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A palaeomagnetic study of Empress 1A, a stratigraphic drillhole in the Officer Basin: evidence for a low-latitude position of Australia in the Neoproterozoic

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Abstract

A palaeomagnetic study of the continuously cored Empress 1A deep stratigraphic drillhole in the Officer Basin, Western Australia, has revealed a stable high-temperature remanence component for the Early Palaeozoic Table Hill Volcanics, and the Neoproterozoic Lupton, Steptoe, Kanpa, Hussar, and Browne Formations. The low inclination of the remanence supports a low-latitude position for Australia in the Neoproterozoic and Early Palaeozoic. These palaeolatitudinal estimates are consistent with the results of previous palaeomagnetic studies of Australian Neoproterozoic rocks, and support a low-latitude position during deposition of glaciogenic rocks in the Marinoan Lupton Formation.

Keywords: Australia; Glaciation; Neoproterozoic; Officer Basin; Palaeomagnetism

1. Introduction

The problem of low-latitude glaciations in the Neoproterozoic has generated much debate in the past decade (Chumakov and Elston, 1989; Schmidt et al., 1991; Meert and Van der Voo, 1994) arising mainly from the Neoproterozoic low-inclination palaeomagnetic data from glacial deposits in Australia and western Laurentia (Schmidt et al., 1991; Park, 1997). A recent analysis of the global palaeomagnetic database (McElhinny and Lock, 1996) also shows that low palaeolatitudes prevailed in the Palaeozoic and Precambrian (Kent and Smethurst, 1998) and it is widely accepted that there were at least two glacial periods between 850 and 550 Ma (Chumakov and Elston, 1989; Meert and Van der Voo, 1994; Kennedy et al., 1998; Grey and Corkeron, 1998). Possible explanations for such a Precambrian palaeoclimatic paradox include a high-obliquity Earth (Williams, 1975), a 'snowball Earth' (Kirschvink, 1992), a Saturn-like equatorial ice ring (Sheldon, 1984), a geomagnetic field significantly different from the geocentric axial

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dipole model (Kent and Smethurst, 1998), incorrect identification of glaciogenic deposits or unreliable palaeomagnetic results (Meert and Van der Voo, 1994), lower mean global temperatures for a 'ringworld' supercontinent (Worsley and Kidder, 1991; Kent and Smethurst, 1998), a cooler global temperature resulting from a lower luminosity of the juvenile Sun (Crowley and Baum, 1993), and a high mean altitude of continents (Sorokhtin and Sorokhtin, 1997)

Although not all low-inclination (corresponding to low palaeolatitude) palaeomagnetic results from glacial deposits are reliable (Evans, 2000), a well-tested case for low-latitude glaciation is presented by the Marinoan glaciation (ca. 600 Ma) in South Australia, where at least two independent studies gave almost identical palaeopole positions that indicate glaciation near the palaeoequator (Schmidt and Williams, 1995; Sohl et al., 1999). The data passed a *syn*-depositional slump fold test (Sumner et al., 1987; Schmidt and Williams, 1995; Sohl et al., 1999) and have dual polarities (Schmidt and Williams, 1995; Sohl et al., 1999) indicating that the remanence is of primary origin, and that the palaeo-secular variation has been averaged out. A recent study of the 'cap dolomite' of the Neoproterozoic Walsh Tillite in the southern Kimberley craton, northwestern Australia, also suggests a palaeolatitude of $25 \pm 12°$ for South Australia (Li, 2000) although the age of the Walsh Tillite is debated.

Ideally, continuous apparent polar wander paths (APWPs) of relevant continents for the entire Neoproterozoic, and not just for the glacial intervals, are required for testing the Earth's geodynamic/geomagnetic/palaeoclimatic behaviour in that time interval. However, only a few reliable palaeomagnetic poles have been obtained so far from Australia, because most studies have been carried out in deformed sedimentary successions within fold belts (McWilliams and McElhinny, 1980). These studies often encounter problems such as widespread orogeny-related magnetic overprints (Powell et al., 1993).

In the absence of well-exposed Neoproterozoic successions, a continuous stratigraphic diamond drillhole, such as Empress 1A in the western Officer Basin (Stevens and Apak, 1999), provides an opportunity to obtain a continuous record of palaeolatitudinal changes from undeformed and unweathered sedimentary rocks. Possible intervals of rapid continental drift between high and low latitudes should be detectable from such data.

Empress 1 and 1A are vertical stratigraphic holes drilled by the Geological Survey of Western Australia (GSWA) in 1997, at 27° 03′ 13″ S, 125° 09 24″ E (Fig. 1). The drillholes are only 6 m apart. Empress 1 and 1A were continuously cored from 105 to 612.9 and 210.5 to 1624.6 m total depth, respectively. The two drillholes here are treated as a composite section (referred to as Empress 1A) because the majority of the samples for the present study were collected from Empress 1A, where core recovery was greater than 98%. This drillhole penetrated the little deformed Palaeozoic and Neoproterozoic rocks, and a more deformed Mesoproterozoic succession (Fig. 2).

2. Geological setting and age constraints

The Officer Basin is a part of the Neoproterozoic Centralian Superbasin that covers about 2 million km^2 (Fig. 1) (Walter et al., 1995). The superbasin was developed over the junction between the North, South, and West Australian cratons at ca. 820 Ma (Walter et al., 1995; Myers et al., 1996) probably because of thermal subsidence related to a mantle plume (Zhao et al., 1994).

Age constraints on Officer Basin strata are generally poor, because reliable geochronology is sparse. However, now there is a wealth of stable isotope and biostratigraphic data (summarised by Calver and Lindsay, 1998; Calver, 2000; Hill and Walter, 2000; Hill et al., 2000; Preiss, 2000) that, when interpolated between the available radiometric dates, allows a detailed reconstruction of a stratigraphic matrix for the Australian Neoproterozoic (Walter et al., 2000). By using a combination of correlation techniques, the Officer Basin succession, and more particularly, the Empress 1A drillhole, can be correlated with successions elsewhere in the Centralian Superbasin and Adelaide Rift Complex (Hill et al., 2000) (Fig. 1). The application of these correlation techniques

