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Palaeomagnetism of Archaean rocks of the Onverwacht Group, Barberton Greenstone Belt (southern Africa): Evidence for a stable and potentially reversing geomagnetic field at ca. 3.5 Ga

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ABSTRACT

Palaeomagnetic data from the Palaeoarchaean Era (3.2–3.6 Ga) have the potential to provide us with a great deal of information about early conditions within, and processes affecting, the Earth's core, mantle, and surface environment. Here we present new data obtained from some of the oldest palaeomagnetic recorders in the world: igneous and sedimentary rocks from the Onverwacht Group of the Barberton Greenstone Belt (Kaapvaal Craton, southern Africa). Our palaeomagnetic measurements strengthen a recently published positive conglomerate test (Y. Usui, J.A. Tarduno, M. Watkeys, A. Hofmann and R.D. Cottrell, 2009) and our new U–Pb date constrains the conglomerate to older than 3455 ± 8 Ma. The new palaeomagnetic data from other units are nontrivial to interpret and are of uncertain reliability when taken individually; similar, we argue, to all other published palaeomagnetic data of this age. Nonetheless, four poles (two new, two derived from published data) produced from high temperature components of magnetisation recorded in the Komati, Noisy, and Hooggenoeg formations exhibit considerably improved clustering when their directions are corrected for differences in attitude resulting from a large fold structure dated at 3.23 Ga. On the basis of this enhanced consistency in stratigraphic coordinates, the positive conglomerate test, and the absence of any clear indications of their remagnetisation from comparison with younger poles, we argue that these are the most trustworthy palaeomagnetic results yet produced from any rocks of Palaeoarchaean age. When taken in conjunction with published data, the new results present the most compelling evidence to date that the Earth had a stable geomagnetic field at ca. 3.5 Ga in addition to presenting tentative evidence that it was undergoing polarity reversals. The data do not appear to support a claim, made previously from Palaeoarchean palaeomagnetic data from the Pilbara Craton (Y. Suganuma, Y. Hamano, S. Niitsuma, M. Hoashi, T. Hisamitsu, N. Niitsuma, K. Kodama and M. Nedachi, 2006), of extremely rapid latitudinal plate motion during this period. Finally, when compared with similarly aged data from the Pilbara Craton (Western Australia), the new data do not rule out the hypothesis that the two cratons were conjoined at this point in their history in the supercraton Vaalbara.

1. Introduction

The Palaeoarchaean Era (3.2–3.6 Ga) represents the oldest period of time for which reliable whole rock palaeomagnetic data can potentially be produced from terrestrial rocks. Two particular areas on Earth have provided most of the palaeomagnetic data from rocks of this age: the Barberton Greenstone Belt (BGB) in the Kaapvaal Craton of southern Africa and the Pilbara Granite-Greenstone terrain in western Australia, both of which are favoured on account of their low degree of metamorphism and excellent preservation of supracrustal rocks. Palaeomagnetic studies focused on these areas have the potential to address key questions about processes operating during early Earth times. Published studies have already argued for the existence of a global, stable geomagnetic field (Hale and Dunlop, 1984; McElhinny and Senanayake, 1980; Yoshihara and Hamano, 2004) of intensity comparable to that observed in the Phanerozoic

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(Tarduno et al., 2010), and rapid latitudinal motion of lithosphere (Suganuma et al., 2006) at ca. 3.5 Ga.

All of these published results are exciting but it is clear that more needs to be done to establish their individual reliability before their claims can be entirely trusted. For example, a recent study by Usui et al. (2009). suggested that the previous studies of Hale and Dunlop (1984), McElhinny and Senanayake (1980), Suganuma et al. (2006), and Yoshihara and Hamano (2004) were all of limited credibility because of the substantial risk that the rocks in question no longer retained their primary natural remanent magnetisations (NRMs). Usui et al. (2009) also published arguably the most important palaeomagnetic result so far from rocks of this age: a positive conglomerate test from a ~3.45 Ga unit in the BGB.

In this study, we present new palaeomagnetic data, supported by rock magnetic and microscopic analyses, as well as a new U–Pb age estimate, from rocks of the 3.30–3.55 Ga Onverwacht Group of the BGB and compare them with published results. Our new data increase confidence that some BGB rocks are likely to record a primary or near-primary magnetisation from ~3.5 Ga. They also provide further tantalising, but as yet preliminary, evidence that the field was undergoing polarity reversals at this stage in its history. Prior to this study, the oldest tentative evidence for their occurrence is dated at 3.2 Ga (Layer et al., 1998) and the oldest well-established reversals (supported by positive reversal tests) are dated at ~2.8 Ga (Strik et al., 2003). Finally, our results also have implications for tectonic processes: suggesting that latitudinal drifts rates at ~3.5 Ga may not have been substantially different from today's and that rocks of the Onverwacht Group may have been conjoined with similar aged rocks in the Pilbara Craton of Western Australia.

2. Geological background and sampling

The Barberton Greenstone Belt is located close to the eastern margin of the Kaapvaal Craton (Fig. 1) and comprises volcanic and sedimentary rocks spanning the period ca. 3550–3230 Ma. It has been interpreted as an accretionary complex of Archaean ocean crust (de Wit et al., 2011; de Wit et al., 1992) comprising meta-basalts of midoceanic, island arc and back-arc origins; it is intruded on all sides by granitoid plutons.

The Onverwacht Group (ca. 3550–3300 Ma) is the oldest group within the belt and is overlain by the Fig Tree and Moodies Groups. The lowermost sequences within the Onverwacht are in the

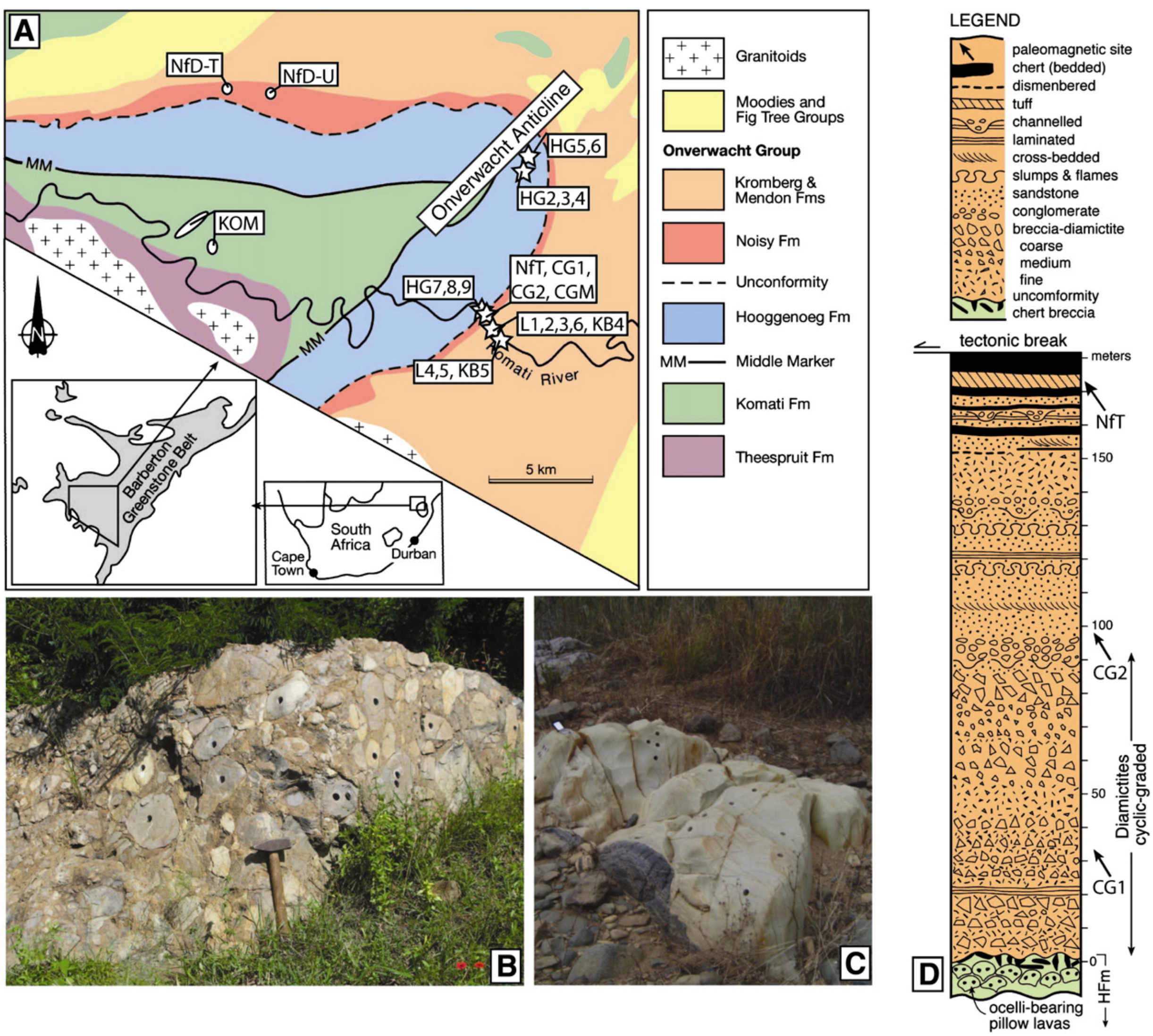

Fig. 1. (A) Location maps and simplified geological map of field area showing sampling sites within the Onverwacht Group. Stars show sites sampled for the purpose of this study (see Table 1 for codes). KOM refers to the location of sampling sites studied by Yoshihara and Hamano (2004). NfD-U refers to the location of Noisy formation dacite studied by Usui et al. (2009). Their conglomerate site was also located in the Komati River section close to our site CG1. NfD-T marks the site studied mainly for palaeointensities by Tarduno et al. (2010) (referred to in that study as BGB). (B,C): Photographs of CG2 and NfT sampling sites showing palaeomagnetic drill holes. (D): Stratigraphic column of the Noisy formation showing levels of sampling sites.

Sandspruit and Theespruit Formations but these are too intensely metamorphosed (penetratively deformed and heated up to 700 °C in parts; Dziggel et al., 2002) to be of use for palaeomagnetism. Above these are the komatiites and basalts of the Komati Formation (ca. 3482–3472 Ma) which have been previously studied using palaeomagnetism (see discussion later). Volcanic rocks of the Hooggenoeg (ca. 3472–3460 Ma) and Kromberg (ca. 3335–3445 Ma) Formations overlie these and were sampled for the purpose of this study (Table 1). The sequence is capped by the Mendon and Weltevreden Formations which were not studied here.

The Noisy formation (ca. 3460–3445 Ma, also referred to as the Buck Ridge volcanosedimentary complex; de Vries and Touret, 2007) is a newly named formation (lower case 'f' indicates its yet informal status) separating the predominantly mafic Hooggenoeg and Kromberg Formations; traditionally it is included as the top of the Hooggenoeg Formation (de Wit et al., 2011). The Noisy formation (Fig. 1D) is an upward fining felsic volcanoclastic sequence intruded by contemporaneous quartz-plagioclase porphyritic sills. The lowermost units comprise a coarse terrestrial clastic sequence, containing coarse fluvial conglomerates and occasional felsic airfall tuffs. This sequence is best exposed in the original type area in a gorge along the Komati River (Viljoen and Viljoen, 1969), where it transgresses unconformably across the Hooggenoeg Formation and is separated from the Kromberg Formation by a tectonic break (de Wit et al., 2011).

