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GR Focus Review

Magma flow in dyke swarms of the Karoo LIP: Implications for the mantle plume hypothesis

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ABSTRACT

The ~183 Ma old Karoo Large Igneous Province extends across southern Africa and is related to magmatism in Antarctica (west Dronning Maud Land and Transantarctic Mountains) and parts of Australasia. Intrusive events, including the emplacement of at least ten dyke swarms, occurred between ~183 Ma and ~174 Ma. We review here the field evidence, structure and geochronology of the dyke swarms and related magmatism as it relates to melt sources and the mantle plume hypothesis for the Karoo LIP. Specifically, the magma flow-related fabric(s) in 90 dykes from five of these swarms is reviewed, paying particular attention to those that converge on triple junctions in southern Africa and Antarctica. The northern Lebombo and Rooi Rand dyke swarms form an integral part of the Lebombo monocline, which converges upon the Karoo triple junction at Mwenezi, southern Zimbabwe. Dykes of the Northern Lebombo dyke swarm (182–178 Ma) appear to have initially intruded vertically, followed later by lateral flow in the youngest dykes. In dykes of the Okavango dyke swarm (178 Ma) there is evidence of steep magma flow proximal to the triple junction, and lateral flow from the southeast to the northwest in the distal regions. This is consistent with the Karoo triple junction and the shallow mantle being a viable magma source for both these dyke swarms. In the Rooi Rand dyke swarm (174 Ma) there is also evidence of vertical and inclined magma flow from north to south. This flow direction cannot be reconciled with the Karoo triple junction, as the northern termination of the Rooi Rand dyke swarm is in east-central Swaziland. The Jutulrøra and Straumsvola dyke swarms of Dronning Maud Land display evidence of sub-vertical magma flow in the north and lateral flow further south. The regional pattern of magma flow is therefore not compatible with direction expected from the Weddell Sea triple junction. The overall flow pattern in Karoo dykes is consistent with the triple junction being an important magma source. However, the Limpopo Belt and Kaapvaal Craton have significantly controlled the structure and distribution of the Lebombo and Save–Limpopo monoclines and the Okavango dyke swarm. The locus of magma flow in dykes of Dronning Maud Land is at least 500 km from the Karoo triple junction, as is the apparent locus for the Rooi Rand dyke swarm. In comparison with recent modelling of continental assembly, the structure and flow of the dyke swarms, linked with geochronology and geochemistry, suggests that thermal incubation during Gondwana assembly led to Karoo magmatism. A plate tectonic, rather than a fluid dynamic plume explanation, is most reasonably applicable to the development of the Karoo LIP which does not bear evidence of a deep-seated, plume source.

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1. Introduction

Globally, mantle plumes have been assumed to have been responsible for the development of voluminous large igneous provinces (LIPs) (Bryan and Ernst, 2008). Because LIPs represent relatively short-lived (<5 Ma), high volume magmatism (covering ~10^6 km^2), many of them are believed to have been formed predominantly by mantle plume processes; including the Siberian and Deccan Traps, Iceland, the Kerguelen Plateau, the Paraná–Etendeka flood basalt province and the ~183 Ma old Karoo LIP (Courtillot and Renne, 2003).

There are limitations, however, in the ability to definitively assess the role played by a mantle plume(s) in the development of an LIP or hot spot. For example, there is a growing body of geophysical evidence that no appreciable thermal anomalies exist beneath specific active hot spots, which is shedding light on the depth from which thermal anomalies may arise, and the relative thermal difference between mantle plumes and ambient mantle (e.g. Foulger and Natland, 2003; Korenaga, 2004; Deuss, 2007; Foulger, 2012).

Some workers have also shown that LIPs may be related to events of magma drainage (Silver et al., 2006) or processes fundamentally similar to the formation of non-LIPs (Cañón-Tapia, 2010). Indeed, the development of subduction zones at supercontinent margins can strongly influence thermal convection beneath the lithosphere (O'Neill et al., 2009) and high volumes and short eruption intervals can be consistent with shallow sources and tectonic triggering (Anderson, 2000). Thermal insulation of the mantle by supercontinent assembly has also been proposed as a viable mechanism for the development of thermal anomalies (Anderson, 1982, 1998; Lenardic et al., 2005; Coltice et al., 2007; Rolf et al., 2012).

A mantle plume origin for the Karoo LIP has been favoured for some time, particularly because of the dyke swarms that radiate outwards from the Karoo triple junction (Burke and Dewey, 1973; Campbell and Griffiths, 1990; Cox, 1992; Storey, 1995; Ernst and Buchan, 1997; Storey and Kyle, 1997; White, 1997; Storey et al., 2001) and the concentrically arranged and compositionally zoned primitive nephelinites and picrites centred on the Karoo triple junction (Bristow, 1984; Cox, 1992; White, 1997). The plume model continues to be supported by some advocates, particularly in the study of associated magmatism in Antarctica (Elliot and Fleming, 2000; Ferraccioli et al., 2005; Riley et al., 2005; Curtis et al., 2008). The study of magma flow directions in dykes associated with the Karoo triple junction also suggests that lateral magma flow from the triple junction has occurred (Ernst and Duncan, 1995; Aubourg et al., 2008; Hastie et al., 2011b), which is consistent with the mantle plume hypothesis because of dynamic uplift, and the high relative magma production rates at the plume centre (Campbell and Griffiths, 1990; Cox, 1992; Storey et al., 2001). There are, however, a number of geochemical (Sweeney and Watkeys, 1990; Sweeney et al., 1994; Jourdan et al., 2007a) and structural features (Le Gall et al., 2005; Klausen, 2009), as well as geochronology, which cannot be reasonably reconciled with a mantle plume origin (Jourdan et al., 2004, 2005, 2006, 2007b).

Lateral magma flow in dyke swarms has been used in the past to infer the presence of mantle plumes because of the assumed hydraulic gradient developed in the crust by uplift, and the high magma production rates at the plume centre (Ernst, 1990; Ernst and Baragar, 1992; Ernst and Duncan, 1995; Ernst and Buchan, 1997; Callot et al., 2001; Chaves and Correia Neves, 2005). Before the reliance on quantitative flow fabrics measured using anisotropy of magnetic susceptibility (AMS) and mineral shape-preferred orientation (SPO), the existence of a mantle plume was inferred by other means—such as the presence of triple junctions. The basic premise is that, similarly to dykes radiating outward from a volcanic edifice, dyke swarms that converge upon a central point (e.g. the Karoo triple junction, or the Central Atlantic magmatic province) were intruded laterally from a centralized mantle plume at that point of convergence (Ernst and Baragar, 1992; Ernst et al., 1995; Ernst and Buchan, 1997). Since the advent of AMS and mineral SPO, however, lateral magma flow in dyke swarms has been used to infer the presence of mantle plumes in a more quantitative way (Ernst and Duncan, 1995; Callot et al., 2001; Chaves and Correia Neves, 2005).

There is little doubt that the eventual dispersion of southern Gondwana at ~167 Ma occurred because of Karoo magmatism (Watkeys, 2002) or alternatively, resulted from incipient supercontinental break-up and dispersal. The vast network of sills and dykes provides good indicators of stresses in the crust at the time of their formation (Uken and Watkeys, 1997; Le Gall et al., 2005) and relative and/or absolute ages of dykes and magma flow determinations may help in piecing together the tectonic history of the Karoo LIP. In this regard we explore the implications of magma flow direction determined in dykes of the Northern Lebombo dyke swarm, Okavango dyke swarm and Rooi Rand dyke swarm in the context of the Karoo triple junction (Aubourg et al., 2008; Hastie et al., 2011b and previously unpublished data). These directions have been determined using AMS and the SPO of plagioclase grains.

In addition, various studies of the Jurassic-age dykes of Dronning Maud Land in Antarctica are discussed because of their apparently shared history of magmatism with the Karoo LIP (Zhang et al., 2003; Riley et al., 2005; Curtis et al., 2008). This provides additional constraint