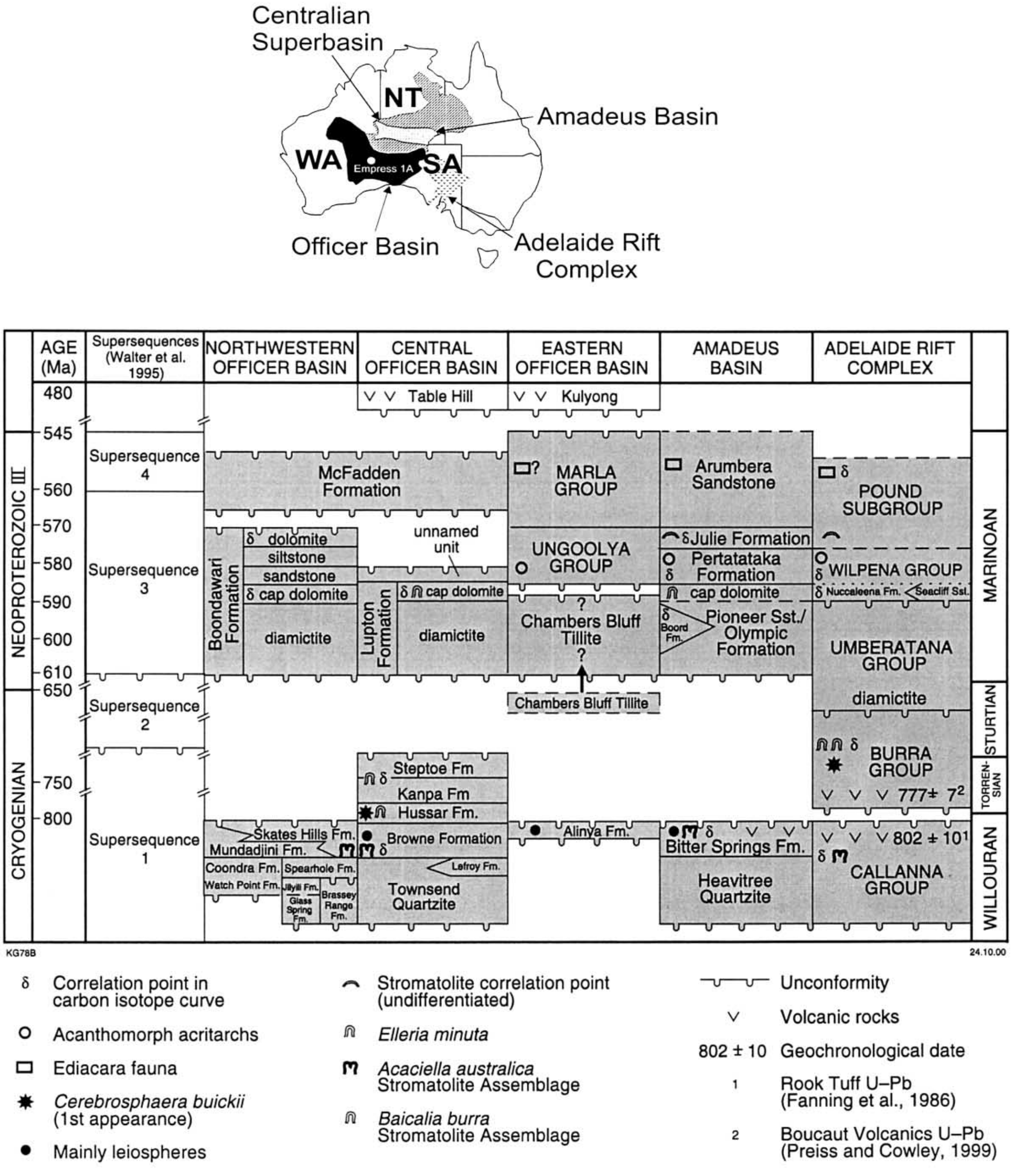

Fig. 1. (a) Probable extent of the Centralian Superbasin, some of its constituent basins, and the Adelaide Rift Complex during the Neoproterozoic (after Walter et al., 1995) showing location of the Empress 1A drillhole. (b) Summary of Neoproterozoic chronostratigraphy and stratigraphic correlations in Australia (after Grey et al., 1999) showing position of key tie points for correlations in Australia (after Grey et al., 1999) showing position of key tie points for correlations based on biostratigraphy (palynology and stromatolite biostratigraphy) and chemostratigraphy (after Hill et al., 2000), together with absolute ages from geochronology.