Rocks of the upper Onverwacht Group are low-grade (<400 °C) metamorphosed with peak metamorphism of the Komati and Hooggenoeg Formations inferred to have been produced immediately following their formation and burial on the sea floor (Knauth and Lowe, 2003; Schoene et al., 2008). Isotopic resetting of Onverwacht cherts is inferred to have taken place a short time later during hydrothermal activity related to the magmatic episode responsible for emplacement of parts of the Noisy formation at ca. 3.45 Ga (de Vries and Touret, 2007; Knauth and Lowe, 2003).

Major thermal events affecting the belt since this time have been inferred at 3.2, ~3.1–3.0, ~2.7 and ~2.1 Ga (de Ronde et al., 1991; Kamo and Davis, 1994; Toulkeridis et al., 1994; Weiss and Wasserburg, 1987). These were not associated with any large scale hydrothermal activity (Tice et al., 2004) and radioisotope data suggests that the BGB has not been significantly reheated since the Palaeoproterozoic (de Ronde and de Wit, 1994; de Wit et al., 1992; Weiss and Wasserburg, 1987).

Samples for this study were taken from 18 sites across the Hooggenoeg, Noisy, and Kromberg Formations (Fig. 1, Table 1). All sites are located on the limbs of steeply plunging folds with a SW–NE trending fold-axes that formed at 3.23 Ga during a major tectonic phase affecting the eastern Kaapvaal Craton (de Ronde and de Wit, 1994; de Wit et al., 2011; Schoene et al., 2008). The tectonic correction applied to the palaeomagnetic data used in this study consists of the following two steps:

1. An anticlockwise rotation of amount α, about a horizontal axis with heading 315°. This corrects for the regional plunge of the fold to the northeast.
2. A site-specific rotation, acting to bring the bedding back to the horizontal. This is a standard tilt correction but made using the bedding orientations (given in Table 2) only after they have first been modified by step 1.

The Noisy and Hooggenoeg formations are separated by an unconformity, whilst the Noisy and Kromberg formations are separated by a fault contact (de Ronde and de Wit, 1994; de Wit et al., 2011). Therefore, the plunge correction for the Hooggenoeg and Noisy formation sites ($\alpha = 92°$) is different from that applied to the Kromberg sites ($\alpha = 115°$) although the axis and sense of rotation remains the same.

The palaeomagnetic sampling of these rocks was performed during trips in May 2001, June 2004, and June 2005 and the palaeomagnetic measurements were made between 2001 and 2010. Samples were drilled with a small petrol-powered portable drill and oriented *in situ*, using both sun and magnetic compasses. Bedding plane measurements (Table 2) were made in the field. In basaltic volcanic sequences, these were obtained using abundant lava withdrawal levels in pillow lavas as the palaeohorizontal reference. In general, between 7 and 18 independently oriented core samples were taken from each of the sites and just one specimen per core was measured. The exception was site NfT (Fig. 1C) which, because of the importance of establishing a robust result for this particular unit (as will become clear), was sampled with 41 cores, from which 71 specimens were measured.

Table 1

Site details and secondary NRM components. Demag refers to type of demagnetisation (T = thermal, AF = alternating field; T + AF = thermal to 430–480 °C followed by AF). Dec is the declination, Inc the inclination, k the estimate of the Fisher dispersion parameter, α_{95} the semi-angle of the 95% cone of confidence around the mean direction, N is the number of site mean directions used to produce the given mean direction. Components are described in Section 3.

Formation	Lithology	Site	Lat (°)	Long (°)	Demag	Comp	T range °C	N/n	Dec (°)	Inc (°)	k	α95 (°)
Hooggenoeg	Basalt dyke	HG2	−25.96549	31.00314	T + AF	Unstable		0/11				
Hooggenoeg	Basalt flows	HG3	−25.96543	31.00305	T + AF	LT2	100–340	5/8	215.4	−66.4	18	14.8
Hooggenoeg	Basalt flows	HG4	−25.96593	31.00345	T + AF	LT1	100–250	6/10	357.6	−47.0	10.6	17.6
Hooggenoeg	Basalt flows	HG5	−25.96037	31.00548	T + AF	Unstable		0/11				
Hooggenoeg	Basalt flows	HG6	−25.96021	31.00549	T + AF	LT1	100–300	7/11	5.1	−56.9	81.7	5.9
Hooggenoeg	Basalt flows	HG7	−26.0209	30.98794	T	Unstable		0/9				
Hooggenoeg	Basalt flows	HG8	−26.02137	30.98791	T + AF	MT	370–430 (+35 mT)	0/10	281.7	76.5	89.4	13.1
Hooggenoeg	Basalt flows	HG9	−26.02161	30.98783	T + AF	LT1	100–400	8/10	347.1	−68.6	52.8	6.8
Kromberg	Pyroxenite dyke	KB4	−26.02898	30.9914	T	Unstable		0/8				
Kromberg	Basalt sill	KB5	−26.03295	30.99442	T, T + AF	Unstable		0/7				
Kromberg	Basalt flows	L1	−26.02928	30.99137	T	LT2	200–400	6/8	163.2	−45.3	31.3	12.2
Kromberg	Basalt flows	L2	−26.02928	30.99137	T	LT2	200–400	9/9[a]	137.7	−61.2	20.5	11.7
Kromberg	Basalt flows	L3	−26.029	30.99127	T, T + AF	LT1	100–250	5/18[a]	338.9	−51.4	33.2	10.9
						LT2	250–440	11/18	135.9	−58.7	15.1	12.1
						MT	370–460	4/18[a]	83.5	78.8	14.8	18.2
Kromberg	Basalt sill	L4	−26.03295	30.99442	T, T + AF	LT2	100–440	12/13	231.9	−67.5	45.5	6.5
						MT	440–480	4/13	338.4	78.1	18.9	21.7
Kromberg	Basalt sill	L5	−26.03295	30.99442	T, T + AF	MT	200–330	6/11[a]	254.5	85.4	113.5	6.3
Kromberg	Basalt flows	L6	−26.02917	30.99107	T	LT2	200–440	8/10	134.2	−61.5	208.5	3.8
Noisy	Conglomerate clasts	CG1	−26.0244	30.98803	T + AF	LT1	20–460	5/8	332.5	−68.2	71.4	9.1
Noisy	Conglomerate clasts	CG2	−26.02489	30.98857	T + AF, T, AF	–	20–550	6/30	340.7	−82.0	1.7	78.5
Noisy	Conglomerate matrix	CGM	−26.02449	30.98844	T + AF	LT1	100–460	4/4	358.7	−68.7	18.4	16.3
Noisy	Tuff	NfT	−26.02539	30.98902	T, T + AF, AF	LT2	100–400	49/71	159.7	−74.0	7.3	8.1

[a] These results had one or more components with MAD > 15°.

Table 2
Characteristic remanent magnetisation (ChRM) components measured in this study. Strike and dip refer to the bedding planes measured in the field. Dec′ and Inc′ refer to the direction of the components in stratigraphic coordinates after applying the plunging fold correction.

Formation	Site	Comp.	T/T + AF range	N/n	Dec (°)	Inc (°)	k	α95 (°)	Strike (°)	Dip (°)	Dec′ (°)	Inc′ (°)
Hooggenoeg	HG2	HT1	340 °C–430 °C + 110 mT	5/11	177.3	66	264	3.9	3	100	64.0	−11.4
Hooggenoeg	HG3	HT1	430 °C + 30 mT–430 °C + 110 mT	6/8	175.5	62.3	153	4.6	3	100	67.7	−12.3
Hooggenoeg	HG4	Unstable		0/10								
Hooggenoeg	HG5	HT2	460 °C–460 °C + 58 mT	6/11[a]	14.8	60.2	113.4	6.3	348	100	12.9	−21.8
Hooggenoeg	HG6	HT2	460 °C + 5 mT–460 °C + 58 mT	7/11	47.3	71.9	44.7	9.1	348	100	31.3	−25.3
Hooggenoeg	HG7	Unstable		0/9								
Hooggenoeg	HG8	HT2	430 °C + 15mT–430 °C + 90mT	6/10	27.4	52.5	92.2	7.0	55	90	8.4	16.4
Hooggenoeg	HG9	HT2	460 °C + 20 mT–460 °C + 103 mT	9/10	66.1	61.5	57.8	6.8	55	90	14.5	−5.3
Kromberg	KB4	Unstable		0/8								
Kromberg	KB5	Unstable		0/7								
Kromberg	L1	HT1	400–550 °C	8/8	171.5	60	44.2	8.4	63	84	28.3	−22.4
Kromberg	L2	HT1	400–570 °C	9/9	176	54.8	79.9	5.8	63	84	33.0	−26.2
Kromberg	L3	HT1	400–570 °C	12/18[b]	156.1	50.4	87.6	4.3	63	84	20.8	−33.5
Kromberg	L4	L4	400–580 °C	7/13	177.3	10.9	64.6	7.6	55	88	84.9	−55.4
Kromberg	L5	HT1*	200–480 °C	5/11[b]	136.7	57.8	36.4	12.8	75	90	346.3	−28.0
Kromberg	L6	HT1	480–580 °C	4/10[b]	167.4	53.8	41.8	14.4	43	86	46.0	−25.5
Noisy CG1	CG1	CG1	460 °C–460 °C + 110 mT	6/8[c]	144.1	−58.3	1.2					
Noisy	CG2	CG2	480 °C + 5 mT–480 °C + 110 mT	19/30[c,d]	216.8	70.1	1.1	134.1				
Noisy	CGM	Unstable		0/4								
Noisy	NfT	NfT	400–580 °C	30/71[e]	167	−33.8	25.1	5.4	46	90	185.3	−45.4

[a] One of these directions was a great circle fit.
[b] One or more of these directions had MAD > 15°.
[c] All of these directions had MAD < 5°.
[d] These numbers refer to the number of specimens, results were obtained from a total 10 clasts and the mean of each of these were used to calculate the overall mean direction.
[e] All of these directions had MAD < 10°.

For the conglomerate tests, eight oriented core samples (CG1) were taken from individual clasts at a site where an *in situ* unit cropped out. A further 19 cores (CG2) were taken from a large loose block (Fig. 1B) from the same diamictite unit some 60–70 m to the southeast. Site CG2 included 3 clasts from which two independently-oriented samples were taken to check consistency of directions. Additionally, measurements were made from multiple specimens from the same cores from CG2 (30 specimens measured from 19 cores). Close to site CG2, four oriented cores (site CGM) were taken from the matrix of the conglomerate unit.