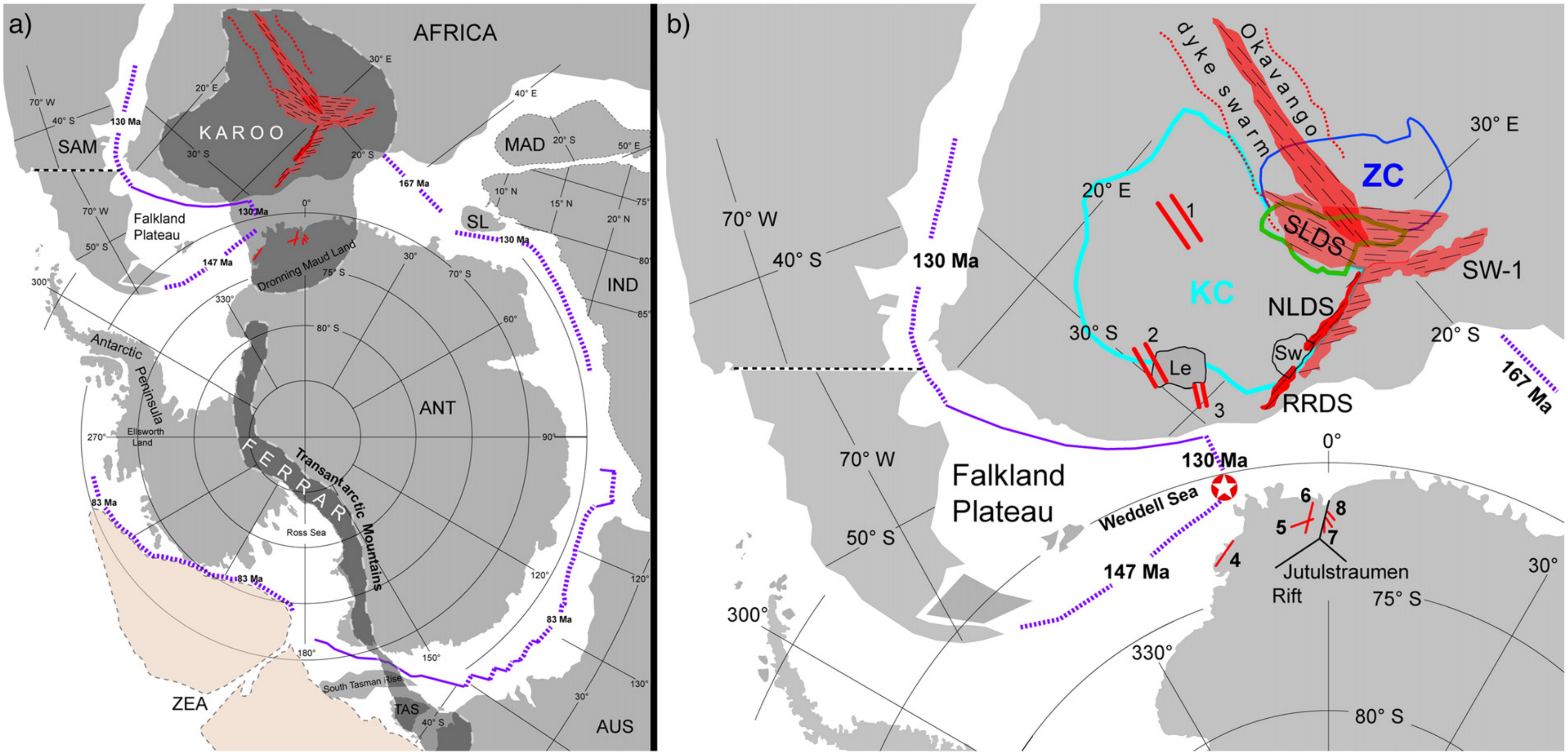

Fig. 1. Maps showing (a) the positions of the continents of southern Gondwana at ~180 Ma with the inferred extent of Karoo and Ferrar magmatism (dark grey with dashed white outlines) across southern Africa, Dronning Maud Land, and the remainder of the Transantartic Mountains and Australasia. The major dyke swarms of southern Africa and Dronning Maud Land are shown. Note that oceanic rifts and transform boundaries are shown in purple, with ages of the first known sea floor anomalies shown in Ma (ANT = Antarctica, AUS = Australia, IND = India, MAD = Madagascar, SAM = South America, SL = Sri Lanka, TAS = Tasmania, ZEA = Zealandia). The map shown in (b) illustrates more detail of the spatial arrangement of the dyke swarms relative to the Kaapvaal Craton (light blue outline, KC), Zimbabwe Craton (dark blue outline, ZC) and the Limpopo Belt (green outline). The dyke swarms shown include the NE trending Save–Limpopo dyke swarm (SLDS), the WNW trending Okavango dyke swarm, the N–S trending Northern Lebombo (NLDS) and Rooi Rand dyke swarms (RRDS) of the Lebombo monocline. The younger (~170 Ma?) SW-1 dyke swarm is also shown to the east of the Lebombo in Mozambique. The trends of the dykes are illustrated with dashes; note that the Save–Limpopo dyke swarm has a "kink" at the juncture where the younger Okavango dyke swarm has intruded. The Okavango dyke swarm has a densely intruded area (shaded) but reaches an overall width of ~300 km (delineated by red dashed line). The other dyke swarms shown are numbered as follows: 1 = Southern Botswana dyke swarm, 2 = Southern Lesotho dyke swarm, 3 = Underberg dyke swarm, 4 = Vestfjella dyke swarm, 5 = Group 1 dyke swarm, 6 = Alhmannryggen dyke swarm, 7 = Jutulrøra dyke swarm, 8 = Straumsvola dyke swarm.
Re-drawn from Fig. 4 of Veevers, 2012 (geometrical, non-historical construction in Lambert azimuthal equal-area projection) and White and McKenzie, 1989; Storey et al., 1992; Encarnación et al., 1996; Storey and Kyle, 1997; Watkeys, 2002; Jourdan et al., 2004; Ferraccioli et al., 2005; Riley et al., 2006; Curtis et al., 2008.

on the cause of magmatism in southern Gondwana as there is overlap in composition and age between magmatism associated with the Karoo triple junction and dyke swarms that occur in Dronning Maud Land.

We attempt to bring together the existing data related to (1) the timing and duration of Karoo and associated magmatism, (2) the mantle plume hypothesis for the development of the Karoo LIP and (3) magma flow directions which have been found in dykes of the Karoo LIP and associated magmatic provinces. We aim, therefore, to constrain magma flow in dykes of the Karoo LIP and the timing of dyke formation in order to critically assess the mantle plume hypothesis in the context of the Karoo LIP.

2. The Karoo Large Igneous Province

2.1. Regional distribution

The Karoo LIP comprises a large number of volcanic and intrusive components, which in South Africa alone have been studied for more than 80 years (e.g. Du Toit, 1929; Cox et al., 1967; Duncan et al., 1984; Ellam et al., 1992; Jourdan et al., 2009). There has not been a collation of the existing and more recent work on the Karoo LIP since Erlank (1984), but readers are referred to Jourdan et al. (2009) who provide an updated synthesis of the current understanding of Karoo magmatism.

The Karoo LIP extends across southern Africa, covering ~3×10^6 km^2 (Eales et al., 1984). The Karoo LIP is contemporaneous with other igneous provinces, including the Ferrar Province of Antarctica (Fig. 1a) (Kyle et al., 1981; Encarnación et al., 1996; Zhang et al., 2003), the associated Kirkpatrick basalts and various intrusive components of west Dronning Maud Land and the Tasman dolerites of Australasia (Elliot, 1975; Fleming et al., 1995).

Geochronology indicates that the bulk of the magmatism occurred between ~183 Ma and ~178 Ma, but continued up to ~174 Ma with the intrusion of the RRDS in the southern Lebombo "monocline" (Duncan et al., 1997; Jourdan et al., 2005, 2007b,c). The Lebombo is in fact a N–S striking, easterly dipping structure comprising a highly magmatic and rifted volcanic margin (Watkeys, 2002; Klausen, 2009).

The voluminous, low-Ti, tholeiitic magmas which characterise much of the Karoo LIP erupted within 3–4.5 Ma as continental flood basalts (CFBs) now preserved mainly in Lesotho and western Botswana (Jourdan et al., 2007a,c). The volcanic sequences of the Lebombo reach a thickness of 1.5–12 km and overlie the uppermost arenaceous rocks of the Karoo Supergroup (Clarens Formation) (Eales et al., 1984; Watkeys, 2002). There is also an associated network of dolerite sills extending throughout the Karoo basin of South Africa (not shown).

The dyke swarms associated with the Karoo LIP have become increasingly studied as they provide structural, geochemical and geochronological information about their own development, and that of the entire Karoo. These include, but are not limited to, the Okavango dyke swarm (Reeves, 1978; Elburg and Goldberg, 2000; Reeves, 2000; Le Gall et al., 2002; Jourdan et al., 2004; Le Gall et al., 2005; Aubourg et al., 2008) and the Northern Lebombo and Rooi Rand dyke swarms of the Lebombo monocline (Saggerson et al., 1983; Armstrong et al., 1984; Watkeys, 2002; Jourdan et al., 2006; Klausen, 2009; Hastie et al., 2011a,b) (Fig. 1a, b). The dyke swarms and Karoo triple junction are discussed in further detail in Section 2.3.

Along with the Okavango dyke swarm and Lebombo monocline, another dyke swarm, the Save–Limpopo dyke swarm (Jourdan et al., 2006) converges on Mwenezi (formerly Nuanetsi) in southern Zimbabwe (Fig. 1). The Save–Limpopo dyke swarm is found primarily within the Sabi monocline (or Save–Limpopo monocline) which developed during normal faulting in the Permian (Watkeys, 2002). The intersection of the monoclines and the Okavango dyke swarm produces a conspicuous triple junction. This geometry has been used (other than compositional and geochemical data) as evidence for the role played by a mantle plume in its formation (Burke and Dewey, 1973; White and McKenzie, 1989; Campbell and Griffiths, 1990; Cox, 1992; Ernst and Buchan, 1997; Storey and Kyle, 1997; White, 1997). Before further discussion of the timing and duration of the eruptive and intrusive phases it is necessary to examine the distribution and relevant petrology of the rock types within the Karoo LIP.

