other means (e.g., biological rafting or gravitational processes; Bennett et al., 1994; Menzies, 1995). Signs of erosion beneath ice sheets are preserved, particularly around basin margins, as rare, glacially striated pavements (Biju-Duval and Gariel, 1969), in some cases associated with valleys tens to hundreds of meters deep (Christie-Blick, 1983; Edwards, 1984; Karfunkel and Hoppe, 1988; Sønderholm and Jepsen, 1991). Periglacial features, such as sand wedges and frost-heaved blocks, are also found in connection with a number of the glacial deposits (e.g., Nystuen, 1976; Deynoux, 1982; Williams and Tonkin, 1985; Zhang, 1994). Considering the extent and regional stratigraphic context of those deposits, it is clear that debris flows or meteorite impacts cannot explain the Neoproterozoic deposits (Young, 1993). Suggestions that the deposits are glacially derived but were produced by glacial activity in an alpine setting (Eyles, 1993) cannot account for all circumstances, because several examples were clearly deposited at some distance from features with significant topographic relief (Christie-Blick, 1983; Deynoux et al., 1989; Moncrieff, 1989; Lemon and Gostin, 1990).

Criticism of the low-latitude glaciation hypothesis also has been based upon the available paleomagnetic data. Meert and Van der Voo (1994) assessed the quality of recent paleomagnetic studies supporting the low-latitude hypothesis, including those studies mentioned here, by applying criteria developed to evaluate the reliability of paleopoles calculated from such data (e.g., Van der Voo, 1990). While not foolproof, reliability criteria provide a more uniform way to evaluate the likelihood that the magnetic remanences measured are primary. In their review, Meert and Van der Voo (1994) found that the majority of the studies they examined had low reliability scores for a variety of reasons. Poor stratigraphic correlations between rocks sampled and the glacial rocks (Congo), possible remagnetizations (Namibia, north China), and improved data reflecting at least mid-latitude depositional settings (Baltica) have cast doubt on the data supporting the low-latitude glaciation hypothesis for nearly all of the locations examined. While most continents are now thought to have been glaciated at middle latitudes or higher during either of the Neoproterozoic glacial intervals (cf. Park, 1997), the Marinoan glacial interval of Australia continues to provide the best evidence for the existence of an unusually cold climate in the Earth's past.

The most recent paleomagnetic studies from Australia (Schmidt et al., 1991; Schmidt and

Williams, 1995) have yielded excellent data supporting the low-latitude glaciation hypothesis. However, these studies still leave open the possibility that the magnetic remanence preserved is not primary, but a remagnetization that occurred perhaps 50–90 m.y. after glaciation, possibly in relation to the Delamerian orogenic event in South Australia (as noted in Meert and Van der Voo, 1994). The Schmidt et al. (1991) study was problematic because the samples were taken from a single section of tidal rhythmites in the southern Flinders Ranges representing as little as 60–70 yr, an interval that is insufficient for averaging out the secular variation of the Earth's magnetic field. Schmidt and Williams (1995) addressed this problem by sampling a number of sites from three separate sections in the central Flinders Ranges, and reported a possible reversal. However, the combination of a small number of sites sampled (five per section), scatter in the data as reflected by small values of k (i.e., the best estimate of Fisher's precision parameter), and relatively small structural variation between sections did not permit the application of a regional-scale fold test to help constrain the age of the magnetization.

Moreover, Schmidt and Williams's (1995) principal argument for the primary nature of the magnetization still rested upon the samples' stable, high-temperature magnetic remanence (a good indication, but not a guarantee), as well as the results of micro-scale fold tests conducted on slabs of the southern Flinders rhythmite (Sumner et al., 1987; Schmidt et al., 1991). Williams (1996) interpreted the small folds sampled as synsedimentary, describing them as part of gravity-slide deposits that formed after the sediments had been transformed into a hydroplastic state. The photographic evidence provided, however, does not demonstrate the sort of disharmonic folding normally considered characteristic of synsedimentary folds (e.g., Ricci Lucchi, 1995). If these small folds are not synsedimentary, their origins—and thus the origins of the magnetization they preserve—remain unclear. These uncertainties indicate a need for additional paleomagnetic work.

NEW PALEOMAGNETIC RESULTS FROM AUSTRALIA

The objective of this study was to acquire paleomagnetic data that could provide the strongest constraints yet on the paleolatitudes of Marinoan glacial deposits, especially by constraining the timing of magnetic remanence acquisition. To accomplish this goal, high-resolution sampling was undertaken in locations that permitted us to apply field stability tests for the age of magnetization (e.g., was the remanence acquired before, during, or after tectonic folding?). (Part A of Data Repository[1] discusses sampling and analytical methods.) Detailed sampling would also permit the construction of a magnetostratigraphy that could be used to correlate between stratigraphic sections, both within and between basins, to investigate whether magnetization was acquired in close association with sedimentation. Preglacial strata were also sampled to determine whether it was likely that Australia had been glaciated at high latitudes, and then drifted into lower latitudes where the glacial rocks had acquired a low magnetic inclination during remagnetization (referred to here as the drift hypothesis; Crowell, 1983; Meert and Van der Voo, 1994). Various regions of Australia with differing tectonic histories were sampled in an effort to gain independent confirmation of our paleomagnetic results for the Marinoan. Not all of these efforts were entirely successful, but several yielded critical data that support the low-latitude glaciation hypothesis.

Regional and Stratigraphic Context for Sampling

Three regions with lithostratigraphic units of roughly equivalent age were selected for this study: the central Flinders Ranges and Stuart shelf of South Australia, and the northeastern portion (Ooraminna subbasin) of the Amadeus basin of central Australia (Fig. 1). Structural deformation of the rocks in these regions varies in both degree and timing. The strata of the central Flinders Ranges were gently warped into a series of doubly plunging folds during the Delamerian orogeny, a mid-Early Cambrian to earliest Ordovician (ca. 526–480 Ma) collisional event that affected the southeastern margin of the Australian paleocontinent (e.g., Preiss, 1987; Jenkins, 1990; Turner et al., 1994; Chen and Liu, 1996; Flöttman et al., 1998; see time scale of Bowring and Erwin, 1998). The Stuart shelf, located on the eastern edge of the Gawler craton, was virtually unaffected by the Delamerian orogeny and has not been subjected to any significant tectonic activity through the Phanerozoic; its strata are essentially undeformed (Sprigg, 1952; Thomson, 1969; Preiss and Forbes, 1981; Drexel and Preiss, 1995). The Ooraminna subbasin of the Amadeus basin underwent some deformation during Neoproterozoic time as a result of salt movement (Field, 1991; Kennedy, 1993, 1994), but Sturtian and younger rocks were not involved in any significant compressional tectonic events until the Alice Springs orogeny in Devonian-Carboniferous time (Oaks et al., 1991; Ding et al., 1992). Of these three regions, only the central Flinders Ranges provided useful paleomagnetic data; these are discussed in detail in following sections. (Paleomagnetic data from the Stuart shelf and Amadeus basin that proved uninterpretable are discussed in Part B of Data Repository; see footnote 1.)

Paleomagnetic sampling in the central Flinders Ranges focused upon four formations: the preglacial Trezona and Yaltipena Formations, and the late glacial to early postglacial Elatina and Nuccaleena Formations (see Fig. 2 for regional stratigraphic context of these units, and Fig. 3 for sampling locations). The Trezona and Yaltipena Formations are important for this study not simply because lithostratigraphic relationships mark them as preglacial, but also because the sedimentology of these two units has been interpreted as typical of a warm, tropical setting (Lemon, 1988). Thus, the Trezona and Yaltipena Formations provide an opportunity to test the "drift hypothesis" (Crowell, 1983). The Trezona Formation consists of stromatolitic and intraclastic limestone layers alternating with calcareous and non-calcareous siltstones that has been interpreted as typical of a warm-water carbonate ramp (Preiss, 1987; Lemon, 1988). Locally preserved at the top of the Trezona Formation is the Yaltipena Formation (formerly referred to informally as the Yaltipena Member of the Trezona Formation; Lemon, 1988), which consists of red siltstones and sandstones that contain mud cracks, dolomitic tepees, and oscillation ripple marks characteristic of very shallow water depths (Lemon, 1988; Lemon and Reid, 1998), as well as rare calcite vugs that may be evaporite replacements.

The glaciomarine Elatina Formation consists largely of feldspathic, silty to fine-grained sandstones that display considerable local facies variation (Lemon and Gostin, 1990). The basal unit is typically a diamictite or pebble conglomerate of variable thickness, with a sandy matrix and pebble to cobble and boulder-sized clasts of diverse lithology. In the areas north of Trezona Bore (Fig. 3), exposures of the basal diamictite and/or conglomerate rest upon an unconformable surface displaying several meters of erosional relief and plastic deformation of the underlying, partially lithified red sandstones of the Yaltipena Formation; these features are suggestive of ice-contact deformation. Additional diamictite beds with silty matrix and cobble-sized clasts of diabase that are most likely derived from xenoliths exposed in nearby diapirs (Fig. 3) are also found in this area (Lemon and Gostin, 1990). The tidal rhythmite member of the Elatina Formation, which was the focus of some of the earlier paleomagnetic studies (Embleton and Williams, 1986; Schmidt et al., 1991), is repre-

[1]GSA Data Repository item 9965, supplemental tables and data, is available on the Web at http://www.geosociety.org/pubs/ftpyrs.htm. Requests may also be sent to Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301; e-mail: editing@geosociety.org.