3. Experimental results

3.1. Palaeomagnetic analyses

Standard-sized palaeomagnetic samples were demagnetised using thermal and alternating field (AF) treatments and measured using a '2G' SQUID magnetometer (noise level 3×10^{-12} Am^2 although the sample holder, whose magnetisation is corrected for, potentially increases this noisy level by up to an order of magnitude). Thermal demagnetisation was performed using an in-house thermal demagnetiser and an ASC oven at the Fort Hoofddijk palaeomagnetic laboratory, Utrecht University (The Netherlands). Alternating field demagnetisation experiments were performed using the unique "robot" at Fort Hoofddijk: a 2G magnetometer with an inline AF demagnetiser attached to an automatic sample handler which measures batches of 96 samples in three orientations. In a large number of cases, (see Table 1), thermal demagnetisation was performed stepwise (using increments of 10–100 °C) up to 430–480 °C and then AF demagnetisation was used to remove (with increments of 5–10 mT up to 110 mT) the remaining NRM. This combination was chosen to make best use of the rapid and precise robot magnetometer while still ensuring that secondary magnetisations were fully removed. The analysis of the palaeomagnetic data was made using standard techniques (Fisher, 1953; Kirschvink, 1980). Examples of Zijderveld plots produced by measurements made in this study are given in Fig. 2. Site mean and component mean directions are given in Tables 1, 2 and 3 and plotted in Fig. 3. Poles associated with these are plotted in Fig. 4A.

Samples from the Hooggenoeg and Kromberg metabasalts generally had NRM intensities in the range of 1–100 mA/m, similar to other basalts of Archaean age (Biggin et al., 2009; Strik et al., 2003, 2007) but orders of magnitude weaker than more recent basalts (presumably reflecting a lower content of iron oxide minerals). These samples generally produced multi-component Zijderveld plots with at least one stable overprint direction removed by the stepwise demagnetisation prior to the characteristic remanent magnetisation (ChRM) component being isolated. Samples from site NfT tended to be weakly magnetised (<1 mA/m) which added noise to the directional data, particularly at high demagnetisation levels. Samples from the clasts of the underlying Noisy conglomerate units had average NRM intensities of 1–2 mA/m and gave generally good quality Zijderveld plots while samples from the matrix (site CGM) were an order of magnitude weaker and unstable at high temperatures (Fig. 2).

Where sample data were excluded from mean directions (as some 50% were), this was generally either because the signal was too noisy for acceptable components to be adequately resolved (often associated with weak magnetisations) or because the entire sample was lightning remagnetised. Samples which had been struck by lightning recorded randomly-directed univectoral components of magnetisation some ~1–3 orders of magnitude stronger than their neighbours. All samples from sites KB4 and KB5 were affected in this way. Additionally, the ChRM isolated in some of the AF experiments clearly veered away from the origin at high (50–110 mT) fields making them unreliable. This type of behaviour is possibly caused by gyroremanent magnetization (GRM) acquisition (Fig. 2G; Dankers and Zijderveld, 1981). Recorded ChRM and overprint components generally had a maximum angular deviation of less than 15° (see Tables 1 and 2 for exceptions). All ChRM components except for four from site HG8 (origin-excluded fit) and one from site L5 (great circle fit) were calculated using an origin-anchored fit.

Rock magnetic and unblocking temperature spectra analyses described in the supplementary information suggested that all analysed units retained their NRMs in pseudo-single domain grains

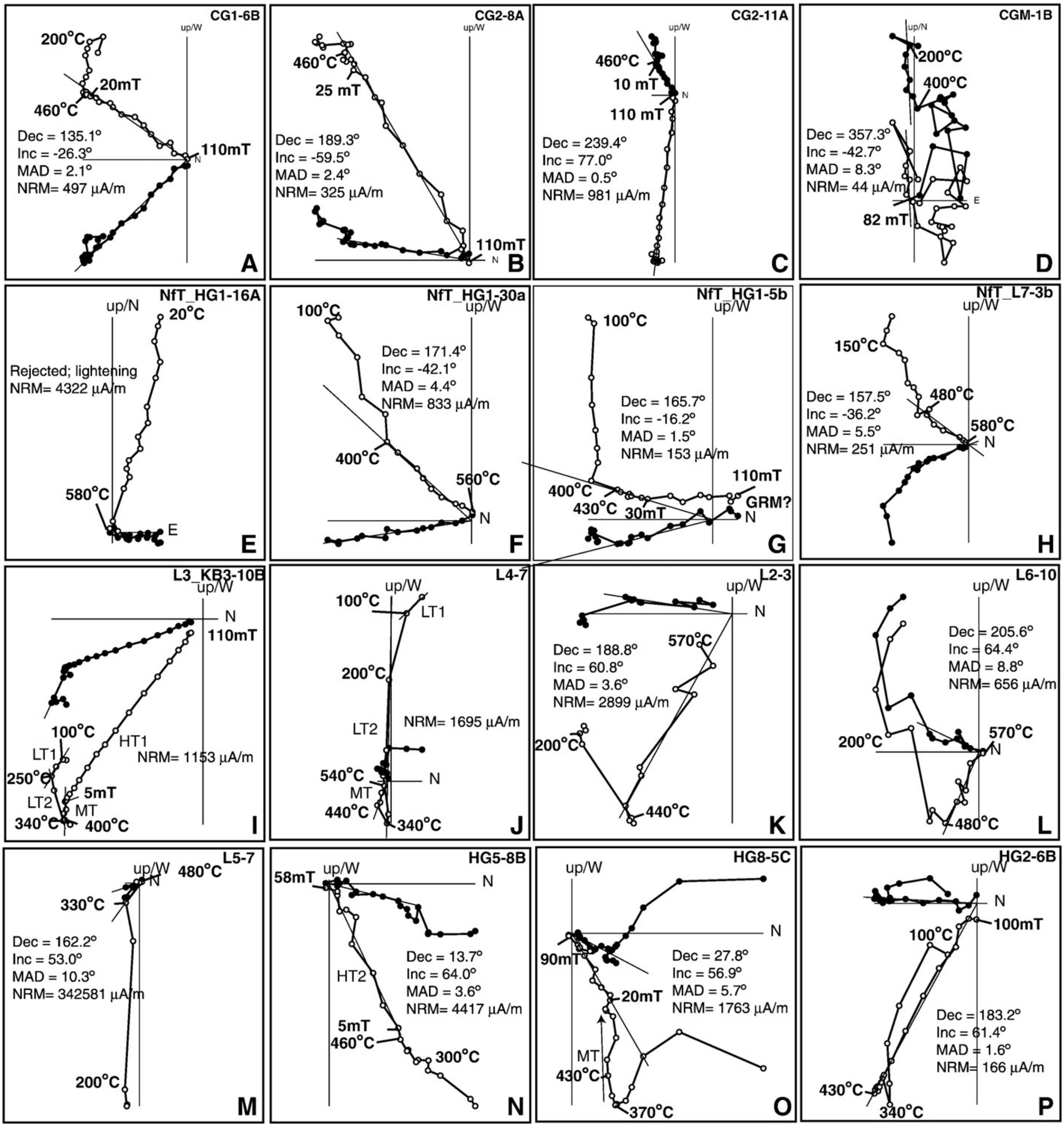

Fig. 2. Examples of Zijderveld plots produced from the Noisy formation conglomerate (A–D); the Noisy formation tuff (E–H); the Kromberg Formation meta-basalts (I–M) and the Hooggenoeg Formation meta-basalts (N–P). Filled (unfilled) points represent the horizontal (vertical) component. Temperature and/or AF demagnetisation field steps are indicated. All plots are in geographic coordinates.

(or a mixture of multidomain and single domain grains) of magnetite or Ti-poor titanomagnetite.

3.1.1. Description of overprint components

The overprint components in the sites HG2, HG5, and HG7 were too scattered between specimens to produce meaningful mean directions. For the other sites, overprint components were clustered in three dominant directions (Fig. 3A, B and Table 1). A few samples from sites KB3 and L4 featured all three of these components as overprints (Fig. 2I,J). In this section and the one that follows, all directions are presented in geographic coordinates except where explicitly stated as tectonically corrected.

Component LT1, observed in six sites (mean direction, declination/inclination (D/I) = 350.7°/−60.6°, α_{95} = 9.6°), was generally removed by thermal demagnetisation to 250 °C though it persisted up to 400 °C in samples from site HG9 and up to 460 °C in sites CGM and CG1 respectively (see Table 1). The LT1 direction produces a pole (in geographic coordinates) that plots between the present-day field (PDF) and geocentric axial dipole (GAD) poles (Fig. 4A) and therefore very likely records a recent field direction.

Table 3
Mean directions and poles used in this study (where two are given, the upper is in geographic coordinates and the lower in stratigraphic coordinates). ΔDec and ΔInc are the 95% uncertainty limits for the mean declination and inclination derived using the distribution of VGPs. Pal-lat is the palaeolatitude associated with the mean direction (assuming a GAD field) and λ and ϕ are the latitude and longitude of the associated (north) poles. A_{95}, dp, and dm are 95% confidence parameters for the mean pole.