2.2. Igneous stratigraphy and petrology

Volcanic rocks of Karoo LIP range from earliest nephelinites and picrites to continental tholeiites and rhyolites (Eales et al., 1984). There has been much geochemical study performed on the compositionally diverse igneous rocks of the Karoo LIP (Cox et al., 1967; Duncan et al., 1984; Sweeney and Watkeys, 1990; Hergt et al., 1991; Ellam et al., 1992; Sweeney et al., 1994; Riley et al., 2006). These findings have a significant bearing on the mantle plume model for the Karoo LIP, and are summarised below.

The most primitive rock types are centred on the Karoo triple junction (Fig. 2) and consist of the incompatible element enriched Mashikiri Formation nephelinites and the Letaba Formation picrites (Bristow, 1982, 1984). The triple junction, at the northern termination (Mwenezi) of the Lebombo monocline, is within the Limpopo Belt (discussed further in Section 2.3) while the remainder of the N–S trending structure occupies a position along the eastern edge of the Archaean Kaapvaal Craton (Fig. 1b).

The Mashikiri Formation overlies the arenaceous Clarens Formation, which in the northern Lebombo directly overlies Precambrian basement. The Mashikiri Formation is not as laterally extensive north-to-south as the overlying volcanics of the Lebombo Group, which include the Letaba Formation picrites and Sabie River Formation basalts. There are units of rhyolite (flows and intrusions) within the basalts of the Sabie River Formation known as the Olifants Beds (Bristow, 1982; Riley et al., 2004). The youngest volcanics of the Lebombo are the Jozini and Mbuluzi Formation rhyolites and rhyodacites (Watkeys, 2002; Klausen, 2009).

The Mashikiri nephelinites display enrichment of incompatible trace elements which reflect melting of an ancient, metasomatically-enriched sub-continental lithospheric mantle (Ellam and Cox, 1989; Hawkesworth et al., 1999; Jourdan et al., 2007a). The Letaba Formation picrites (MgO 10–24%) of the northern Lebombo monocline are more voluminous than the nephelinites and reach a maximum thickness of 4 km. They may have been derived from the sub-

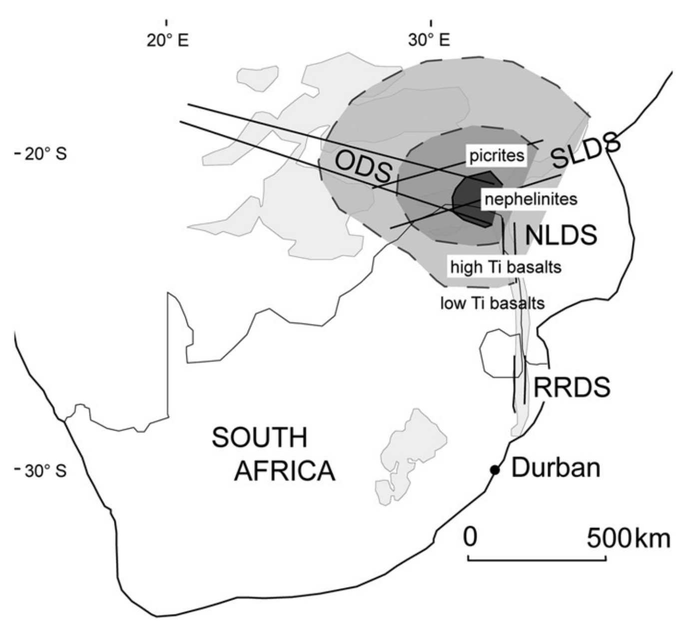

Fig. 2. Schematic map of the outcrop of the Karoo LIP (lightest grey) in southern Africa, illustrating the concentric distribution of nephelinites, picrites and high- and low-Ti basalts with respect to the Karoo triple junction centred on Mwenezi, southern Zimbabwe. Re-drawn from Jourdan et al., 2004; Klausen, 2009.

continental lithospheric mantle (Bristow et al., 1984; Ellam and Cox, 1989) or from mixing between asthenospheric mantle and sub-continental lithospheric mantle (Ellam and Cox, 1991; Sweeney et al., 1991) or from a heterogeneous source (Ellam, 2006).

The voluminous continental flood basalts and associated dykes and sills which followed the primitive volcanics at ~182 Ma are low in MgO (2% < MgO < 9%) and possess a variation in Ti and Zr across southern Africa. Specifically, basalts in the northern Lebombo and further north into Zimbabwe are high-Ti ($TiO_2 > 2\%$), while basalts of the southern Lebombo, Lesotho, central Botswana and central Namibia are low-Ti (Fig. 2) (Cox et al., 1967; Duncan et al., 1984; Jourdan et al., 2007a) and those of the Lebombo have calc-alkaline affinities (low-K tholeiites) (Duncan, 1987). This distinction is attributed to differences in melting, resulting in 30–40% of the incompatible elements in the high-Ti basalts being derived from the lithospheric mantle (Sweeney and Watkeys, 1990; Sweeney et al., 1994). Indeed Ellam and Cox (1989) have shown, using Sm–Nd systematics that the Karoo yields a Proterozoic eruption "isochron" which demonstrates mixing in the melt source region. This is also evident from the background levels of crustal contamination and generally MORB-like signature of incompatible elements (Cox and Bristow, 1984; Hawkesworth et al., 1984). Bristow et al. (1984) also found evidence of a heterogeneous mantle source in the bulk of the Karoo basalts in the $^{87}Sr/^{86}Sr$ ratios, although these authors found no convincing evidence for interaction of the melts with crustal material. The low-Ti basalts were possibly derived from an asthenospheric source which equilibrated with refractory lithospheric mantle (Sweeney et al., 1991). The voluminous basaltic volcanism was followed by the eruption of the Jozini and Mbuluzi Formation rhyolites which cap the Karoo volcanics of the Lebombo monocline (Bristow, 1982). This was followed by the intrusion of the Rooi Rand dyke swarm in the southernmost region of the Lebombo (Armstrong et al., 1984) at ~174 Ma.

The Ferrar Province is exposed along the length of the Transantarctic Mountains and comprises the Dufek intrusion as well as massive, laterally extensive dolerite sills (Elliot et al., 1999) and pyroclastics capped by the Kirkpatrick basalts (Kyle et al., 1981; Elliot and Fleming, 2000). The Ferrar Province is dominated by low-Ti tholeiitic basalts which were most likely derived from a lithospheric mantle source (Hergt et al., 1991). The extensional rift-like structure along which the Ferrar magmas intruded has been suggested as the reason for its linear extent (Storey et al., 1992) (Fig. 1a). It has also been suggested that the Ferrar is indicative of a linear melting anomaly caused by a long-lived subduction zone on the southern margin of Gondwana (Cox, 1992; Storey, 1995). However, some workers argue that the Ferrar and related magmas were emplaced from the Weddell Sea triple junction (Fig. 1b) (Elliot and Fleming, 2000, 2004; Leat, 2008; Luttinen et al., 2010). The Weddell Sea triple junction only developed oceanic crust at ~130 Ma, but may have been a point source for earlier injection of dykes in Dronning Maud Land [the Weddell Sea triple junction may correlate with the present Bouvet or Shona triple junctions, see Storey et al. (2001) for discussion]. Leinweber and Jokat (2012) have shown that the Weddell Sea triple junction also resulted in the formation of the Agulhas plateau, following which it moved south-westward as South America rotated away from Africa. The role of the Weddell Sea triple junction is discussed in further detail in Section 4.

The mafic rocks of Dronning Maud Land are comparable to those in southern Africa with two varieties, DM1 and CT1 (Luttinen et al., 1998; Riley et al., 2005 respectively) being similar to the low-Ti compositions of southern Africa (DM = Dronning Maud, CT = chemical type). Luttinen et al. (1998) provide general geochemical and isotopic data from the basalts and dykes of Vestfjella, in Dronning Maud Land. Using Sr and Nd isotopes, they differentiate 4 magma types (CT1–CT4) which can be identified in related lavas and dykes. Dolerite dykes belonging to the CT4 magma type are the only OIB-like (ocean island basalt) magmas known from the Karoo LIP. The CT1 magma type is tholeiitic, most likely derived from an Archaean, sub-lithospheric mantle source. The CT2 and DM2 magma types have MORB affinities similar to the Rooi Rand dyke swarm and most likely derive from an asthenospheric source. There is also a high-Ti basalt composition (Group 4 basalts of Riley et al., 2005). The ferro-picrite CT3 magma type does not have a comparable composition within the Karoo LIP of southern Africa and is thought to be derived from early mantle plume melts (Riley et al., 2005).

