Emplacement temperatures of pyroclastic and volcaniclastic deposits in kimberlite pipes in southern Africa

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Palaeomagnetic techniques for estimating the emplacement temperatures of volcanic deposits have been applied to pyroclastic and volcaniclastic deposits in kimberlite pipes in southern Africa. Lithic clasts were sampled from a variety of lithofacies from three pipes for which the internal geology is well constrained (the Cretaceous A/K1 pipe, Orapa Mine, Botswana, and the Cambrian K1 and K2 pipes, Venetia Mine, South Africa). The sampled deposits included massive and layered vent-filling breccias with varying abundances of lithic inclusions, layered crater-filling pyroclastic deposits, talus breccias and volcaniclastic breccias. Basalt lithic clasts in the layered and massive vent-filling pyroclastic deposits in the A/K1 pipe at Orapa were emplaced at >570°C, at 200–440°C in the pyroclastic crater-filling deposits, and at <180°C in crater-filling talus breccias and volcaniclastic breccias.

The results from K1 and K2 pipes at Venetia are suggestive of emplacement temperature estimates for the vent-filling breccias of 260°C to >560°C, although interpretation of these results is hampered by the presence of Mesozoic magnetic overprints. These temperatures are comparable to the estimated emplacement temperatures of other kimberlite deposits and fall within the proposed stability field for common interstitial matrix mineral assemblages within vent-filling volcaniclastic kimberlites. The temperatures are also comparable to those obtained for pyroclastic deposits in other silicic volcanic systems. Because the lithic content of the studied deposits is 10–30%, the initial bulk temperature of the pyroclastic mixture of cold lithic clasts and juvenile kimberlite magma could have been 300–400°C hotter than the palaeomagnetic estimates. Together with the discovery of welding and agglutination of juvenile pyroclasts in some pyroclastic kimberlites, the palaeomagnetic results indicate that there are examples of kimberlites where phreatomagmatism did not play a major role in the generation of the pyroclastic deposits. This study indicates that palaeomagnetic methods can successfully distinguish differences in the emplacement temperatures of different kimberlite facies.

Keywords: Kimberlite, Emplacement temperature, Palaeomagnetism, Pyroclastic deposits, Thermoremanent magnetization, Explosive eruption

Introduction

Kimberlites are mantle-derived ultramafic volcanic rocks preserved in dykes, volcanic pipes and craters (Dawson, 1971; Mitchell, 1986). Over 5000 kimberlite occurrences are known (Kjarsgaard, 1996), but are confined to the ancient cratonic regions of continents (e.g., south, central and western Africa; Canada, Australia, Russia). Their emplacement ages vary from early Proterozoic through to early Tertiary and because no kimberlite eruptions have ever been witnessed many aspects of kimberlite volcanism are unclear (Sparks et al, 2006). In addition, kimberlite rocks are usually highly altered, particularly the matrix – though less so for incorporated lithic clasts, and contaminated with mantle and crustal debris (Mitchell, 1986; Sparks et al., 2006; Stripp et al., 2006; Buse et al., 2010). This has made it difficult to reconstruct the fundamental properties of the magma, such as its chemistry, temperature, viscosity and volatile content (e.g., Sparks et al., 2006).

Two principal theories have been put forward as the driving force for kimberlite eruptions: (1) the exsolution of magmatic volatiles (e.g., Dawson, 1971; Clement and Reid, 1989; Field and Scott Smith, 1999; Sparks et al., 2006; Wilson and Head, 2007), and (2) the interaction of rising kimberlite magma with ground water (e.g., maar-diatreme model of Lorenz, 1975; Kurszlaukis et al., 1998; Lorenz and Kurszlaukis, 2007). Recent dynamical models propose a volatile-driven eruption mechanism similar to other types of explosive volcanic eruptions (see Sparks et al., 2006), and the similarities between steep-sided kimberlite pipes and maars and diatremes, formed during hydrovolcanic explosions, are striking. As in other varieties of volcanism the two models are not mutually exclusive.

The ability to estimate the emplacement temperatures of pyroclastic deposits using palaeomagnetic methods has proved useful in distinguishing between magmatic and phreatomagmatic modes of eruption, and in discriminating between pyroclastic deposits and epiclastic deposits (Aramaki and Akimoto, 1957; Wright, 1978; Hoblitt and Kellogg, 1979; Downey and Tarling, 1991; Bardot et al., 1996). The technique, pioneered by Aramaki and Arimoto (1957), has been successfully applied to many deposits around the world (Kent et al., 1981; Hoblitt and Kellogg, 1979; Clement et al., 1993; Mandeville et al., 1994; De'Gennaro et al., 1996; Cioni et al., 2004; McClelland et al., 2004; Porreca et al., 2008; Paterson et al., 2010). Palaeomagnetic emplacement temperature determinations of pyroclastic deposits have also been shown, in certain cases, to be as accurate as direct measurements taken shortly after eruption (e.g., the 1980 eruption of Mt. St. Helens; Paterson et al., 2010).

Several attempts have been made to estimate the emplacement temperatures of kimberlite deposits. The general absence of thermal metamorphic effects on entrained xenoliths and in adjacent country rocks has led some authors to propose emplacement temperatures of <500°C for volcaniclastic kimberlite deposits (e.g., Watson, 1967; Mitchell, 1986; Skinner and Marsh, 2004). Sosman (1938) deduced intrusion temperatures of 340°C for North American kimberlites from thermal effects on coal inclusions (see Watson, 1967). Stasiuk et al. (1999) used reflectance values in dispersed organic matter inclusions within Canadian kimberlites to deduce temperatures of 150–200°C for pipe-facies and <100°C for crater-facies volcaniclastic kimberlites. Palaeomagnetic studies by McFadden (1977) on wall rock samples and accidental inclusions close to the contacts of four South African pipes suggested emplacement temperatures of ~300°C. The stability fields of the common alteration assemblages in kimberlite pyroclastic rocks give minimum temperatures of 250–400°C (Stripp et al., 2006; Buse et al., 2010).

Here we present the results of palaeomagnetic measurements of the emplacement temperatures of a range of pyroclastic and volcaniclastic kimberlite deposits within three contrasting kimberlite pipes in southern Africa, for which the internal geology is well constrained by recent geological studies (Field et al., 1997; Brown et al., 2009; Gernon et al., 2009a). Measurements of the thermoremanent magnetization (TRM) of included lithic clasts show that the range of emplacement temperatures varies from around ambient for volcaniclastic deposits, interpreted as talus breccias and debris flow deposits, through to >200 to >570°C for pyroclastic deposits – temperatures which are similar to those obtained for pyroclastic deposits in other volcanic systems. We discuss the implications of these results in the context of the supporting geological evidence for the mode of formation of the kimberlite deposits and for the nature of kimberlite eruptions.

Geological setting

Orapa A/K1 kimberlite pipe, Botswana

The Orapa A/K1 kimberlite (Fig. 1a and 2a) comprises two steep-sided coalescing pipes (north and south) that were intruded at ~92.1 Ma through Archaean basement (granitegneisses and tonalites) and Phanerozoic sediments and basalt lavas of the Karoo Supergroup (Field and Scott-Smith, 1999). The geology of the Orapa A/K1 pipe is described by Field et al. (1997) and Gernon et al. (2009a). The upper portions are dominated by a layered pyroclastic kimberlite lithofacies (Northern Pyroclastic Kimberlite, NPK; Fig. 1a) which is characterised by abundant basement and basalt fragments (≤10 m diameter). These fragments can be concentrated in crude, reverse- to normally graded layers. The matrix comprises abundant serpentinised olivine macrocrysts and phenocrysts and juvenile lapilli. Interstitial pore-space is filled with serpentine-diopside cement. At depth the layering disappears and the deposit becomes massive (Field et al., 1997).

The south pipe is larger than the north pipe and its crater cuts the north pipe (Field et al., 1997). Talus deposits outcrop along the western margin of the south pipe and comprise crudely bedded clast-supported basalt breccias and bedded crystal-rich grain flow deposits (Fig. 1a and 2a), both of which dip towards the centre of the pipe at the angle of repose (Field et al., 1997). Field et al. (1997) propose that these deposits formed by the post-eruption decrepitation of the pipe walls and a surrounding tephra cone. Inner crater lithofacies (Southern Volcaniclastic Kimberlite, SVK; Fig. 1a and 2a) comprise a sub-horizontally layered sequence of poorly sorted basalt-bearing lapilli-tuffs and breccias and stratified olivine tuffs and grits. These are interpreted as a series of ignimbrites derived from neighbouring pipes and sheet-flood deposits (Gernon et al., 2009a).

K2 and K1 pipe, Venetia Mine, South Africa

The Venetia kimberlite cluster is located in the Limpopo region of South Africa (Fig. 1b). It comprises 14 pipes outcropping over ~4 km^2 (Seggie et al., 1998; Kurszlaukis and Barnett, 2003). The pipes were intruded at ~519 Ma into complex Proterozoic basement comprising biotite gneiss, biotite schists and amphibolite gneiss, quartzo-feldspathic gneiss and metasediments (Phillips et al., 1999). Shale and lava clasts within the pipes indicate that Waterberg Formation rocks covered the basement at time of emplacement (Kurszlaukis and Barnett, 2003).

K2 is a steep-sided volcanic pipe which tapers at depth (Kurszlaukis and Barnett, 2003; Brown et al., 2009). It is 250 m by 300 m wide and is broadly divisible into two parts (Fig. 1b and 2b). K2 East is dominated by massive volcaniclastic kimberlite (MVK). K2 West is filled with crudely bedded coarse-grained country rock breccias (Br) and matrix- or clast-supported volcaniclastic kimberlite breccias (mVKBr, cVKBr) with variable amounts of lithic lapilli, blocks and boulders. The contact between these two halves is marked by a shear surface dipping ~64° westwards (Fig. 1b and 2b). The breccias in K2 are thought to result from two competing processes: (1) gravitational collapse of pipe margins that generated abundant brecciated country rock (Br lithofacies) and (2) proximal fallout of country rock clasts and pyroclasts from eruption jets during explosive eruptions (VKBr lithofacies; Brown et al., 2009). The breccias pre-date the MVK in K2 East, which represents the deposits of a later stage of explosivity in the pipe (Kurszlaukis and Barnett, 2003; Brown et al., 2009). K1 is an irregular-shaped kimberlite pipe that is mainly filled with MVK similar to that found in K2 (Fig. 1c and 2c; Kurszlaukis and Barnett, 2003; Walters et al., 2006).

Palaeomagnetic determination of emplacement temperatures

Palaeomagnetic determination of the emplacement temperatures of volcanic deposits relies on the fact that lithic clasts incorporated into a pyroclastic (or volcaniclastic) deposit will have originally been magnetized *in situ* prior to eruption and will thus possess a natural remanent magnetization (NRM) aligned with the Earth's magnetic field during their formation or during some metamorphic event. The nature of this pre-eruption magnetization is not critical to the interpretation. During the eruption these clasts are thrown into the air and 'jumbled up' such that the direction of magnetization will now vary from clast to clast when they come to rest. If the deposits are emplaced above ambient temperatures, the lithic clasts are heated during their incorporation into the deposit and will cool to ambient temperature in their present position. If the clast contains a population of magnetic grains with a spectrum of grain sizes, a portion of the original magnetization with unblocking temperatures (T_{ub}) less than or equal to the emplacement temperature (T_e) will be reset by this heating and replaced or overprinted by a new partial thermoremanent magnetization (pTRM). Thus, the clasts will now contain two components of magnetization – a low unblocking temperature component that will be parallel to the Earth's magnetic field at the time of cooling, and the original high unblocking temperature magnetization that will have random orientations from clast to clast. The emplacement temperature (T_e) of the lithic clasts can, therefore, be determined by progressive thermal demagnetization of the components of magnetization present within the clast. The estimate of T_e is the temperature above which the overprinted magnetization is removed and the randomized high temperature magnetization is uncovered (e.g., McClelland and Druitt, 1989; Bardot, 2000).