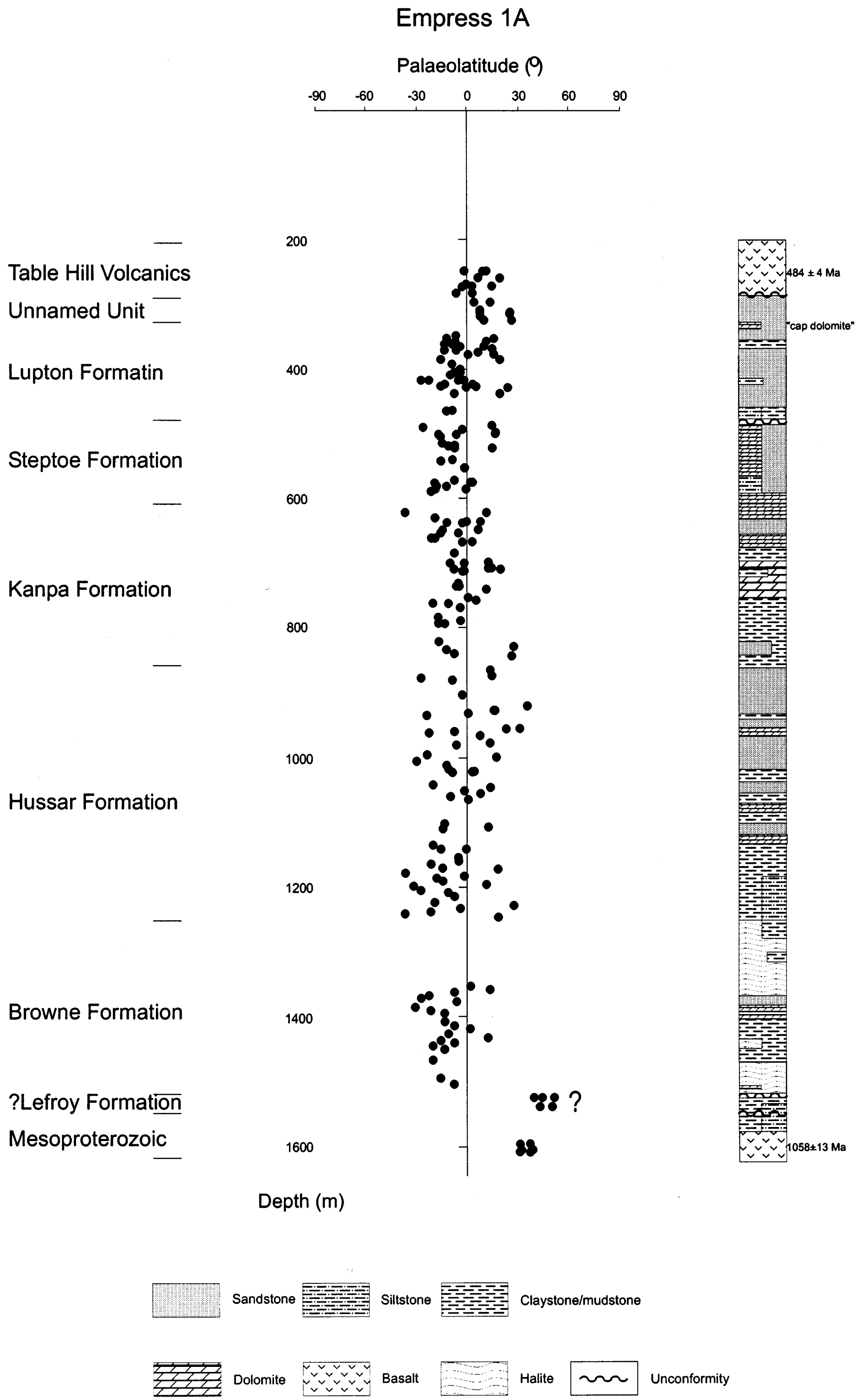

Fig. 2. Express 1A drillhole showing lithostratigraphy, measured palaeolatitudes, and generalised lighology. The results from the ?Lefroy Formation are considered preliminary and are queried on the diagram because complete thermal demagnetisation could not be performed. The sign of the palaeolatitudes are arbitrary, corresponding to that of the inclinations.

has allowed the identification of tie-points between the basins, confirming most of the previously proposed lithostratigraphic correlations.

Empress 1A (Fig. 2) penetrated nearly 290 m of Phanerozoic rocks (including about 84 m of Ordovician Table Hill Volcanics), nearly 200 m of late Neoproterozoic sandstone, dolomite, and diamictite (assigned to an unnamed unit and the Lupton Formation), more than 1000 m of Cryogenian rocks (assigned to the Steptoe, Kanpa, Hussar, and Browne Formations), about 20 m of siltstone with thin basalt interbeds (tentatively assigned to the Cryogenian Lefroy Formation, but which could be part of the underlying undifferentiated Mesoproterozoic succession), and 85 m of Mesoproterozoic siltstone and basalt, in which the drillhole terminated.

The section sampled for palaeomagnetic studies starts at 201.3 m in fine- to coarse-grained basalts of the Table Hill Volcanics, which yielded an Early Ordovician whole-rock K–Ar age of 484 ± 4 Ma (Stevens and Apak, 1999), from a depth of 215.9–216.49 m in Empress 1A.

An unnamed unit (Stevens and Apak, 1999) unconformably underlies the Table Hill Volcanics at 286 m and consists of a fine- to coarse-grained sandstone (Fig. 2). This unit has very shallow (4–8°) bedding dips. It is interpreted as the initial phase of a marine transgression following a glaciogenic phase of deposition (Carlsen et al., 1999; Grey et al., 1999).

The Lupton Formation gradationally underlies the unnamed unit (Stevens and Apak, 1999; Grey et al., 1999) at 317.1 m, and consists of light-brown sandstones and mudstones. The upper 15 cm of the Lupton Formation is dolomitic and is lithologically similar to the 'cap dolomite' at the top of Marinoan glaciogenic rocks in the Amadeus and other basins of the Centralian Superbasin (Preiss et al., 1978). The thin, dolomitic interval in Empress 1A is correlated with the Marinoan 'cap dolomite' because the light carbon-isotopic composition, with values around -8 to -9 $\delta^{13}C_{carb}$ (PDB) '...is distinctive and consistent with the identification of this unit as the upper marker cap dolomite...' (Walter and Hill, 1999), and is supported by a tentative stromatolite correlation (Grey, 1999a). The remainder of the Lupton Formation is predominantly a diamictite that contains some striated pebbles, and is correlated with other Marinoan glacigenic units in Australia, such as the Boondawari, Olympic, and Nuccaleena Formations (Grey et al., 1999). It is almost horizontal, with dips generally less than 5° (Stevens and Apak, 1999).

Absolute dating of late Neoproterozoic successions is poor, and the available data and their significance were recently reviewed by Preiss (2000). Based on this data together with dates extrapolated from the global stable-isotope record, Walter et al. (2000) estimated the age of the Marinoan glaciation to be 595–605 Ma. Both the 'unnamed unit' and the Lupton Formation are a part of the Supersequence 3 of Walter et al. (1995). A sample of Lupton Formation diamictite (from 457.5 m in Empress 1A) produced detrital zircon ages using the U–Pb Sensitive High-Resolution Ion Microprobe (SHRIMP) method. Thirty-nine analyses were obtained from 35 zircons. One detrital grain gave a weighted mean of 791 ± 18 and 788 ± 26 Ma for $^{207}Pb/^{206}Pb$ and $^{206}Pb/^{238}U$ ages, respectively. These are interpreted as the maximum possible ages for the deposition of diamictite (Nelson, 1999). Although confirming a Neoproterozoic age for the sample, the maximum age does not differentiate between the Sturtian and Marinoan glacial events.

The Steptoe Formation unconformably underlies the Lupton Formation at 483 m (Apak and Moors, 2000). It consists of sandstone, mudstone, and dolomite, and was deposited in a range of restricted shallow-marine or lagoonal environments. The upper part of the Steptoe Formation has been significantly karstified, and there appears to be a major time break between the Lupton and Steptoe Formations that probably corresponds to all of the Sturtian Epoch, together with the early part of the Marinoan Epoch of the Adelaide Rift Complex. Geochronological age control for the Sturtian Epoch is poor, but it is estimated to be about 720–680 Ma by Preiss (2000, Table 2), and the Sturtian glaciation is inferred to have occurred at about 690–700 Ma by Walter et al. (2000, Fig. 8) and about 700 Ma by Preiss (2000, Table 3).

The Kanpa Formation conformably underlies the Steptoe Formation at 617 m. It predominantly consists of dolomite, grey mudstone, and sandstone. It was probably deposited in a sabkha to shallow-marine environment under oxidizing to slightly reducing conditions (Carlsen et al., 1999).

The Hussar Formation conformably underlies the Kanpa Formation at 860.8 m. It comprises red and grey sandstone, dolomite, mudstone, and minor, locally developed conglomerate. It contains some minor erosional contacts and was deposited mainly in a tidal-flat to sabkha and shallow-marine environments (Carlsen et al., 1999).