The evidence for structural control on the distribution of basaltic compositions in southern Africa is found in compositional changes (e.g. low-Ti to high-Ti) that occur across basement boundaries through which the basalts erupted (Proterozoic mobile belts vs. Archaean cratons) (Cox et al., 1967; Duncan et al., 1984; Watkeys, 2002; Jourdan et al., 2004, 2006). Field relationships and geochronology show that the lithospheric architecture has also controlled the development of the Karoo triple junction and associated dyke swarms (Watkeys, 2002; Jourdan et al., 2004; Le Gall et al., 2005).

This has also been shown in Dronning Maud Land by Luttinen et al. (2010) who identify Grunehogna province magmas and Maud province magmas on the basis of geochemistry and Nd and Sr isotope data. In brief, the authors indicate that the Grunehogna magma type shows evidence of recycled oceanic crust in the parent magma, and that the magmas most likely derived from the partial melting of eclogite-bearing asthenospheric mantle. Contamination has occurred, with high-Ti types showing evidence of lithospheric mantle contamination and low-Ti types showing evidence of crustal contamination. The Maud province magma type, however, derived from relatively low-pressure partial melting of lithospheric mantle associated with rifting in the region of the Weddell Sea triple junction. This is a broadly similar pattern to the distribution of magma types across the southern African cratonic regions (high-Ti) and the central (Lesotho) Karoo and Botswana areas (low-Ti).

2.3. Dykes of the Karoo and associated LIPs

There are three main dyke swarms associated with the Karoo triple junction (Save–Limpopo, Northern Lebombo and Okavango dyke swarms; Jourdan et al., 2004), four more isolated swarms (Rooi Rand, Underberg, Southern Lesotho and Southern Botswana dyke swarms; Armstrong et al., 1984; Riley et al., 2006; Jourdan et al., 2007a) and at least five dyke swarms in Dronning Maud Land (Curtis et al., 2008) (Fig. 1b).

The Karoo triple junction at Mwenezi is situated within the Limpopo Belt and comprises both the earliest (nephelinites, picrites) and latest (Mwenezi igneous complex) manifestations of Karoo magmatism. The Limpopo Belt comprises poly-metamorphosed rocks that represent an Archaean collision between the Kaapvaal and Zimbabwe Cratons. These deformed cratonic and volcano-sedimentary rocks have a strong ENE-trending fabric which has been exploited by both faulting (Watkeys, 2002) and dyking (Uken and Watkeys, 1997; Jourdan et al., 2004) since the Proterozoic.

The ENE-trending Save–Limpopo dyke swarm comprises predominantly fine to medium grained dolerite dykes emplaced within the central and northeastern regions of the Limpopo Belt, extending for ~600 km from SE Botswana (the Tuli basin) to the NE of the Limpopo Belt. The Save–Limpopo dyke swarm is 50–100 km wide and comprises vertical to sub-vertically dipping dykes. These dykes have been dated to 180.4 ± 0.7–178.9 ± 0.8 Ma although a significant proportion of the dykes are Proterozoic in age (Le Gall et al., 2002; Jourdan et al., 2005, 2006). Field relationships indicate that this dyke orientation (ENE to NE) predate dykes of the Northern Lebombo and Okavango dyke swarms, particularly evident from the picritic dykes of the Mwenezi region having intruded in this orientation (Watkeys, 2002).

The Northern Lebombo dyke swarm is hosted by the basaltic and other volcanic units, as well as the Clarens Formation, of the Lebombo. Dykes of the N–S trending Northern Lebombo dyke swarm comprise several generations of feeder dykes which can be directly correlated with volcanic units of the Sabie River Formation basalts (D1 and D2 generation), the Jozini Formation (D3 generation) and possibly the Movene Formation basalts (D4 dykes) within the Lebombo monocline (Klausen, 2009). In other areas, such as the Okavango and Rooi Rand, the dykes intrude the basalts of the Karoo, and have no known extrusive equivalents. The D1 to D4 nomenclature is indicative of relative dyke ages, such that D4 dykes are the youngest, and tend to strike NW, similar to the Okavango dyke swarms. Two radiogenic ages of 181.4 ± 0.7 and 182.3 ± 1.7 Ma have been found for the Northern Lebombo dyke swarms (Jourdan et al., 2005). As yet, no Proterozoic age dykes have been recognised in the Northern Lebombo dyke swarm, suggesting that the N–S dyke trend developed in the Lebombo monocline during the Jurassic in response to E–W extension (Watkeys, 2002).

The WNW-trending Okavango dyke swarm is ~1500 km in length, reaching ~300 km in width where it overlaps with the northernmost Lebombo and the Save–Limpopo dyke swarm (Jourdan et al., 2006) (Fig. 1). It converges with the Lebombo monocline and Save–Limpopo dyke swarm at Mwenezi (Jourdan et al., 2004). Dykes of the Okavango dyke swarm have intruded Precambrian basement (central and northwestern regions of the Limpopo Belt) and sedimentary and volcanic rocks of the Karoo Supergroup within the Tuli basin (Smith, 1984; Elburg and Goldberg, 2000; Le Gall et al., 2005; Aubourg et al., 2008). The orientation of the Okavango dyke swarm has been recognised as essentially a reactivation of an older structural trend evident in both the Kaapvaal and Zimbabwe Cratons (Watkeys, 2002). From field evidence, Le Gall et al. (2005) inferred a NNW–SSE dilation direction for the dykes, a similar direction to that inferred for the Save–Limpopo dyke swarm.

The Okavango dyke swarm was first described as a post-Karoo (Cretaceous) dyke swarm associated with a failed rift axis (Reeves, 1978, 2000), i.e. an aulacogen, while Uken and Watkeys (1997) considered the Okavango dyke swarm to be Karoo in age. Indeed, subsequent work has shown a predominantly Karoo-age with ~13% of the dykes being Proterozoic in age (Elburg and Goldberg, 2000; Le Gall et al., 2002, 2005). Dykes of the Okavango dyke swarm have been dated to 179 ± 1.2–178.4 ± 1.1 Ma, with the Proterozoic component providing ages of 851 ± 6–1672 ± 7 Ma (Jourdan et al., 2004). The dykes are doleritic in composition; dominated by plagioclase (35–45%), clinopyroxene (20–35%) and Fe–Ti oxides. Both high-Ti and low-Ti varieties occur (Elburg and Goldberg, 2000; Aubourg et al., 2008).

The youngest dyke swarm in the Karoo LIP is the Rooi Rand dyke swarm (173.9 ± 0.7 Ma) which post-dates the main Karoo flood basalts (Jourdan et al., 2007b,c). The MORB-like Rooi Rand dyke swarm is a N–S trending dyke swarm found in the southern Lebombo monocline, extending ~180 km from the Msunduze River in KwaZulu-Natal northwards to central Swaziland. The 10–22 km thick swarm intruded the Sabie River Formation and Beaufort Group, just to the west of the Lebombo monocline (Marsh, 1987, 2002; Watkeys, 2002). The steeply dipping (>80°), generally N–S striking dykes of the Rooi Rand dyke swarm in the central area give way to more shallowly dipping (50°–70°) NNE–SSW striking dykes in the north (in Swaziland). The Rooi Rand dyke swarm most likely originated from the melting of an upwelling asthenosphere, as is typical during lithospheric rupture in the early stages of continental break-up (Saggerson et al., 1983; Armstrong et al., 1984; Meth, 1996). There is evidence that a set of seaward-dipping reflectors developed at this time in the region between southern Africa and Antarctica with a later set during the development of the Explora Wedge at ~130 Ma during the parting of Antarctica and the Mozambique Ridge (Cox, 1992; Luttinen and Furnes, 2000; Leinweber and Jokat, 2012). As previously mentioned, the CT2 and DM2 magma types of Dronning Maud Land have MORB affinities similar to the Rooi Rand dyke swarm (Luttinen et al., 1998) which is consistent with the coeval development of the Lebombo monocline and the aforementioned seaward-dipping reflectors.

There are additional dyke swarms in southern Africa which require introduction, as they are relevant to the tectonomagmatic characteristics of the Karoo LIP. These are the southern Botswana and southern Lesotho dyke swarms and the NW–SE striking Underberg dyke swarm (Jourdan et al., 2004) (Fig. 1b). The southern Botswana and Southern Lesotho dyke swarms are shown only in the interest of completeness, because magma flow and geochronological studies have not been conducted on these swarms.

Dykes of the Underberg dyke swarm (Fig. 1b) are fine- to medium-grained dolerites with intergranular and/or sub-ophitic textures. The Underberg dyke swarm intruded sedimentary sequences of the Triassic Beaufort Group and the overlying Molteno, Elliot and Clarens Formations. The Underberg dyke swarm is geochemically similar to the low-Ti basalts of Lesotho (and the Southern Lesotho dyke swarm); although field relationships and geochronology confirm that the dykes are younger than the Lesotho basalts. Riley et al. (2006) provide an age of ~176 Ma for the intrusion of the Underberg dyke swarm. These authors have found that the strike of the dykes is remarkably uniform (130–140°), which differs slightly from the Okavango and Southern Botswana dyke swarms (110°–120°). The Underberg dyke swarm was derived from sub-lithospheric melts involving some crustal contamination (Riley et al., 2006). AMS measurements have been undertaken on only three dykes, and therefore we remain sceptical of the regional significance of the measurements.