Emplacement temperature validation

The palaeomagnetic technique may give erroneous results if the magnetic mineralogy of a lithic clast has altered by being heated, either during the eruption or during the laboratory experiment. Alteration of the ferromagnetic mineralogy of a lithic clast may also have occurred post-eruption during hydrothermal alteration of the deposits, and by low temperature diagenesis and weathering. The growth of a new magnetic phase during the eruption or during later alteration or laboratory heating could produce a chemical remanent magnetization (CRM) which would parallel the Earth's magnetic field at the time of alteration and may partly or completely replace the existing magnetization (Bardot and McClelland, 2000). The low unblocking temperature component in some samples could therefore have two origins; thermal activation (pTRM), which cannot exceed the T_e, or a CRM retained in newly formed or chemically altered grains resulting from alteration events. A secondary CRM may not demagnetize in the laboratory until the Curie temperature of the new magnetic phase, and this temperature would not be related to the emplacement temperature of the deposit (McClelland et al., 2004). In order to test the reliability of the emplacement temperature estimates, we monitored the variation of magnetic susceptibility with temperature to determine the Curie temperature (T_c) of the magnetic-mineral assemblages in the lithic clasts. To determine if the ferromagnetic mineralogy of the lithic clasts was altered during syn- or post-eruption processes, the mineralogy of lithic clasts was compared with corresponding country rock samples.

An additional complication arises from the possible presence of Viscous Remanent Magnetisations (VRM). This is the magnetization that is gradually acquired by very small magnetic grains with very short relaxation times – they tend to rapidly realign with the most recent magnetic field, and yield a low-temperature component of magnetization that parallels the most recent field and can mask older components of magnetization. A VRM can be removed by thermal demagnetization at a specific time-temperature combination related to that at which the rock acquired the VRM. In essence this means that if the rock acquired a VRM over a period of 780,000 years (the time since the last magnetic reversal) we would need to heat it to a substantially higher temperature on a laboratory time scale (50 minutes) to remove it. This means that there is a lower limit on the emplacement temperature that we can determine using the palaeomagnetic technique, and for rocks older than 780,000 years this has been determined empirically to be 163°C (Bardot and McClelland, 2000). We have chosen to be conservative, given that there will also be some error on the temperature settings in our furnace (±10°C), and have not attributed any geological significance to components fitted below 180°C.

Methods

Sampling strategy

The sampling of the lithic clasts followed that outlined by McClelland and Druitt (1989) and Bardot (2000). Given that the clasts were irregularly shaped, rigid plastic plates were glued onto the surface of *in-situ* clasts and the strike and dip of the plate was recorded. This was done so that we would have a perfectly flat surface for precisely orienting the clasts. Clasts ranged in size from 4 to 24 cm diameter. Cores with a diameter of 1.9 cm were drilled from Orapa A/K1 basalt clasts. Standard 2.5 cm diameter cores were drilled from Venetia K1 and K2 clasts because they possessed weaker NRMs. Multiple core specimens were obtained from individual lithic clasts to detect any magnetic inhomogeneity or thermal gradients within the samples.

In the Orapa A/K1 pipe 110 basalt lithic clasts were sampled from the north pipe (NPK) and the south pipe (SVK and talus breccias; Fig. 1a and 2a). Fifty five basalt lithic clasts were sampled from the NPK from two localities at depths of 155 m and 140 m below present surface (Sites 1 and 2 respectively; Fig. 1a and 2a). Five clasts were sampled from talus breccias (Site 3; 95 m below present surface (bps)). Eight clasts were sampled from inner crater SVK deposits at Site 4 (110 m bps) and 42 clasts from Site 5 (140 m bps). All samples are tholeiitic Stormberg Formation basalt of Jurassic age. They are generally fine- to medium-grained, holocrystalline and feldspar-phyric, with a seriate groundmass of feldspars and pyroxenes. The feldspars exhibit embayed grain boundaries and breakdown to oxides and the groundmass is altered to clay and chlorite.

Thirty-five lithic clasts were collected from the K1 and K2 pipes, Venetia mine. Thirty-three clasts were collected from the two main lithofacies in K2: 28 from the matrix-supported volcaniclastic kimberlite breccias (mVKBr) in K2 West and 5 from the massive volcaniclastic kimberlite deposits (MVK) in K2 East (Fig. 1b and 2b). Because the incorporation of lithic clasts can significantly cool pyroclastic mixtures (Marti et al., 1991) only localities with <35 vol.% lithic clasts were chosen for sampling. Sampling localities (Fig. 1b) aimed to provide a good radial and azimuthal spread, collecting six to eight clasts from a site spread over 5 to 10 m of outcrop. Sites 1 to 8 are exposed at 90 m bps, and sites 9 and 10 at 75 m bps (Fig. 1b). Two clasts were collected from MVK deposits in K1 (Site 1; Fig. 1c) at a depth of 170 m below present surface. Amphibolite gneiss clasts (13 clasts) were preferentially taken because they are the most likely to carry a strong and stable magnetization due to a suitable ferromagnetic mineral content and a competent massive structure. At some localities amphibolite clasts were scarce and biotite gneiss (5 clasts) and garnetiferous biotite schist (5 clasts) were sampled instead. Three argillite clasts, two basalt clasts, and seven clasts of Proterozoic dolerite were also sampled. At sites 6 and 7 there were no clasts that could be sampled.

Laboratory methods

Measurements of the natural remanent magnetization (NRM) were made in a magnetically shielded laboratory using a 2-G Enterprises cryogenic magnetometer. The samples were demagnetized using a Magnetic Measurements thermal demagnetizer with a residual field less than 50 nT in heating steps of 20°C or 40°C until the remaining intensity was less than 5% of the NRM. Occasionally more detailed steps were used where demagnetization occurred in a narrower temperature window. Demagnetization results were visually inspected using orthogonal vector component plots (Zijderveld, 1967) and stereographic projections. Remanence directions were determined from stable end-points using principal component analysis (Kirschvink, 1980). Magnetic components were considered stable where they were defined by at least three points on vector component diagrams and had a maximum angular deviation (MAD) not exceeding 15°. Statistical analysis of the magnetization components and directional data were evaluated using Fisher (1953) spherical statistical parameters. The significance of groupings of vector components from each site was assessed using Watson's (1956) test for randomness.

The Curie temperature T_c of representative samples was determined by taking measurements of low-field susceptibility versus temperature, using a Agico KLY-2 Kappabridge magnetic susceptibility meter with furnace attachment, on 1–2 cm^3 of powdered sample taken from lithic clasts prior to thermal demagnetization. Measurements of susceptibility were made every 15–20 s as the sample was heated from 40–700°C, and then as it cooled back to 40°C, with a typical heating-cooling cycle taking about 2 hours. The Curie temperatures were determined using the inverse-susceptibility method of Petrovský and Kapička (2006). Polished sections were analysed using back-scattered imagery on a JEOL JSM-840A scanning electron microscope (SEM) and energy dispersive X-ray spectroscopy (EDS) using an Oxford Instruments ISIS 300 system.

Results

Orapa A/K1 kimberlite pipe, Botswana

The 110 basalt samples displayed NRM intensities from 0.20 to 48.7 amperes/meter (A/m). North pipe (NPK) samples displayed the highest intensities (1.05 to 48.7 A/m); with the majority displaying intensities >20 A/m. The lowest NRM intensities are displayed by talus breccia samples (<1 A/m). Fifty six samples were fully demagnetized after the 590°C heating step with the remaining samples demagnetized over heating steps between 600 and 700°C, indicating that the samples contain ferromagnetic grains with T_c >590°C. All samples exhibited stable behaviour during thermal demagnetization and displayed well-defined (MAD<15°) single or two-component magnetizations. Homogeneous thermal demagnetization behaviour is displayed between samples taken from the same clast in all but three clasts – where different emplacement temperature estimates were obtained from the rim and interior of the clasts.

Thermal demagnetization behaviour

Representative thermal demagnetization data for north pipe (NPK) and south pipe (SVK and talus breccias) samples are shown in Fig. 3. Groupings of magnetization directions on equal-area stereonets are shown in Fig. 4; statistical parameters are detailed in Table 1. Well-defined emplacement temperature determinations are obtained from clasts where two components of magnetization are identified through thermal demagnetization. In clasts where the primary magnetization has been overprinted by thermal activation (pTRM) of the magnetism, the two components are separated by a change in the direction of the magnetization, which occurs at the emplacement temperature. This behaviour is depicted in Figs. 3a and 3b for samples AK1-38a1 and AK1-51a1 respectively (SVK; Site 5). In sample AK1-38a1 (Fig. 3a) the initial magnetic vector points northwest (D = 330.5°) and upwards (I = -60.6°). As demagnetization progresses (up to 240°C) the declination of the vector rotates westwards and the inclination steepens. At subsequent heating steps the direction of the vector remains similar, but the intensity decreases until the sample is fully demagnetized between 570 and 590°C. These changes in the magnetization vector are mapped by two separate lines. All points between room temperature and 240°C can be fitted by a well-defined line (MAD = 2.9°) with a direction (D = 342.3°, I = -54.9°) that is similar to the Cretaceous Earth's field direction (D = 350°, I = - 69°) for the region (Hargreaves and Onstott, 1980). The points from 280°C to the origin of the plot lie on a high-temperature line (MAD = 10.0°) with a direction (D = 245.1°, I = -65.9°) that is different to that of the low-temperature component and the Cretaceous Earth's field direction. This line represents the original magnetization of the clast which has moved to a different orientation during transport in the deposit. An emplacement temperature (T_e) estimate uses the temperature range between the last point on the low-temperature line and the first point on the high-temperature line, in this case T_e = 240–280°C. In another example AK1-51a1 (Fig. 3b) all points from room temperature to 360°C lie on a low-temperature line with a direction (D = 359.6°, I = -32°) that is similar to the Cretaceous direction, and all points from 400°C to the origin lie on a line with a direction (D = 291.7°, I = -17.4°). In this case the emplacement temperature estimate is 360–400°C. We define this demagnetization behaviour as 'type-1'. Three clasts display a variation in T_e estimates between specimen cores taken from the interior and exterior of the clast. Sample AK1-25a1 (Fig. 3c) taken from the exterior of the clast displays a two-component magnetization with the separation of the two-components occurring at T_e = 280–320°C. In sample AK1-25a2 (Fig. 3d) taken from the interior of the clast only a single component is present which is identical to the high-temperature component of the exterior sample. This is the original magnetization of the clast and indicates the core has not been heated above the minimum temperature which can measured by palaeomagnetic methods.

If lithic clasts are heated to temperatures greater than the maximum Curie temperatures (T_c) of the minerals they contain, the original magnetization will be completed overprinted by a new pTRM component, and the magnetization will be a single-component parallel to the Earth's magnetic field. In these examples the maximum T_c of the minerals will provide minimum emplacement temperature estimates. This is defined as 'type-2' behaviour and two examples with different maximum T_c are shown in Fig. 3e, f. Sample AK1-76b2 (Fig. 3e) is fully demagnetized between 660°C and 680°C and a single component of magnetization (D = 1.9°, I = -67.4°) is defined by a well-fitted line (MAD = 4.5°) through all data points from room temperature to the origin. The single-component (D = 337.2°, I = -71.4°) in sample AK1-101a2 (Fig. 3f) is fully demagnetized between 570 and 590°C. Both directions are similar to the Cretaceous Earth's field direction and the samples provide minimum T_e estimates of >660°C and >570°C respectively. Samples where the natural magnetic grain size distribution is extremely restricted and no grains with low unblocking temperatures (T_{ub}) are present are defined as 'type-3' behaviour. No thermal overprint would be recorded in these clasts if heated to temperatures less than the minimum T_{ub}. In sample AK1-2a2 (Fig. 3g) no demagnetization occurs in heating steps up to 360°C, after which the magnetization is removed between 400 and 590°C. Similar behaviour is shown by sample AK1-5a1 (Fig. 3h) where demagnetization only occurs between 440 and 620°C. The magnetizations have a random direction and indicate the clasts have been emplaced at temperatures less than the minimum T_{ub}. The samples therefore provide maximum T_e estimates of <400°C and <440°C respectively.