Component	Estimated age (Ma)	Dec (°)	Inc (°)	k	α95 (°)	ΔDec (°)	ΔInc (°)	N	Pal-lat (°)	λ (°)	ϕ (°)	A95 (°)	dp	dm	Ref.	Age ref.
HT1[a]	<3230	169.8	58.1	121.9	6.1	9.7	6.9	6	38.8	−74.6	62.4	7.5	6.6	9.0	This study	9
		44.1	−22.9	16.8	16.8	16.9	29.3		−11.9	46.2	110.8	16.5	9.5	17.8		
HT2[a,b]	3472–3455	36.9	62.9	37.3	15.3	31.9	18.0	4	44.4	12.0	57.0	22.2	18.9	24.0	This study	This study, 6
		16.5	−9.2	14.6	24.9	16.1	31.5		−46	63.4	70.3	16.1	12.7	25.1		
L4 ChRM[a,b]	?	177.3	10.9	64.6	7.6	6.1	11.8	7	5.5	−69.3	203.4	6.1	3.9	7.7	This study	
		84.9	−55.4	64.6	7.6	12.1	9.6		35.9	18.8	152.6	9.8	7.7	10.8		
NfT ChRM[a,b]	~3455	167.0	−33.8	25.1	5.4	5.1	7.4	30[d]	−18.5	−43.7	193.8	4.8	3.1	5.5	This study	This study
		185.3	−45.4	25.1	5.4	6.4	7.1		−26.9	−36.8	216.9	5.7	4.4	6.9		
LT1[c]	~0–250	350.7	−60.6	50.0	9.6	16.1	10.3	6	−41.6	72.6	234.9	12.0	11.1	14.6	This study	This study
		282.3	61.5	56.9	9.0	15.0	9.2		42.7	−9.0	344.3	11.0	10.7	13.9		
LT2[c]	1800–2054	162.8	−66.0	18.8	14.3	35.0	16.4	7	−48.9	−13.6	199.5	22.2	19.3	23.5	This study	7
		109.8	64.2	11.9	18.3	41.5	21.5		45.9	0.1	14.8	27.4	15.2	19.0		
MT[c]	~2.7 Ga ?	321.8	85.6	50.4	13.1	99.9	13.0	4	81.2	−19.0	25.3	25.6	25.7	26.0	This study	
		18.9	4.0	17.6	22.5	21.5	42.8		2.0	56.6	67.0	21.5	11.3	22.5		
NfD-U ChRM[b]	~3455	267.4	44.5	16.4	31.5	39.7	44.1	3	26.2	−13.3	323.8	34.9	25.0	39.6	1	This study
		0.8	9	16.4	31.5	23.7	46.4		4.5	59.5	32.5	23.6	16.0	31.8		
NfD-T ChRM[b]	~0–250?	9.5	−48.6	128.2	22.2	–	–	2[d]	−29.6	80.8	146.4	–	19.2	29.3	2	This study
		238.7	−44.5	128.2	22.2	–	–		−26.2	−13.1	262.9	–	17.6	28		
KOM ChRM[b]	3472–3482	272.3	72.6	26.3	12.0	38.4	12.2	7	57.9	−20.7	356.4	19.3	18.9	21.3	3	6
		22.4	14.8	16.9	15.1	13.0	24.6		7.5	50.0	66.9	12.9	7.9	15.5		
MB1	3458–3471	336.7	−50.5	19.8	6.5	8.6	8.2	26[d]	−31.2	67.4	358.7	7.4	5.4	8.1	4	8,9
MB2		333.9	−42.9	21.2	6.1	6.9	8.1	28[d]	−25.0	65.7	15.9	6.2	4.4	7.1		
MB3		327.1	−34.6	16.3	6.8	6.7	9.6	29[d]	−19.0	59.1	27.9	6.3	4.2	7.4		
MB4		333.1	−34.9	17.8	5.2	5.5	7.8	44[d]	−19.2	64.7	29.4	5.2	3.3	5.8		
MB5		328.8	−40.4	48.4	2.7	3.1	3.9	57[d]	−23.1	61.1	20.2	2.8	2.0	3.3		
MB6		319.9	−26.0	32.5	3.7	3.1	5.2	48[d]	−13.7	51.1	34.1	3.0	2.0	3.8		
MB7		314.1	−19.6	25.9	5.5	4.7	8.4	28[d]	−10.1	44.6	36.5	4.6	2.8	5.4		
DUF	3464–3471	335.5	31.8	30.6	8.4	7.6	11.6	11	17.2	43.9	86.3	7.3	5.3	9.4	5	8,9

1 = Usui et al. (2009); 2 = Tarduno et al. (2010); 3 = Yoshihara and Hamano (2004); 4 = Suganuma et al. (2006); 5 = McElhinny and Senanayake (1980); 6 = Armstrong et al. (1990); 7 = Scoates and Friedman (2008); 8 = McNaughton et al. (1993); 9 = "Thorpe et al. (1992).
[a] Age of component discussed in Section 3.1.2.
[b] Age of component discussed in Section 4.2.
[c] Age of component discussed in Section 3.1.1.
[d] Denotes cases where N refers to the number of samples rather than the number of sites.

Component LT2 was somewhat scattered but directed broadly to the south and moderate-steep upwards (D/I = 162.8°/−66.0°, α_{95} = 14.3°); it was demagnetised dominantly between 100 °C and 440 °C in seven sites (five of them from the Kromberg Formation). In site NfT, it was found that the overprints, though having a mean direction consistent with LT2, were smeared out between the ChRM direction and the present day field (Fig. 5A). The LT2 mean direction produces a pole (λ = −13.6°, ϕ = 199.5°, A_{95} = 22.2°) which is well away from those from the Mesozoic and younger. This pole does, however, overlap with some poles recently (Letts et al., 2009) produced from rocks of the 2054.4 ± 1.3 Ma Bushveld Complex (Fig. 4A). This ~0.5–1 × 10^6 km^3 igneous intrusion is located approximately 200 km to the northwest of the BGB, with large satellite sills that intrude rocks flanking the northwest margin of the BGB. It is thought to have been associated with a regionally widespread thermal event affecting the area at this time. The youngest Ar–Ar and Sr–Rb ages in both sediments and plutonic rocks from the BGB record this date (de Ronde and de Wit, 1994; de Wit et al., 1992; Weiss and Wasserburg, 1987) also providing evidence for partial thermal resetting. The presence of the LT2 component in seven of the sites analysed here could suggest that this event reset the NRM of these rocks up to unblocking temperatures of between 340 and 440 °C. It would then be very encouraging that the most recent and intense thermal event recorded in the isotopic systems of BGB rocks is manifested in their palaeomagnetic signals as an overprint., Most of the palaeomagnetic directions recorded by the Bushveld Complex itself, however, are of the opposite polarity to LT2 which would imply that this overprint was acquired during a relatively short period of time associated with the emplacement of the Upper Zone of the complex (Letts et al., 2009). An alternative explanation is that the overprint is related to slightly later intraplate magmatism observed in the Waterberg and Soutpansberg Groups to the north of the Bushveld Complex at 1.8–1.9 Ga. Fig. 4 shows that the (dual polarity) pole produced by these intrusions (Gose et al., 2006; Hanson et al., 2004) lies very close to that produced by LT2. Either of these alternatives would imply that components of NRM recorded at higher unblocking temperatures than LT2 are likely older than 1.8–2 Ga.

Component MT was directed near vertical downwards (D/I = 321.8°/85.6°, α_{95} = 13.1°) and observed between 370 °C and 480 °C in sites L3 and L4, between 370 °C and 430 °C + 35 mT in site HG8 and as the dominant component between 100 °C and 330 °C in site L5 (Fig. 2M). Since individual samples from sites L3 and L4 record MT at higher unblocking temperatures than LT2 (Fig. 2I,J), it is likely that it represents a magnetisation acquired prior to 1.8 Ga. Comparing the mean pole (λ = −19.0°, ϕ = 25.3°, A_{95} = 25.6°) with others older than this age from the Kaapvaal Craton, it best agrees with those from the late Archaean: the Westonara Basalt (Ventersdorp Supergroup, part of a regionally widespread large igneous province; Strik et al., 2007) dated at 2714 ± 8 Ma (Armstrong et al., 1991) and the Modipe Gabbro intrusion (constrained only to be older than the adjacent Gabarone Complex at 2783 ± 2 Ma; Grobler and Walraven, 1993) and the 2782 ± 5 Ma Deerdepoort Basalt (Wingate, 1998). One possible scenario is that the MT component was acquired during a thermal event at ~2.7 Ga (de Ronde et al., 1991; Toulkeridis et al., 1998) and that the signal of this was later entirely overprinted in most rocks by the later thermal event responsible for imparting the LT2 component.

Overprint components isolated in samples from site CG2 persisted up to high temperatures (550 °C in one specimen) and were reasonably consistent between specimens taken from the same clasts

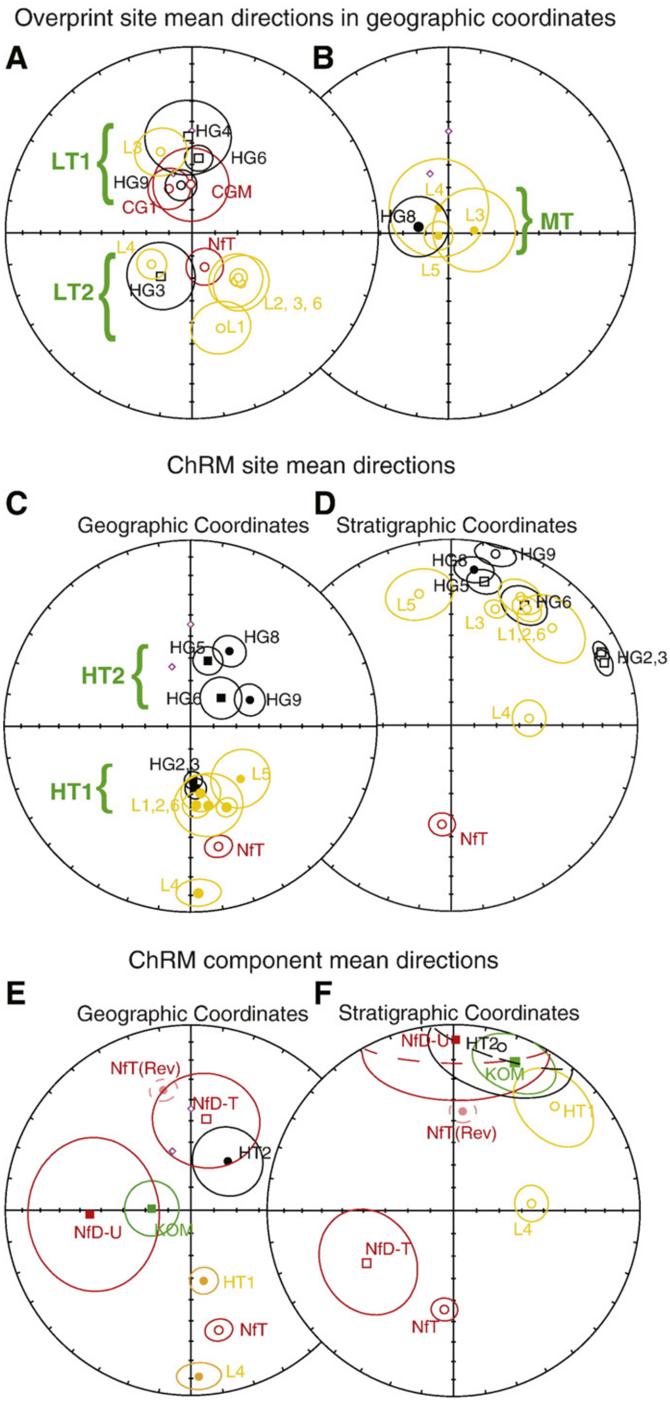

Fig. 3. Equal area plots showing site mean directions (A–D) and overall averages (E, F) of NRM components with filled (unfilled) points representing the lower (upper) hemisphere and ellipses indicating α_{95} cones of confidence. Orange, red, black, green points represent results from the Kromberg, Noisy, Hooggenoeg, Komati formations respectively. Tables 1, 2, and 3 provide the actual data shown in these six plots.

(Fig. 5F). However, from clast to clast, these overprint directions were very scattered (though clearly not random).

3.1.2. Description of ChRM components

In sites CG1 and CG2, ChRM components (all with MAD<5°) were isolated after thermal demagnetisation to 460 °C and then subsequent AF demagnetisation to between 0 and 35 mT (dominantly 10 mT). In the two cases where pure AF demagnetisation was successfully applied, the peak demagnetisation steps required to isolate the ChRM were 60–70 mT. The directions of the ChRM components at both sites were random (see Section 4.1).