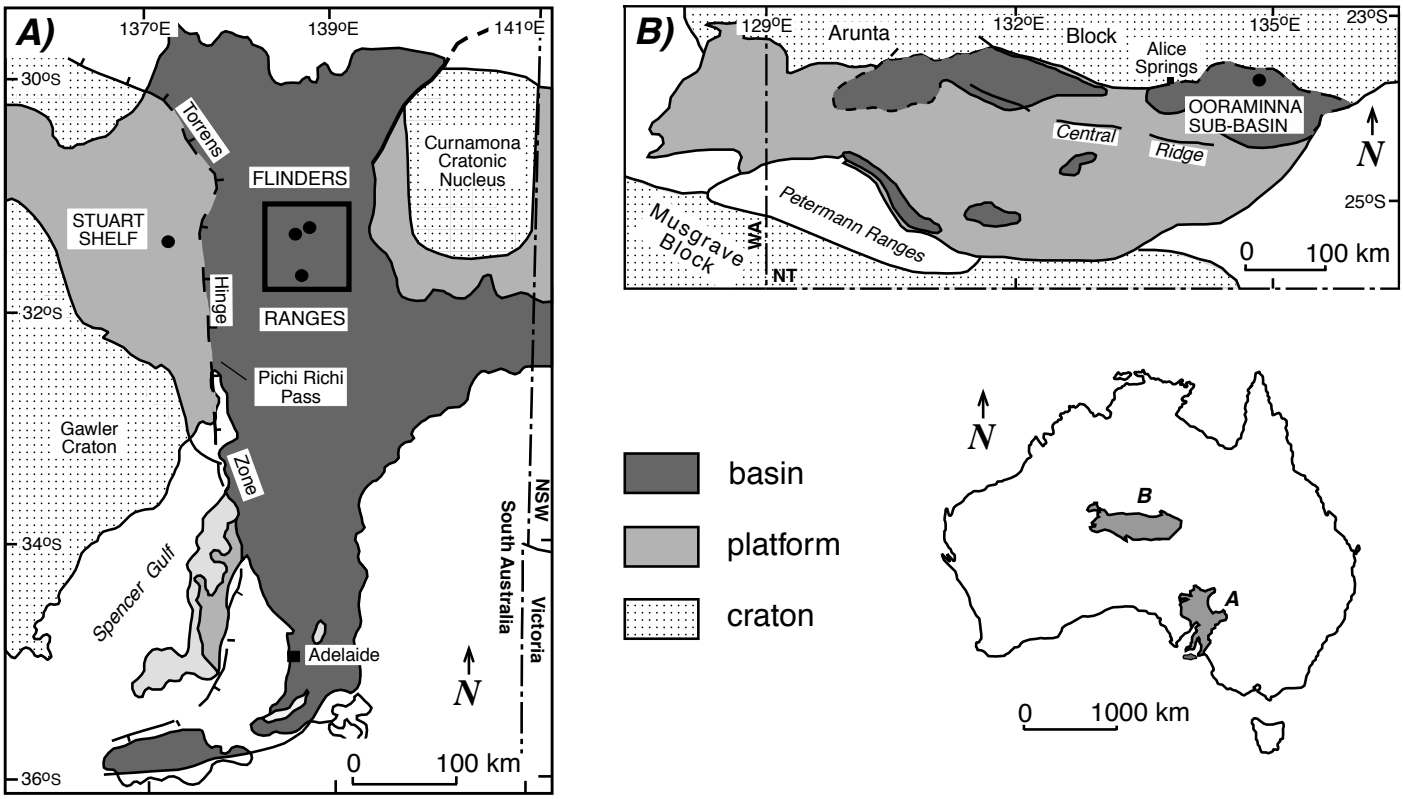

Figure 1. Locations of sections sampled for paleomagnetic study. (A) Adelaide geosyncline and Stuart shelf. (B) Amadeus basin (adapted from Preiss, 1987; Lindsay et al., 1987; Lindsay and Korsch, 1991). Area inside box in A is shown in greater detail in Figure 3.

sented in the central Flinders Ranges by a single interval <1 m thick at the northern end of Arkaba diapir, ~36 km south-southwest of the Elatina Formation section sampled at Trezona Bore (Fig. 3); it is interbedded with the very fine grained feldspathic sandstone common elsewhere near the top of the formation. The Elatina Formation is overlain disconformably by the Nuccaleena Formation. The Nuccaleena is the cap carbonate of the central Flinders Ranges; it is typically a finely laminated, buff-weathering pink dolomite, although some facies variations have been observed (Preiss, 1987; Kennedy, 1996). Although the presence of cap carbonates like the Nuccaleena have been interpreted in the past as a return to warm tropical conditions (e.g., Lemon and Gostin, 1990), the proposed unusual geochemical conditions associated with these carbonates as well as their subtidal depositional settings (Kennedy, 1996; Hoffman et al., 1998) make the determination of temperate vs. tropical depositional environments for the cap carbonates difficult at present.

The Elatina and Nuccaleena Formations were sampled along three sections that afforded the best opportunities to conduct a fold test (see Figs. 1A and 3). Two of these sections, the Bennett Spring section (31°13′S lat, 138°48′E long) and the Trezona Bore section (31°17′S lat, 138°39′E long) are ~18 km apart and are on opposite sides of a salt diapir. The third section at Warcowie (31°47′S lat, 138°40′E long) is located ~55 km south-southeast of Trezona Bore. The average bedding attitude of the Elatina and Nuccaleena Formations at the Bennett Spring Section is strike 165°, dip 14°NE; at the Trezona Bore section, strike 170°, dip 16°SW; and at Warcowie, strike 055°, dip 85°NW. The Trezona and Yaltipena Formations were sampled along only two sections, because at Warcowie the exposure of the Trezona Formation is too incomplete and the Yaltipena Formation is completely absent. The average bedding attitude for the Trezona and Yaltipena sections at Bennett Spring is strike 168°, dip 14°NE; at the Northern Park section, located a few hundred meters north of the Elatina Formation's Trezona Bore section to take advantage of superior outcrop, it is strike 167°, dip 16°SW.

Analysis of Paleomagnetic Results

The analysis of the paleomagnetic data illustrates well the difficulties in obtaining reliable results from shallow-water carbonates and diamictites. Remanence directions from the carbonates of the Trezona and Nuccaleena Formations are widely scattered and poorly defined. Their relatively low unblocking temperatures (<600 °C) and isothermal remanent magnetization (IRM) tests of the magnetic mineralogy in these formations (method of Lowrie, 1990) suggest that magnetite and goethite are the principal remanence carriers. The scattering and poor definition of the magnetization directions, as well as the common presence of goethite, suggest that the carbonates carry unstable magnetizations unsuitable for use in paleomagnetic studies. Thus, the carbonates have not been included in subsequent analysis.

The Yaltipena and Elatina Formations yielded much better results overall. Two components of natural remanent magnetization (NRM) could be consistently identified as linear thermal demagnetization trajectories in these samples (Fig. 4): a low-temperature component A (0–300 °C), and a high-temperature component C (600–675 °C). In many samples there is also some indication of an intermediate temperature component B (350–575 °C; see, e.g., Fig. 4I). The remanence directions recorded by the A components of the Yaltipena and Elatina Formations are northerly with moderately steep inclinations in in situ coordinates, similar to (al-

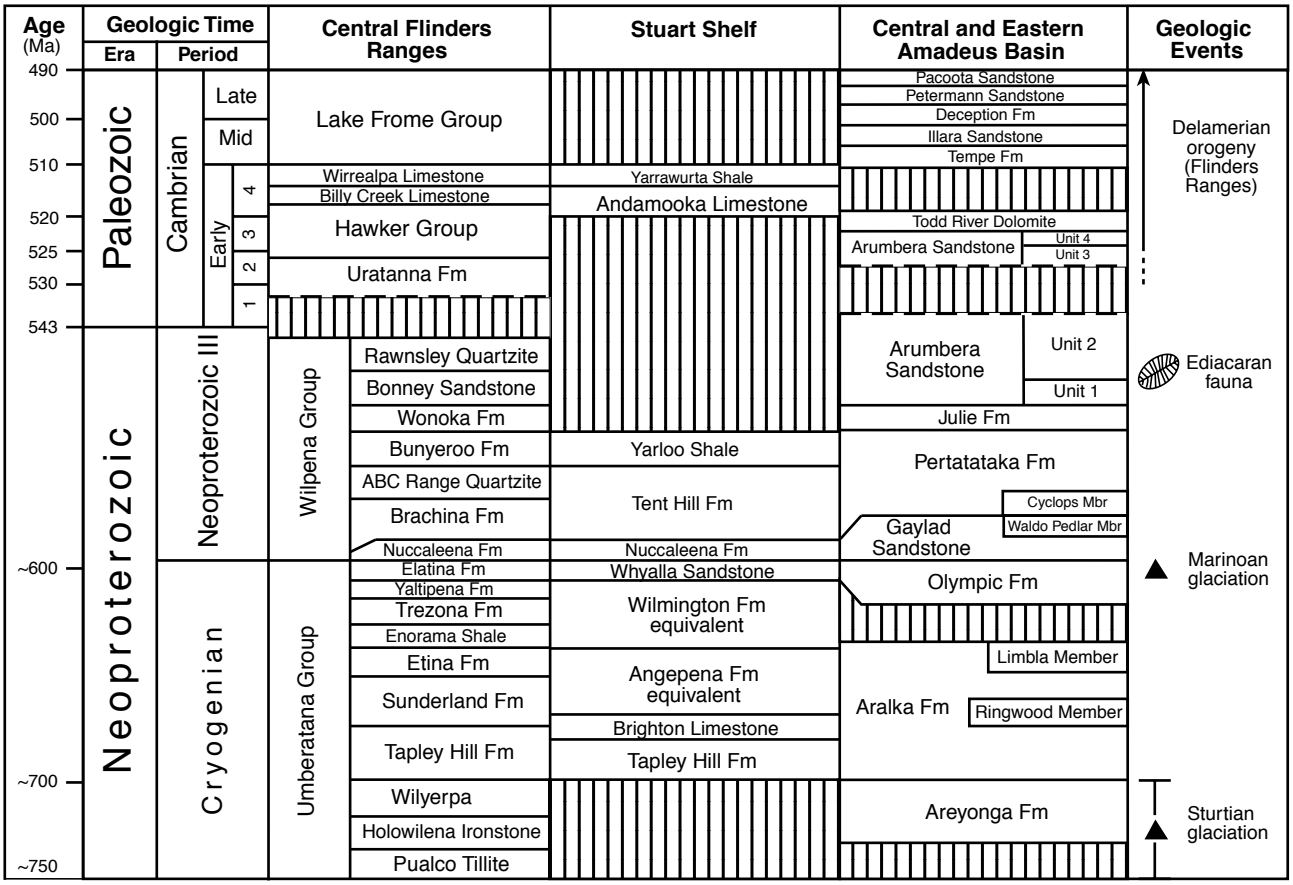

Figure 2. Stratigraphic correlation chart for the regions sampled in this study, from mid-Neoproterozoic through Cambrian time (adapted from Preiss, 1987; Kennard and Lindsay, 1991; Shergold et al., 1991; Gravestock, 1993; Callen and Reid, 1994; Preiss et al., 1998). The time scale for the Cambrian Period was adapted from Bowring and Erwin (1998); subdivisions of the Early Cambrian shown are: 1 = Nemakit-Daldynian (Manykaian), 2 = Tommotian, 3 = Atdabanian, and 4 = Botomian. Neoproterozoic age estimates for the Sturtian and Marinoan glaciations are from Knoll and Walter (1992). The dotted lines marking the bottom of the Uratanna Formation, as well as the bottom of the Arumbera Sandstone unit 3 and top of the Arumbera unit 2, reflect the current uncertainty in the ages of these boundaries (e.g., Mount and McDonald, 1992).

though distinct from) the orientation of the present-day magnetic field in the region (Fig. 5, A and C). The A component is likely a thermoviscous magnetization acquired during the Tertiary uplift of the Flinders Ranges (Wellman and Greenhalgh, 1988).

In both the Yaltipena and Elatina samples, the site mean C component directions have moderate to shallow inclinations and either north-northwest or southwest declinations (Fig. 6), although there are slight differences in the distribution of the site mean directions from section to section (see Table 1 for summary of data). The overall remanence direction for the Elatina Formation in this study (site mean D = 212.1°, I = –16.9° for 58 sites) is similar to those directions previously published (e.g., D = 197.4°, I = –7.1° for 10 sites; Schmidt and Williams, 1995).

In contrast to the A and C components, the B component is not well defined (Fig. 5, B and D); on the site level, the remanence directions are spread out in a broad band with generally northwest directions. The B component vectors typically make a small angle with either the A or C component, or even show signs of curvature (e.g., Fig. 4I). These characteristics combined suggest two possible origins for the B component: (1) it is an artifact resulting from overlap between the low-temperature, thermoviscous A component and the high-temperature C component; or (2) it represents a magnetization component, the demagnetization spectrum of which overlaps that of the C component to such an extent that it is difficult to isolate. The latter origin is supported by aspects of the magnetic mineralogy and stability tests performed on the samples (see following discussions).