In addition, there is the Olifants River dyke swarm (not shown) which was once thought to be related to the Karoo LIP, but has been shown to be older than ~800 Ma (Marsh, 2002; Jourdan et al., 2006). The Olifants River dyke swarm extends south-westward from the Lebombo monocline following an Archaean/Proterozoic dyking direction (Uken and Watkeys, 1997; Watkeys, 2002). It is common to find that NE–SW striking dykes are truncated by NW–SE striking dykes—the same orientation of the Okavango dyke swarm. For example, the Save–Limpopo dyke swarm (Fig. 1b) overlaps with the Okavango dyke swarm and has been dated to 180.4 ± 0.7–178.9 ± 0.8 Ma (Le Gall et al., 2002; Jourdan et al., 2005, 2006) although Proterozoic ages have been found (728 ± 3–1683 ± 18 Ma, n = 14). Karoo dykes generally do not occur in the 0°–40° orientation, which is best explained by the SW–NE strike of the Proterozoic age dykes of the Save–Limpopo dyke swarm. The geochronology of the Karoo-age dykes is considered in further detail in Section 3.

Karoo-age dyke swarms in Dronning Maud Land include the Alhmannryggen (178 Ma), Straumsvola (178–176 Ma) and Vestfjella (177 Ma) dyke swarms. There are two older dyke swarms, the Group 1 dykes of the Alhmannryggen region (~190 Ma) (Riley et al., 2005) and the Jutulrøra dyke swarm (~205 Ma). The Jutulrøra dyke swarm consists predominantly of low-Ti tholeiitic dykes that trend NNW–SSE (Fig. 1b). The Straumsvola dyke swarm, which is mainly restricted to the Straumsvola area (Fig. 1b), is predominantly doleritic in composition and was emplaced between 178.5 Ma and 174.8 Ma (concordant ^{40}Ar/^{39}Ar isochron age of 174.8 ± 2.0 Ma). Heinonen et al. (2010) have shown that the Vestfjella dyke swarm comprises two geochemically distinct suites—an enriched, OIB-like type and a depleted, MORB-like type.

The youngest dyke phase of the Straumsvola dyke swarm yielded an age of 170.9 ± 1.7 Ma (younger than the Lebombo rhyolites). These younger dykes, including phonolitic and lamprophyric compositions, cross-cut the dolerite dykes. Curtis et al. (2008) propose two distinct phases of mafic dyke emplacement rather than a protracted magmatic history. From field, structural and AMS data these authors assert that the dykes were sourced locally. For example, power-law distribution of the dyke thicknesses and spacing indicates that the dykes generally become more widely spaced ~25 km south of Straumsvola. This may reflect a spatial variation in magma pressure, such that the Straumsvola dyke swarm was emplaced under higher magmatic pressure than the Jutulrøra dyke swarm; the restricted orientations thereof indicating that the Jutulrøra dyke swarm was bound to the stress field of the host rock and not the magma pressure driving emplacement.

Despite attributing the origin of these dykes to a mantle plume (Curtis et al., 2008), it is evident that the majority of these dykes are low-Ti tholeiites, similar to the bulk of the Karoo mafic dykes and lava

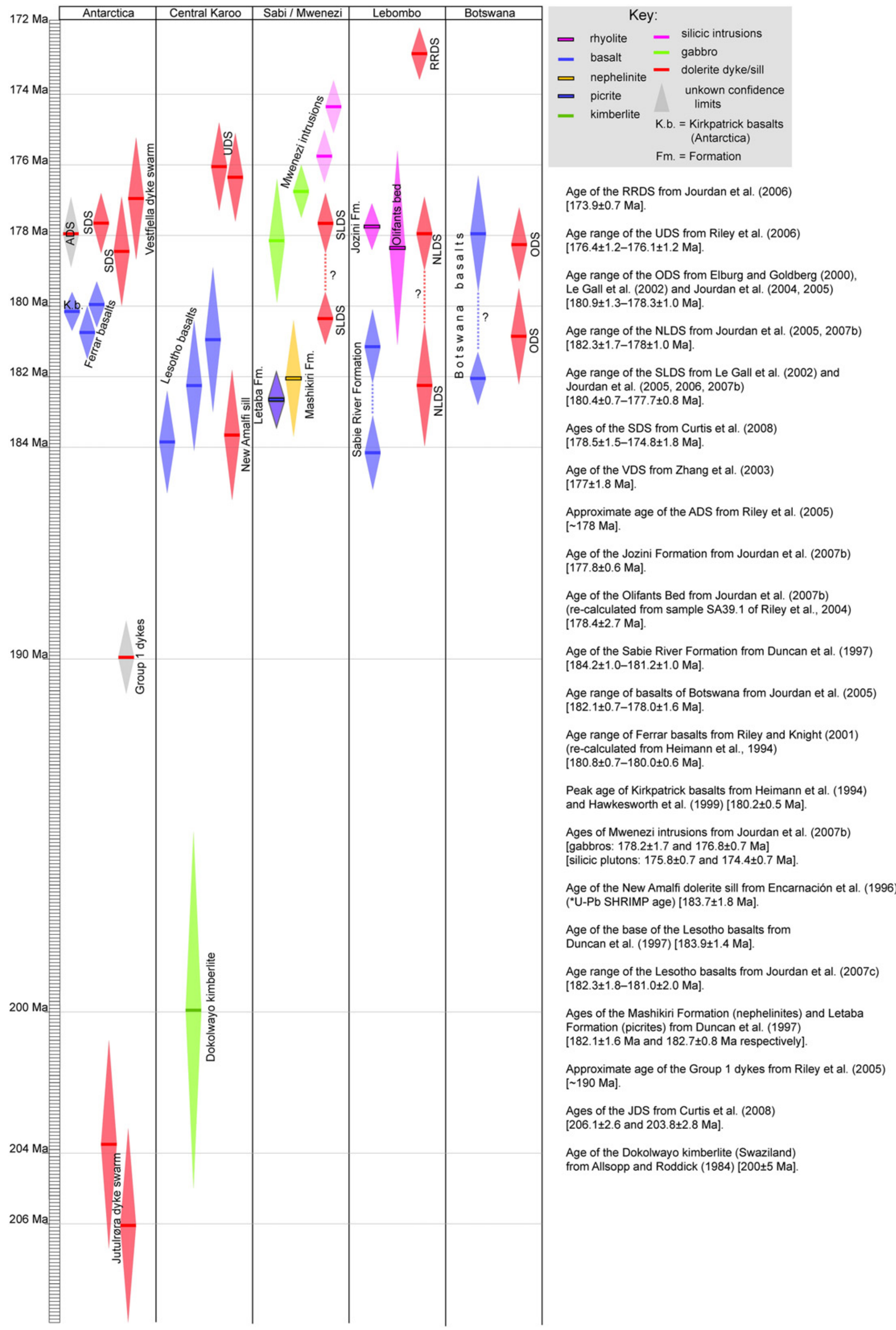

Fig. 3. Chronostratigraphy of the major regions of the Karoo LIP and associated igneous provinces. Note that ages are shown in solid colours with lighter shaded error bars. The smallest age increments are 0.1 Ma. RRDS = Rooi rand dyke swarm, UDS = Underberg dyke swarm, ODS = Okavango dyke swarm, SLDS = Save–Limpopo dyke swarm, NLDS = Northern Lebombo dyke swarm, SDS = Straumsvola dyke swarm, VDS = Vestfjella dyke swarm, ADS = Alhmannryggen dyke swarm, JDS = Jutulrøra dyke swarm. See Section 3.2 for discussion of age reliability.

flows. Interestingly, these compositions are also found off-craton in Antarctica in the same manner as the Grunehogna magma type (Luttinen et al., 2010) and the low-Ti tholeiites of the southern Lebombo (Sweeney and Watkeys, 1990).

An important structural feature of this region of Dronning Maud Land is the Jutulstraumen rift system (triple junction) first described by Ferraccioli et al. (2005). It was discovered beneath the Antarctic ice sheet in west Dronning Maud Land and converges on Neumayerskarvet, which is near the southern termination of the Alhmannryggen dyke swarm (Fig. 1b). The convergent rifts of this triple junction are occupied by Jurassic-age volcanic rocks and alkaline and tholeiitic intrusions (Ferraccioli et al., 2005; Curtis et al., 2008). Riley et al. (2005) observed in the Ahlmannryggen region that initial crustal dilation caused by dyking was N–S oriented (190 Ma) followed at 178 Ma by regional NW–SE oriented dilation. Ferraccioli et al. (2005) demonstrate, however, that extension in these rifts cannot be accounted for by dyke dilation alone, and suggest considerable crustal thinning prior to, or post-dyking. These authors also argue that the triple junction geometry and the ferro-picrite composition of certain dykes are consistent with derivation from an early mantle plume (~190 Ma?), although the Jurassic dyke swarms described above do not coincide geometrically with these rifts. The same is clear in the dyke swarms of the Karoo LIP (e.g. the Rooi Rand, Southern Lesotho and Underberg dyke swarms).