All 110 basalt lithic clasts displayed well-defined single or two-component magnetizations of one of the three demagnetization behaviour types (types 1, 2 or 3). Twenty-five samples display a two-component magnetization (type-1) delineated by sharp changes in the direction of the magnetization vectors. Statistical grouping of the low-temperature components in these clasts is only observed for Site 5 (SVK; n = 23) with a mean direction (D = 330°, I = -43.9°, α95 = 12.2; Fig. 4a; Table 1) which is similar to the Cretaceous field direction. High-temperature components from the same clasts are scattered (Fig. 4b) indicating these are the original magnetizations of the samples. These samples were used to determine emplacement temperatures. Two samples from Site 4 displayed a two-component magnetization where the low-temperature components are different to the Cretaceous field direction. These are considered clasts which have either been moved within the deposit or have been emplaced at ambient temperatures, and the two-component magnetizations predate the emplacement. Due to this ambiguity, they have not been used to provide emplacement temperature estimates. Sixty-seven samples display a single-component (type-2) magnetization. All samples from NPK (n = 55; Sites 1 and 2) display single-component magnetizations that are significantly grouped and have a mean direction (D = 22.1°, I = -64.5°, α95 = 4.2°; Fig. 4c; Table 1) sub-parallel to the Cretaceous field direction. These are interpreted to have been emplaced at temperatures greater than the

Curie temperature of the minerals present. The remaining type-2 behaviour clasts (n = 12) exhibit single-component magnetizations that have random directions. These samples should acquire a pTRM component parallel to the Cretaceous field direction because of the well-distributed range of blocking temperatures. These are clasts interpreted to have been emplaced at ambient temperatures (i.e. the single-component is the original magnetization of the clasts which predates the eruption of the pipe). No groupings of single-component and two-component magnetizations are observed at Site 3 (talus breccias; Fig. 4d) and Site 4 (SVK; Fig. 4e) and these deposits are interpreted to have been emplaced at ambient temperatures. Eighteen samples display type-3 behaviour single-component magnetizations with restricted blocking temperature spectra where no low-T_{ub} grains are present. A thermal overprint would not be recorded in these clasts if they were emplaced at temperatures less than the minimum T_{ub}. Therefore, where the overprint in these clasts does not parallel the Cretaceous field direction, the clasts provide minimum T_e estimates.

K2 and K1 pipes, Venetia Mine, South Africa

The thirty-five collected samples displayed weak remanent magnetizations with NRM intensities ranging from 0.043 to 85 mA/m. The majority exhibited intensities of < 1 mA/m. A well-distributed range of NRM intensities is displayed by amphibolite, biotite gneiss, argillite, and basalt clasts, with amphibolite samples showing both the lowest (0.043 mA/m) and highest (85 mA/m) measured NRM. All five garnetiferous biotite schist (GBS) samples displayed NRM intensities < 0.5 mA/m. Dolerite clasts displayed stronger intensities up to 82 mA/m, with five of the seven dolerite clasts displaying intensities > 10 mA/m. Twenty-four samples displayed stable magnetic behaviour in which a single or two-component magnetization could be identified over a range of heating steps. Homogeneous thermal demagnetization behaviour is displayed between all cores taken from the same clast, except for biotite gneiss sample V53 from Site 4 (mVKBr) which gave conflicting T_e estimates. The remaining eleven samples displayed erratic magnetizations, with strongly fluctuating intensities, in which no stable magnetization components could be identified. This behaviour is observed in all clast lithologies but restricted to the most weakly magnetized samples (intensities < 0.5 mA/m), and probably results from low signal-to-noise ratios within the samples. Garnetiferous biotite schist (GBS) samples displayed the most erratic behaviour, exhibiting massive intensity spikes (increases) that occur throughout the demagnetization process. Considerable mineralogical alteration occurred in GBS clasts during the heating process and no samples from Site 10 were including in any emplacement temperature determinations.

Of the twenty four samples that provided acceptable demagnetization results sixteen samples were fully demagnetized to <5% of the original NRM intensity at a peak temperature of 590°C. The remaining eight samples experienced unstable behaviour during the heating process above ~400°C, most likely a result of mineralogical alteration. Low temperature components could be identified and fitted in these samples.

Thermal demagnetization behaviour

Representative thermal demagnetization data for samples displaying stable single or two-component magnetizations are shown in Fig. 5. Grouping of magnetization directions on equal-area stereonets is shown in Fig. 6; statistical parameters are detailed in Table 1. Nine samples display a well-defined two-component (type-1) magnetization. An example is depicted in Fig. 5a for sample V2a (amphibolite; Site 8). In this example all points between room temperature and 260°C can be fitted by a well-defined line (MAD = 8.1°) with a direction (D = 322.7°, I = 31.7°), which is similar to a Cambrian reference palaeomagnetic direction (D = 317.1°, I = 42.5°) for the Venetia area (Meert, 2003). The points from 300°C to the origin lie on a high-temperature line (MAD = 7.2°) with a direction (D = 68.3°, I = -11.4°), different from the low-temperature component. The emplacement temperature (T_e) estimate uses the temperature range between the last point on the low-temperature line and the second point on the high-temperature line, in this case T_e = 260–300°C. The remaining eight type-1 samples display a well-defined two-component magnetization but where the low-temperature component does not parallel the Cambrian field direction. An example is shown for Fig 5b for sample V34b (biotite gneiss; Site 8). In this case all points between room temperature and 420°C can be fitted by a well-defined line (MAD = 8.3°) with a direction (D = 358.3°, I = -63.1°) that does not parallel the Cambrian field direction, but, is parallel to the Mesozoic field direction for the region. All points between 460°C and the origin can be fitted by a high-temperature line (MAD = 5.0°) with a direction (D = 277.5°, I = -34.7°) that is different from both the low-temperature line and the Cambrian field direction. Hence three interpretations are possible: a) the clast may have been emplaced at elevated temperatures (420−460°C) within the deposit, with a Cambrian overprint direction, but then moved within the deposit after it had cooled, with the overprint direction fortuitously rotating to a Mesozoic field direction; b) the two-component magnetization may have pre-existed in the clast, and it was emplaced at ambient temperature; or c) the clast was emplaced at elevated temperature, but the Cambrian overprint has itself later been overprinted by a Mesozoic field direction. Such a Mesozoic overprint would likely be of chemical origin, and hence the sample cannot be used to estimate a Cambrian emplacement temperature. Given that eight of the nine samples with a stable two-component magnetizations yield low-temperature components that parallel the Mesozoic field direction this is our preferred interpretation.

Two samples displayed stable single-component (type-2) magnetizations. An example is depicted in Fig. 5c for sample V60a (amphibolite; Site 5). In this example a single magnetization vector is defined by a well-fitted line (MAD = 6.1°) through all data points from room temperature to the origin. The direction of the line (D = 308.4°, I = 22.3°) is similar to the Meert (2003) Cambrian reference palaeomagnetic direction, implying the clast was emplaced at temperatures greater than the maximum Curie temperatures of the minerals it contains (in this case T_e >560°C). In three samples the natural magnetic grain size distribution is extremely restricted and no grains with low unblocking temperatures (T_{ub}) are present (type-3 behaviour). This is illustrated in Figs. 5d, e for Site 1dolerite samples V18b and V19a respectively. In these samples little or no demagnetization occurs in heating steps below 460°C, after which 90 % of the magnetization is removed. In sample V18b (Fig. 5d) over 90% of the remanent magnetization is lost in heating between 500 and 590°C. As a TRM can only be recorded by high-T_{ub} grains in these samples, no thermal overprint would be recorded in these clasts if heated to temperatures less than the minimum T_{ub}. In sample V19a (Fig. 4e) all points between 460°C and the origin lie on a well-fitted line (MAD = 5.0°) with a direction (D = 253.0°, I = 25.5°) which is sub-parallel to the Cambrian field direction. This implies the clast was emplaced at a temperature greater than the maximum Curie temperatures of the minerals it contained (T_e > 560°C). However, in sample V18b (Fig. 5d) the direction of the magnetization (D = 274.1°, I = -69.9°) is different to the Cambrian field direction, and, instead parallels the Mesozoic field direction. This may suggest that this clast has been completely overprinted with a Mesozoic direction. The remaining ten samples ("type-4" behaviour) display a restricted range of blocking temperatures where no high-T_{ub} grains are present. This behaviour is independent of clast type. Representative samples displaying this behaviour are shown in Figs. 5f, g for samples V10a (argillite; Site 1) and V30a (biotite gneiss; Site 3) respectively. In these samples 80% of the remanent magnetization is removed by heating to 220–340°C. Well-defined lines can be fitted to a low-T_{ub} component, but no high-T_{ub} components can be defined after 80% of the remanence is removed because the signal-to-noise ratio becomes too low. Some samples are completely demagnetized at temperatures as low as 300°C (e.g. V10a; Fig. 5f) which indicates no ferromagnetic grains with T_{ub} >300°C are present. However, only one of these samples yields an over print direction that is not parallel to the Mesozoic field. In Site 1 amphibolite sample V23a (Fig 5h) a well-defined (MAD = 9.6°) component with a direction D = 47.6°, I = 61.4° can be fitted to all points from room temperature to 340°C, after which the magnetization of the sample becomes erratic.

For the Venetia samples the emplacement temperature estimations are complicated by the lack of samples displaying stable magnetic behaviour, preventing unbiased statistical analysis of the magnetization components. As noted above there is a prevalence of low-temperature components that parallel the Mesozoic field direction (Fig. 6a; Table 1). They are likely of chemical origin and these samples cannot be used for emplacement temperature determinations. This leaves us with only 4 samples, which have overprint directions that might be Cambrian in age: one type-1 (V2a); one type-2 (V60a); one type-3 (V19a); and one type-4 (V23a). When these are combined they yield a mean-direction declination of 308° and a mean inclination of 46° (Fig. 6b; Table 1), but the statistical parameters are very poor, only just passing the Watson (1956) test for non-randomness (Table 1). Nevertheless these four samples indicate emplacement temperatures of 260°C to >560°C. The values are similar to that obtained at Orapa, and are taken to be supportive of a range of emplacement temperatures of the kimberlite lithofacies, rather than being diagnostic in their own right.

Rock magnetism results

Spurious emplacement temperature estimates may result from lithic clasts where the magnetic mineralogy has been altered during or after emplacement. Newly formed or chemically altered grains will acquire a chemical remanent magnetization (CRM) parallel to Earth's magnetic field which may partly or completely overprint the existing magnetization. The unblocking temperatures of these components may not coincide with T_e but instead record the Curie temperatures (T_c) of new grains formed in a later (possibly low-temperature) alteration event. To test the reliability of the emplacement temperature results we determined the range of T_c values by measuring the variation in magnetic susceptibility with temperature in 32 powdered samples from lithic clasts which provided T_e estimates from the thermal demagnetization study. The Curie temperatures were determined using the inverse-susceptibility method of Petrovský and Kapička (2006).

Representative magnetic susceptibility-temperature curves for 9 basalt lithic clasts sampled from Orapa A/K1 and 2 amphibolite samples from Venetia K2 are shown in Fig. 7; Demagnetization vector plots are also shown for three samples with T_e estimates ranging from 240–280°C to 360–400°C. We find most basalt samples (n = 26) are dominated by magnetite or Ti-poor titanomagnetites ($x_{Ti} \leq 0.1$) with T_c values of 559-595°C (e.g., Fig. 7a, b, and f, h). Little or no alteration has occurred during heating or cooling of the samples, although 11 samples show a higher susceptibility (but no change in T_c) after the experiment (e.g., Fig. 7j, l). This is interpreted to result from the annealing of defects within the crystal lattice of ferromagnetic minerals during heating (Bardot, 1997). The remaining basalt samples (n= 4) with T_c values of 595-613°C could result from the oxidation of magnetite to maghemite (or titanomagnetite to titanomaghemite) during the laboratory heating. The main ferromagnetic minerals within amphibolite samples (n = 2) is titanomagnetite ($x_{Ti} \sim 0.2$) with T_c values of ~500°C (Fig. 6l, m). No spurious T_c values are observed to coincide with the emplacement temperature estimates obtained during thermal demagnetization which supports the validity of the emplacement temperature results.