Magnetic Mineralogy

Thermal demagnetization of the Yaltipena and Elatina samples and IRM tests of the magnetic mineralogy (method of Lowrie, 1990) have identified hematite as the dominant magnetic remanence carrier, although traces of goethite (a recent weathering product) and rare relict magnetite are also present (Fig. 7). A scanning electron microscope (SEM) survey revealed numerous silt- to very fine sand-sized opaque detrital grains; these grains include various Fe-Ti oxides displaying exsolution lamellae, as well as grains that are now hematitic in composition, but that retain the relict crystallographic structure of magnetite (see also Embleton and Williams, 1986). The ultimate source of the opaque detrital grains must be igneous or metamorphic, because only these sources would have achieved the high tempera-

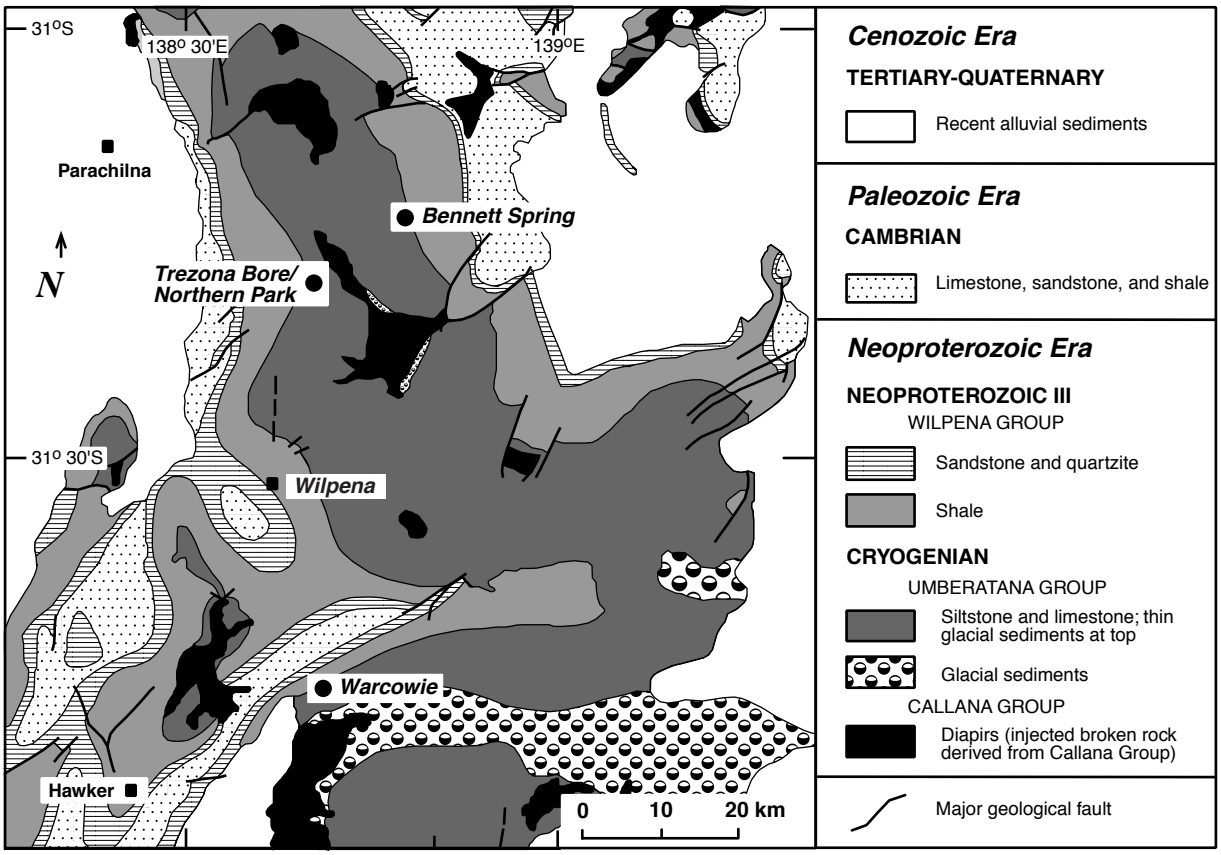

Figure 3. Map of the central Flinders Ranges showing the locations of sections sampled for this study (after Selby, 1990).

tures necessary (>600 °C) for the formation of exsolution lamellae. Appropriate potential sources for these grains can be found in the Mesoproterozoic volcanic and metasedimentary rocks of the Gawler craton, west of the central Flinders Ranges (see Fig. 1; Drexel et al., 1993). Embleton and Williams (1986) reported that many Fe-Ti oxide grains in the Elatina Formation around Pichi Richi Pass (Fig. 1A) had been martitized, i.e., hematite had replaced magnetite. Many of these titano-magnetite grains were likely oxidized to titano-hematite prior to deposition, and thus the C component magnetization carried by the Fe-Ti oxides could be primarily a detrital or early post-depositional remanent magnetization.

In addition to the detrital Fe-Ti oxide grains, there is abundant hematite in the form of submicron-sized crystals that coat the larger grains. The origins of this ultrafine hematite are not certain; it could be the product of diagenetic alteration of ferric oxyhydroxides originally deposited within the glacial sediment. A diagenetic origin for the hematitic coating would mean that any magnetic remanence carried by the coating is a chemical remanent magnetization (CRM) younger than the remanence carried by the detrital Fe-Ti oxide grains. The strength of the magnetizations in the samples examined (~10^{-2}–10^{-3} A/m; see Fig. 4 for examples) suggests that the detrital grains, rather than the hematitic coating, carry most of the remanent magnetization present. However, a CRM carried by the coating may be the source of the overprint (i.e., the B component) observed in many of the samples.

Stability Tests

To help determine whether the site mean C component data from the central Flinders Ranges represent a primary magnetic remanence (i.e., a detrital or an early diagenetic remanence) or some later magnetization, we applied geological or field stability tests to the central Flinders data to link the magnetizations to the geologic history of the rocks. Four tests commonly used for sedimentary rocks are the fold test, the reversals test, the conglomerate test, and the unconformity test (see review in Butler, 1992).

Two of the tests we applied, the conglomerate and unconformity tests, yielded indeterminate results. The diamictites in the Trezona Bore section of the Elatina Formation were targeted for a conglomerate test, because the clasts within the diamictite are composed of diverse lithologic types and are large enough to sample with a rock drill. A conglomerate test is positive if the stable directions of individual clasts have a random distribution, indicating that the entire unit (and perhaps other stratigraphic units in the region) has not been remagnetized. Unfortunately, the clasts tested yielded unstable magnetizations that could not be compared with those of the matrix.

The contact between the Yaltipena and Elatina Formations at the Trezona Bore–Northern Park sections was selected for an unconformity test (Kirschvink, 1978a), since there are several meters of relief on the unconformable surface between the two units. Had we been able to demonstrate that the polarity stratigraphy of the Yaltipena beneath the unconformity was different from laterally adjacent but stratigraphically superjacent polarity stratigraphy of the Elatina Formation above the unconformity, the magnetizations of both the Yaltipena and Elatina rocks could have been considered primary. Unfortunately, the crucial intervals of both formations yielded unstable magnetizations, and so could not be used for the test.

Other tests proved to be more revealing. Definitive inferences regarding the reliability of magnetizations can be made from the results of

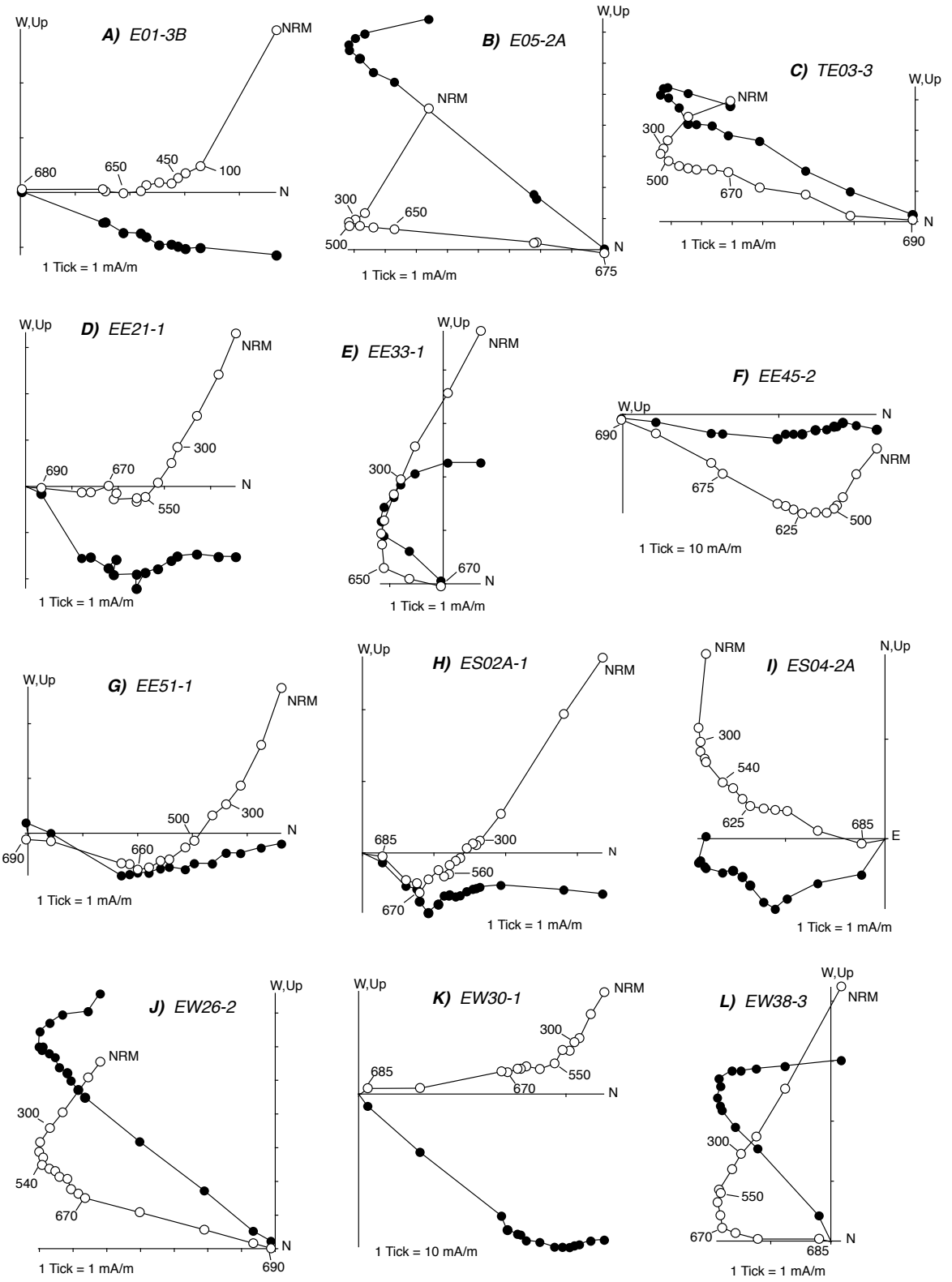

Figure 4. Representative vector component diagrams (Zijderveld plots) showing progressive thermal demagnetization (in bedding tilt–corrected coordinates) of selected samples from the Yaltipena and Elatina Formations. These are among the samples that best illustrate the low-temperature A component and high-temperature C component. Yaltipena Formation: (A, B) Northern Park section; (C) Bennett Spring section. Elatina Formation: (D, G) Bennett Spring section; (H, I) Warcowie section; (J, L) Trezona Bore section. Axes scale indicates measured intensity of the sample's characteristic remanent magnetization. Numbers on the plots indicate particular temperature steps. Open circles and filled circles represent the vertical and horizontal components, respectively, of the magnetic vector for each temperature step given. NRM—natural remanent magnetization.