The Steptoe and underlying Kanpa and Hussar Formations contain abundant stromatolites belonging to the *Baicalia burra* Stromatolite assemblage (Stevens and Grey, 1997; Grey, 1999a). Stromatolite taxa are identical to those recorded from several outcrops in the Officer Basin, as well as in the Burra Group of South Australia (Preiss, 1972, 1974; Stevens and Grey, 1997; Hill et al., 2000) especially in the Skillogalee Dolomite of the Adelaide Rift Complex. Moreover, a palynomorph assemblage, dominated by the distinctive acritarch *Cerebrosphaera*, occurs in Empress 1A at the same stratigraphic level in several other Officer Basin drillholes, and in the Burra Group of South Australia (Grey and Cotter, 1996; Cotter, 1999; Grey, 1999b; Hill et al., 2000) and in the Svanbergfjellet Formation in Spitsbergen (Butterfield et al., 1994). Biostratigraphic correlation is further supported by carbon-isotope analysis. The carbon-isotope curve from Empress 1A (Walter and Hill, 1999) can be matched to curves from several sections throughout the Australian Cryogenian, as well as to curves elsewhere in the world for the same interval (Hill and Walter, 2000; Walter et al., 2000).

Rhyolites from the Boucaut Volcanics at the base of, or within, the Rhynie Sandstone (the basal unit of the Burra Group) have U–Pb SHRIMP zircon ages of 777 ± 7 Ma (Preiss and Cowley, 1999; Preiss, 2000). The age of the Burra Group is estimated to range from 700 to 780 Ma (Preiss, 2000, Tables 2 and 3), and the Hussar, Kanpa and Steptoe Formations are regarded as being approximately of the same age. They are equivalent to the upper part of the Supersequence 1 of the Centralian Superbasin (Walter et al., 1995).

The Browne Formation conformably underlies the Hussar Formation at 1247.1 m (Stevens and Apak, 1999). It comprises red, fine-grained siliciclastic rocks, evaporites and stromatolitic dolomite. It was deposited in environments ranging through aeolian, sabkha to tidal-flat, hypersaline, lacustrine and lagoonal (Carlsen et al., 1999).

The Browne Formation in Empress 1A contains stromatolites of the *Acaciella australica* Stromatolite assemblage (Grey, 1999a). Taxa of this assemblage are widespread in older Supersequence 1 successions throughout Australia, and the Browne Formation can be correlated with the Skates Hills Formation of the northern Officer Basin, the Bitter Springs Formation of the Amadeus Basin, the Yackah beds of the Georgina Basin, and the Coominaree Dolomite of the Callanna Group in the Adelaide Rift Complex, using stromatolite biostratigraphy (Grey, 1995, 1999a; Stevens and Grey, 1997; Hill et al., 2000).

Palynological assemblages are more restricted than in the overlying Supersequence 1 succession, and lack distinctive marker species, but they are similar to those recorded previously elsewhere in the Centralian Superbasin, in particular in the Bitter Springs Formation of the Amadeus Basin and the Alinyah Formation of South Australia (Zang and Walter, 1992; Zang, 1995; Grey and Cotter, 1996; Cotter, 1997, 1999; Grey and Stevens, 1997; Grey, 1999b; Hill et al., 2000).

Isotope chemostratigraphy also strongly supports the correlation of the Browne Formation in Empress 1A with the formations listed above (Walter and Hill, 1999; Hill and Walter, 2000; Hill et al., 2000; Walter et al., 2000) (Fig. 1).

Geochronological constraints are available for this part of the succession in the Adelaide Rift Complex, and correlations between the Browne Formation and the Coominaree Dolomite have been proposed based on both the stromatolite and stable-isotope curves, and indicate that the Browne Formation is about 800 Ma old (Hill and Walter, 2000; Hill et al., 2000). The Coominaree Dolomite lies stratigraphically below the Rook Tuff, which has a concordant U–Pb age of 802 ± 10 Ma from single zircon grains (Fanning et al., 1986; Preiss, 2000). A constraint on the maximum age for Supersequence 1 is provided by dating on the volcanic rocks at, or near, the base of the succession. Recent publications (Preiss, 2000; Walter et al., 2000) regard the Gairdner Dyke Swarm, Little Broken Hill gabbro, a dolerite dyke in the Western Musgrave Complex, and the Wooltana Volcanics to be cogenetic, indicating an age of about 824–830 Ma.

The stratigraphic unit between 1521.8 and 1540.2 m in Empress 1A, tentatively assigned to the Lefroy Formation by Stevens and Apak (1999) is overlain with a low-angle unconformity by the Browne Formation. It consists of brownish and light-grey silty mudstone, basalt, and a basal conglomerate. Its stratigraphic position remains uncertain. It may be equivalent to the Lefroy Formation (Lowry et al., 1972), but similarities in lithology and the presence of thin, highly altered basalt layers, indicate a similarity to the underlying Mesoproterozoic succession. Limited data suggest an increase in thermal maturity (Stevens and Apak, 1999) between the Browne Formation and the ?Lefroy Formation. Although the conglomerate at 1540.2 m was interpreted as a basal conglomerate to the Officer Basin succession, it is possible that it is intraformational. In this case, the unit would be of an age of about 1060 Ma similar to the unnamed pre-Officer Basin succession.

The pre-Officer Mesoproterozoic succession consists of mudstone, with thin silty and sandy mudstone horizons from 1540.2 to 1570 m, and fine-grained basalt from 1570 to 1624.6 m. A whole-rock K–Ar age of 1058 ± 13 Ma was obtained from this basalt (Stevens and Apak, 1999).

Although direct absolute dating is poor for the succession described above, a 'correlation-point' approach, using a combination of geochronological reference points, stable isotope values for carbon, oxygen and strontium, stromatolite biostratigraphy, and palynology, as well the more traditional lithostratigraphic approach, has provided a detailed stratigraphic matrix for the Australian Neoproterozoic (Walter et al., 2000). As more data become available, the ages assigned to the various stratigraphic components may vary somewhat, but the consistencies observed so far suggest that this will not represent the major readjustments, and will be within the constraints required for palaeomagnetic analysis. We have therefore followed the timeframes proposed by Preiss (2000) and Walter et al. (2000) as the basis for our age determinations.

According to Powell et al. (1994) and Myers et al. (1996), the Australian craton was assembled at ca. 1100 Ma, and the Centralian Superbasin developed > 200 Ma after this, so the palaeomagnetic results reported here should be representative for the whole craton.