There is a "new" dyke swarm that has been recognised geophysically along the eastern length of the Lebombo monocline and Mozambique coastal plain that is known as the SW-1 dyke swarm (or Mozambique dyke swarm) (Mekonnen, 2004). There is virtually nothing known about or published regarding this swarm, although its prominent NNE strike direction (Fig. 1b) and presence within thinned crust of the Mozambique plains suggests that it intruded later (~170 Ma) than the Karoo LIP in response to extension between SE Africa and Antarctica (Jokat et al., 2003; Mekonnen, 2004). Note that the Gap dyke swarm and Southern Malawi dyke swarms (Jourdan et al., 2004, 2006) are not discussed here as their ages and genesis are not established.

3. Geochronology of the Karoo LIP

3.1. Overview

As illustrated already, the Karoo LIP comprises many components; volcanic and intrusive, with contemporaneous and overlapping volcanism spread across southern Africa and Antarctica. Handling the observed and resolved field relationships and the radiogenic age-data can therefore be somewhat cumbersome. Thus, an overview of the salient and reliable dates from the growing number of data on the Karoo is provided, which better constrain the ages of particular lithologies and events, and the duration of Karoo magmatism.

Although Jourdan et al. (2007b) provide a thorough analysis and review of the development of the Karoo LIP based on $^{40}Ar/^{39}Ar$ geochronology, there are new data (Curtis et al., 2008) and previously determined ages on the Antarctic components of the Karoo LIP which have not been discussed (Heimann et al., 1994; Zhang et al., 2003; Riley et al., 2005). Thus, a concise chronostratigraphy of the major components of the Karoo LIP is presented, along with the source references for the ages shown (Fig. 3). The assembly of ages shown is not exhaustive, but is restricted to ages determined by the $^{40}Ar/^{39}Ar$ and U–Pb methods on single plagioclase and zircon grains respectively (i.e. no bulk sample/bulk rock/groundmass ages). A synopsis of the reliability of these ages, and the omission of others, is presented below because this impacts on what geologically significant conclusions can be drawn from this chronostratigraphy.

3.2. Reliability of ages

Firstly, whole-rock ages (including those determined using K/Ar and Rb/Sr) are excluded because their meaning is highly debatable (Jourdan et al., 2005). Also, only Jurassic ages are shown for dyke swarms in which Proterozoic ages have also been found (the Save–Limpopo and Okavango dyke swarms).

There is some concern about the ages found by Riley et al. (2005) for the Group 1 dykes and the Alhmannryggen dyke swarm. These authors indicate that the criteria necessary to define an age plateau during step-heating in the $^{40}Ar/^{39}Ar$ method were generally not met and the effects of alteration and/or excess ^{40}Ar make distinguishing any age differences uncertain.

These criteria are: [1] at least 70% of the ^{39}Ar must be released, [2] a minimum of three successive steps in the age plateau must be evident and [3] the integrated age of the plateau should (within 2σ confidence limits) agree with each apparent age increment of the plateau (Jourdan et al., 2007b).

This probably relates to making age determinations on the groundmass of 2 samples which can be problematic because of mineralogical alteration (Jourdan et al., 2007c). These ages have, therefore, only been included in the interest of completeness and they are shown with "unknown" confidence limits.

The age of 180.2 ± 2.7 Ma found for the lowermost Olifants Bed (rhyolite flow within the Sabie River Formation basalt) is a reliable regression age (Fig. 3a of Riley et al., 2004). The other ages determined (U–Pb zircon ages) are discordant. Unfortunately, this includes the ages determined for the Jozini Formation rhyolites, which also have provided older ages (182.1 ± 2.9 Ma) than the rhyolites stratigraphically beneath them. Thus, these ages are not included here.

Jourdan et al. (2007b) re-calculated the U–Pb age of Riley et al. (2004) in order to compare it to $^{40}Ar/^{39}Ar$ ages, and found the age to be 178.4 ± 2.7 Ma. This is essentially indistinguishable from the age of 177.8 ± 0.6 Ma found for the Jozini Formation rhyolites (Jourdan et al., 2007b), although these evidently overlie the rhyolite flow within the Sabie River Formation. Duncan et al. (1997) determined $^{40}Ar/^{39}Ar$ ages of the Jozini rhyolites, but these may represent cooling ages (Riley et al., 2004) and are therefore not considered here.

The re-calculation of ages by Riley and Knight (2001) on the basalts of the Ferrar Province was done because ages calibrated to older standards (e.g. McClure Mountain Hornblende, MMhb-1 at 523.1 ± 2.6 Ma) cannot be compared reliably to those calibrated to newer standards (e.g. Hb3gr hornblende at 1072 Ma). See Jourdan et al. (2007c) for further discussion about the intercalibration of radiometric ages within the Karoo LIP.

3.3. Progression & duration of the Karoo LIP

If the oldest (~184 Ma) and youngest (~174 Ma) ages are considered as reliable, there is an approximate duration of 10 Ma for the Karoo LIP sensu stricto. The older age (~205 Ma) for the Jutulrøra dyke swarm is most likely a legitimate pre-Karoo LIP age. However, there is evidence of older volcanism in southern Africa from this time. For example, there are small volumes of volcaniclastic material associated with diatreme-like vents within the Molteno, Elliot and Clarens Formations and andesitic to dacitic dome complexes (McClintock et al., 2008). Although no absolute ages are known, the field relationships do suggest that they pre-date the earliest mafic eruptions of Karoo LIP proper. Furthermore, it has been shown that the Dokolwayo kimberlite (Swaziland) intruded during the deposition of the Karoo Supergroup, but is older than the uppermost sedimentary units (Molteno and Clarens Formations). Dokolwayo has been dated to 200 ± 5 Ma ($^{40}Ar/^{39}Ar$ age on phlogopite) (Allsopp and Roddick, 1984).

It is clear from Fig. 3 that there was an overlap of basalt eruption (Lesotho and Lebombo) and sill intrusion with volcanic and intrusive activity in the northern Lebombo (such as the dyke swarms). If the dyking activity in the northern Lebombo is assumed to have been continuous, it overlaps consistently with the age ranges of both the Save–Limpopo and Okavango dyke swarms. The voluminous outpourings of flood basalts in northern Botswana occurred over this time, from ~182–178 Ma, while those of southern Botswana erupted from ~185–181 Ma (Jourdan et al., 2005).

At approximately the same time volcanic activity that gave rise to the Ferrar Province began in Antarctica, with clearly later dyking of the Vestfjella and Straumsvola dyke swarms (Zhang et al., 2003; Curtis et al., 2008) and a remarkable lack of concomitant basalt eruption. This is consistent with the earlier findings of Riley and Knight (2001), who showed that the bulk of Ferrar magmatism occurred at ~180 Ma, approximately 3 Ma after the bulk of Karoo magmatism in southern Africa.

There is a dearth of ages for the Sabie River Formation which, on the basis of the interbedded Olifants Bed, suggests that it may have erupted until ~178 Ma, although Jourdan et al. (2007b) suggests that it may have only erupted over a 2–3 Ma period. The Sabie River Formation was fed from the D2 dyke generation of the Northern Lebombo dyke swarm (Klausen, 2009).

By this time in the Lebombo volcanism had progressed to more rhyolitic outpourings and intrusive activity in the Sabi/Mwenezi region, culminating in the youngest intrusive activity (silicic and syenite intrusions) of the northern Karoo LIP. This was likely contemporaneous with the intrusion of the Okavango and Save–Limpopo dyke swarms (Jourdan et al., 2007b). The ~176 Ma old Underberg dyke swarm intruded at this time in the central Karoo region, cross-cutting the basalts of Lesotho, while the youngest Karoo LIP-related magmatism is represented in the southern Lebombo by the intrusion of the Rooi Rand dyke swarm at ~174 Ma.

Besides the D2 dyke generation of the Northern Lebombo dyke swarm, there is little evidence to suggest that the dykes were feeders flowing from the Karoo triple junction to the now preserved basaltic volcanic pile. While many of the sills may be comparable in age to the basalts (see Jourdan et al., 2007c), the dyke swarms are not. For example, to the southeast of the main Lesotho basalts is the Underberg dyke swarm which is ~176 Ma in age compared to ~183 Ma for the basalts. The same is true of the Alhmannryggen and Vestfjella dyke swarms in Antarctica, and the Save–Limpopo and Rooi Rand dyke swarms.

There is also confirmation of diachronous magmatism in field evidence (Watkeys, 2002) and the broad geochronological data (Jourdan et al., 2004). Early primitive melts erupted at the Karoo triple junction, followed by volcanism and early dykes (earliest dykes of the Northern Lebombo dyke swarm) in the south and central (Lesotho basalts) regions, followed by further dyking in the north (Okavango dyke swarm) and Antarctica (early Straumsvola dyke swarm), rhyolitic volcanism in the Lebombo virtually synchronous with Antarctic magmatism and finally an apparent return to activity in the south (Underberg and Rooi Rand dyke swarm). Magmatism of the Mwenezi Igneous Complex appears to span this period of late dyking (178–174 Ma).