High temperature ChRM components were isolated and gave consistent directions in six sites each from the Kromberg and Hooggenoeg Formations (Table 2, Fig. 3C). Eleven of these twelve ChRM site mean directions fell into two directional clusters in geographic coordinates (Fig. 3C). The first component is labelled HT1, has a southerly moderately-steep down direction (D/I = 169.8°/ 58.1°, α_{95} = 6.1°), and was observed in samples from five Kromberg sites (on the southeastern limb of the Onverwacht Fold) and two Hooggenoeg sites (close to the fold axis, Fig. 1A). This component was generally isolated above 340 °C. The mean ChRM direction in samples from site L5 was slightly offset from the other HT1 site means and these samples also had an anomalously low unblocking temperature spectrum (Fig. 2M). This site mean was therefore excluded from the overall HT1 mean calculation. The six site mean directions comprising HT1 clearly group far better in geographic than in stratigraphic coordinates (k decreases from 121.9 to 16.8 upon application of the tectonic correction; Fig. 3C and D). This suggests a post-folding (i.e. younger than 3.23 Ga) age for the magnetisation and, in fact, the antipode of the pole in geographic coordinates (λ = 74.6°, ϕ = 242.4°, A_{95} = 7.5°) is consistent with a Mesozoic–Cenozoic age of magnetisation. One potential cause of Mesozoic age remagnetisation is the Karoo large igneous province at ca. 180 Ma which produced widespread partial remagnetisation in other parts of the Kaapvaal Craton (Strik et al., 2007). If this component is indeed relatively recent then it requires that a chemical remagnetisation process occurred that did not affect components of magnetisation (LT2, MT) recording a much more ancient field at lower unblocking temperatures. Another possibility is that HT1 is both post-folding and older than LT2 and MT, and that its pole's proximity to recent ones is purely coincidental. These two conditions would be met if it were recording the 3.1 Ga thermal event reported by Toulkeridis et al. (1994), or the massive igneous events in the BGB region between 2.9 and 3.0 Ga (de Wit, in press; Schoene et al., 2008).

The second cluster of ChRM directions, named HT2, is directed northeast and moderately-steep down (D/I = 36.9°/62.9°, α_{95} = 15.3°) and was observed in samples from four Hooggenoeg sites at higher unblocking temperatures (>460 °C) than HT1. Although the mean HT2 direction is moderately close to the MT direction (see Table 3 and Fig. 3), it is statistically distinguishable using a Monte Carlo common true mean direction test (McFadden and Jones, 1981) (angular distance = 26.3°, critical distance = 17.5°, P = 0.008) and is therefore considered distinct. The mean pole in geographic coordinates is close to the Bushveld Complex poles but also distinguishable from their mean pole (ΔP = 26.2°, ΔPc = 21.9°, P = 0.017, where ΔP refers to great circle rather than angular distance).

Four site mean directions are too few to allow a reliable fold test to be made for component HT2. Interestingly, although these site mean directions become less well clustered on application of the tectonic correction (k decreases from 37.3 to 14.6, Fig. 3D, Table 3), their virtual geomagnetic poles (VGPs) become more clustered (K increases from 18.1 to 33.7). This difference is caused by the increase in site-pole distance when moving from geographic to stratigraphic coordinates. This process tends to decrease the amount of scatter in the poles for a given distribution of directions (or conversely, to increase the scatter of directions for a given pole distribution). This disagreement introduces some ambiguity into the results of the fold test and consequently, it is not possible to determine whether HT2 was acquired pre- or post-folding.

Site L4 produced somewhat scattered ChRM directions although the higher quality components were well-clustered and the scattering was a product of the weaker samples. Excluding the results of those six samples (46% of total), the mean ChRM direction is distinct from all of those previously discussed: south and shallow down (D/I = 177.3°/

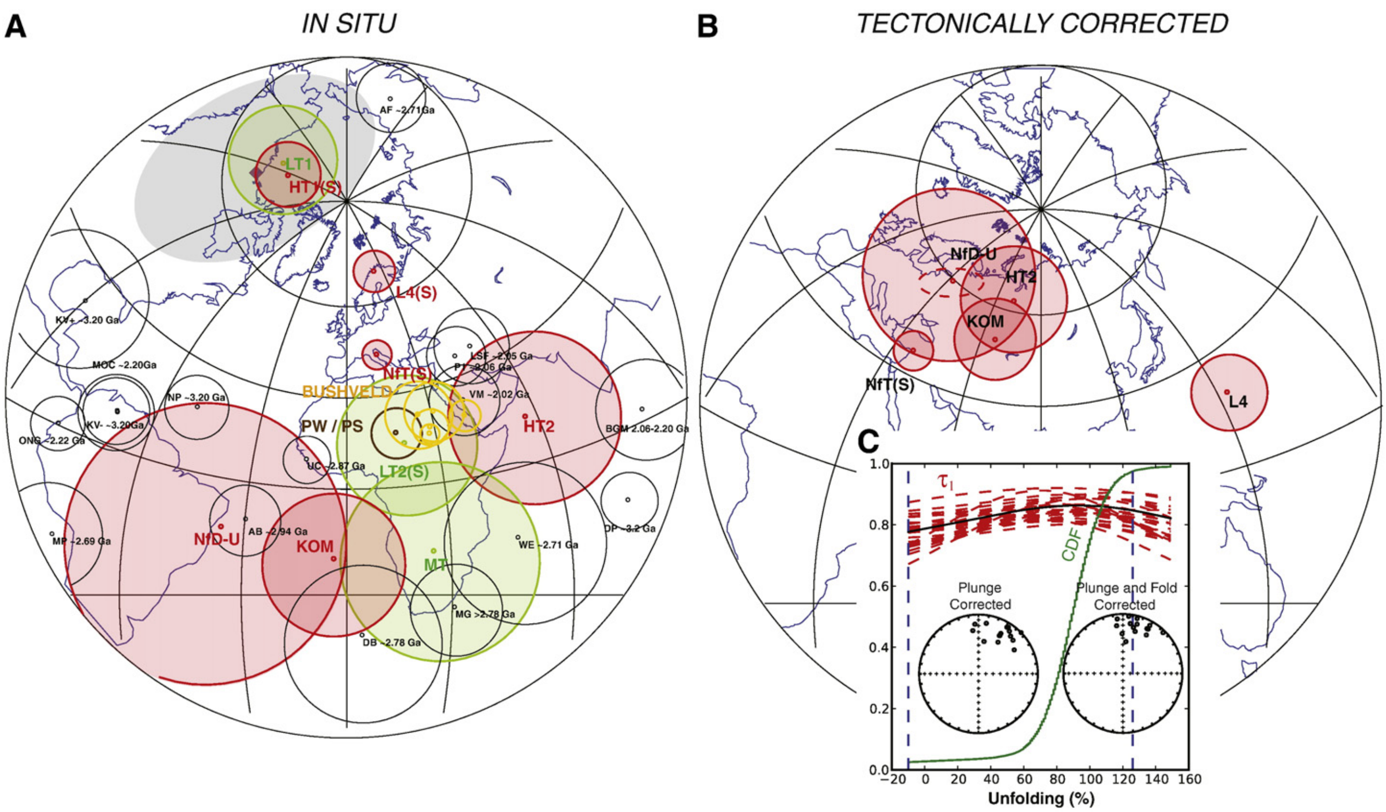

Fig. 4. (A) Average poles produced from studies of the Onverwacht Group (see Table 3 for data (Shaded circles, red = ChRM components, green = low and medium temperature components; all these poles are north unless marked with an S). Also shown for comparison as unfilled circles are other published poles older than 2 Ga from the Kaapvaal Craton (black except for the Bushveld Complex poles (Letts et al., 2009) in thick orange and those from the post-Waterberg/post-Soutpansberg mafic intrusions (Gose et al., 2006; Hanson et al., 2004) in thick brown). The shaded area shows the approximate extent of Mesozoic and younger poles from Africa. The purple diamond is the pole obtained from the present-day field (PDF) direction at the Barberton Greenstone Belt. AF = Allanridge Fm (de Kock et al., 2009); BGM = Basal Gamagara/Mapedi (Evans et al., 2002); DB = Deerdepoort Basalt (Wingate, 1998); DP = Dalmein Pluton (Tarduno et al., 2007); KV+/− Kaapvaal Pluton (Tarduno et al., 2007); LSF = Lower Swaershoek Fm (de Kock et al., 2006); MG = Modipe Gabbro (Evans and Mcelhinn, 1966); MOC = Mamatwan Ore Complex (Evans et al., 2001); MP = Mbabane Pluton (Layer et al., 1988, 1989); NP = Nelshoogte Pluton (Layer et al., 1998); ONG = Ongeluk lava (Evans et al., 1997); P1 = Phalaborwa 1 (Morgan and Briden, 1981); UC = Usushwana Complex (Layer et al., 1998); VM = Vredefort Mean (Salminen et al., 2009); WE = Westonaria Basalt (Strik et al., 2007). (B) The poles associated with the five magnetic components derived from Onverwacht rocks (those that are the least likely to be much later remagnetisations — see text) shown after tectonic correction. Note that four of them cluster far better than in geographic coordinates suggesting a pre-folding (older than 3.23 Ga) age for these. The dashed ellipse shows the allowed position for the mean pole of NfD-U allowing for ±10° uncertainties in the bedding plane strike and dip at this location. (C) Results of a bootstrap foldtest (Tauxe and Watson, 1994) applied to the individual site mean directions constituting components HT2, NfT, KOM and NfD-U (see Table 3). τ_1 is the major eigenvalue of the orientation matrix shown as a function of the unfolding (red lines are examples of bootstraps and the black line is the average of all 10,000; blue dashed lines indicate 95% confidence limits for the maximum of τ_1). The green line is the cumulative density function for the maxima in τ_1 for the bootstraps. The equal area plots show directions before and after unfolding with open (closed) points indicating upper (lower) hemispheres. Panels (A) and (B) were produced using GMAP software (Torsvik and Smethurst, 1999); plot (C) was produced using the pmagpy-2.49 software provided by L. Tauxe.

10.9°, α_{95} = 7.6°). Its pole in geographic coordinates is distinct from any of the other new data and other published poles of Archaean or Palaeoproterozoic age from the Kaapvaal Craton (Fig. 4A). It is also reasonably well-removed from Mesozoic and younger poles from Africa.

Since the smear of NfT overprint directions overlaps slightly with the distribution of NfT ChRM directions (Fig. 5A), it is not possible to be entirely sure that the latter has been fully isolated in every case. Fig. 5B shows that the group of samples which were most dominated by the overprint component (i.e. those in which the ratio of the ChRM to the overprint magnitude was below 0.5) tended to have higher inclinations than the rest. This suggests that, in this group of samples, the ChRM may not have been fully isolated. However, we point out that some of the other samples shown in Fig. 5B that had much smaller relative overprint components also had higher inclinations and that the regression shown in Fig. 5B is not statistically significant. We nonetheless remove those samples with ChRM/overprint ratios <0.5 before calculating the mean. No discernable difference from the overall mean direction (D = 167.0°, I = −33.8°, α_{95} = 5.4°) is found if the directions from specimens from the same core samples are first averaged before combining into the overall mean, (D = 165.8°, I = −35.2°, α_{95} = 5.8°). Like L4, its pole is distinct from any of the others shown in Fig. 4A.