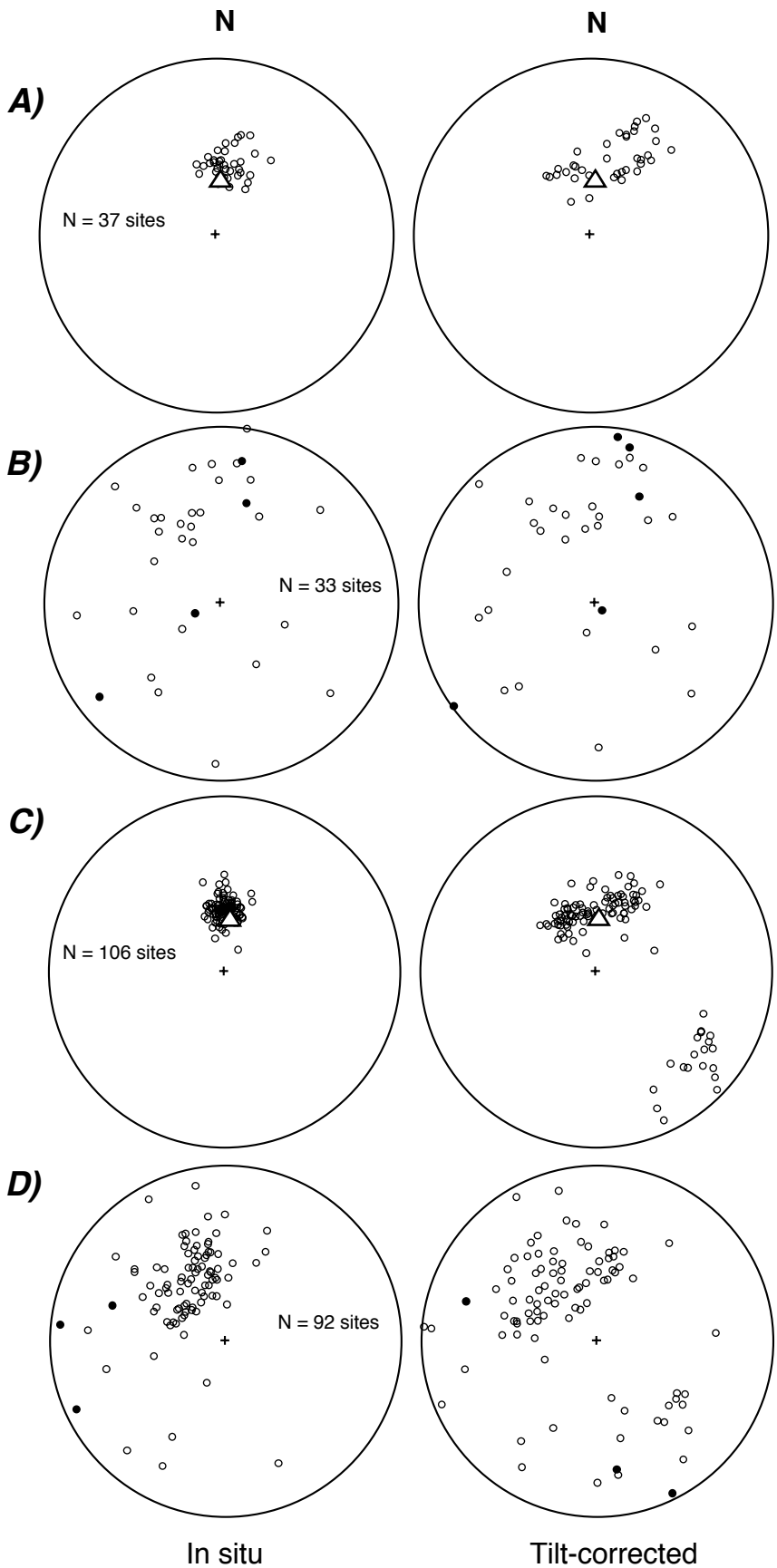

Figure 5. Equal-area projections of A and B component site mean directions in both in situ and bedding tilt-corrected coordinates. Yaltipena Formation, all sites: (A) A component and (B) B component. Elatina Formation, all sites: (C) A component and (D) B component. Open circles and filled circles represent negative and positive inclinations, respectively. The orientation of the present-day magnetic field in the central Flinders Ranges with respect to the A component directions is indicated by a white triangle in A and C.

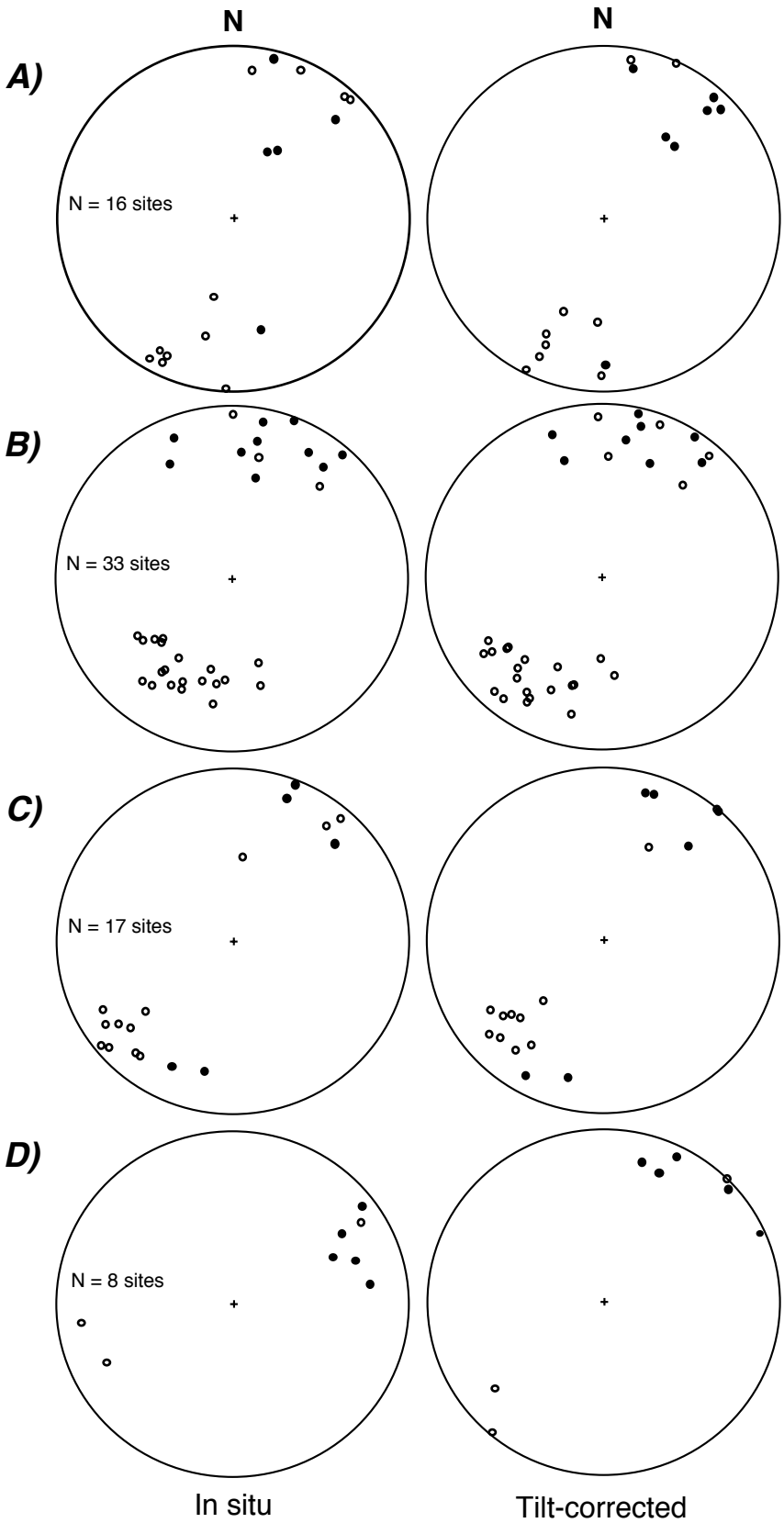

Figure 6. Equal-area projections of C component site mean directions in both in situ and bedding tilt-corrected coordinates. Yaltipena Formation: (A) Northern Park and Bennett Spring sections combined. Elatina Formation: (B) Bennett Spring section; (C) Trezona Bore section; and (D) Warcowie section. Open circles and filled circles represent negative and positive inclinations, respectively.

TABLE 1. PALEOMAGNETIC DATA FROM THE CENTRAL FLINDERS RANGES

Formation/ section	N	In situ coordinates				Tilt-corrected coordinates								
		Dm (°E)	Im (°N)	k_1	α_{95} (°)	Dm (°E)	Im (°N)	k_2	α_{95} (°)	λ (°N)	ϕ (°E)	dp (°)	dm (°)	Plat (°)
Yaltipena Formation														
Bennett Spring (YBS)	5	209.8	–38.7	13.8	21.4	215.5	–27.4	14.2	21.0	–33.0	0.9	12.5	22.9	–14.5
Northern Park (YNP)	11	200.6	0.2	12.1	13.7	199.2	–11.2	14.2	12.6	–48.8	348.4	6.5	12.8	–5.7
Yaltipena Formation total	16	203.0	–11.9	8.0	13.9	204.0	–16.4	12.2	11.0	–44.2	352.7	5.9	11.4	–8.4
Elatina Formation														
Bennett Spring (EBS)	33	201.8	–28.1	9.0	8.8	206.9	–19.7	9.2	8.7	–41.2	355.1	4.8	9.1	–10.2
Trezona Bore (ETB)	17	219.9	–4.9	9.8	12.0	218.0	–14.7	10.5	11.6	–36.9	8.4	6.1	11.9	–7.5
Warcowie (EW)	8	246.3	–18.2	18.4	13.3	219.5	–9.6	19.0	13.0	–37.5	11.8	6.7	13.2	–4.8
Elatina Fm total (sites)	58	213.9	–20.6	7.1	7.6	212.1	–16.9	9.9	6.2	–39.7	1.9	3.3	6.4	–8.6
Elatina Fm total (sections)	3	223.1	–17.9	11.5	38.1	214.9	–14.7	94.9	12.7	–38.9	5.6	6.7	13.0	–7.5

Notes: N—number of sites or sections; Dm and Im—mean declination, inclination of N sites or sections; k—the best estimate of Fisher's precision parameter; α_{95}—radius of 95% circle of confidence for the mean direction; λ and ϕ—latitude and longitude of paleopole for mean direction; dp and dm—semi-minor and semi-major axes of 95% polar error ellipse; Plat—paleolatitude.

the fold test. The reversals test shows the necessity of averaging over both normal and reversed polarities to suppress bias of residual overprint contamination. The presence of distinct polarity intervals, combined with the magnetic mineralogy, provides further supportive evidence of early magnetization acquisition. These analyses are described in the following.

Fold Test

Graham's (1949) fold test can determine whether magnetization was acquired before or after folding by comparing the clustering of site mean directions uncorrected for bedding tilt (geographic, or in situ directions) to clustering of site mean directions with bedding restored to horizontal (McElhinny, 1964; McFadden, 1990; McFadden and Jones, 1981). A significant improvement of clustering upon tilt correction is evidence that the magnetization was acquired prior to folding, and the directions are said to have passed the fold test. If the clustering worsens upon tilt correction, the magnetization was acquired after folding (the directions fail the fold test). On occasion, the clustering of site means reaches a maximum at some point <100% unfolding; such a result might indicate magnetization acquired during folding (e.g., Sumner et al., 1987; Schmidt et al., 1991; but see Halim et al., 1996).