Petrography and textural relations

The basalt lithic clasts contain ilmenite and titanomagnetite grains in roughly equal proportions (total proportion 5–15%). Grain diameters vary from <1 μm to ~200 μm. Ilmenite typically occurs as elongate grains (<100 μm length) and titanomagnetites as subhedral to euhedral grains (10–200 μm diameter). Titanomagnetites invariably display coarse or fine ilmenite lamellae along {111} crystallographic planes, and occasionally contain inclusions of ilmenite with irregular or sharp boundaries parallel to the crystallographic axes (Fig. 8a, b). The host titanomagnetite is enriched in Fe with compositions approaching magnetite. These features are typical of deuteric oxidation of host titanomagnetite grains (Ulvospinel-Magnetite solid solution) in igneous rocks (Haggerty, 1991). The intergrowth textures dramatically reduce the effective grain size of the host titanomagnetites (Fig. 8c, d) and it is likely a proportion of the thermal remnant magnetization (TRM) is carried by these magnetite or Ti-poor titanomagnetite intergrowth grains. These textures are mirrored in country rock samples (Fig. 8e, f), although finer grained and skeletal titanomagnetites are observed in samples from basalt flow tops. Collectively these observations indicate that there is no evidence of pervasive hydrothermal alteration of primary igneous ferromagnetic grains or the growth of new ferromagnetic species in the lithic clast samples. The oxides observed under the SEM, however, are larger than those which typically carry stable magnetic remanence, which are normally in the sub-μm size range and hence below the resolution of the SEM, but the fact that original igneous textures are preserved gives us confidence that the primary magnetic mineralogy is likely also preserved within the lithic clasts.

The amphibolite gneiss lithic clasts in Venetia K1 and K2 are upper amphibolite metamorphic facies rocks of the Venetia Klippe unit (Barton et al., 2003). They show compositional variability but typically display granoblastic textures containing hornblende ± garnet + biotite ± feldspar ± quartz + ilmenite. Fe-Ti oxide proportions vary from <1% to 3%. Oxides occur as acicular grains in garnets and biotites (<100 μm length), or as larger (≤300 μm) anhedral inclusions within hornblende grains (Figs. 8g, h). The grains are typically host illmenite grains with bladed intergrowths of rutile and titanite, or fine grained mottled intergrowths of Ti-poor titanomagnetite, rutile and illmenite. These textures and assemblages are mirrored in samples of *in situ* amphibolite country rock. Therefore little alteration of the metamorphic and ferromagnetic mineralogy of the lithic clasts is considered to have occurred during their emplacement. The paucity of magnetite and titanomagnetite in the samples accounts for weak magnetizations observed in the lithics.

Emplacement temperatures and their interpretation

Given the relative paucity of data from the Venetia pipes we focus on the results from the Orapa A/K1 kimberlite pipe, Botswana. Palaeomagnetically determined T_e estimates determined from individual lithic clasts are shown in Fig. 9. Single-component (type-2; n = 55) magnetizations in NPK basalt clasts provide minimum T_e estimates of >570°C to >660°C for vent-filling pyroclastic breccias (Fig. 9a). Maximum T_e estimates of <180°C (which is the lower limit of the palaeomagnetic method for these samples) are provided by randomly orientated single-component and low-temperature component magnetizations (n = 13) in SVK and talus breccia samples (Fig. 9b). These are interpreted to have been emplaced at ambient temperatures. Well-defined two-component magnetizations (n = 23) provide T_e estimates of 200–440° for lithic clasts within inner crater SVK deposits (Fig. 9b).

The results from NPK samples provide minimum deposit equilibrium temperatures (T_{dep}) of >570°C which implies that the NPK is a primary pyroclastic deposit emplaced at significantly elevated temperatures, as proposed by Field at al. (1997) based on field and petrographic evidence. NPK rocks contain the typical metamorphic alteration assemblage of serpentine and diopside which is constrained to be formed at temperatures <400°C (Stripp et al., 2006). This lower estimate on the emplacement temperature of the deposits is consistent with the deposits being emplaced at higher temperature (according to the TRM results) with olivine, as the major component of the deposits, being stable. Later as the deposits cooled down in a hydrothermal system olivine is replaced by serpentine to generate the lower temperature assemblage. T_e estimates for SVK samples of <180°C in talus breccias and <180°C to 200–440°C in volcaniclastic breccias implies that clasts were emplaced in the south pipe at both ambient and elevated temperatures. This indicates both epiclastic and pyroclastic depositional processes occurred in the south pipe. The result from talus deposits is consistent with emplacement by epiclastic processes around the margin of the pipe, as proposed by Field et al. (1997). Volcaniclastic breccias emplaced at ambient temperatures within the inner crater are consistent with emplacement as rock-fall and debris-flow deposits (Field et al., 1997; Gernon et al. 2009a). Clasts emplaced at elevated temperatures (Site 5; Fig.1a and 2a) are located within a laterally extensive, 15–20 m thick kimberlite pyroclastic flow unit described by Gernon et al. (2009b). The unit lies unconformably over crater-filling deposits, which postdate the eruption of the north pipe. Gernon et al. (2009b) infer that the pyroclastic flow originated from a neighbouring kimberlite pipe.

Volcanological implications

Results from Orapa A/K1 provide minimum T_{dep} values of >570°C for vent-filling pyroclastic rocks. These samples were heated to temperatures greater than the Curie temperatures of the minerals present and are observed to have reached thermal equilibrium within the deposit. It is inferred that most, if not all, of the lithic clasts will have been incorporated into the vent-filling pyroclastic deposits at ambient temperatures because the conduit wall rock will have been cold prior to onset of volcanism. The mixing of cold lithic clasts with erupting kimberlite magma will cool the erupting pyroclastic mixture (Marti et al., 1991) as heat is shared out between the juvenile kimberlite pyroclasts and lithic clasts. An idealised equilibrium temperature of the erupting mixture of pyroclastic and lithic components, T_M, can be calculated from the heat conservation equation:

$$T_M = \frac{C_K T_K (1-x) - x C_L T_L}{C_L x + C_K (1-x)},$$

where T_K is the temperature of the kimberlite pyroclasts (assumed to be similar to kimberlite magmatic temperatures of 1273-1473 K (Ferdochouk and Canil, 2004; Mitchell, 2008; Sparks et al., 2009), T_L is the initial temperature of the lithic clasts (293 K), x is the mass fraction of lithic clasts entrained in the mixture, C_K is the specific heat capacity of kimberlite pyroclastics ($C_{olivine}$ at 1000 K = 1150 J kg^{-1} K^{-1}; Gillet et al., 1991), and C_L is the specific heat capacity of lithic clasts (C_{basalt} at 800 K = 1100 J kg^{-1} K^{-1}; Bouhifd et al., 2007). The conversion of lithic clast volume to weight % was made using standard densities for lithic clasts (ρ_{basalt} = 2900 kg m^{-3}), and kimberlite pyroclasts ($\rho_{olivine}$ = 3270 kg m^{-3}). The average proportion of lithic clasts within vent-filling pyroclastic breccias is ~10–20% (Walters et al., 2006; Brown et al., 2009). The bulk mixing of a juvenile magmatic component with 10–20 % cold lithic clasts (x = 0.08–0.16) would reduce the temperature of the mixture from magmatic temperatures of 1000°C to ~920-850°C. These values can be considered maximum estimates of the T_{dep} of vent-filling deposits and therefore the vent-filling deposits in A/K1 could have been 200–300°C hotter than the TRM estimates (>570°C). T_e estimates for individual lithic clasts within the pyroclastic flow deposit in the Orapa south pipe crater range from 200–440°C. The average lithic content of the pyroclastic flow deposit is ~30% (Gernon et al., 2009b) and the ideal equilibrium temperature for a mixture of juvenile magma at 1000°C and 30 vol.% (x = 0.25) cold lithic clasts is ~760°C. The deposit is interpreted as a lithic breccia deposited rapidly upon deceleration of a pyroclastic flow (Gernon et al., 2009b). The scatter in the observed T_e estimates can be explained if some of the lithic clasts do not equilibrate with the flow temperature during rapid transportation. Since the clasts were originally cold and were heated up in the deposit, the upper T_e estimates of 400–440°C most closely represent the equilibrium temperature of the deposit. Further cooling of the pyroclastic flow from source temperatures of ~760°C to deposit equilibrium temperatures of 400–440°C can be explained by air entrainment during transport, and does not necessitate the input of water (Sparks et al., 1978).

The amount of cooling due to the incorporation and vapourisation of water into the pyroclastic flow can be determined using the heat balance equation:

$$T_{dep} = T_M - \left[\frac{yC_{LW}(T_{100}-T_W)+yL_W+yC_{VW}(T_K-T_{100})}{C_M(1-y)}\right],$$

where T_M is the initial temperature of the pyroclastic mixture, T_W is the initial water temperature (293 K) which will be heated to boiling point T_{100} (373K), y is the mass fraction of water, L_W is the latent heat of vaporization of water (2.3×10^6 J kg^{-1}), C_{LW} is the specific heat capacity of liquid water (4228 J kg^{-1} K^{-1}), C_M is the heat capacity of the pyroclastic flow (assumed to be similar to kimberlite pyroclasts; $C_{olivine}$ at 1000 K = 1150 J kg^{-1} K^{-1}), C_{VM} is the specific heat capacity of steam (2040 J kg^{-1} K^{-1}). Cooling of approximately ~300°C from eruption to deposit equilibrium temperatures would require 10 wt. % liquid water.

Conclusions

Thermoremanent magnetism studies of lithic inclusions have successfully differentiated the emplacement temperatures of a variety of volcaniclastic lithofacies preserved within the Cretaceous Orapa kimberlite pipes (Fig. 10). Data from the Cambrian-aged Venetia pipe are of a much lower quality, but they do not contradict the results from Orapa. Obtained T_{dep} values are consistent with the field evidence of the various deposits (Field et al., 1997;

Brown et al., 2009; Gernon et al., 2009a). T_{dep} estimates for pyroclastic deposits are consistent with previous estimates of kimberlite emplacement temperatures (≤340°C; Sosman 1938; Watson 1967; McFadden, 1977; Stasiuk et al., 1999). The T_{dep} values obtained for pipe-filling pyroclastic deposits (260–300°C to >570°C) are similar to the emplacement temperatures of extrusive pyroclastic rocks erupted from more silicic volcanic systems (Hoblitt and Kellogg, 1979; Banks and Hoblitt, 1981; Kent et al., 1981; Cioni et al., 2004; McClelland et al., 2004; Scott and Glasspool, 2005). Pyroclastic deposits become welded at high temperatures and this is particularly common in pyroclastic deposits of low-viscosity magmas (Sumner 1998). Welding textures have recently been recognised in some rocks in Venetia K2 pipe (Brown et al., 2008), providing further evidence of high emplacement temperatures for some kimberlite pyroclastic deposits. Low T_{dep} estimates for clasts in the SVK and talus breccias (<180°C), which show a single-component magnetization, are consistent with emplacement at ambient temperatures by epiclastic processes (rock fall and debris flow; Fig. 10; see Field et al., 1997; Gernon et al., 2009a).

Many kimberlite pipes exhibit evidence for multiple phases of eruptive activity (e.g., nested and cross-cutting pyroclastic and breccia units (Venetia K2 and K1; Kurszlaukis and Barnett, 2003; Brown et al., 2009; Koffiefontein, South Africa, Naidoo et al., 2004). The periods between these phases of activity are unknown, but the presence of clasts with two-component magnetizations may indicate that individual kimberlite pipes are active over limited periods. The well-defined, two-component magnetizations could only have been acquired if they had cooled in a stationary position, after the deposits had completely stabilised and were not subsequently rotated or moved. Cooling times by conduction for kimberlite pipe fills are estimated at decades to centuries (Sparks et al., 2006), thus, late-stage disruptive volcanic activity must have taken place while the deposits were still hot or after the deposits had cooled and lithified (by diagenesis and serpentinisation). The results from are consistent with high temperature mineral assemblages of serpentine-diopside-calcite-chlorite observed in these deposits, which indicate temperatures of 300-340°C (Stripp et al., 2006).