3.2. U–Pb age determination

U–Pb dating of magmatic zircons from the NfT site was done by LA–MC–ICP–MS, using AEON's High Resolution multicollector inductively coupled plasma mass spectrometer (Nu Instruments) coupled to a UP 193 solid-state laser system (New Wave Research) and a desolvation nebulizer system (DSN-100, Nu Instruments). The supplementary information contains more details about the experimental and analytical approach and tabulates the results summarised in Fig. 6. Accuracy was tested by repeat analyses of zircons from the Kaap Valley pluton — an isotopically well-characterized tonalite (Kamo and Davis, 1994; Schoene et al., 2006). The weighted average of the obtained $^{207}Pb/^{206}Pb$ dates for the Kaap Valley pluton zircons is 3225.4 ± 5.8 Ma (Fig. 6A and B), which is indistinguishable from the TIMS (thermal ionisation mass spectrometry) date of 3227 ± 1 Ma obtained by Kamo and Davis (1994).

Twenty-five analyses were performed on 20 grains. The zircons generally yielded high U concentrations. Therefore, the lattice of many

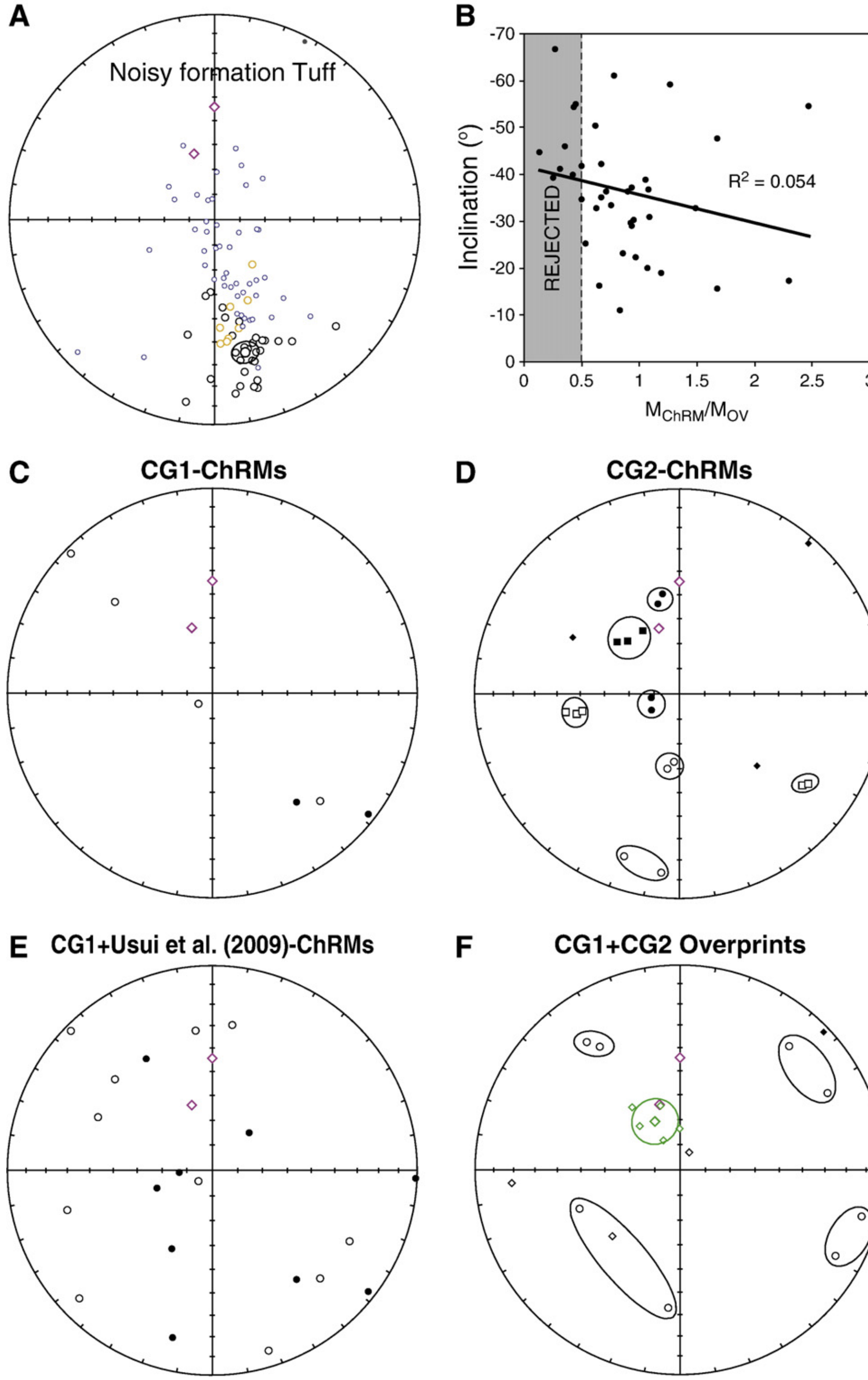

Fig. 5. Specimen-level results from the Noisy formation tuff (NfT) and conglomerate clasts (CG1 and CG2). Filled (unfilled) points indicate lower (upper) hemisphere. Purple diamonds show the present day field in the BGB and the recent GAD field. (A) NfT accepted (black) and rejected (orange) ChRM directions and NfT overprint directions (blue). (B) Plot showing ratio of magnitudes of ChRM to overprint components against inclination of the ChRM for results from site NfT. ChRM components in shaded area were rejected (see text). (C) ChRM directions for CG1. (D) ChRM directions from CG2; squares are independently oriented cores from the same clast; circles are specimens from the same core and diamonds are single specimens; specimens from the same clasts are encircled. (E) Clast mean directions from both CG1 and Usui et al., (2009) combined. (F) Overprint directions from site CG1 (green) and CG2 (black with ellipses grouping specimens from the same core sample). In all plots, the purple diamonds show the directions of GAD and the PDF.

grains suffered variable degrees of radiation damage, resulting in partial Pb loss. Eleven analyses yielded concordant dates, with a combined concordia date (Ludwig, 1998) of 3455.2 ± 7.5 Ma. (Fig. 6C). The weighted average of the $^{207}Pb/^{206}Pb$ dates of 14 analyses is 3452.7 ± 4.0 Ma (Fig. 6D). One grain (2 spots) has an older date of 3546 ± 12 Ma (concordia age ± 2σ), which is interpreted as inherited from older felsic crust of that age common along the southern and eastern margins of the Barberton Greenstone belt (e.g. Schoene et al., 2008).

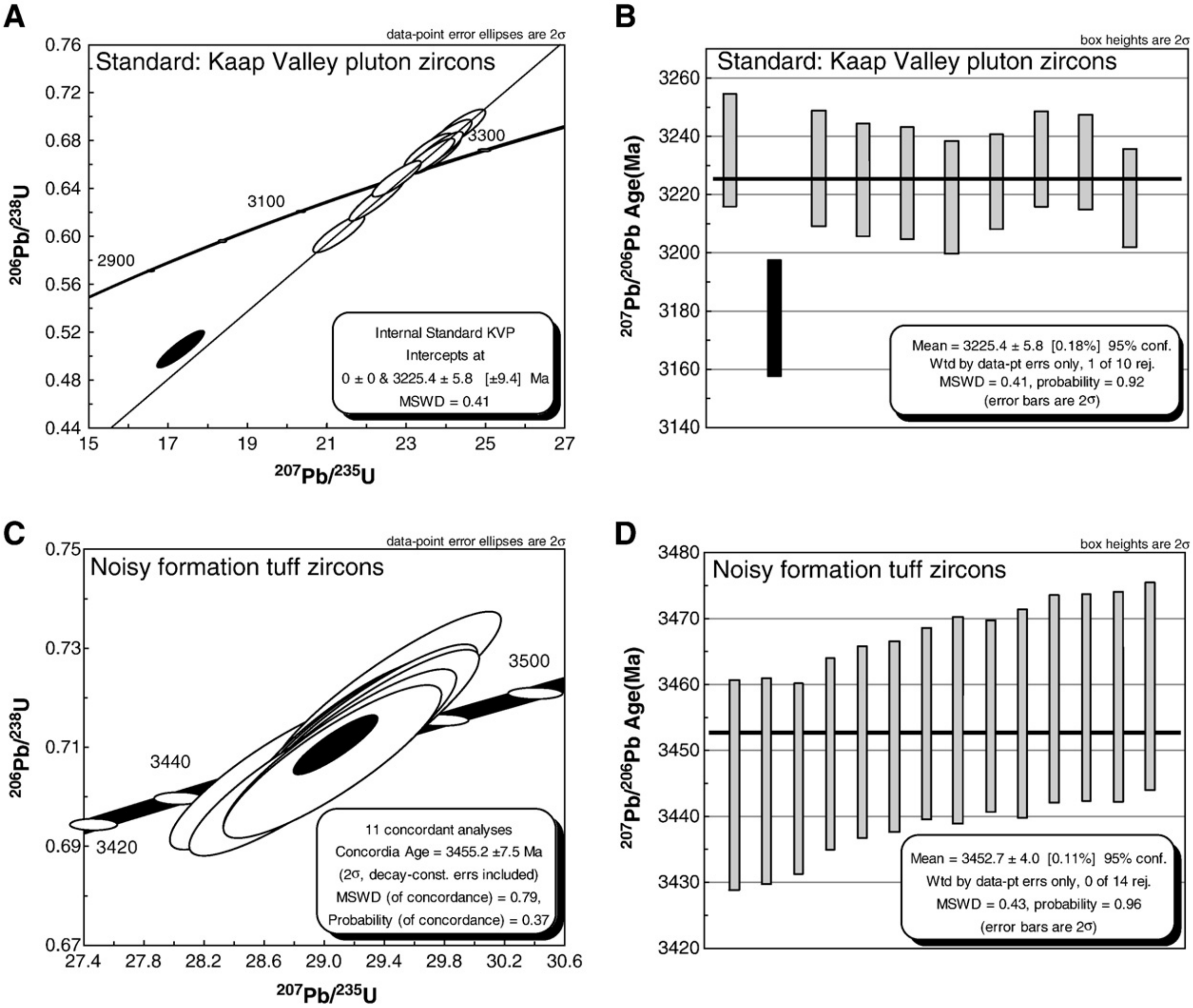

Fig. 6. Results of U–Pb LA–MC–ICP–MS dating of zircon from the Noisy formation tuff. (a, b) For comparison of accuracy, results for the internal standard (zircon from the Kaap Valley pluton); A) Upper intercept date of 3225.4 ± 9.4 Ma; B) the weighted mean of the $^{207}Pb/^{206}Pb$ dates acquired on zircons. These results are indistinguishable from the previously published TIMS date of 3227 ± 1 Ma (Kamo and Davis, 1994). (C) 11 concordant analyses of zircon from the Noisy formation tuff define a concordant date (Ludwig, 1998) of 3455.2 ± 7.5 Ma, which we interpret as the most accurate age of this crystal tuff; D) Weighted mean of 14 $^{207}Pb/^{206}Pb$ ages is 3452.7 ± 4.0 Ma.