4. Mantle plume origin for the Karoo LIP

4.1. Proponent hypotheses

Burke and Dewey (1973) were the first to postulate a link between mantle plumes and the Karoo triple junction based on the assumption that the development of a triple junction within a continent was caused by uplift above an upwelling mantle plume. To account for the geometry of the triple junction, and volcanism further to the northeast, they proposed two possible mantle plume positions: one at the triple junction (Fig. 4a) and one in the lower Zambezi valley (Fig. 4b). White and McKenzie (1989), however, proposed a much broader (~2000 km diameter) plume situated at the juncture between Antarctica and the Lebombo (~450 km ENE of Maputo) to account for the contiguous magmatism of Dronning Maud Land and the Karoo (Fig. 4c).

Campbell and Griffiths (1990) and Storey (1995) upheld the idea of a plume head impacting directly beneath the triple junction itself, to account for its geometry in the same manner as Burke and Dewey (1973) and the composition and distribution of the nephelinites and picrites (Fig. 2). Cox (1989) showed that drainage patterns of rivers of the northern Lebombo break along an arc that coincides with the circumference of the plume position proposed by White and McKenzie (1989). This area may have been underplated by a hot, low density upwelling from the mantle, an idea upheld by White (1997) who also suggest that uplift across an area of up to 2000 km in diameter occurred above the plume head.

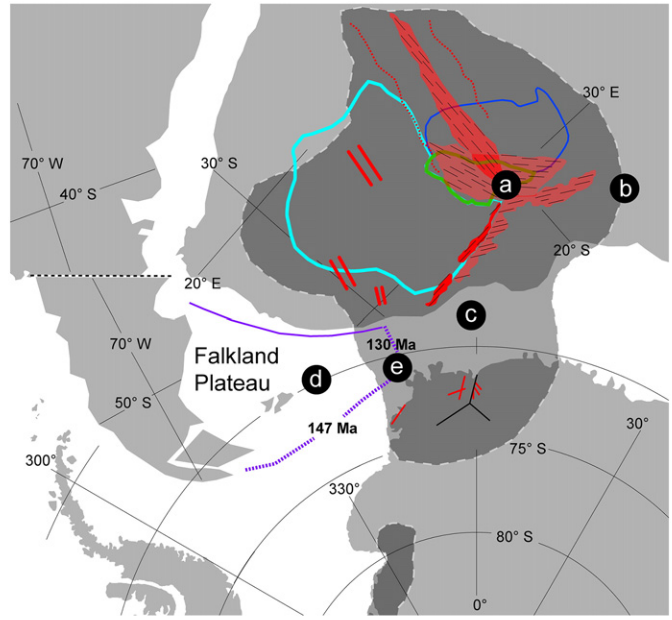

Fig. 4. Map of southern Gondwana showing the extent of the Karoo LIP and associated magmatic provinces and dyke swarms (see also Fig. 1), as well as previously proposed mantle plume positions. The plume positions shown are: (a) from Burke and Dewey (1973); Storey (1995), (b) from Burke and Dewey (1973), (c) from Cox (1989); White and McKenzie (1989); Storey et al. (1992); White (1997); Curtis et al. (2008), (d) centre of "megaplume" from Storey and Kyle (1997), (e) Weddell Sea triple junction from Elliot and Fleming (2000).
Modified from Cox, 1992; Riley et al., 2006; Veevers, 2012.

The mantle plume origin for the Karoo LIP suggested by White (1997) is a modified version—with stretching of the lithosphere from ~150 km to 50–60 km depth. This would have allowed for decompression melting brought about by an upwelling mantle plume. White (1997) favours a plume centred off the Kaapvaal Craton because a plume would have thermally perturbed the sub-continental lithosphere; an unlikely scenario given the eruption of Proterozoic and older diamonds through the Craton during the emplacement of post-Karoo Jurassic and Cretaceous-age kimberlites. Storey and Kyle (1997) suggest that a "megaplume" (≥2000 km diameter) positioned on the pre-break up position of the Falkland Plateau (Fig. 4d) can account best for the contiguous magmatism across the region.

There are also compositional factors of the Karoo LIP that are consistent with mantle plume melting, based on the inherent assumption that ambient upper mantle is homogeneous and isothermal, and therefore anomalies stem from deep sources. The earliest magmatism in the Karoo is composed of picrites and nephelinites (Bristow, 1984) which are geochemically compatible with uncontaminated plume-derived melt (White, 1997). Furthermore, Sweeney et al. (1994) have suggested that the more evolved high-Fe basalts in the central region of the Lebombo are consistent with mantle plume melting. This is echoed by Riley et al. (2005) who suggest that the ferro-picrite CT3 magma type is derived from a mantle plume melt. Indeed, the early nephelinites and picrites which centre on the Karoo triple junction may be accounted for in this fashion.

More recent work has been accommodating of modified plume models suggested previously. For example, Jourdan et al. (2007a) indicate that the isotope and trace element geochemistry of the Karoo magmas are compatible with a combination of isolation of mantle source regions and enrichment processes that may have involved a mixed mantle plume contribution. A similar idea was put forward earlier by Cox (1992) who suggested that there may have been interplay between a hinterland plume (Karoo) and more distal subduction along the Pacific margin of southern Gondwana (Ferrar, Antarctica).

O'Neill et al. (2009) have explored the possibility of an upwelling mantle plume being triggered by an earlier subduction event, an idea originally proposed by Cox (1992).

4.2. Opponent hypotheses

There are features of the Karoo LIP and the triple junction that do not appear compatible with a mantle plume origin. Triple junction formation itself does not require a plume or an active upwelling; they are intrinsic geometric manifestations of plate tectonics (Anderson, 2002). There is also no necessitation for a deep-seated source as implied by a mantle plume, particularly if it is considered that ambient mantle, particularly beneath large, long-lived plates (at ~150 km depth), can be hotter than the mantle under ridges (Anderson, 2011).

The proposed uplift in the Mwenezi region inferred to have been caused by a mantle plume (Cox, 1989) has been shown to be a much longer-lived palaeo-high (a horst-like structure) in existence since the Permo-Carboniferous (Watkeys, 2002). This is a feature of other LIPs, such as the Siberian Traps. It has become evident from studies of the Siberian Traps, and the surrounding country rocks, that significant subsidence (>3500 m in 1 Ma) occurred during and after the Late Permian eruption of the traps (Lind et al., 1994; Kamo et al., 2003). This contradicts one of the key tenets of the mantle plume hypothesis (Czamanske et al., 1998).

It has also been suggested that the axis of the proposed plume was not necessarily responsible for the primitive rocks centred on Mwenezi (Cox, 1992). The ENE-striking metamorphic fabric and relative structural weakness of the Limpopo Belt has most likely influenced the position of the primitive lithologies, an idea upheld by current geochronological and structural studies (Le Gall et al., 2002; Watkeys, 2002; Jourdan et al., 2004; Le Gall et al., 2005).

Geochronological studies have demonstrated two important features of the radiating dyke swarms of the Karoo triple junction. Firstly, there are a significant number of dykes in the Okavango and Save–Limpopo dyke swarms that are Proterozoic in age, suggesting that the orientation of these two swarms was determined prior to Karoo magmatism (Uken and Watkeys, 1997; Le Gall et al., 2002; Watkeys, 2002; Jourdan et al., 2004; Le Gall et al., 2005). Secondly, dating of basalts shows a clear progression in age from south to north (Lesotho, 183 Ma; Botswana, 178 Ma) while the intrusion of Jurassic dykes in southern Africa appears to have progressed from the Northern Lebombo, to the Save–Limpopo and Okavango dyke swarms and finally to the Underberg and Rooi Rand dyke swarms (Watkeys, 2002).

The Karoo triple junction, despite its apparent activity during Karoo magmatism, is also linked firmly to lithospheric architecture and its pre-Jurassic past (Watkeys, 2002; Jourdan et al., 2004). For example, the Lebombo monocline occupies the eastern edge of the Kaapvaal Craton and the structural trend of the Limpopo Belt has been exploited by the Save–Limpopo dyke swarm. Indeed, the area of maximum basalt thickness in Lesotho lies conspicuously on the southern boundary of the Kaapvaal Craton where it is adjacent to granite–gneiss terranes of the Proterozoic Namaqua–Natal Metamorphic Province.

Compositionally, the majority of volcanic and intrusive rocks within the Karoo LIP are not necessarily plume-related. There are few OIB-like compositions in the Karoo (except the CT4 magma type of Luttinen et al., 1998), a melt composition generally regarded as being plume-derived. It is only igneous provinces younger than the Karoo, such as the Paraná, Etendeka and the Deccan Traps that display isotopic and major and trace element signatures comparable to OIB (Hawkesworth et al., 1999). Regardless, some models show that OIB can originate from sub-lithospheric fractionation (Anderson, 1985).