The T_{dep} estimates for the pyroclastic deposits allow some constraints to be placed on the eruption dynamics of studied kimberlite pipes. Minimum T_{dep} values of 260–300°C to >570°C in the Venetia and Orapa pipes are consistent with volatile-driven explosive eruptions. The temperatures of kimberlite magmas is now thought to be in the range of 1200-1000°C (Ferdochouk and Canil, 2006; Mitchell, 2008; Sparks et al., 2009). The incorporation of 10–20% cold lithic clasts would cool the source pyroclastic mixture from magmatic temperatures to 920–850°C. Therefore the actual temperatures of the primary components of the vent-filling deposits may have been 200–300°C higher than the TRM estimates. Similarly the Orapa south pipe contains a pyroclastic flow deposit with, on average, 30 % lithic clasts (Gernon et al., 2009b) and so the original pyroclastic component may have hotter by up to 300°C than the TRM estimates. The amount of cooling from magmatic temperatures to the estimated deposit temperatures is typical of many other pyroclastic deposits worldwide, and can be explained by the entrainment and heating of air and cold lithic material. It is difficult to reconcile these elevated emplacement temperatures with phreatomagmatic explosivity. Explosive interaction of ground or surface water with rising magmas commonly cools erupting flows close to ambient (Thomas, 1993; Bardot et al., 1996). Quantitative measurements of the emplacement temperatures of phreatomagmatic deposits indicate values of <140°C (Thomas, 1993; Bardot et al., 1996; Porreca et al., 2008) to as much as 275°C (De'Gennaro et al., 1996). Low temperatures in phreatomagmatic deposits are also suggested by many features such as the inclusion of uncharred wood in phreatomagmatic deposits (e.g., Giordano et al., 2002) and by the plastering of wet ash onto obstacles (e.g., Cole et al., 2001). However, we stress that our findings do not rule out phreatomagmatic explosivity during kimberlite eruptions.

This research demonstrates that thermoremanent studies of lithic inclusions can successfully differentiate between epiclastic and pyroclastic deposits within ancient kimberlite crater-fill successions. This may have important economic implications for the mining and exploration of kimberlite bodies - not all deposits preserved within a volcanic crater above a kimberlite pipe may be sourced directly from that pipe.

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Figure Captions

Figure 1. Simplified geological maps of the studied kimberlites, with major lithofacies identified and localities of samples indicated by crosses. **a** Summary geological map of the Orapa A/K1 kimberlite body (modified from Field et al., 1997 and Gernon et al., 2009a). The body comprises two steep-sided pipes (north and south). **b** Simplified geological map of the K2 kimberlite pipe, Venetia Mine, South Africa (from Brown et al., 2008). **c** Geological map of K1 kimberlite pipe, Venetia Mine, South Africa. **Inset**: map of southern Africa and locations of mines.

Figure 2. **a** Photograph of the Orapa A/K1 mine (looking south), showing the localities and stratigraphic context of the sampled deposits (sampling sites are numbered). The wall rock comprises Stormberg Formation basalts and Ntane Formation sandstones, the latter of

which crop out as a septum between the north and south pipes. **b** Photograph of the Venetia K2 mine (looking northwest) showing the westward dipping shear surface which separates the two halves of the pipe (K2 east and K2 west). Marked lithological contrasts occur across the shear zone with massive volcaniclastic kimberlite (MVK) in contact with matrix-supported volcaniclastic kimberlite breccias (mVKBr) on the lower bench, and with clast-supported volcaniclastic kimberlite breccias (cVKBr) on upper bench. The sampling sites which are within the field of view are numbered. **c** Photograph of the Venetia K1 mine (looking northwest). K1 is predominately filled with MVK and the wall rock comprises Proterozoic basement gneisses

Figure 3. Representative thermal demagnetization vector plots for basalt clasts from Orapa A/K1. Solid squares represent the magnetization vector for each sample (projected on to the horizontal plane) at different laboratory temperatures (in °C); open circles represent the vector projected onto a vertical plane at the same laboratory temperature. **a, b** Type-1 two-component magnetizations. **c, d** Two-component and single-component magnetization obtained from exterior and interior of the same lithic clast. **e, f** Type-2 single component magnetizations. **g, h** Type-3 (restricted T_{ub} spectra) single component magnetizations.

Figure 4. Equal area stereographic projections of palaeomagnetic directions recorded at each sampling site at Orapa A/K1. Open symbols denote upper hemisphere projections; solid symbols denote lower hemisphere projections. Small circle is the mean direction of the data and the larger circle around the mean shows the 95% confidence limit of the mean (expressed as an angular radius). Open star is the Cretaceous reference palaeomagnetic

direction (D = 350°, I = -69). **a** Low-temperature components in type-1 clasts from Site 5 (pyroclastic flow deposit). **b** High-temperature components in type-1 clasts from Site 5. **c** Type-2 single component magnetizations from NPK clasts (Sites 1 and 2). **d** Single-component magnetizations from Site 3 (Talus deposit). **e** Low-temperature and single components from Site 4 (Debris flow deposit).

Figure 5. Representative thermal demagnetization vector plots for lithic clasts from Venetia K1 and K2 pipes. Solid squares represent the magnetization vector for each sample (projected on to the horizontal plane) at different laboratory temperatures (in °C); open circles represent the vector projected onto a vertical plane at the same laboratory temperature. **a,b** Type-1 two-component magnetization. **c** Type-2 single component magnetization. **d, e** Type-3 (restricted T_{ub} spectra) single component magnetizations. **f, g, h** Type-4 (restricted T_{ub} spectra) single component magnetizations.

Figure 6. Equal area stereographic projections of palaeomagnetic directions recorded at each sampling site at Venetia K1 and K2. Open symbols denote upper hemisphere projections; solid symbols denote lower hemisphere projections. Small circle is the mean direction of data and the large circle around the mean shows the 95% confidence limit of the mean (expressed as an angular radius). Solid star is the Cambrian reference palaeomagnetic direction (D = 317.1°, I = 42.5°). Open star is the Cretaceous reference palaeomagnetic reference direction (D = 350°, I = - 69°). **a** Mesozoic overprint directions in Venetia K1 and K2 samples. **b** Cambrian overprint directions observed in 4 samples from Venetia K1 and K2.

Figure 7. Typical magnetic susceptibility-temperature curves for samples from **a, b** Orapa A/K1 SVK (Site 4), **c, d, f** Orapa A/K1 SVK (Site 5), **h, i** Orapa A/K1 NPK (Site 2), **j, k** Orapa A/K1 SVK (Site 3), **l, m** Venetia K2. **e, g, n** Representative thermal demagnetization plots for samples providing T_e estimates. The Curie temperatures (T_c) were determined using the inverse-susceptibility method of Petrovský and Kapička (2006).

Figure 8. SEM micrographs of Fe-Ti oxides within basalt and amphibolite samples. **a** Host titanomagnetite grain in basalt sample AK1-51a (T_e = 360-400°C); Ilmenite lamellae (dark grey) are observed along {111} cystallographic axes of the host titanomagnetite (light grey). **b** Host titanomagnetite grain in basalt sample AK1-76b2 (T_e ≥660°C). **c, d** Higher magnification view of Ti-poor titanomagnetite intergrowths in AK1-51a and AK-76b2 respectively. **e, f** Country rock basalt samples. **g** Ilmenite grain in amphibolite gneiss sample V60a (T_e ≥590°C) showing ilmenite, rutile and titanite intergrowths. The mottled regions are fine-grained intergrowths of Ti-poor titanomagnetite, rutile and illmenite. **h** Ilmenite grain in country rock amphibolites sample. *ilm* ilmenite, *rut* rutile, *ti* titanite, *tm* titanomagnetite.

Figure 9. T_e estimates obtained from individual lithic clasts. **a** Minimum T_e estimates obtained from Orapa A/K1 NPK lithic clasts provided by the maximum T_c of the minerals in the samples. The range of maximum T_c values and the number of individual samples displaying these values are shown. **b** T_e for individual samples in Orapa south pipe deposits. Downward pointing arrows indicate the data point is a maximum T_e estimate.

Upward pointing arrows indicate the data point is a minimum T_e estimate. Error bars indicate the upper and lower limits of each T_e estimate obtained from clasts displaying a two-component magnetization. Open symbols are used to indicate where different T_e estimates have been obtained from a single clast.

Figure 10. Schematic illustration of the different T_{dep} estimates for the various deposits in the Orapa and Venetia kimberlite pipes. Vent-filling pyroclastic breccias provide minimum T_{dep} values of >570°C. Deposits interpreted as ignimbrites have intermediate T_{dep} values and epiclastic deposits (debris flows and talus breccias) have been emplaced at ambient temperatures (<180°C). Estimates of the source temperature of the pyroclastic deposits are ~760–920°C.

Table 1: Statistics of magnetization components from lithic clasts from the Orapa A/K1and the Venetia K1 and K2 kimberlite pipes.

N, number of magnetization vectors; n, total number of samples studied; **N**/n is the number of stable magnetization vectors identified per number of studied samples; Dec., declination of the mean direction of magnetization of N vectors; Inc., inclination of the mean direction of magnetization of N vectors; R, vector resultant of N vectors; k best estimate of precision parameter; α_{95}, 95% confident limit; R_0 Watson's (1956) parameter; grouping is random for N vectors if $R<R_o$; asterisk*, too few data are available to calculate accurate statistical parameters (N<4)

Table 1

Orapa A/K1	Low-T_b component								High-T_b component							
Group or site	**N**/n	Dec.	Inc.	R	k	α_{95}	R_0	Ran.	N	Dec.	Inc.	R	k	α_{95}	R_0	Ran.
Site 1	**4**/4	11.7	-64	3.87	23.99	19.1	3.10	no								
Site 2	**51**/51	22	-64.5	48.67	21.49	4.4	11.53	no								
NPK	**55**/55	22.1	-64.5	52.54	21.93	4.2	11.97	no								
Site 3	**5**/5	83.6	-70.8	2.73	1.77	85.6	3.50	yes								
Site 4	**8**/8	3.7	-40.5	3.39	1.52	74.2	4.48	yes	1	213.6	-43.7		1.34		*	*
Site 5 (type-1)	**23**/23	330	-43.9	19.9	7.09	12.2	7.74	no	23	338.7	-66	11.86	1.97	30.4	7.74	no
Site 5 (type-2)	**5**/5	16.3	-37.6	1.47	1.13	>90	3.50	yes								
Site 5 (type-3)	**14**/14	201.9	60.6	3.58	1.25	75.7	5.98	yes								
Venetia	Low-T_b component								High-T_b component							
Group or site	**N**/n	Dec.	Inc.	R	k	α_{95}	R_0	Ran.	N	Dec.	Inc.	R	k	α_{95}	R_0	Ran.
K2 site 1	**12**/14	330.8	-49.7	7.87	2.7	33.3	5.52	no	3	299	0.5	1.42	1.3	>90	*	*
K2 site 2	**2**/2	356.5	-57.2	1.76	4.2	>90	*	*								
K2 site 3	**2**/4	4.9	-78.6	1.99	70.9	30.1	*	*	1	81	-55.4				*	*
K2 site 4	**1**/1	294.8	-38.8				*	*	1	338.8	25.7				*	*
K2 site 5	**1**/1	311.8	25.1				*	*								
K2 site 8	**2**/2	334.9	-16.4	1.3	1.4	>90	*	*	2	11.5	-58	0.9	0.9	>90	*	*
K2 site 9	**2**/4	223.2	-57.9	1.37	1.6	>90	*	*	1	8.3	-7.6				*	*
K2 site10	**0**/5															
K1 (site 1)	**2**/2	301.1	-74.1	1.7	3.4	>90	*	*	1	350.5	-21.6				*	*
Cambrian	**4**	308.2	46.2	3.13	3.4	58.5	3.10	no								
Mesozoic	**20**	329.8	-64.4	16.69	5.7	14.9	7.17	no								
All clasts	**24**/35	324.8	-54.1	15.79	2.8	21.9	7.91	no	9	347	-21.9	4.41	1.7	58.2	4.76	Yes