When the fold test is applied to the A component site mean directions for both the Yaltipena and Elatina Formations, the clustering of site mean directions clearly worsens upon tilt correction (Fig. 5, A and C). To confirm these observations, we applied the McFadden and Jones (1981) fold test to the Yaltipena directions, and the more conservative McElhinny (1964) fold test to the Elatina directions (where the large number of sites makes application of the McFadden and Jones (1981) test impractical). In both cases, the observed test values for the in situ directions exceed the critical test values at the 99% level; thus both the Yaltipena and Elatina Formation A components are postfolding in origin, consistent with the Cenozoic age inferred from its mean direction. The same fold test applied to the B component for both formations produces similar results, i.e., the B component is also postfolding in origin. The diffuse distribution of B component directions (Fig. 5, B and D) makes it difficult to identify the precise age of the component, but the northwesterly directions and moderate inclinations most closely resemble the Early Jurassic paleomagnetic data of Australia (e.g., Schmidt, 1976, 1990).

We then applied the fold test to the C component directions for the two units. We compared the two Yaltipena sections to each other, and each of the three Elatina sections to one another, using the in situ and tilt-corrected mean directions calculated for each section from the individual C component site means. For the Elatina Formation, mean directions were calculated from the section means as well (Table 1). In all cases, the clustering of data improved when bedding was restored to horizontal (Figs. 6A and 8). The improvement in clustering of the Elatina section mean directions (Fig. 8B) is obvious, and is significant at the 95% confidence limit according to the conservative test of McElhinny (1964) (i.e., $k_2/k_1 = 8.25$, exceeding the critical test ratio of 6.39). Using the more sensitive fold test of McFadden and Jones (1981), the improvement in clustering of site means for both the Yaltipena and the Elatina is significant at the 99% level (Table 2). An additional examination of clustering at various levels of unfolding of the Yaltipena and Elatina Formations does not indicate the presence of synfolding magnetization (Fig. 9). Thus both the Yaltipena and Elatina Formations cleanly pass the fold test. However, such a prefolding remanence could have been acquired at any point during the lengthy interval between the Marinoan glaciation and the Delamerian orogeny (~50–90 m.y.); there is also a question of magnetic overprint acquisition (see next section). Further evidence for the timing of remanence acquisition is required.

Reversals Test

The presence of dual polarities (i.e., reversals) within the site mean directions of the Yaltipena Formation, and particularly within the Elatina Formation, is an encouraging indicator for the preservation of a primary magnetization (Fig. 6). Van der Voo (1990) used the presence of reversals as one of the criteria for evaluating the reliability of paleomagnetic data. We performed a reversals test to determine whether the mean direction of one polarity in a section is statistically indistinguishable, at the 5% level of significance, from the antipode of the opposite polarity; if the two directions are indistinguishable, the section passes the reversal test (McFadden and Lowes, 1981; McFadden and McElhinny, 1990).

For the Yaltipena and Elatina Formations, the means of the north-northeast cluster of directions were compared with the means of the southwest cluster of directions. In all cases, the north-northeast cluster means are statistically distinguishable from the antipodes of the southwest cluster means; accordingly, both formations fail the reversals test (Table 3). This failure may be due to either a wide separation in age between the north-northeast and southwest directions, or to some magnetic overprint that was not removed by the thermal demagnetization procedures outlined here. A fold test applied to the north-northeast and southwest site mean directions separately for both the Yaltipena and Elatina Formations suggests that both polarities predate Delamerian folding (Table 2). Since much of the magnetization is apparently carried by detrital Fe-Ti oxide grains, the north-northeast and southwest directions are likely close in age to each other as well as the time of deposition.

However, a CRM overprint carried by the hematitic coating—perhaps partly represented by the ill-defined B component noted here—would be difficult to separate completely from the remanence carried by the detrital grains because their demagnetization spectra overlap signifi-

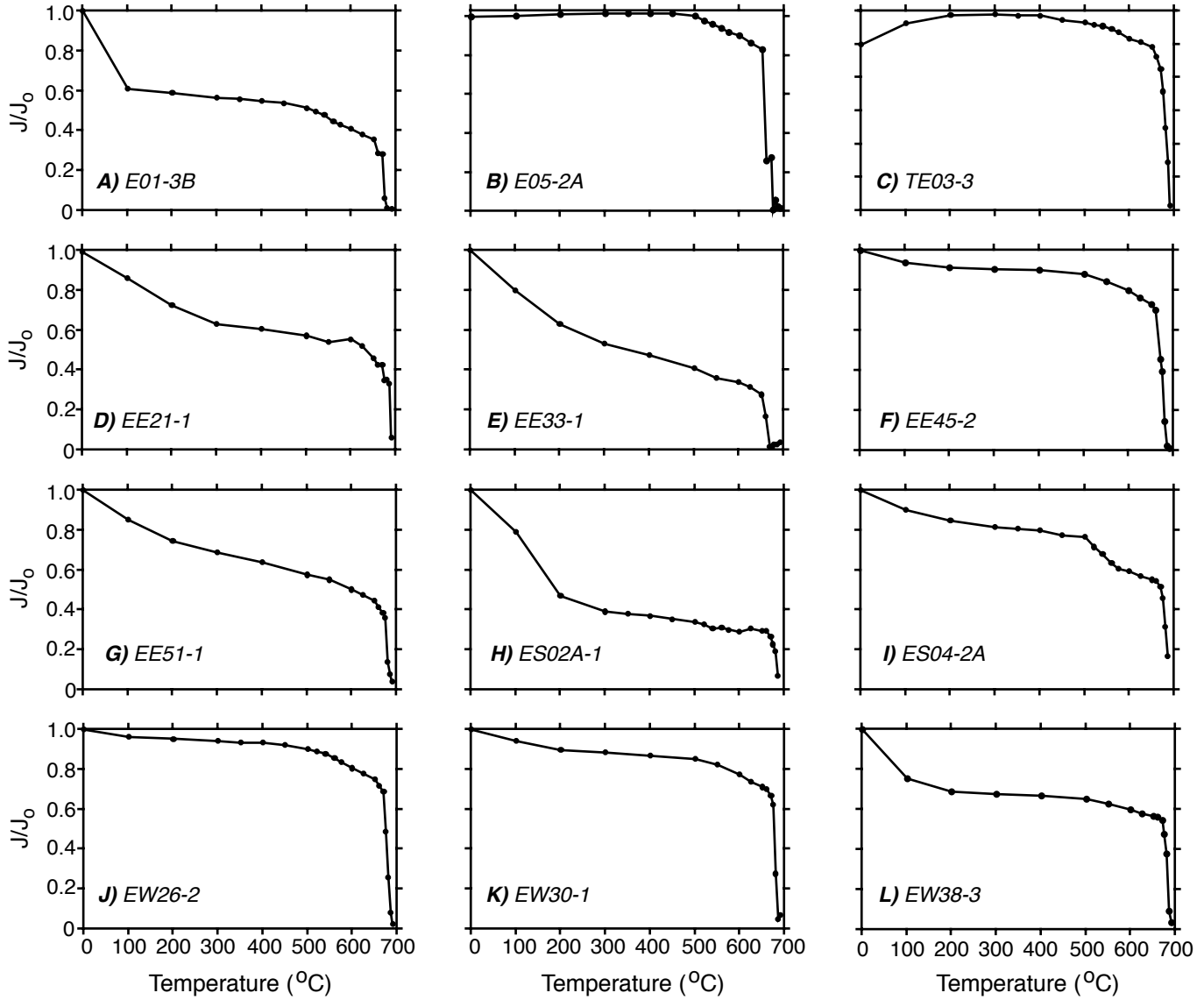

Figure 7. Plots of the change in magnetic intensity J/J_0 with progressive thermal treatments for selected samples. Yaltipena Formation: (A, B) Northern Park section; (C) Bennett Spring section. Elatina Formation: (D, G) Bennett Spring section; (H, I) Warcowie section; (J, L) Trezona Bore section. Note that in most cases, a significant portion of the sample's measured intensity is retained until a temperature of 660–680 °C is reached, indicating that hematite carries the high-temperature C component. Sharp decreases in J/J_0 from 0°–100° or 200 °C (e.g., samples A, H, and L) indicate the presence of goethite; the abrupt dip in J/J_0 at ~560 °C in sample I likely indicates the presence of magnetite.

cantly. It is quite probable that the paleomagnetic results of previous studies have been similarly affected by such a CRM overprint. The fewer data in these studies, combined with unequal numbers of normal and reversed polarity sites, may have prevented the clear identification of the overprint (see following section on comparison with other Neoproterozoic paleomagnetic results).

Magnetostratigraphy

The chief difference between this study and previous studies of Marinoan age rocks becomes apparent when the mean site data are plotted (in modified form) in their stratigraphic context: it is clear that distinct polarity magnetozones have been recorded in both the Yaltipena and Elatina Formations (Fig. 10). We make tentative interpretations of polarity here based upon previous paleomagnetic work that covers the post-Marinoan through Cambrian interval in the Amadeus basin (Kirschvink, 1978b), as well as previously published apparent polar wander (APW) paths for east Gondwana (e.g., Powell et al., 1993). According to these interpretations, the Earth's magnetic field in late Neoproterozoic time corresponds to reversed polarity for southwest and up directions and normal polarity for north-northeast and down directions; Australia thus is right-side-up near the equator, but mostly in the Northern Hemisphere (see Table 1 and Fig. 11). Stratigraphic intervals labeled as being of uncertain polarity are so marked for either of two reasons: (1) the section was not sampled owing to poor exposure and/or unpromising lithology (e.g., white sandstone); or (2) samples were taken that did not yield reliable paleomagnetic data according to the criteria established for this study. In counting the number of reversals per section, we assume no

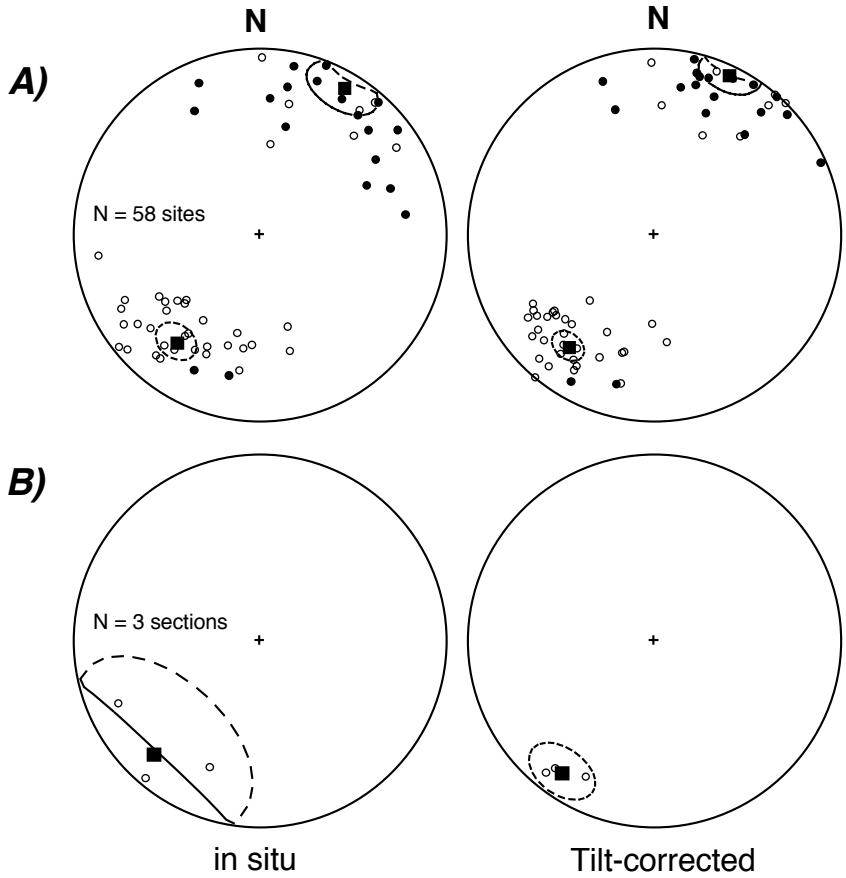

Figure 8. Comparison of the Elatina's in situ and tilt-corrected (A) site mean directions and (B) section mean directions for the fold test. Mean directions for north-northeast and southwest clusters in (A), and overall mean directions in (B), are indicated by the solid black squares with their corresponding α_{95} confidence limits. Improvement in the clustering of directions upon tilt correction is most obvious for the section means, although it is also evident in the site mean directions through the decreasing diameter of α_{95} confidence limits for north-northeast and southwest tilt-corrected cluster means.