4. Discussion

4.1. The conglomerate test

For site CG1 (Fig. 5C), each of the 6 specimens which gave ChRM directions were from separate clasts and these directions were randomly oriented at the 95% significance level according to the Watson (Watson and Beran, 1967) test ($R_{crit} = 3.85$, $R = 1.71$). In CG2 (Fig. 5D), the ChRM directions were consistent within clasts but randomly directed between clasts ($N = 10$, $R_{crit} = 5.03$, $R = 1.89$). We do not combine data from sites CG1 and CG2 as the latter of these was sampled in a loose block. However, we can combine the CG1 data with the results published by Usui et al. (2009) as both sites were from *in situ* exposure along the Komati River section. In addition to the Watson test (positive, $N = 20$, $R_{crit} = 7.17$, $R = 1.88$), we also performed the Shipunov et al. (1998) conglomerate test on this combined dataset. This more recent test compares the observed data to a known reference direction, for which we used all the components described above (i.e. LT1, LT2, MT, HT1, HT2, and the ChRMs from the NfT and L4 sites; Table 3). In every case, the test was positive which strongly suggest that some rocks of the Onverwacht Group have not been remagnetised since the deposition of the conglomerate and therefore may record a primary or near-primary remanence of Palaeoarchaean age. The new U–Pb date of 3455.2 ± 7.5 Ma for the NfT unit may be considered a minimum age for the conglomerate unit (which directly underlies it) and for the ChRM components recorded in its clasts.

Rock magnetic and microscopy analyses described in the supplementary information do not suggest that the clasts of the conglomerate differ strongly in their magnetomineralogy or thermochemical histories from any of the other units from the Onverwacht Group which retain stable high temperature ChRM components. In particular, the microscopic analyses indicate that pervasive hydrothermal alteration may well have reset the magnetisation of all of the rocks studied here (including the clasts of the conglomerate). However, the palaeomagnetic results from the conglomerate support earlier claims (de Vries and Touret, 2007; de Wit et al., 1982; Knauth and Lowe, 2003; Schoene et al., 2008; Tice et al., 2004) that this alteration occurred very early in the history of each of the units. I.E. the clasts must have been remagnetised prior to their emplacement in the sedimentary unit and likely close to the time their source units were emplaced a few Myr earlier.

The inferred secondary hydrothermal origin of the magnetic remanence in many of these rocks may be useful in ensuring that each rock unit averages geomagnetic secular variation (SV) to some extent (although this may have been reduced in the Archaean at low latitudes in any case; Biggin et al., 2008). If the rocks are recording chemical remanent magnetisations then these were likely acquired over a prolonged period of time (centuries to millennia) which may help with providing some time averaging of the field.

4.2. Comparison with previously published palaeomagnetic data from the Onverwacht Group

We will now proceed by discussing three published palaeomagnetic studies on the Onverwacht Group (Tarduno et al., 2010; Usui et al., 2009; Yoshihara and Hamano, 2004) in the light of the new findings. Two of these, whose results we will refer to as NfD-U (Usui et al., 2009) and NfD-T (Tarduno et al., 2010), were studies of potential source rocks for the Noisy formation conglomerate: dacitic intrusions also from the Noisy formation but located ~13 km away on the other (northwestern) limb of the Onverwacht antiform (Fig. 1). The third, whose results we will refer to as KOM, is the most recent palaeomagnetic study (Yoshihara and Hamano, 2004) of units from the 3.48 Ga Komati Formation.

The NfD-U (Usui et al., 2009) mean palaeomagnetic directions comprise high temperature (>550 °C) components from three sites that together make up only five samples (we did not consider results from site das4 since the authors stated that ChRM components may not have been completely isolated in that site). These three directions were not tightly clustered ($k = 16.4$) and therefore the uncertainties in the mean direction and associated pole position are large ($\alpha_{95} = 31.5$, $A_{95} = 34.9$). Nonetheless, this pole is entirely distinct from any of the poles produced by this study in geographic coordinates (Fig. 4A) although it does overlap with the Mesoarchaean (ca. 2.94 Ga) pole obtained from a study of the Agatha Basalt of the Pongola Supergroup (Strik et al., 2007) (Fig. 4A). We interpret this as coincidence because no other Onverwacht rocks record this direction as an overprint. Furthermore, other evidence presented below hints at a pre-folding (i.e.>3.23 Ga) origin for this component.

The NfD-T palaeomagnetic direction was calculated from measurements made from high temperature (>510 °C) components from just two oriented single quartz crystals. These were taken from another dacitic intrusion approximately 2 km to the west of NfD-U (Fig. 1) and were used primarily to obtain single crystal palaeointensity estimates (Tarduno et al., 2010). We calculated the mean direction of these in geographic coordinates by using the direction as published in stratigraphic coordinates and reversing the tectonic correction provided by the authors (Tarduno et al., 2010) (a plunging fold correction very similar to that outlined in Section 2). The resulting *in situ* direction plots within 95% confidence limits of both the recent GAD field and the PDF (Fig. 5E).

Usui et al. (2009) did not attempt to make a tectonic correction to the NfD-U directions. Here we apply the same plunging fold correction as for our own data. From regional field and aerial observations, we estimate the bedding in that area to be 280°/90° and conservatively estimate uncertainties to be 10° on both of these (Fig. 4B). The NfD-T and NfD-U directions and poles do not fall close to one another in geographic coordinates (Pole great circle distance, $\Delta P = 112.5°$ or 67.5° if an antipode is taken) and application of the tectonic correction to them both does little to improve this ($\Delta P = 120.7°$ or 59.3°). Since the NfD-T direction is close to our LT1 directions, we suspect that it may be a recent overprint. If this were true, it would cast doubt onto the associated palaeointensity estimate (Tarduno et al., 2010) but would not necessarily invalidate it. Their direction is based on just two of the twelve single crystal specimens that were used to produce the palaeointensity. It would, however, invalidate the palaeolatitude value used to calculate the virtual dipole moment (VDM). The NfD-U direction after tectonic correction produces a mean palaeolatitude of 4.5° instead of 29.6° obtained from the NfD-T direction. If this new value is used in place of the old, it has the effect of changing the mean calculated VDM from $5.8 \pm 0.9 \times 10^{22}$ Am^2 to $7.2 \pm 1.1 \times 10^{22}$ Am^2.

The KOM results (Table 3) generate a pole that, in geographic coordinates, falls between those of NfD-U and MT (Fig. 4A). The KOM direction in stratigraphic coordinates was recalculated for this study (Table 3) using the same plunging fold correction as applied to the Hooggenoeg and Noisy data (Section 2) and the bedding planes provided by Yoshihara and Hamano (2004). This was necessary because these directions were inappropriately transformed using a simple tilt correction in the original publication. The individual site mean data cluster slightly better in geographic than in stratigraphic coordinates (k falls from 26.3 to 16.9; Table 3) and the fold test of Tauxe and Watson (1994), although very poorly constrained, supports a post-folding origin (the optimal clustering occurs with 95% bounds between −63% to 75% of unfolding). Nonetheless, the clustering of the VGPs improves on application of the tectonic correction (K increases from 10.8 to 22.8) and Yoshihara and Hamano (2004)) gave several arguments as to why they thought the remanence was likely to be near-primary (pointing out that their high temperature component, being more than 50% contained in the temperature range of 570–590 °C, resides well above peak temperatures of metamorphism having affected this part of the BGB). Below, we present some evidence that the Komati Formation rocks do indeed record remanences that are likely to be pre-folding (>3.23 Ga) in age.

For reasons given in Section 3 and above, we consider components LT1, LT2, MT, and HT1 as well as the ChRM of NfD-T as likely to be remagnetisations acquired subsequent to the folding of the BGB at 3.23 Ga. Fig. 4 compares the poles associated with the remaining components (HT2, NfT ChRM, and L4 ChRM) with the poles derived from the two remaining published studies (NfD-U and KOM) discussed above. These six poles are derived from rocks formed in a period of time spanning 20–30 Myr so some similarity can be reasonably expected once the tectonic correction has been applied. In Fig. 4B, it can be seen that four of these (HT2, KOM, NfD-U, and the antipode of NfT ChRM) cluster together far better in stratigraphic ($K = 24.3$) than in geographic ($K = 3.0$) coordinates, supporting a pre-folding origin. However, this encouraging result is more convincing in pole-space than in direction space (k only increases from 5.3 to 11.0 on application of the tectonic correction). This can be seen in particular for the NfT (reversed) and NfD-U ChRM components which, in stratigraphic coordinates, do not overlap in direction space (Fig. 3F) but which do in pole-space (Fig. 4B). These differences are again due to changes in the site-pole distance between geographic and stratigraphic coordinates as explained with reference to the HT2 directions in Section 3.1.2. The overall result does not have the backing of a formal positive fold test when all the site mean directional data are included. A bootstrap fold test does clearly favour a better grouping in stratigraphic coordinates (with the maxima of the eigenvalue being most commonly found when unfolding is in the range of 60% to 120%; Fig. 4C) but the limitations of the present dataset preclude a statistically significant outcome. Note finally that the convergence is not due to differential plunge corrections applied to the data; the same rotation in step 1 ($\alpha = 92°$, Section 2) was applied to all of these data.

The caveats outlined above notwithstanding, on the basis of their improved consistency after tectonic correction, the positive conglomerate test, and the absence of any clear indications of their remagnetisation resulting from comparison with younger poles, we argue that the units recording the components HT2, NfT ChRM, NfD-U, and KOM are the most likely of any rocks of the Onverwacht Group, or indeed any rocks of Palaeoarchaean age to record primary or near-primary palaeomagnetic components. The pole derived from the site L4 does not cluster with the others in stratigraphic coordinates and is therefore considered less reliable.

4.3. Potential implications for the early Earth

In the discussion to follow, we are implicitly assuming that the geocentric axial dipole (GAD) hypothesis holds true for the whole of the Archaean. Palaeomagnetic and palaeoclimatic data have been used to argue for its validity through most of the last 2 Gyr (Evans, 2006) and there appears to be no theoretical reason to suspect that it would not hold for earlier periods too [see e.g. Aubert et al., 2009; Roberts and Glatzmaier, 2001]. Therefore, in the absence of evidence to the contrary, this seems a reasonable approach to take.

The palaeolatitudes between 3.45 and 3.49 Ga for the BGB from the four most reliable results identified in Section 4.2 are within error (albeit in a large range: −22° to +36° once uncertainties are allowed for; Fig. 7A and B). Net latitudinal drift of the rocks over this period therefore appears to have been limited. While the large uncertainties preclude any detailed discussion on drift rates, this result would seem to suggest modest latitudinal drift rates or a tendency for oscillatory motion (caused by true polar wander, for example) to constrain the crustal block to low latitudes. Taking the BGB data shown in Fig. 7A at face value, the most rapid latitudinal drift required is that between the lower bound of palaeolatitude given by NfT(Rev) (21.6°) and the upper bound of palaeolatitude given by HT2 (11.5°). Given the average 9 Myr difference in age estimates between the HT2 and NfT units, this ~1100 km motion translates to a latitudinal velocity of ca. 12 cm/year. This is fast by today's standards but well within the range of plate velocities observed in the Phanerozoic (Klootwijk et al., 1992) and only consistent with the lowest limits of drift rates suggested previously for the Palaeoarchaean by Suganuma et al. (2006).