3. Sampling and rock magnetism

Two hundred and fifty-two samples were collected from the drillcore of Empress 1A with sampling spaces between 1 and 15 m. Two to five cylindrical specimens, each with a diameter of 2.5 cm and length of 2.2 cm, were drilled from each sample. All specimens are oriented vertically, but unoriented azimuthally. The maximum drillhole deviation measured from the vertical over the sampled interval in Empress 1A is 1.4°.

Palaeomagnetic study of core samples was carried out in the palaeomagnetic laboratory at the University of Western Australia (UWA). Magnetic remanence was measured using a 2G755R cryogenic magnetometer. Specimens were subjected to a stepwise thermal demagnetisation using MMTD1 furnaces manufactured by Magnetic Measurements (~ 10 nT residual field), and stepwise alternating field (AF) demagnetisation using a 2G600 automated degaussing system. Magnetic susceptibility was measured after each heating step with a MS2 susceptibility meter. Isothermal remanent magnetisation (IRM) experiments were carried out using a MMPM9 pulse magnetiser.

The Natural Remanent Magnetisation (NRM) of most of the sedimentary rock samples ranges from 0.01 to 60 mA/m and magnetic susceptibility (κ) from 0.1 to 100 ($\times 10^{-5}$) SI units, with the ?Lefroy Formation being the only exception: NRM = (15–114) mA/m, $\kappa = (42\text{–}87) \times 10^{-5}$ SI units. Igneous rocks are much more magnetic, with NRM = (23–265) mA/m and $\kappa = (100\text{–}2300) \times 10^{-5}$ SI units for the Table Hill Volcanics, and NRM = (3000–4600) mA/m and $\kappa = (4500\text{–}4600) \times 10^{-5}$ SI units for the Mesoproterozoic basalts.

IRM experiments show that hematite is present in all the sedimentary formations studied (Fig. 3b–f). However, a low-coercivity mineral (probably magnetite and/or maghemite) is also evident, especially in the Steptoe, Kanpa (Fig. 3c), and Browne (Fig. 3e) Formations. Magnetite is proba-

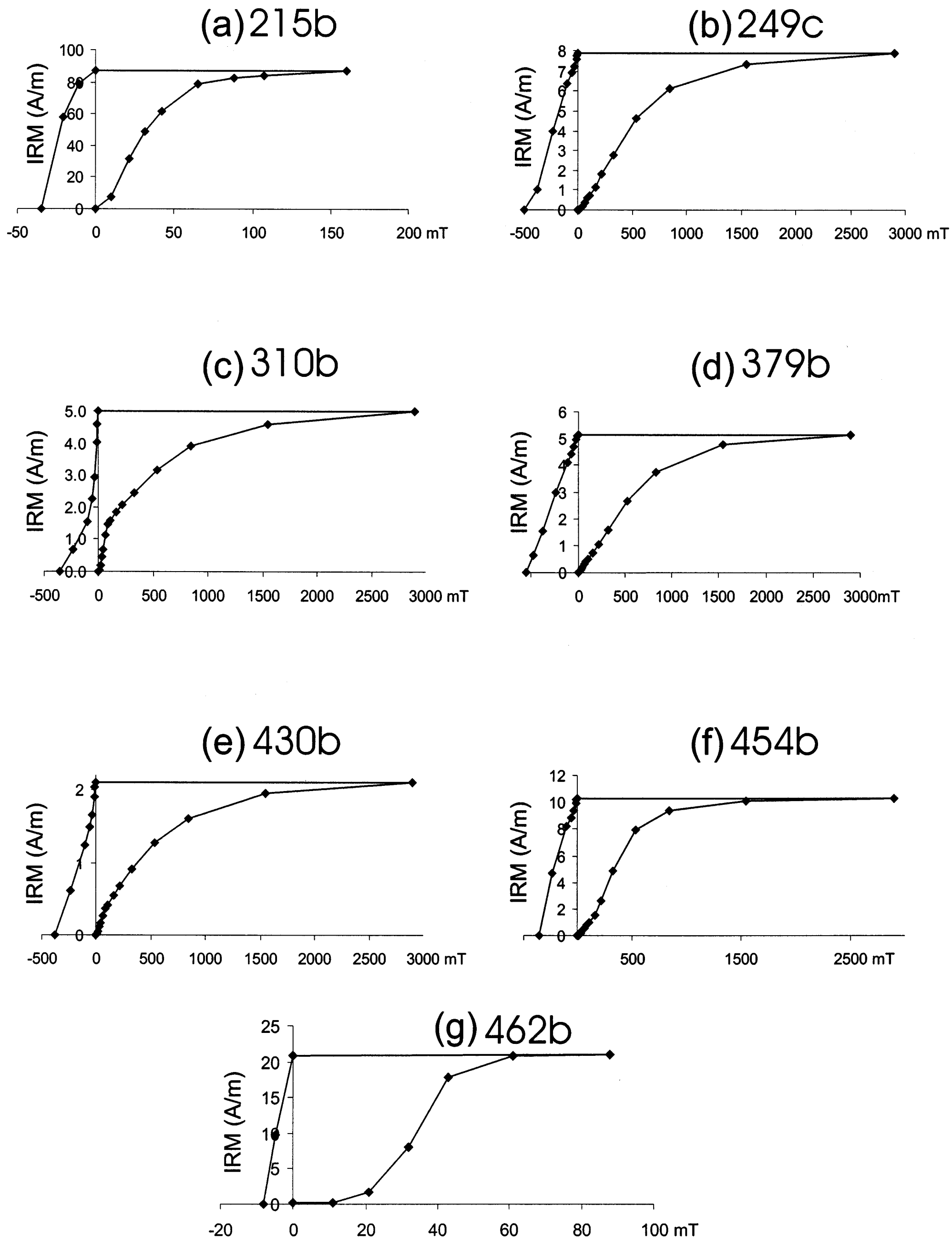

Fig. 3. IRM saturation curves for: (a) the Table Hill Volcanics; (b) the Lupton Formation; (c) the Kanpa Formation; (d) the Hussar Formation; (e) the Browne Formation; (f) the ?Lefroy Formation; and (g) Mesoproterozoic basalts.

bly the main magnetic mineral in the Table Hill Volcanics and the Mesoproterozoic basalts in Empress 1A (Fig. 3a and g).