TABLE 2. FOLD TEST RESULTS FOR C COMPONENT, CENTRAL FLINDERS RANGES*

Formation	N	Critical test value at 1% level	Observed test value (in situ)	Observed test value (tilt-corrected)	Passes fold test?
Yaltipena Formation					
All site mean directions	16	0.3893	0.3959	0.2441	Yes
North-northeast site mean directions	8	1.1548	1.9152	0.7520	Yes
Southwest site mean directions	8	1.1548	0.2434	0.0673	†
Elatina Formation					
All site mean directions	58	3.5082	12.0464	1.8081	Yes
North-northeast site mean directions	25	3.7922	6.7320	1.6027	Yes
Southwest site mean directions	33	3.6490	12.0045	1.5885	Yes

Notes: N—number of sites.

*Fold test calculated as per McFadden and Jones (1981). The fold test is passed if the observed test value (in situ) exceeds the critical test value, while the observed test value (tilt-corrected) does not. The fold test is failed if the opposite holds true.

†When neither the observed in situ or tilt-corrected test values exceed the critical test value, the results of the fold test are indeterminate. Such a result indicates that the structural distortion of the beds was too small to yield a significance with the available sample (McFadden and Jones, 1981).

polarity changes within uncertain intervals. Note that several polarity intervals, particularly those near the bases of the Yaltipena Formation Northern Park section and the Elatina Formation Trezona Bore section, are tentatively identified here by single data points only.

Unfortunately, an obvious feature of the polarity intervals in Figure 10 is the lack of consistency in the pattern of polarity changes from section to section. For example, the Bennett Spring section of the Yaltipena Formation shows only two polarity intervals over a thicker interval than the Northern Park section, which shows seven possible polarity intervals. The discrepancy in magnetostratigraphy between these two sections may be partly an artifact of the wider spacing of sampling sites, since sites were located 15–20 m apart in the lower two-thirds of the section. An additional obstacle is the limited regional preservation of the Yaltipena Formation within salt-withdrawal synclines (Lemon, 1988); precise physical stratigraphic correlation between the two sections is not possible, and there are few constraints on how the polarity intervals can be matched. The relatively small numbers of paleomagnetically reliable data points defining the polarity intervals, particularly within the Bennett Spring section, also cloud the issue.

A similar correlation problem is apparent within the Elatina Formation, as the Trezona Bore section displays seven polarity intervals compared to five at Bennett Spring and only two at Warcowie, over stratigraphic intervals of roughly comparable thickness. In the Elatina Formation, however, the fundamentally discontinuous nature of glacial strata is most likely responsible for much of our inability to correlate magnetostratigraphy between sections. Glaciomarine deposits like the Elatina are typically highly variable, both along depositional strike and crossing from shallow-water to deeper water environments. Such variability is largely a function of their proximity to sediment sources, such as subglacial tunnel mouths in grounded ice margins (e.g., Eyles et al., 1985; Menzies, 1995). This variability is clearly illustrated in the Elatina Formation lithologic columns (Fig. 10B). For example, the Trezona Bore section (the closest section to an inferred grounded-ice margin) has several diamictite units and disconformities and is overall coarser grained in comparison to the Bennett Spring section, which has only one distinct diamictite layer and only one apparent disconformity. The Warcowie section, which is farthest from the inferred grounded ice margin and close to the transition to deep-water glacial sediments of the southern Flinders Ranges, includes only a very thin, discontinuous diamictite bed at its base. It is important to note, however, that our inability to make magnetostratigraphic correlations

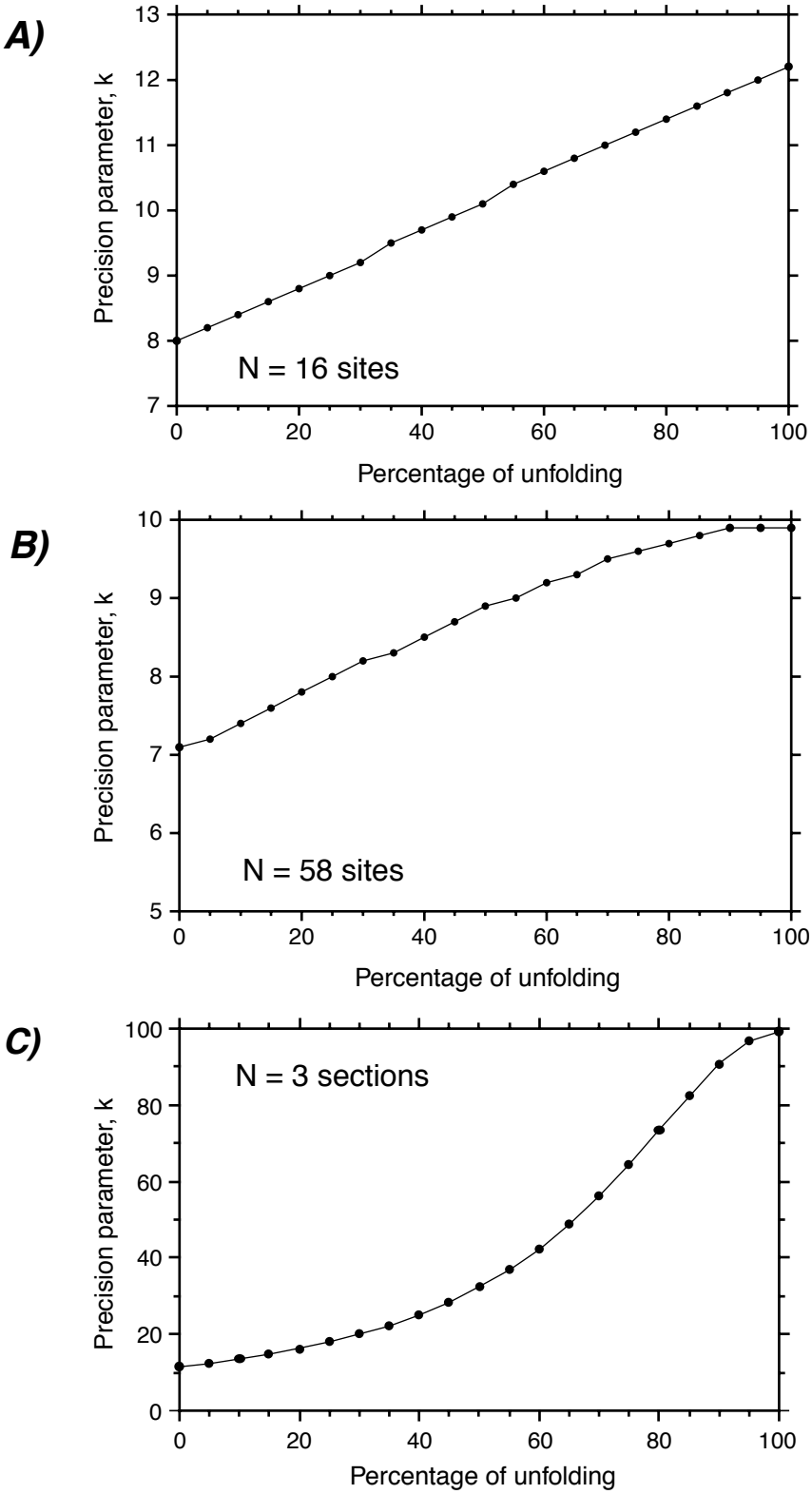

Figure 9. Test for the presence of synfolding magnetization. The precision parameter k for an overall mean direction would reach a peak value at somewhat less than 100% bedding tilt correction (i.e., unfolding to a horizontal position) if magnetization were acquired during folding. The value of k has been calculated at 5% increments of unfolding. (A) Yaltipena Formation (mean direction calculated from site means). (B) Elatina Formation (mean direction calculated from site means). (C) Elatina Formation (mean direction calculated from section means). The steady increase in k in all three plots suggests that magnetization was not acquired during folding.

TABLE 3. ANGULAR SEPARATION BETWEEN NORTH-NORTHEAST AND SOUTHWEST MEAN DIRECTIONS

Formation/section	Angle before tilt correction (°)	Angle after tilt correction (°)
Yaltipena Formation - YBS	104	125
Yaltipena Fm - YNP	165	154
Yaltipena Formation total	168	165
Elatina Formation - EBS	129	142
Elatina Formation - ETB	158	153
Elatina Formation - EW	144	159
Elatina Formation - total	159	156

Note: Abbreviations as in Table 1.

between sections does not automatically invalidate our interpretation of the magnetic polarity intervals as primary, or the Elatina Formation as a low-latitude deposit; it simply means that the precise depositional relationship between sections remains unresolved.

Comparison with Other Neoproterozoic Paleomagnetic Results

The combination of multiple polarity reversals, a positive regional-scale fold test, and detrital or early postdepositional magnetic remanence revealed in this study contributes to a very robust paleomagnetic survey of the Marinoan age glacial deposits. To compare previous paleomagnetic studies more easily to this one, we calculated time-averaged paleopoles for the Yaltipena and Elatina Formations from the section mean directions (Table 1). These poles are then plotted along with other late Neoproterozoic–Cambrian paleopoles published for Australia (Fig. 12; see also Table DR1 [see footnote 1]). The overall similarities between the Yaltipena (YA = Yaltipena average) and Elatina (EA = Elatina average) poles and several other late Neoproterozoic poles (LA, BR1, E1, and E2) are clear; their clustering agrees well with the presumed close age relationships of their corresponding rock units.

There is a small but significant difference (16.3°) between the previously published average paleopole for the Elatina Formation (Schmidt and Williams, 1995) and the average Elatina paleopole in this study, but it can be readily explained by overprint polarity bias. At both the site mean and individual sample levels, Schmidt and Williams' (1995) data contain a ratio of north-northeast:southwest directions ("normal:reversed" polarity) of ~2:1 (see also Table DR2 [see footnote 1]). Thus, the mean direction (and corresponding paleopole) derived either from their sample or site mean data has been somewhat biased by an overprint caused by the abundance of north-northeast directions. The presence of this bias can be confirmed by comparing the antipode of the mean north-northeast direction for the site means in this study (Dm = 202.9°, Im = –4.9°, α_{95} = 10.0) to the overall Schmidt and Williams (1995) mean direction for their site mean data (Dm = 197.4°, Im = –7.1°, α_{95} = 15.2). They are only 5.9° apart and statistically indistinguishable. In contrast, the overall mean direction for the site mean data of this study (Dm = 212.1°, Im = –16.9°) has been calculated from a subequal number of north-northeast and southwest directions (ratio of ~1:1.3); as such, we argue that this overall mean direction, and its corresponding paleopole, should better average out the overprint and is less biased than Schmidt and Williams' previously published result.