Figure 1

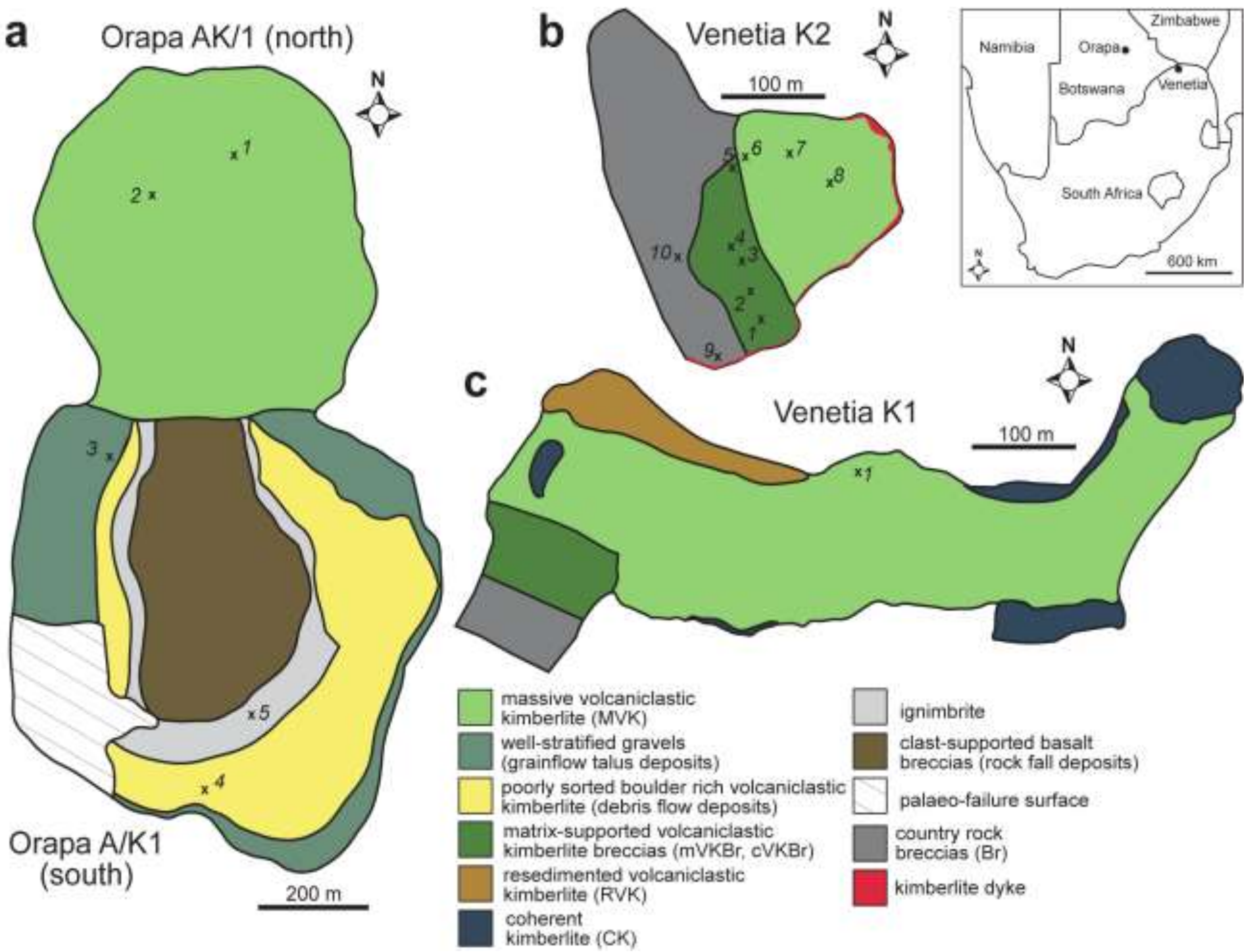

a
Orapa AK/1 (north)
N
1
2
3
5
4
Orapa A/K1 (south)
200 m
b
Venetia K2
N
100 m
5
6
7
8
4
3
10
2
1
9
Namibia
Orapa
Zimbabwe
Botswana
Venetia
South Africa
600 km
c
Venetia K1
N
100 m
1
massive volcaniclastic kimberlite (MVK)
well-stratified gravels (grainflow talus deposits)
poorly sorted boulder rich volcaniclastic kimberlite (debris flow deposits)
matrix-supported volcaniclastic kimberlite breccias (mVKBr, cVKBr)
resedimented volcaniclastic kimberlite (RVK)
coherent kimberlite (CK)
ignimbrite
clast-supported basalt breccias (rock fall deposits)
palaeo-failure surface
country rock breccias (Br)
kimberlite dyke

Figure 2

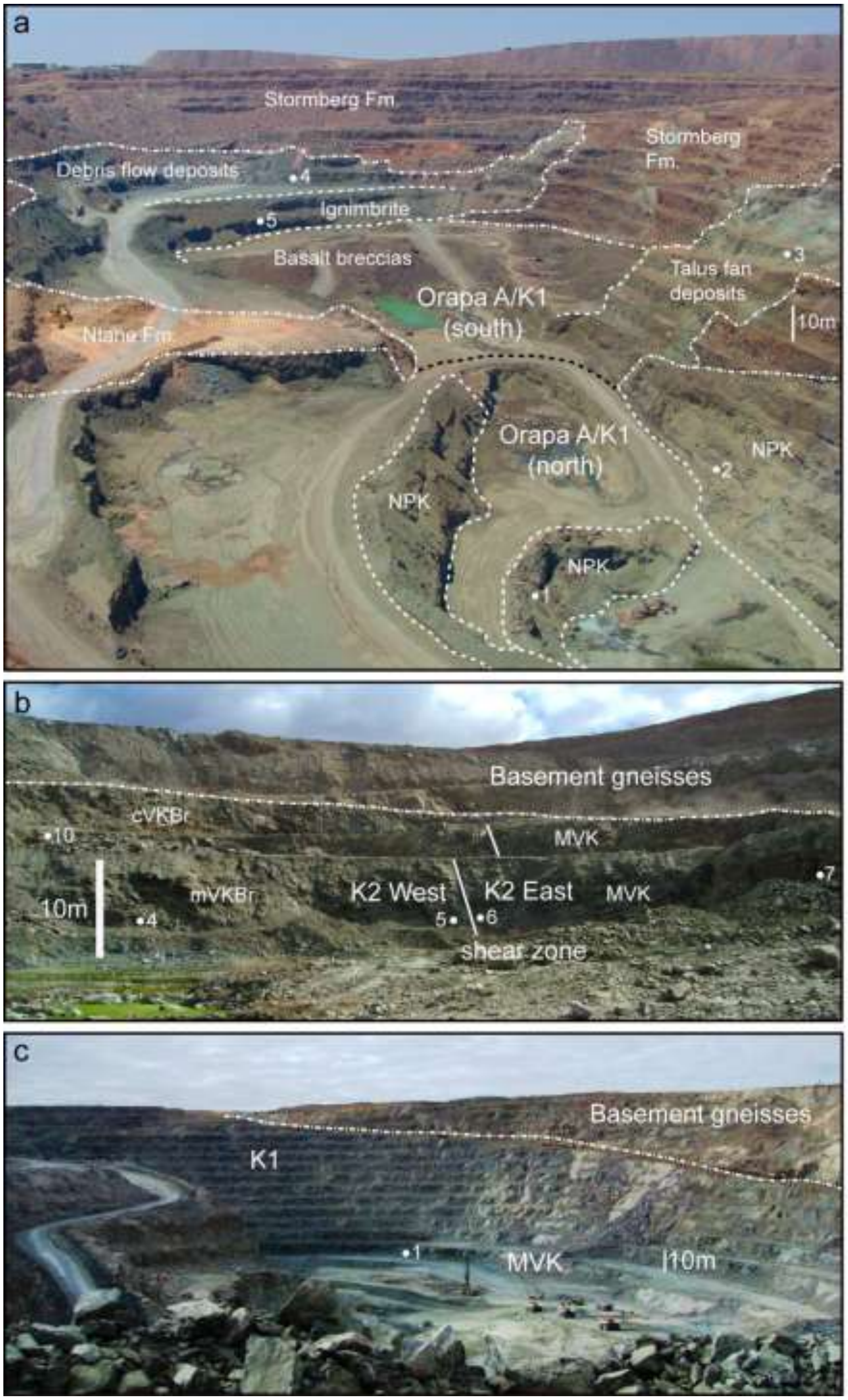
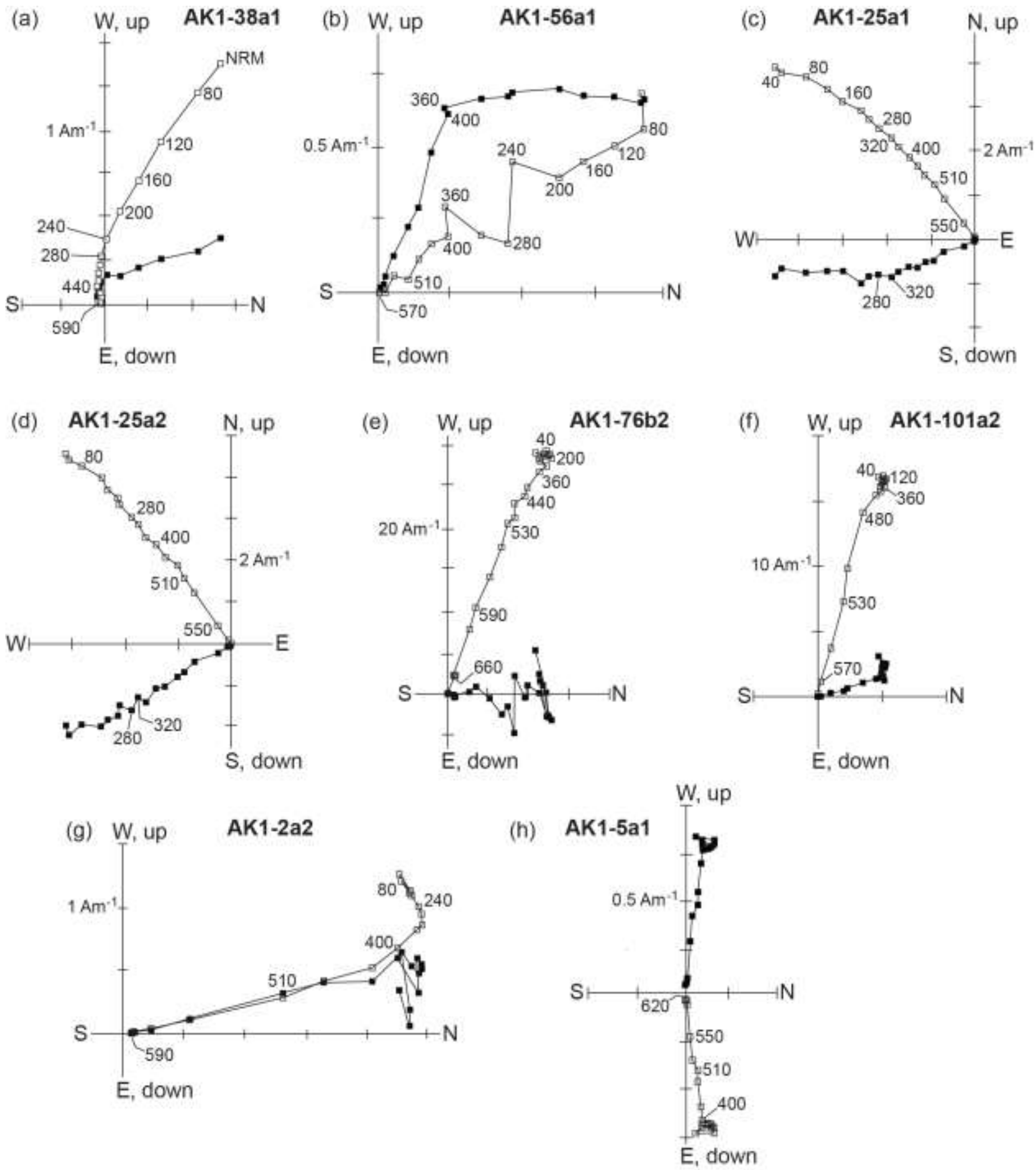

a
Stormberg Fm.
Stormberg Fm.
Debris flow deposits
•4
•5
Ignimbrite
Basalt breccias
•3
Talus fan deposits
10m
Orapa A/K1 (south)
Ntane Fm.
Orapa A/K1 (north)
NPK
•2
NPK
NPK
•1
b
Basement gneisses
cVKBr
•10
MVK
•7
10m
mVKBr
K2 West
K2 East
MVK
•4
5•
•6
shear zone
c
Basement gneisses
K1
•1
MVK
10m