The magnetic susceptibility of most sedimentary rock samples from the Lupton, Steptoe, Kanpa, and Hussar Formations did not change significantly during thermal demagnetisation, implying that heating did not cause any significant change in magnetic mineralogy. However, in some samples from the Kanpa Formation, there was some decrease in the magnetic susceptibility between 360 and 400°C. This change can be explained by the high-temperature oxidation of maghemite and/or magnetite. The grey sedimentary rocks of the ?Lefroy Formation show a significant increase in magnetic susceptibility above 400°C, sometimes causing random 'jumps' of the magnetisation because of remanence contamination in the process. The susceptibility of the red beds in the Browne Formation also increased after 400°C, but did not cause any change in the palaeomagnetic directions (Fig. 4g). The increased susceptibility could have resulted from the generation of magnetite owing to the breakdown of some non-magnetic minerals such as pyrite (Dunlop and Ozdemir, 1997). The Table Hill Volcanics and Mesoproterozoic basalts show a small decrease in magnetic susceptibility after 350–400°C, possibly because of high-temperature oxidation of maghemite and/or magnetite.

4. Palaeomagnetic analysis

The characteristic components of NRM (ChRM) were isolated using a least-squares algorithm (Torsvik, 1986) combined with the analysis of stereograms. Thermal demagnetisation was the main demagnetisation method for sedimentary rock samples because of the widespread presence of hematite. In most samples, 1–2 remanence components were isolated (Fig. 4). AF demagnetisation was ineffective for samples from the Lupton, Hussar, Browne, and ?Lefroy Formations. In the Table Hill Volcanics (Fig. 4b), Mesoproterozoic basalts, and some grey sedimentary rocks of the Steptoe and Kanpa Formations, AF demagnetisation results were identical to those of thermal demagnetisation.

A majority of specimens contains a steep upward component that persists up to 300°C, and in some cases (Fig. 4f and h) to even higher temperatures. This is probably a drilling-induced remanent magnetisation (DIRM), which is largely demagnetised by AF cleaning to 20 mT or heating to about 400°C (Dunlop and Ozdemir, 1997, p. 293). This component is absent in some red beds of the Hussar and Browne Formations (Fig. 4e and g).

The high-temperature component has unblocking temperatures between 400 and 680°C for sedimentary rocks, and between 300 and 580°C for the Table Hill Volcanics and the Mesoproterozoic basalts. In the Palaeozoic and Neoproterozoic rocks of Empress 1A, the component is very stable and mostly has shallow inclinations (Fig. 4). The corresponding palaeolatitudes are shown in Fig. 2.

These low inclinations make the delineation of magnetic polarities difficult because the samples are unoriented azimuthally. We have therefore made two estimates of the mean palaeolatitude for each chosen interval using the individual core directions. The first estimate is based on the assumption of unipolar data for each formation, and this gives a lower limit for absolute palaeolatitude. The second estimate was calculated after the inversion of all data with negative inclinations, and this gives an upper limit for palaeolatitude. We used the algorithm of McFadden and Reid (1982) for both estimates. The results are shown in Table 1 and Fig. 5.

Samples from the ?Lefroy Formation show relatively high downward inclinations and would thus correspond to higher palaeolatitudes. However, we are not fully convinced that this component was properly isolated. It was impossible to carry out the complete thermal demagnetisation of these samples because they exploded in the furnace at temperatures above 400°C. AF demagnetisation was ineffective. Consequently, the results from the ?Lefroy Formation have not been included in our conclusions, although the corresponding palaeolatitudes are shown in Fig. 2.

Results from the Mesoproterozoic basalt are internally coherent and well defined (Fig. 4h). The mean inclination of the high-temperature compo-