Paleolatitude of Marinoan Glacial Deposits

The paleolatitudes of the Yaltipena and Elatina Formations can be determined from the mean tilt-corrected inclinations of each unit (see Table 1), as inclination, I, and latitude, λ, can be related through this simple equation, assuming a geocentric axial dipole:

$$\tan I = 2 \tan \lambda. \quad (1)$$

Error limits based upon the α_{95} values for the mean inclinations can then be assigned to the paleolatitude estimates as follows:

$$\lambda_{maximum} = \tan^{-1} \{0.5[\tan (I + \alpha_{95})]\} \quad (2)$$

$$\lambda_{minimum} = \tan^{-1} \{0.5[\tan (I - \alpha_{95})]\} \quad (3)$$

Thus, the paleolatitude estimate for the Yaltipena Formation (N = 16 sites) is –8.4°, with error limits of –2.7° and –14.5°. For the Elatina Formation (N = 58 sites), the paleolatitude estimate is –8.6°, with error limits of –5.4° and –12.0°; averaged over three sections instead, the Elatina's paleolatitude estimate is –7.5°, with error limits of –1.0° and –14.5°. Based upon polarity interpretations previously published for slightly younger data in the Amadeus basin, we interpret these values as representative of reversed field conditions, so that Australia lies within the Northern Hemisphere.

The paleomagnetic results from the Elatina Formation provide ample evidence in favor of the low-latitude glaciation hypothesis for Australia ca. 600 Ma. In addition, the low paleolatitude for the Yaltipena Formation is consistent with the warm, arid setting inferred for these rocks and implies that Australia occupied a near-equatorial position just prior to glaciation as well, making the "drift hypothesis" rather unlikely. Although we were unable to use our own data from the Amadeus basin and Stuart shelf to verify independently the low paleolatitudinal position of the Marinoan glacial deposits, we note that a calculated average paleolatitude of ~8° for the central Flinders Ranges during the Marinoan glaciation is consistent with a paleolatitude of ~18° (from pole LA) for the younger Lower Arumbera Sandstone of the Amadeus basin (Kirschvink, 1978a) and a paleolatitude of ~3° (from pole BR1) for the Brachina Formation of the Mount Lofty Ranges (Karner, 1974). These secondary constraints also serve to support the low-latitude glaciation hypothesis for the Marinoan glaciation in Australia (Fig. 11).

A fundamental assumption we make for this study is that the Earth's magnetic field behaved in Neoproterozoic time much as it has in the recent past, i.e., that the time-averaged paleofield can be well approximated by a dipole located at the center of the Earth and aligned with the Earth's geographic (rotational) axis (the geocentric axial dipole hypothesis, or GAD). However, some researchers have suggested that the Earth's magnetic field may have behaved differently in the distant past. One of us (Kent) has collaborated on a statistical study of magnetic inclinations in Earth history that found a greater number of low magnetic inclination data from Precambrian and Paleozoic rocks than would be expected if the Earth's magnetic field were behaving dominantly as a GAD (Kent and Smethurst, 1998). That study has suggested that either (1) the magnetic field contained a much larger non-dipole component than it did in Mesozoic and Cenozoic time (~30% compared to <10%) that exerted a "flattening" influence on magnetic inclinations, or (2) the GAD was valid but (super)continents had a tendency to cycle into low latitudinal positions owing to true polar wander. We are unable to exclude the first hypothesis in relation to this study, but we prefer and assume a GAD field for the Marinoan pending additional paleomagnetic data from other continents that could further constrain the magnetic field's behavior.

In summary, the paleomagnetic data from the Yaltipena and Elatina Formations support a low paleolatitude for Australia both prior to and during the Marinoan glaciation. The new Elatina pa-

leopole in this study is robust, and can be used with confidence to constrain the APW path for Australia in late Neoproterozoic time. We now consider what the new paleomagnetic data, combined with stratigraphic information about the Elatina Formation and other Marinoan glacial deposits, imply about the duration of low-latitude glaciation in Australia.

DURATION OF LOW-LATITUDE GLACIATION IN AUSTRALIA

One of the continuing vexations in Neoproterozoic geology is the dearth of reliable radiometric dates to constrain the timing of geologic events. The age of the Marinoan glaciation is, not surprisingly, poorly constrained; the age limits of 610 and 575 Ma are based upon the presumed relationship of the Marinoan glaciation to Varanger glacial intervals elsewhere in the world (e.g., Hambrey and Harland, 1981; Knoll and Walter, 1992), and estimated sedimentation rates of strata between the top of the glacial rocks and the Precambrian-Cambrian boundary, respectively (Conway Morris, 1988; Grotzinger et al., 1995). The duration of glaciation within those age limits has been correspondingly difficult to quantify, particularly where stratigraphic discontinuities have interrupted the sedimentary record. The presence of multiple magnetic polarity intervals in the Elatina Formation permits us, for the first time, to make some semiquantitative estimates of the duration of low-latitude glaciation in Australia.

We assume that the Earth's magnetic field behaved then much as it has in the recent past: (1) the time-averaged magnetic paleofield can be well approximated by the GAD hypothesis (allowing latitude to be calculated from inclination using the dipole formula; see equation 1); (2) the sequence of polarity reversals is stochastic to first order, varying from several per million years to one per several million years, to long intervals of several million to several tens of millions of years with no reversals (Opdyke and Channell, 1996); (3) the Earth's magnetic field typically takes on the order of several thousand years to complete the transition from one polarity state to another (e.g., Clement and Kent, 1984); and (4) once the reversal is complete, the paleomagnetic field tends to be inhibited from reversing again for a minimum of 30–50 k.y. (McFadden and Merrill, 1984, 1993).

The polarity stratigraphy for the Elatina Formation, as shown in Figure 10B, suggests that there were a minimum of four polarity reversals during its period of deposition, and possibly several more (if polarity intervals currently defined by only one data point are included and/or the time spans represented by each stratigraphic section do not completely overlap). The Elatina Formation should thus represent a minimum

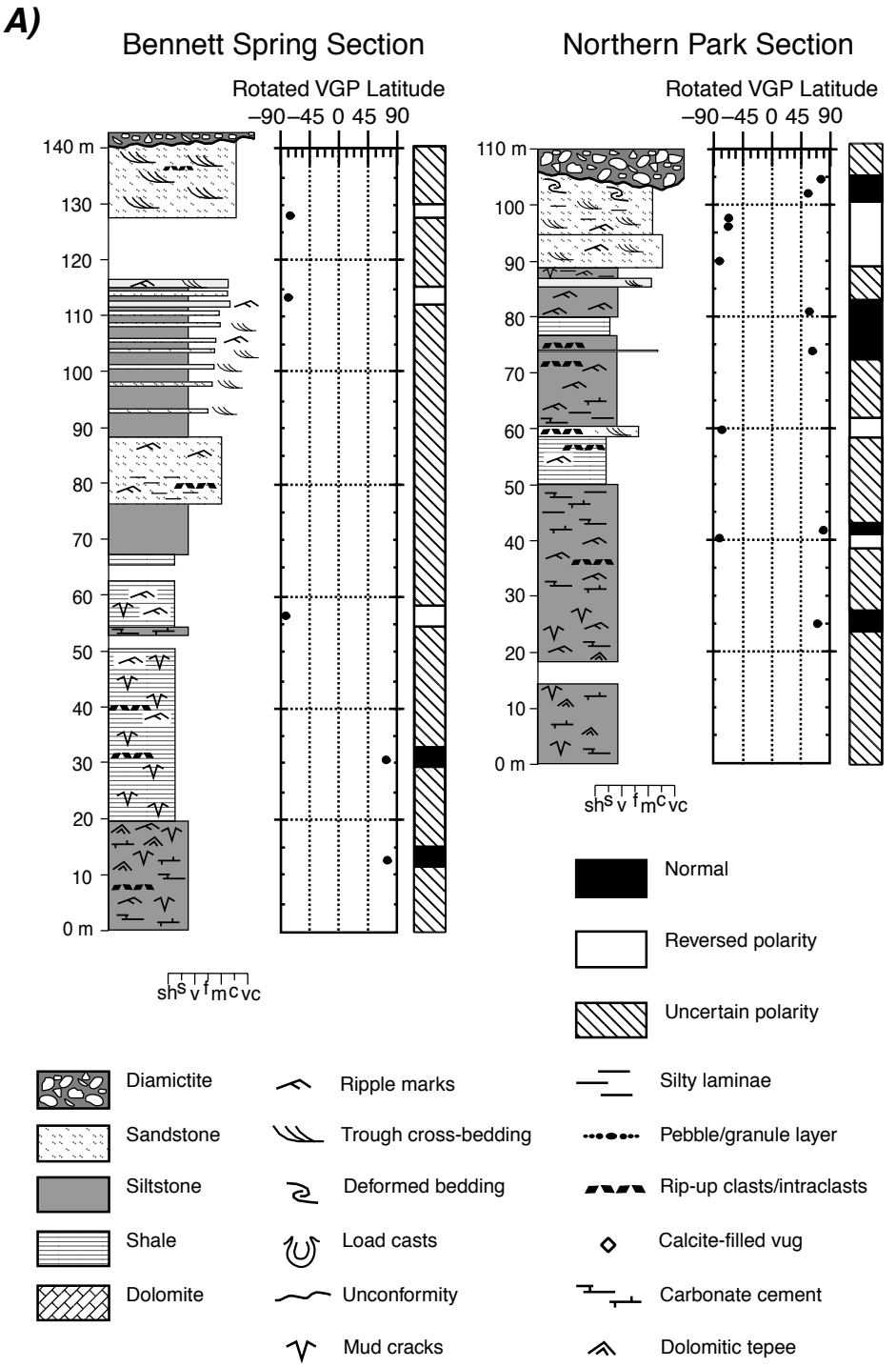

Figure 10. Stratigraphic sections with paleomagnetic data and polarity interpretations for (A) the Yaltipena Formation and (B) the Elatina Formation. The site mean directions are shown here as rotated virtual geomagnetic pole (VGP) latitudes to simplify the comparison between individual site mean directions and the time-averaged paleopole for each section. A VGP is calculated for each site mean, and then both the VGPs and the mean paleopole are rotated until the mean paleopole is at 90° latitude; the new latitude of the VGP is the rotated VGP latitude. The plots show the latitudinal distance between the VGP latitudes and the paleopole; VGP latitudes more than 45° away from the mean paleopole would have been considered unrelated to the paleopole, and would not have been included in this analysis.