Figure 3



(a) AK1-38a1
W, up
E, down
S
N
1 Am⁻¹
NRM
80
120
160
200
240
280
440
590
(b) AK1-56a1
W, up
E, down
S
N
0.5 Am⁻¹
360
400
80
120
160
200
240
280
510
570
(c) AK1-25a1
N, up
S, down
W
E
2 Am⁻¹
40
80
160
280
320
400
510
550
(d) AK1-25a2
N, up
S, down
W
E
2 Am⁻¹
80
280
400
510
550
320
(e) AK1-76b2
W, up
E, down
S
N
20 Am⁻¹
40
200
360
440
530
590
660
(f) AK1-101a2
W, up
E, down
S
N
10 Am⁻¹
40
120
360
480
530
570
(g) AK1-2a2
W, up
E, down
S
N
1 Am⁻¹
80
240
400
510
590
(h) AK1-5a1
W, up
E, down
S
N
0.5 Am⁻¹
620
550
510
400

Figure 4

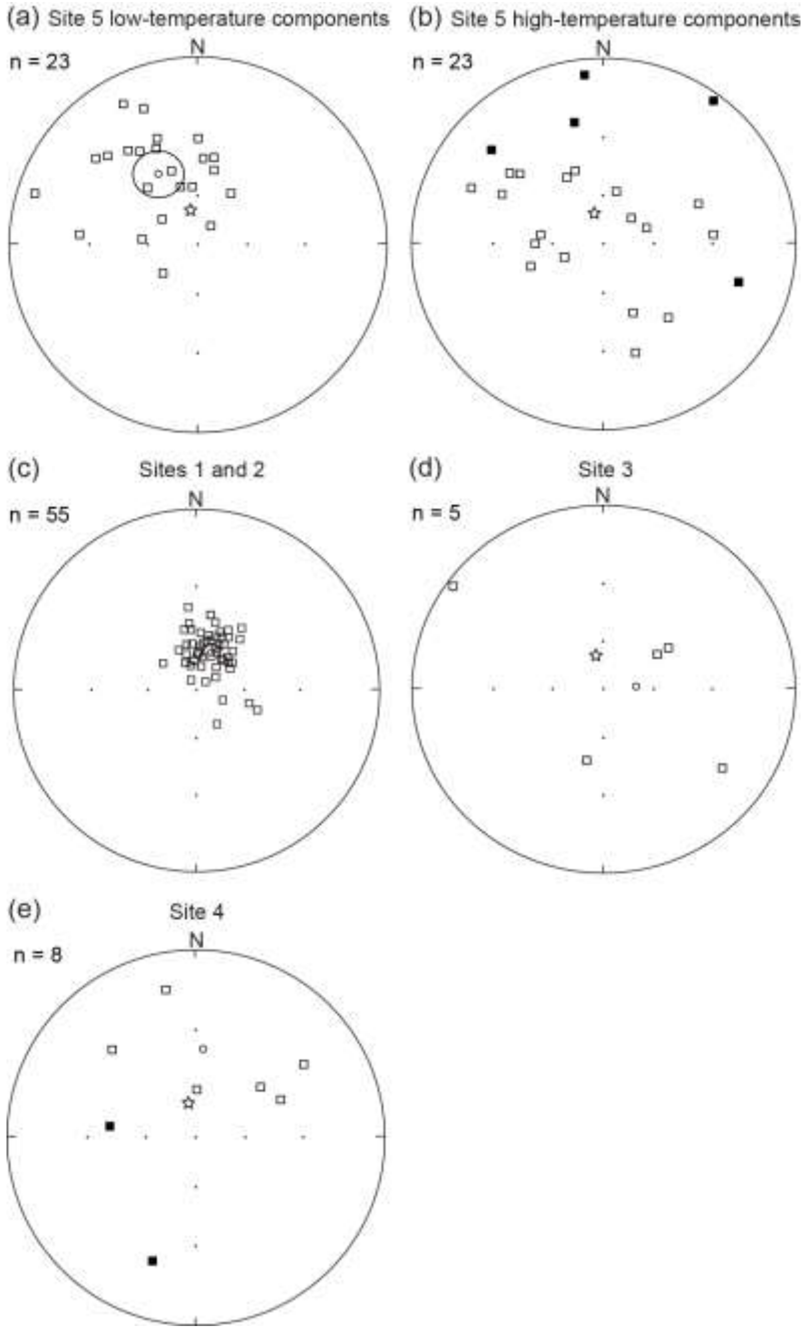

(a) Site 5 low-temperature components
n = 23
N
(b) Site 5 high-temperature components
n = 23
N
(c)
Sites 1 and 2
n = 55
N
(d)
Site 3
n = 5
N
(e)
Site 4
n = 8
N

Figure 5

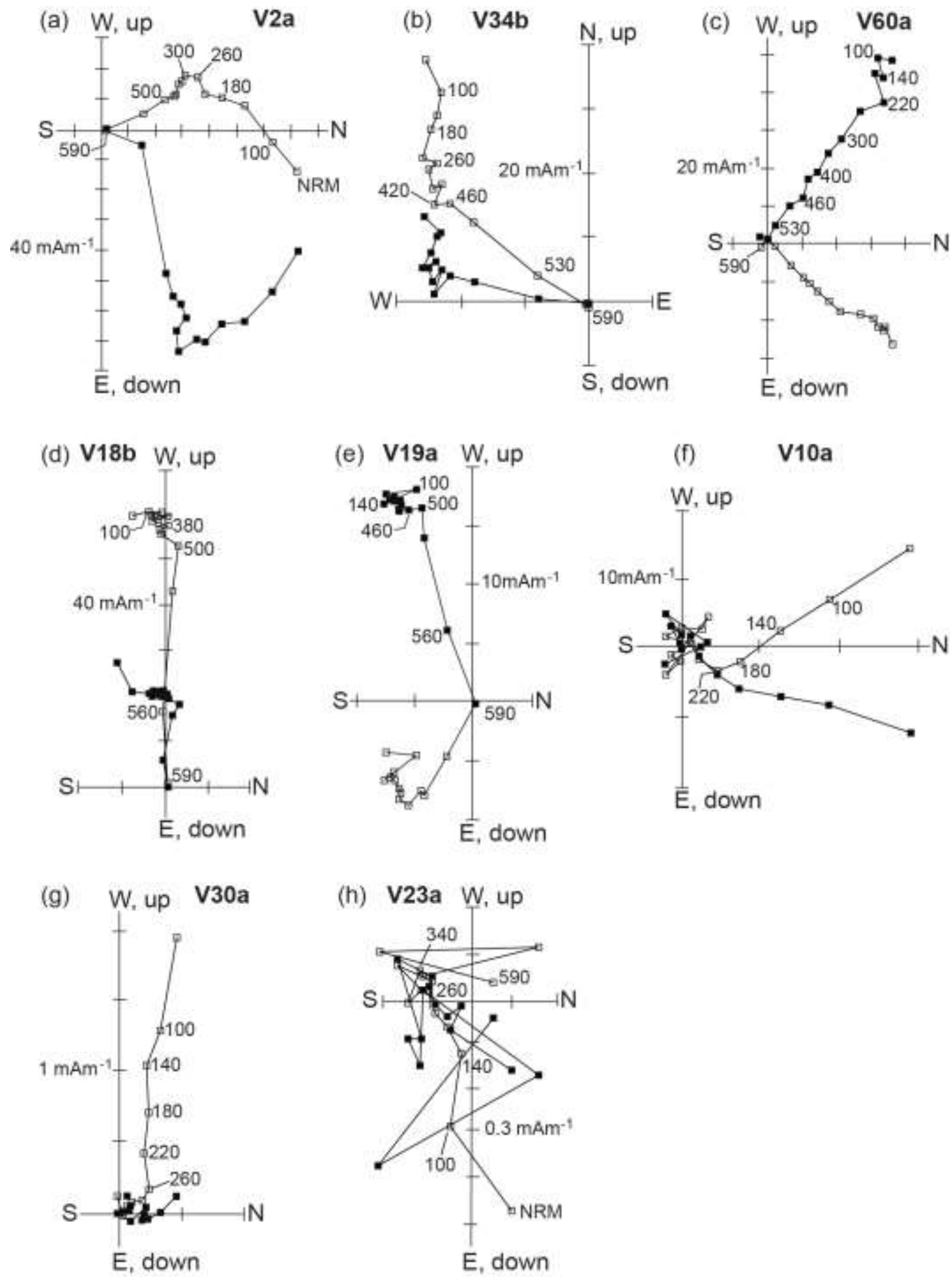

(a) V2a
W, up
E, down
S
N
300
260
500
180
590
100
NRM
40 mAm-1
(b) V34b
N, up
S, down
W
E
100
180
260
420
460
530
590
20 mAm-1
(c) V60a
W, up
E, down
S
N
100
140
220
300
400
460
530
590
20 mAm-1
(d) V18b
W, up
E, down
S
N
100
380
500
40 mAm-1
560
590
(e) V19a
W, up
E, down
S
N
100
140
500
460
10mAm-1
560
590
(f) V10a
W, up
E, down
S
N
10mAm-1
100
140
180
220
(g) V30a
W, up
E, down
S
N
100
140
1 mAm-1
180
220
260
(h) V23a
W, up
E, down
S
N
340
590
260
140
0.3 mAm-1
100
NRM

Figure 6

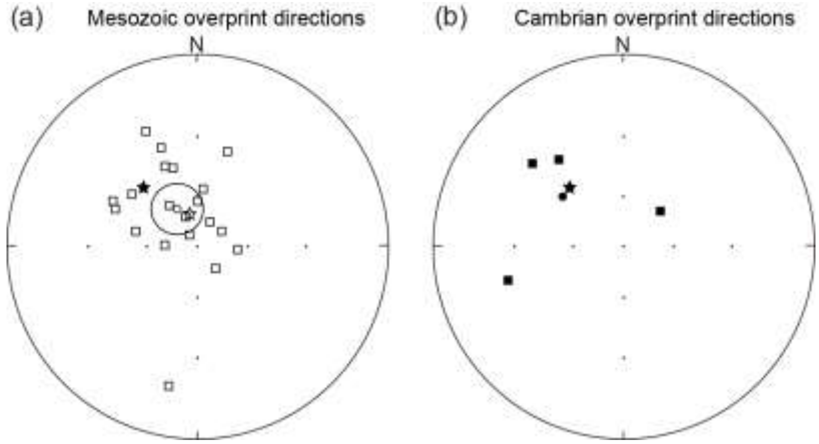
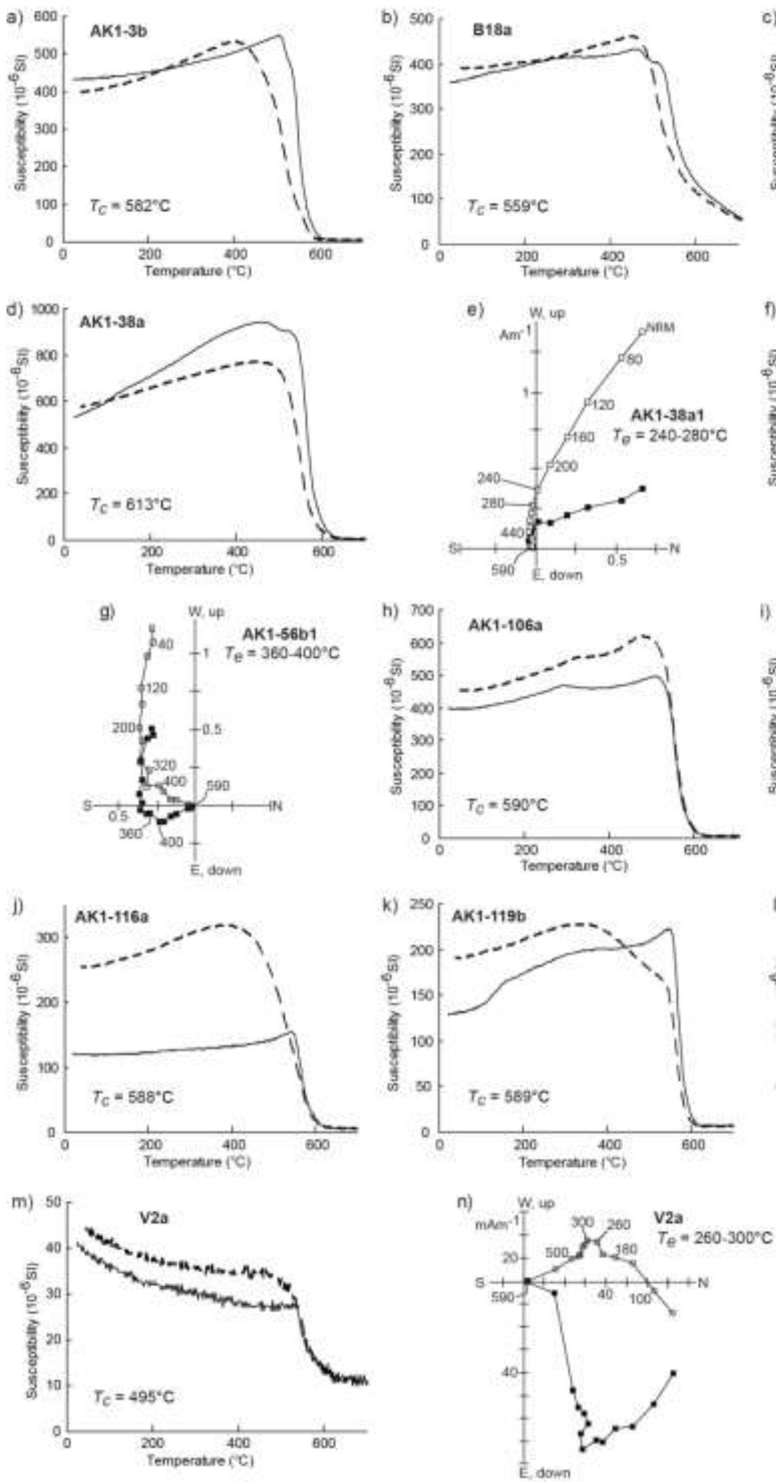