Published plaeomagnetic results from rocks of the same age as the BGB data from the Pilbara Craton in Western Australia are also plotted in Fig. 7A and (with polarities reversed) in Fig. 7B. On the basis of their

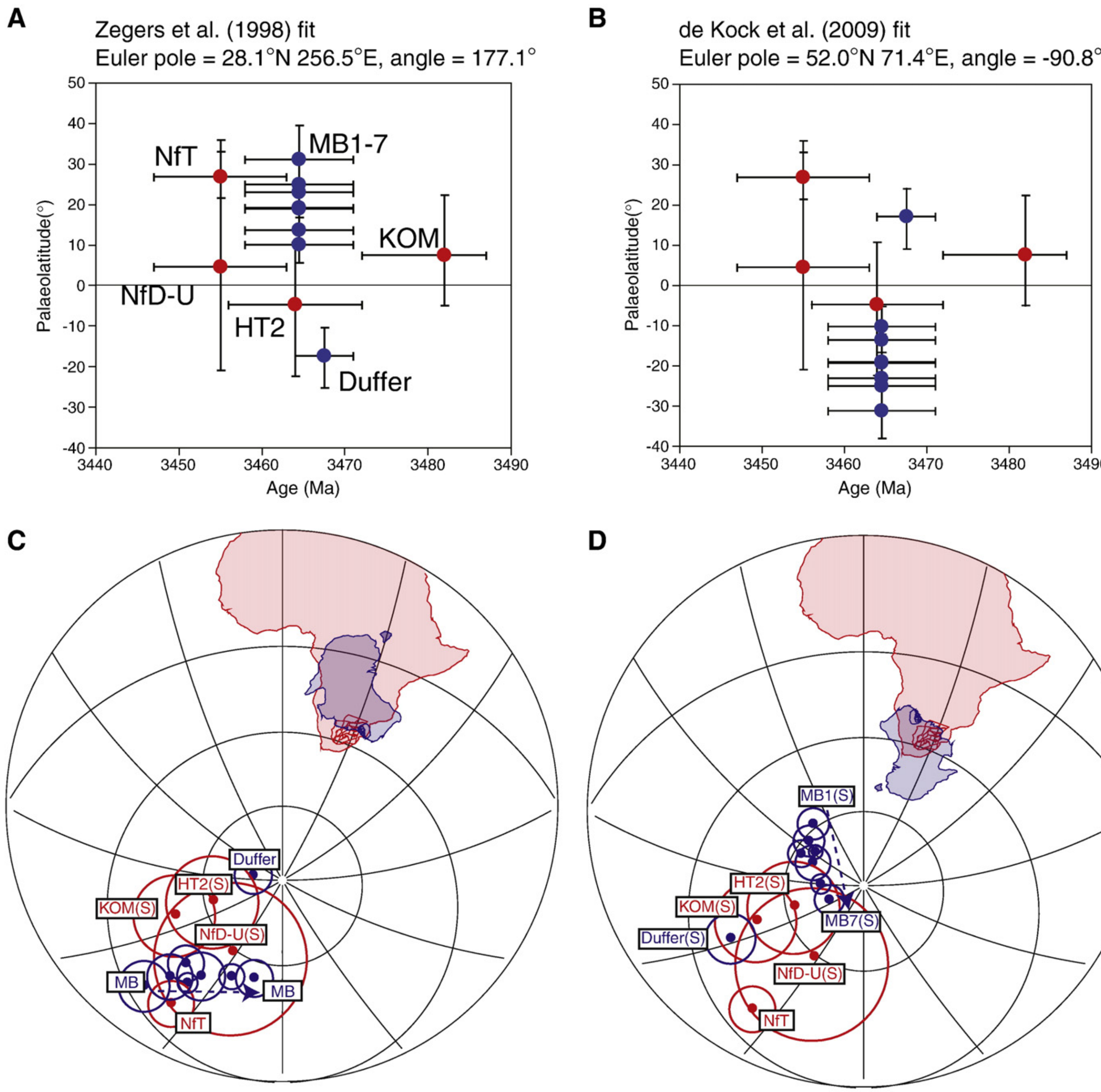

Fig. 7. Tests of two previously proposed configurations of the Vaalbara supercraton (de Kock et al., 2009; Zegers et al., 1998). The directions/poles used are given in Table 3 and discussed in the text. (A,B) comparisons of absolute palaeolatitudes (calculated from mean magnetic inclinations assuming a GAD field) from BGB (red) and Pilbara (blue) rocks of Palaeoarchaean age. (C, D) Comparisons of poles (north unless marked with an S) made in the present-day Africa frame of reference after the Euler rotations given were applied to the Pilbara (blue) poles. The outlines of the present day continents of Africa and Australia and the Kaapvaal (red) and Pilbara (blue) cratons are given for reference. Panels (C) and (D) were produced using GMAP software (Torsvik and Smethurst, 1999).

similar volcano-sedimentary stratigraphy (Button, 1976; Cheney, 1996) as well as correlations in their structural (Zegers et al., 1998) and environmental (de Kock et al., 2009) histories, it has been hypothesised that parts of the Kaapvaal and Pilbara cratons were conjoined in an supercraton referred to as Vaalbara for at least a part of their history prior to 1.8 Ga. An alternative hypothesis is that the two cratons were not proximal but were rather affected simultaneously by global-scale processes (Nelson et al., 1999).

Two studies are available from rocks of the Pilbara Craton from the time range discussed here. The first involved dacites and pillow basalts of the Duffer Formation (DUF; 3470 ± 6 Ma; McElhinny and Senanayake, 1980; McNaughton et al., 1993; Thorpe et al., 1992). It produced a paleomagnetic pole that was associated with a positive fold test which seems to constrain the age of magnetisation to older than 2.8 Ga. The second was produced from the slightly younger (3456–3476 Ma; McNaughton et al., 1993; Thorpe et al., 1992) Marble Bar Chert (MBC) using samples taken from a 23.5 m long section of an oriented drill core (Suganuma et al., 2006). The ChRM directions recovered from these samples were clustered into seven groups defining a quasi-continuous apparent polar wander path. Uncertainties in the duration of time covered by the sampled section allowed for a large range of lateral drift rate estimates from moderately rapid to exceptionally fast (12–673 cm/yr).

As discussed in detail elsewhere (Usui et al., 2009), the reliability of the data emerging from both of these studies is far from certain but there are no grounds for their outright dismissal and a comparison with the BGB data is warranted. The Pilbara results produce palaeolatitude estimates for the Pilbara that are within error of those derived for the BGB (there is greater overlap if we take the antipodes of the Pilbara directions; as we do in Fig. 7A but not Fig. 7B). The palaeolatitudes associated with these magnetic components would agree better if DUF were assumed to have been acquired in a period of reverse polarity with respect to the MBC components. However, this would then require a near-180° rotation of the Pilbara between the two magnetisations being acquired which seems less likely than 25–30° of latitudinal drift. The generally low and within-error palaeolatitudes inferred from all of these units leaves open the possibility of the Vaalbara hypothesis being valid during a significantly (600–700 Myr) earlier time period than that for which it has been previously tested with palaeomagnetic data (de Kock et al., 2006; Strik et al., 2003; Zegers et al., 1998).

Two configurations of Vaalbara (de Kock et al., 2009; Zegers et al., 1998) have been proposed using late Archaean palaeomagnetic data as the primary constraint and both of these are tested for ca. 3.45–3.48 Ga in Fig. 7C and D. The area spanned by poles from both cratons is large and extended further by the polarity ambiguity so that the predictive capacity of this test is poor. Nonetheless, the degree of overlap between the Kaapvaal and Pilbara Palaeoarchaen poles for the Zegers et al. reconstruction, in particular, is striking. These comparisons in no way establish the existence of a proto-Vaalbara supercraton in the Palaeoarchaean in either of these configurations but they do appear to leave open the possibility that it existed in some form.

Leaving aside the issue of Vaalbara, the fact that any apparently meaningful palaeomagnetic results have been obtained from the rocks discussed here suggests that the Earth had a stable and likely global magnetic field at 3.5 Ga and that the intensity of this at the Earth's surface was not orders of magnitude weaker than today's. This in turn further strengthens the case, put forward previously (McElhinny and Senanayake, 1980; Tarduno et al., 2010; Usui et al., 2009), that the conditions in the Earth's core were such that there was sufficient power available to produce a self-sustaining dynamo even if the inner core had not nucleated at this point (as seems likely; Aubert et al., 2009).

In order to produce some agreement of the poles shown in Figs. 4B and 7c, it is necessary to assume that the NfT, MBC, and DUF directions were acquired when the field had a polarity reversed relative to that when the other three sets of rocks were magnetised. Consequently, these data together constitute the first tentative evidence for geomagnetic reversals in the Palaeoarchaean. If future studies can verify these observations by collecting data sufficient in quality and quantity to demonstrate a positive reversal test, then these would be the oldest established geomagnetic reversals by some 700 Myr (Strik et al., 2003).

5. Conclusions

New palaeomagnetic and radiometric data from igneous rocks of the 3.3–3.5 Ga Onverwacht Group have been presented and compared to published datasets from rocks of the same age. As might be expected for such old rocks with such complex structural and thermochemical histories, the results are not straightforwardly interpretable. A positive conglomerate test from the Noisy formation has been strengthened and newly dated with a minimum age of 3455.2 ± 7.5 Ma. These results clearly indicate that rocks of the Onverwacht Group have at least the potential to record a near-primary direction of remanence. Furthermore, the dated tuff layer yields a characteristic palaeomagnetic component which, in stratigraphic coordinates, produces a pole within 95% confidence limits of the antipode of a similar-aged dacitic intrusion on the other limb of a large plunging antiform. High temperature components isolated from volcanic rocks of the older Hooggenoeg (~3.46 Ga) and Komati (~3.48 Ga) Formations are associated with poles which also plot within 95% uncertainty limits of the Noisy formation Dacite in stratigraphic coordinates, further supporting a pre-folding (>3.23 Ga) age for these components. Finally, the uncertainties associated with these four poles overlap with those produced from published studies of ~3.46 Ga rocks from the Pilbara Craton (Western Australia) when reconstructions of the Vaalbara Supercraton previously-proposed for the late Archaean are used. These results provide intriguing preliminary evidence that a stable and reversing geomagnetic field was up and running at ca. 3.5 Ga, that continental drift rates were not excessively fast relative to today, and leaves open the possibility that parts of the Kaapvaal and Pilbara Cratons may have been conjoined in the Palaeoarchaean. Together, the results discussed here also provide a framework in which the results of future palaeomagnetic studies performed in the BGB can be interpreted.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at doi:10.1016/j.epsl.2010.12.024.

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