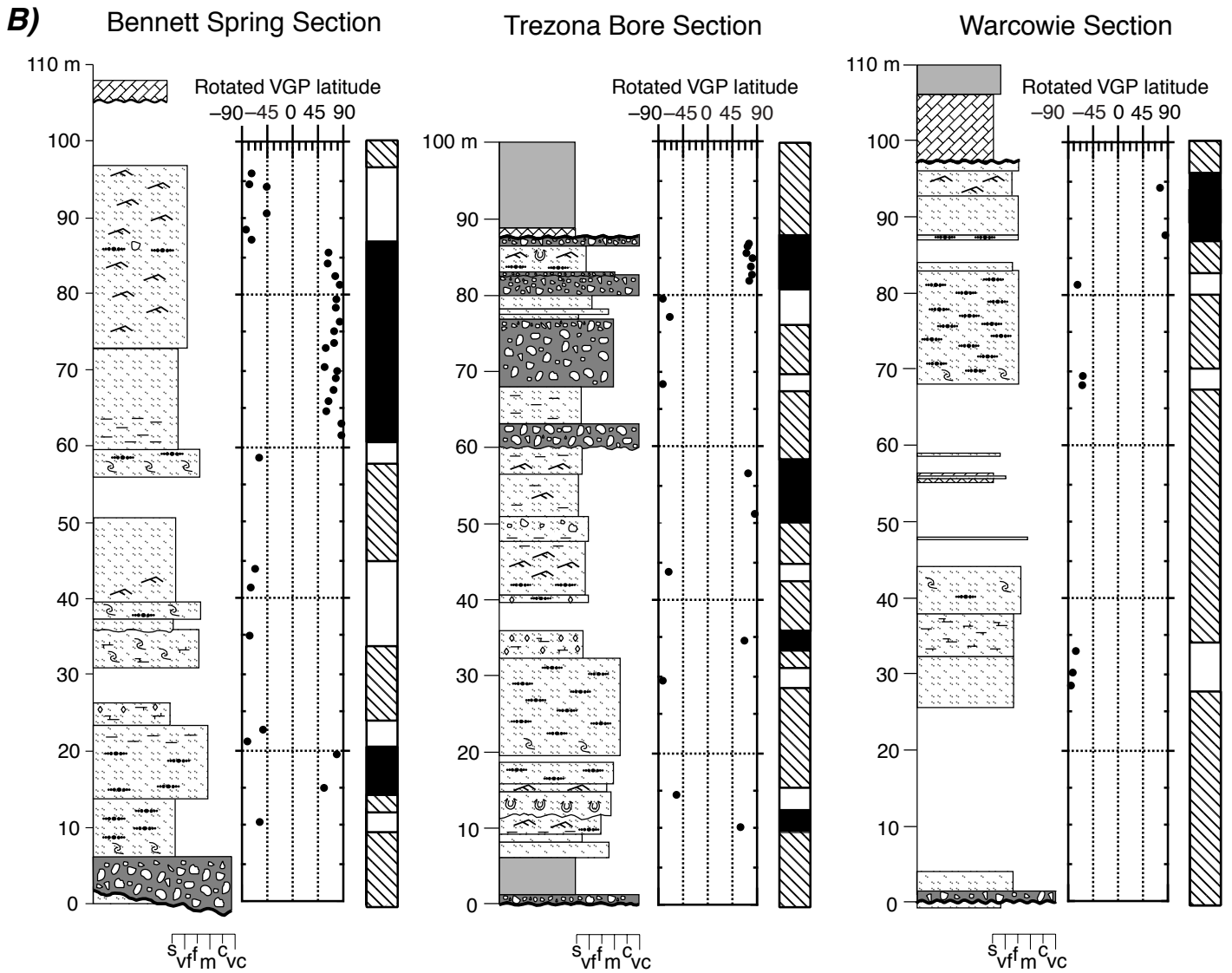

Figure 10. (*Continued*).

span of several tens of thousands of years on the basis of assumed reversal duration, and hundreds of thousands of years if inhibition was generally operative. The unit could represent as much as several million years, depending on the (unknown) rate of reversals in Neoproterozoic time.

In estimating the duration of the Marinoan glaciation, it is important also to consider the stratigraphic relationship of the Elatina Formation to other Marinoan deposits in South Australia. The thinness of the Elatina Formation (generally <100 m) has been used to suggest that it represents the waning stages of glaciation in the region (Lemon and Gostin, 1990). The Elatina Formation contains a number of stratigraphic discontinuities of indeterminate duration, and probably represents multiple advances and retreats of the local ice front. In subbasins to the north and south of the central Flinders Ranges, the Marinoan glacial deposits are considerably thicker, as much as ~1000 m to the north and more than ~1500 m to the south (Preiss, 1987). These subbasins contain thick successions of dropstone-bearing siltstones containing dropstones that presumably accumulated over a considerable amount of time, and stratigraphic relationships suggest that these basinal deposits predate the Elatina Formation in large part (Preiss, 1987). Between the polarity data and the stratigraphic data, it is plain that glaciation in Australia may have persisted for a few million years, and perhaps considerably longer.

As the sole glacial deposits confirmed to have been deposited at low paleolatitudes, the Elatina Formation and related Marinoan units represent the nadir of the late Neoproterozoic cooling trend; they may also be the youngest glacial deposits of the Varanger glacial interval. The potentially lengthy time span encompassed by these deposits suggests that the global cooling trend leading first to the development of ice sheets at higher latitudes was an interval of long duration, perhaps similar in time scale (if not severity) to the Cenozoic cooling trend (e.g., Frakes, 1986; Miller et al., 1987; Ruddiman, 1997). If so, then climate forcings that operate on ~10^6 yr time scales, such as changes in ocean circulation induced by shifting continental masses and/or changes in sea-floor topography, would have been important in setting the stage for severe cooling in equatorial regions. However, the apparent swift collapse of glaciation inferred from models for cap carbonate deposition (e.g., Kennedy, 1996; Hoffman et al., 1998) suggests

that shorter-term "trigger" events also played a role in destabilizing the climatic system. Further work is required to identify which climate forcing mechanisms operating at a variety of time scales are the key controls governing Earth's Proterozoic climate.

CONCLUSIONS

A paleomagnetic investigation of the Elatina Formation of the central Flinders Ranges, South Australia, has yielded a statistically significant, positive regional-scale fold test, as well as revealed the presence of multiple polarity intervals. These findings confirm that the Elatina's magnetization predates tectonic folding during Cambrian time, thus providing the strongest evidence to date in favor of low-latitude glaciation of Australia in late Neoproterozoic time. In fact, with the Elatina Formation located at ~8°, all of Australia's Marinoan glacial deposits must be below 30°. Meert and Van der Voo's (1994) review of the Neoproterozoic paleomagnetic results concerning the low-latitude glaciation hypothesis suggested that the paleomagnetic data from the Elatina Formation, if valid, did not necessarily imply either global glaciation or a non-uniformitarian climate; they cited the proposition (Eyles, 1993) that Marinoan glacial deposits accumulated in the elevated regions of an active tectonic setting, so that the deposits could be low latitude but not low elevation. However, the glaciomarine nature of the Elatina Formation and its correlative basinal strata, plus the fact that its depositional setting had been a passive margin for ~100 m.y., precludes such a possibility (e.g., Preiss, 1987; Powell et al., 1994). Meert and Van der Voo's suggestion that glacial-interglacial intervals occurred as Australia (among other continents) drifted into, or out of, latitudes = 25° is also not supported by the paleomagnetic data from the preglacial Yaltipena Formation; these data suggest that Australia maintained a low latitudinal position prior to, as well as during, the Marinoan glaciation. Thus, our new paleomagnetic data support the existence of an unusually cold climate in late Neoproterozoic time. The number of magnetic polarity intervals present within the Elatina Formation, considered along with its stratigraphic relationship to other Marinoan glacial deposits, suggests that glaciation persisted at low latitudes in Australia for at least several hundreds of thousands to possibly millions of years.

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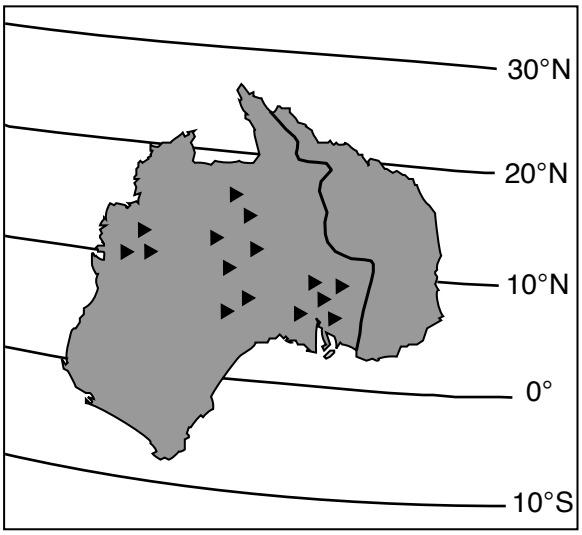

Figure 11. Inferred paleolatitude of Australia during deposition of the Elatina Formation ca. 600 Ma (this study). Triangles mark the locations of Marinoan glacial deposits. The Tasman Line (heavy solid line) marks the approximate eastern limit of the Australian paleocontinent prior to Phanerozoic time (Veevers, 1984).

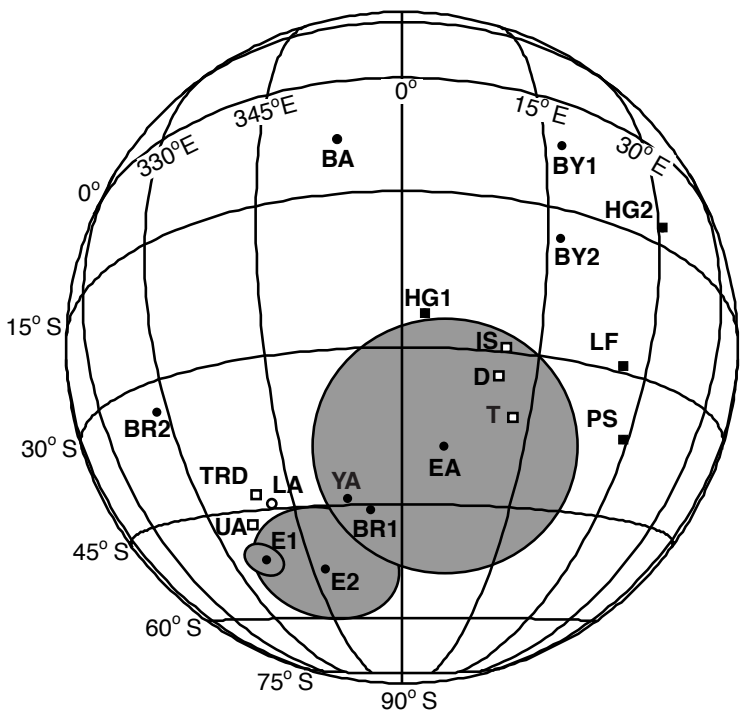

Figure 12. Comparison of Yaltipena and Elatina Formation paleopoles with other Neoproterozoic and Cambrian paleopoles from Australia (see Fig. 2 for their stratigraphic context). Filled circles represent late Neoproterozoic poles from the Flinders–Mt. Lofty Ranges (Amadeus Basin). YA, EA, this study; E1, Embleton and Williams (1986); E2, Schmidt and Williams (1995); BR1, Karner (1974); BR2, BY1, McWilliams and McElhiny (1980); BY2, Schmidt and Williams (1996); BA, Schmidt and Williams (1991); LA, Kirschvink (1978a). Open squares represent Cambrian poles from the Flinders–Mt. Lofty Ranges (Amadeus basin). HG1, HG2, LFG, T, IS, PS, D, Klootwijk (1980); UA, TRD, Kirschvink (1978a). The mean Elatina Formation paleopole from this study, as well as the two previously published Elatina poles E1 and E2, are shown with their α_{95} confidence limits in gray.

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