(a)
Mesozoic overprint directions
N
(b)
Cambrian overprint directions
N

Figure 7



a) AK1-3b
T_C = 582°C
b) B18a
T_C = 559°C
c) AK1-36b
T_C = 602°C
d) AK1-38a
T_C = 613°C
e) AK1-38a1
T_e = 240-280°C
f) AK1-56b
T_C = 588°C
g) AK1-56b1
T_e = 360-400°C
h) AK1-106a
T_C = 590°C
i) AK1-109a
T_C = 569°C
j) AK1-116a
T_C = 588°C
k) AK1-119b
T_C = 589°C
l) V60a
T_C = 505°C
m) V2a
T_C = 495°C
n) V2a
T_e = 260-300°C
Susceptibility (10^{-6}SI)
Temperature (°C)
W, up
E, down
S
N

Figure 8

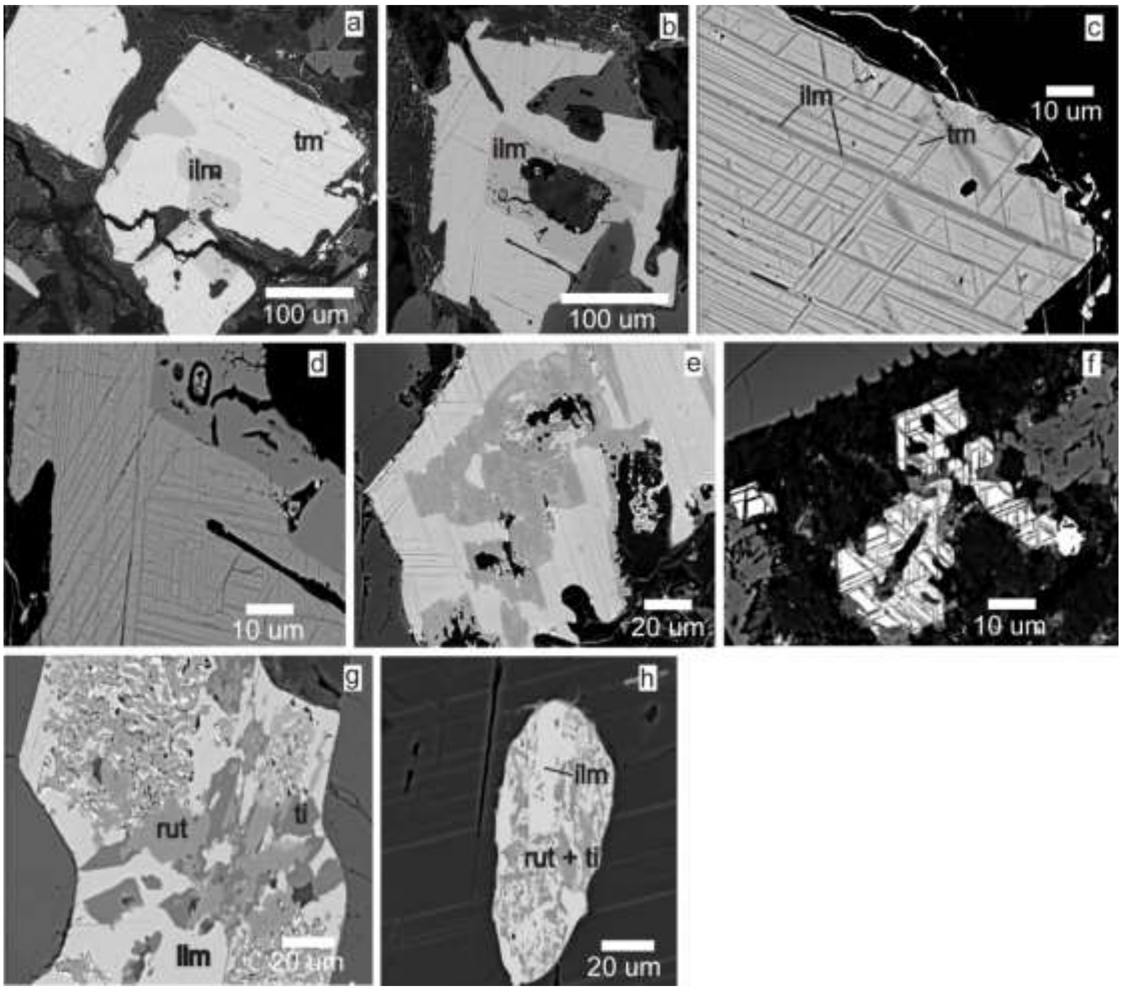
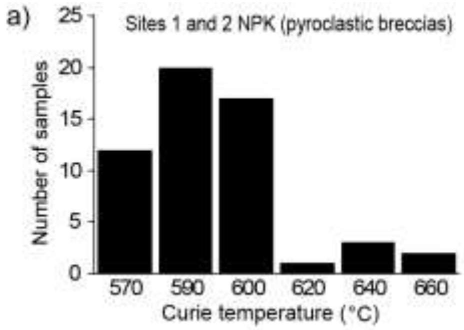

a
tm
ilm
100 um
b
ilm
100 um
c
ilm
tm
10 um
d
10 um
e
20 um
f
10 um
g
rut
ti
ilm
20 um
h
ilm
rut + ti
20 um

Figure 9


a)
Sites 1 and 2 NPK (pyroclastic breccias)
Number of samples
0
5
10
15
20
25
570
590
600
620
640
660
Curie temperature (°C)

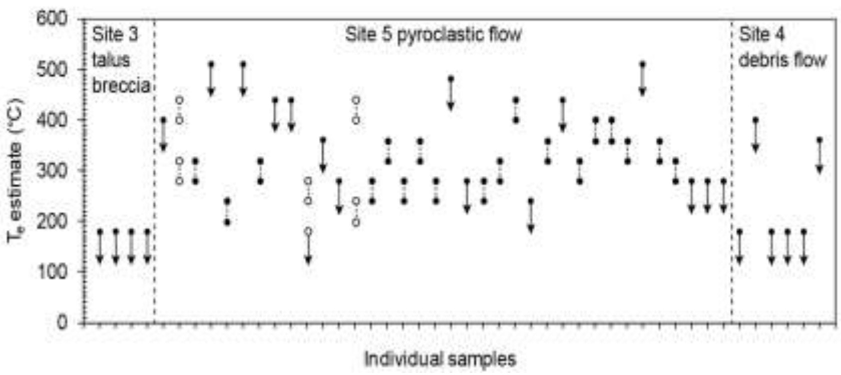
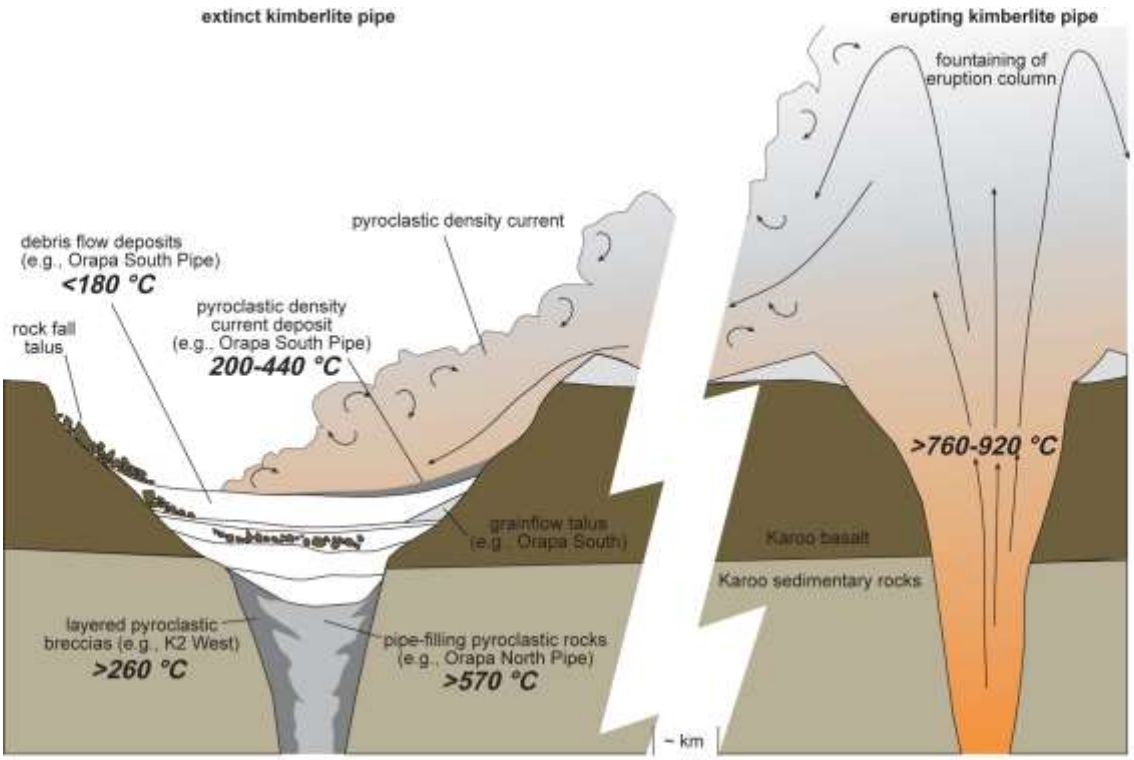

Site 3 talus breccia
Site 5 pyroclastic flow
Site 4 debris flow
T_e estimate (°C)
0
100
200
300
400
500
600
Individual samples

Figure 10



extinct kimberlite pipe
erupting kimberlite pipe
fountaining of eruption column
pyroclastic density current
debris flow deposits (e.g., Orapa South Pipe)
<180 °C
pyroclastic density current deposit (e.g., Orapa South Pipe)
200-440 °C
rock fall talus
grainflow talus (e.g., Orapa South)
>760-920 °C
Karoo basalt
Karoo sedimentary rocks
layered pyroclastic breccias (e.g., K2 West)
>260 °C
pipe-filling pyroclastic rocks (e.g., Orapa North Pipe)
>570 °C
~ km