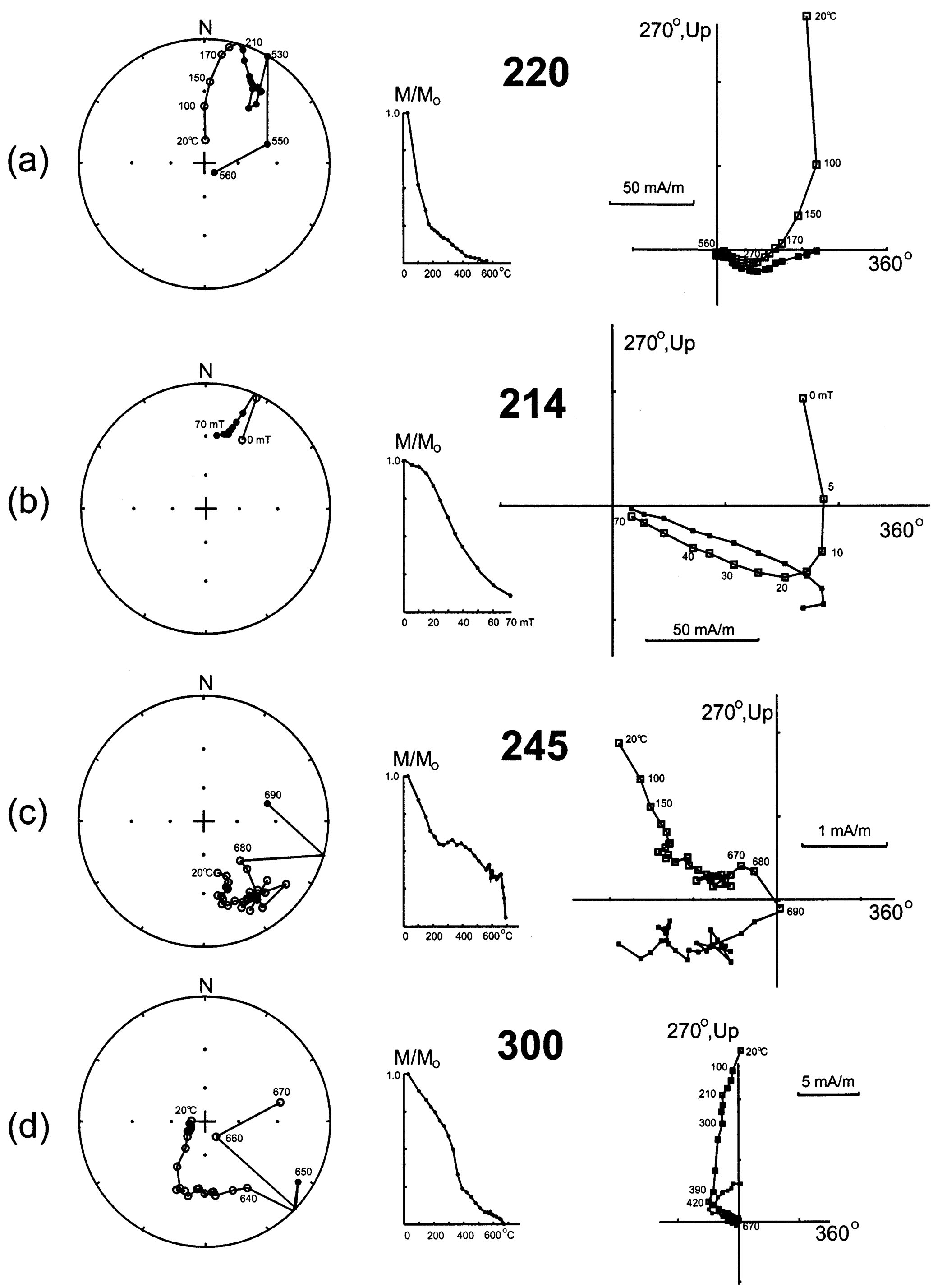

Fig. 4. Examples of thermal and AF demagnetisation results (equal-angle projection, orthogonal plots and normalised magnetisation decay) of samples from Express 1A for the following formations: (a) Table Hill (thermal); (b) Table Hill (AF); (c) Lupton; (d) Kanpa; (e) Hussar (one component); (f) Hussar (two components); (g) Browne; and (h) Mesoproterozoic basalts. Solid and open symbols on the stereographic projections represent downward and upward directions, respectively; solid and open symbols on the Zijderveld (1967) diagram denote horizontal and vertical projections, respectively.

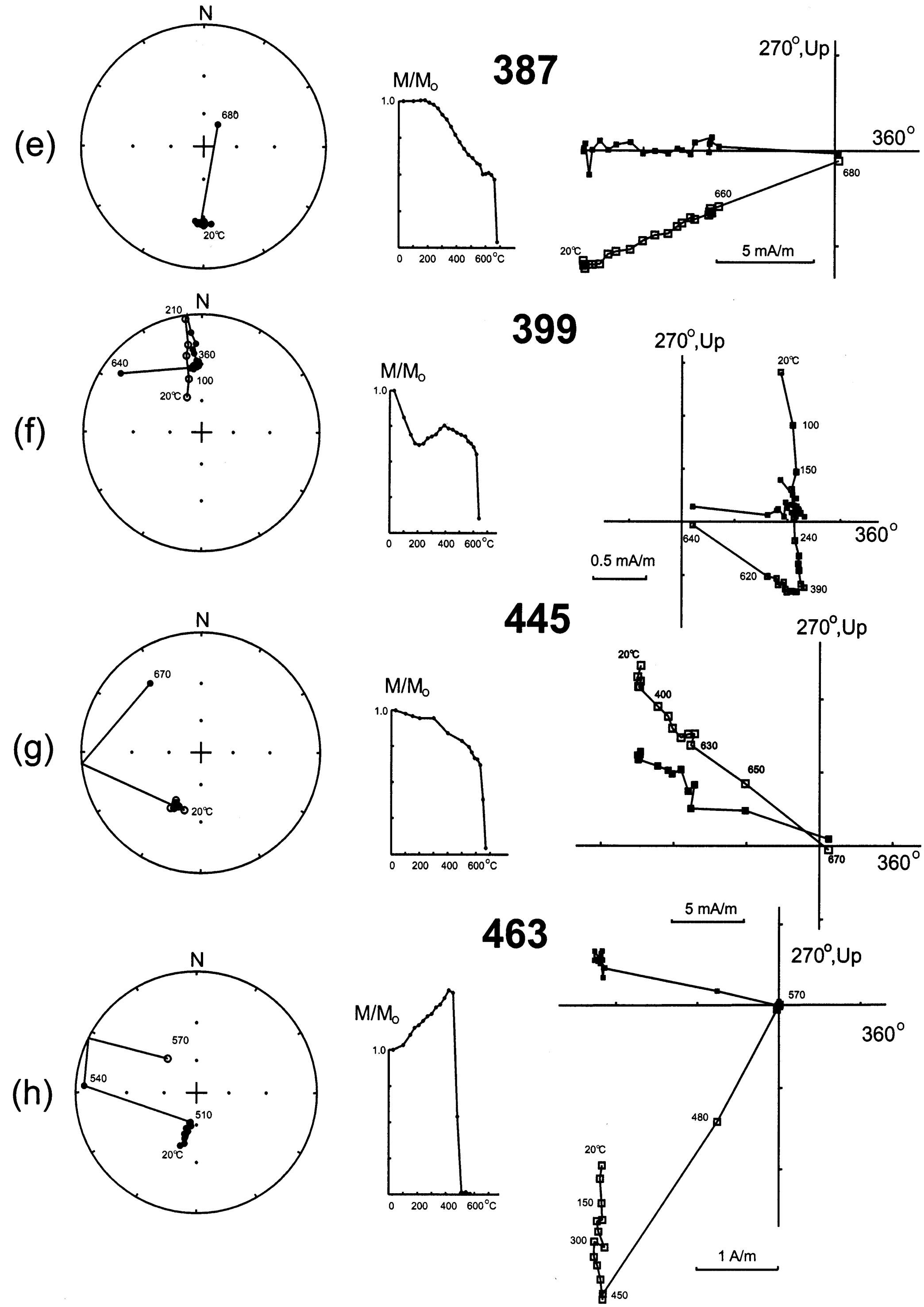

Fig. 4. (*Continued*)

Table 1
Estimated palaeolatitudes of Australia in Neoproterozoic–Early Palaeozoic at Empress 1A, location 27°03′13″ S, 125°09′24″ E

Formation	Interval thickness (m)	Estimated age (Ma)	Plat 1 (°)	Plat 2 (°)
Table Hill	(201–286)	484	3.6 ± 4.5	7.2 ± 3.1
Lupton	(317–483)	600	0.6 ± 4.7	10.9 ± 2.5
Steptoe	(483–589)	750	6.5 ± 5.8	11.6 ± 3.4
Kanpa A	(621–667)	753	7.3 ± 8.0	11.1 ± 5.7
Kanpa B	(670–731)	757	1.5 ± 8.2	8.6 ± 4.4
Kanpa C	(735–785)	760	4.9 ± 9.2	8.9 ± 5.8
Kanpa D	(789–857)	762	3.5 ± 17.5	15.0 ± 8.0
Hussar A	(860–1041)	767	0.1 ± 7.8	14.8 ± 4.1
Hussar B	(1046–1101)	770	0.8 ± 12.8	7.5 ± 7.1
Hussar C	(1106–1151)	780	7.4 ± 21.9	12.5 ± 12.3
Hussar D	(1153–1246)	790	9.6 ± 9.2	16.5 ± 5.3
Browne	(1247–1522)	810	10.6 ± 6.1	13.4 ± 4.1
Mesoproterozoic				
Basalt	(1570–1625)	1058	34.7 ± 4.0	–

Plat 1: palaeolatitude with the assumption of a single polarity; Plat 2: palaeolatitude with the assumption of dual polarities. All palaeolatitudinal estimates were made using the algorithm outlined by McFadden and Reid (1982).

nent (54°) is reasonably close to that of the Kulgera Dyke Swarm of approximately the same age (Rb/Sr 1054 ± 14 Ma, Camacho et al., 1991). However, we regard these results as preliminary because of the small number of samples.

5. Discussion

The palaeomagnetic analysis shows that the Lower Palaeozoic Table Hill Volcanics and Neoproterozoic sedimentary rocks of the unnamed unit, Lupton, Steptoe, Kanpa, Hussar, and Browne Formations all contain a stable high-temperature remanence component with shallow inclination. The remanence is carried by magnetite in the Table Hill Volcanics, and by hematite in the sedimentary rocks. There are three possible explanations for these consistent shallow inclinations: (1) the entire section was remagnetised; (2) Central Australia was in low palaeolatitude during most of the Late Neoproterozoic to the Early Palaeozoic interval (ca. 820–480 Ma); (3) the geomagnetic field in the Neoproterozoic–Early Palaeozoic was significantly different from that of the present.

Remagnetisation could have occurred during one of the two possible thermal events during the Phanerozoic. However, post-Early Carboniferous remagnetisation is an unlikely explanation for the low-inclinations because Australia stayed at a relatively high-latitude position during that time interval (Baillie et al., 1994). The palaeoposition of Australia between the Cambrian and the Early Carboniferous (540–340 Ma) was mostly at low latitudes (Li et al., 1993), so remagnetisation at this time interval is possible. However, there is abundant evidence for dry and warm environments during the deposition of the Browne Formation, e.g. the presence of salt (Carlsen et al., 1999), which indicates a depositional to early diagenetic age for the hematite in these sedimentary rocks. The red sedimentary rocks of the Hussar Formation are palaeomagnetically similar to those of the Browne Formation. The difference in inclinations between the Mesoproterozoic and younger rocks does not favour an overprint either. We therefore see no reason for the low-inclination results to represent a complete remagnetisation.

A low-latitude position for Central Australia, as implied by the low inclination results from the Neoproterozoic formations studied in the Officer Basin, is consistent with the previous palaeomagnetic data (Embleton and Williams, 1986; Schmidt and Williams, 1995; Sohl et al., 1999; Li, 2000).

Table 1 and Fig. 5 show the maximum and minimum estimates of palaeolatitude for the intervals studied. Fig. 5 also shows the palaeolatitudinal estimations from the previous palaeomagnetic results (grey lines) as summarised in Table 2 of Li (2000) for age windows of ca. 510, 540, 600 (recalculated to include the two new results in Sohl et al., 1999), 755, and 1054 Ma. Note that the Walsh Tillite (dashed line) is shown in two possible age positions, but a Marinoan age would contradict the other palaeomagnetic data. The results for the Walsh Tillite suggest an intermediate-latitude position for Australia at that time. Unfortunately there is a hiatus in the Empress 1A core for both the Sturtian glaciation, and Sturtian Epoch, and no other well-dated palaeomagnetic information is available for that time. Only future studies can resolve this problem.

All these data are consistent with the hypothesis that the western Officer Basin mainly occupied a near-equatorial position, probably within the 20°N to 20°S latitudinal band, throughout much of the Neoproterozoic. Although there are a few direct radiometric age determinations for Officer Basin successions, lithostratigraphic units can be correlated with the well-dated formations in other parts of the Centralian Superbasin and Adelaide Rift Complex (Walter et al., 1995, 2000; Preiss, 2000; Hill and Walter, 2000; Hill et al., 2000). The low-latitude result for the Lupton Formation is particularly interesting, because this unit has been correlated with the Marinoan glaciation.

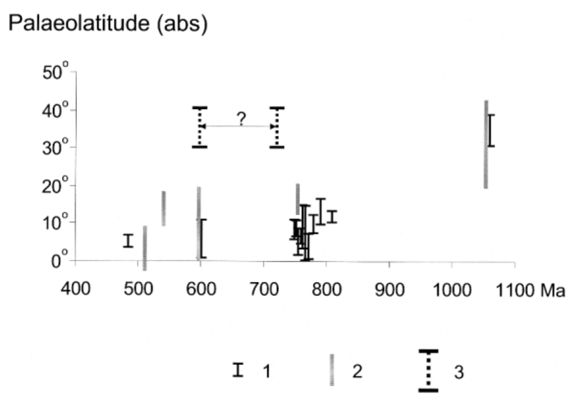

Fig. 5. Estimations of Neoproterozoic palaeolatitudes for Australia (reference point 27° 03′ 13″S, 125°09′24″E, location of the Empress 1A drillhole). (1) data from Table 1 (this study); (2) results summarised in Table 2 of Li (2000) with new data of Sohl et al. (1999) included in the 600 Ma estimation; (3) results from the Walsh Tillite (Li, 2000), with two possible ages.

The absence of palaeomagnetic data for the ca. 1000–820 Ma interval (Fig. 5) represents a regional hiatus in sedimentation (Walter et al., 1995), which may partly be because of the rapid unroofing related to a ca. 830–820 Ma mantle plume (Li et al., 1999). The higher palaeolatitude from the Mesoproterozoic basalt is based on the preliminary results only, but it agrees with other palaeomagnetic results (Camacho et al., 1991). As all data from the Mesoproterozoic basalt are of one polarity, only one estimation was made for this unit.

We cannot totally rule out the possibility that the geomagnetic field in the Neoproterozoic–Early Palaeozoic time frame was significantly different from that of the present. In this scenario, the magnetisation can be primary, but instead of reflecting a low-latitude position for Australia, it may be owing to a modest contribution from the axial geomagnetic dipole field of quadrupole and octupole fields (Kent and Smethurst, 1998).

6. Conclusions

Stable high-temperature remanence components were isolated from the Early Palaeozoic Table Hill Volcanics, and the Neoproterozoic Lupton, Steptoe, Kanpa, Hussar, and Browne Formations, which were penetrated by the stratigraphic drillhole Empress 1A in the Officer Basin. There is circumstantial evidence for a primary origin for these components. The low inclination results support low-latitudinal positions for Australia for selected intervals in the Neoproterozoic (post ca. 820 and pre-700, and about 600 Ma) and in the Early Palaeozoic. The shallowly inclined stable magnetisation of the Lupton Formation supports the low-latitude glaciation hypothesis for the Marinoan glaciation in Australia.

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