The vast majority of the Karoo magmas are low-Ti tholeiites (Hawkesworth et al., 1999) and it is quite consistent with a mantle plume that a progression from primitive to tholeiitic compositions exists. Although high-Ti compositions are indeed widespread in the Karoo, there is also a conspicuous link between crustal architecture and the occurrence of these compositions: low-Ti compositions tend to occur over mobile belts, while high-Ti compositions occur over cratonic basement (Cox et al., 1967). This has been attributed to sub-continental lithospheric mantle control on the melts produced (Sweeney and Watkeys, 1990; Sweeney et al., 1994).

The siliceous, high field strength element-depleted signature in the Ferrar Province is believed to be the result of sedimentary contamination of the mantle source, most likely because of subduction (Hergt et al., 1991). Cox (1992) also indicates that the linear extent of the Ferrar Province (Fig. 1) in Antarctica is most likely beyond the extent of a proposed mantle plume(s). Encarnación et al. (1996) provided the first evidence for the geochronological link between the Ferrar Province and the Karoo LIP, indicating a subduction related process to account for the linear extent and geochemistry of the basalts. The suggestions that a plume may be more linear in extent (Storey and Kyle, 1997; White, 1997), or that a number of non-interacting plumes operated simultaneously in different regions (Storey and Kyle, 1997) seems possible, but speculative in light of other evidence. Some workers suggest that the Ferrar magmas were emplaced from the Weddell Sea triple junction (Elliot and Fleming, 2000) (Fig. 4e) and Leat (2008) also suggests that the magma source for the laterally emplaced Ferrar magmas was in the rift between SE Africa and Antarctica.

The dyke swarms of the Karoo LIP around southern Africa do not form part of a singular volcanic feeder system, as most of the dyke swarms (Okavango, Rooi Rand, Southern Lesotho and Underberg dyke swarms) cross-cut the basalts. This is reflected in the geochronology which shows that the Okavango and Rooi Rand dyke swarms post-date the main Karoo volcanic event by at least 3 Ma. This seems counterintuitive given the classic model of radial injection of dykes from a triple junction (Ernst and Baragar, 1992), simultaneously feeding overlying volcanics. Furthermore, short-lived (~2–5 Ma), high volume magmatism is not strictly applicable to the Karoo LIP; although the volumetrically significant tholeiitic, CFB component did erupt relatively rapidly (Jourdan et al., 2005).

4.3. Implications of the duration of the Karoo LIP

Karoo magmatism extended over the period 183–174 Ma, almost twice the duration that has been suggested as typical for other LIP's. Such a duration is uncharacteristic of the classical model of short lived, rapid magmatism driven by an active mantle plume (Jourdan et al., 2004, 2006). This relatively long duration is suggestive of long-term magma storage and a lack of preservation of primitive melt compositions. The south-to-north migration of magmatism also appears to have been too rapid for it to have occurred as a result of the crust migrating over a plume head (Jourdan et al., 2007b). Magmatism at Mwenezi and in the Lebombo monocline appears to be the longest-lived (183–174 Ma) compared to the other arms of the triple junction (Watkeys, 2002; Jourdan et al., 2007b). The 174 Ma old Rooi Rand dyke swarm most likely represents an isolated dyking event, given its restricted occurrence, age and MORB-like composition; unique in the Karoo LIP. The youngest Karoo volcanic activity is preserved as the predominantly felsic Mwenezi igneous complex which intrudes rhyolites in the Mwenezi trough (Watkeys, 2002). Such diversity in distribution and age is difficult to reconcile with a mantle plume.

It has been asserted that vigorous output of large volumes of magma is linked to active crustal extension caused by plumes (Richards et al., 1989; White, 1997). This would suggest, in the case of the Karoo, a relatively short period of magmatism followed rapidly by continental break-up, i.e. within ~10 Ma as is typical of other LIPs. Jourdan et al. (2005) indicate that the bulk of Karoo magmatism occurred over ~6 Ma, resulting in calculated eruption volumes of ~0.3 $km^3 \cdot yr^{-1}$, approximately a third of the rate for the Central Atlantic Magmatic Province (Marzoli et al., 1999). If, for example, full oceanisation occurred at this time (consistent with rapid rifting caused by a plume), it would still imply a long-lived thermal incubation of at least 10 Ma, which is odds with a mantle plume. In addition, the first sea-floor anomalies of southern Gondwana are ~155 Ma old (Goodlad et al., 1982). This puts ~28 Ma between the onset of magmatism and oceanisation, which seems too protracted a time for a plume to have heated the lithosphere without it rupturing (Storey et al., 1992). However, if new sea-floor magnetic data are considered (Leinweber and Jokat, 2012), which puts the earliest sea-floor spreading at ~167 Ma, there is a shorter period of ~7 Ma following the intrusion of the RRDS before sea-floor spreading initiated.

It is of interest, therefore, to investigate whether the dyke swarms of the Karoo LIP and those of Dronning Maud Land can provide any additional insight into this plume debate. A scenario involving lateral magma flow away from the Karoo triple junction would typically be expected in the event of an impinging mantle plume beneath the triple junction—an idea explored with regard to the Northern Lebombo, Rooi Rand and other dyke swarms of the Karoo LIP.

5. Methodology

5.1. Introduction

Measuring the AMS of samples from mafic dykes has become a standard method of quantifying flow-related petrofabric (Khan, 1962; Ellwood, 1978; Knight and Walker, 1988; Ernst and Baragar, 1992; Poland et al., 2004; Kissel et al., 2010). The susceptibility of mafic igneous rocks derives from Fe-bearing minerals such as magnetite and titanomagnetite. The anisotropy may be controlled by the distribution of grains within the rock (distribution anisotropy, Hargraves et al., 1991) or by magnetocrystalline anisotropy (e.g. hematite or pyrrhotite). For multidomain magnetite, however, it is more typically shape controlled (shape anisotropy) and AMS can therefore be representative of the shape and orientation of the petrofabric (Rochette et al., 1999).

Measurements are typically made on cylindrical drill-core samples (22 mm × 25 mm) collected along opposing chilled margins of a dyke in order to determine if any imbrication of the fabric elements (lineation and/or foliation) can be found (Tauxe et al., 1998; Aubourg et al., 2002; Geoffroy et al., 2002). This is regarded as local scale because data are determined per margin, and per dyke. In the case of a dyke swarm it is useful to view the "bulk" fabric however. This approach is explained further in Section 5.2.

AMS is represented by an ellipsoid with principal axes $K1$, $K2$ and $K3$ (where $K1 > K2 > K3$). The magnetic lineation ($K1$) and foliation (to which $K3$ is a pole) are most important for our purposes, and are typically shown stereographically—either for a single dyke (usually shown as separate margins), or grouped in order to visualise the bulk fabric. The ellipsoid is further described by the degree of anisotropy (P') and the shape of the ellipsoid is indicated by the factor T (where oblate fabrics have $T > 0$ and prolate fabrics have $T < 0$).

There are essentially three types of magnetic fabric (Rochette et al., 1999). Type-A fabric is characterised by the $K3$ axes clustering perpendicular to the dyke wall, with $K1$ and $K2$ lying along the magnetic foliation (sub-) parallel to the dyke plane. This is often referred to as "normal" fabric, and is considered to form by magma flow (Ellwood, 1978; Knight and Walker, 1988; Rochette et al., 1999; Cañón-Tapia and Chávez-Álvarez, 2004). Type-B, or "inverse", fabric is characterised by the interchange of $K1$ for $K3$ axes, with clustering of $K1$ axes perpendicular to the dyke plane (Potter and Stephenson, 1988; Tarling and Hrouda, 1993; Rochette et al., 1999). The third type of magnetic fabric is known as "intermediate", with $K2$ interchanged with the expected orientation of $K3$ in type-A fabric. Intermediate fabric is most commonly attributed to the mixing of fabric types (Rochette et al., 1999) and/or the effects of late-stage compaction (Park et al., 1988; Philpotts and Philpotts, 2007). In addition, AMS fabrics can be highly scattered or fabrics may bear no relation to the intrusion plane at all. The use of type-B fabrics in determining magma flow direction has been well illustrated by Callot et al. (2001) and Aubourg et al. (2008) where inverse $K1$ axes are treated as $K3$ axes in density stereoplots of the bulk AMS fabric.

It must be noted here that the AMS and SPO data and interpretations pertaining to the Northern Lebombo dyke swarm are from Hastie et al. (2011b) and data for the Okavango dyke swarm are from Aubourg et al. (2008). Magnetic fabric data for dykes of Dronning Maud Land are from Curtis et al. (2008) and we present previously unpublished data from the Rooi Rand dyke swarm in conjunction with the prior work of Hastie et al. (2011a).

5.2. Inferring flow direction

While many studies continue to use the magnetic lineation as an indicator of flow direction Henry (1997) has shown that the lineation may be meaningless if the intersection of magnetic foliations is correlative with the magnetic lineation. This has been demonstrated in further studies as well (e.g. Dragoni et al., 1997; Callot and Guichet, 2003; Hastie et al., 2011b) where the $K1$ axis can be at right angles to the flow direction. Our approach here is to focus on the imbrication of the

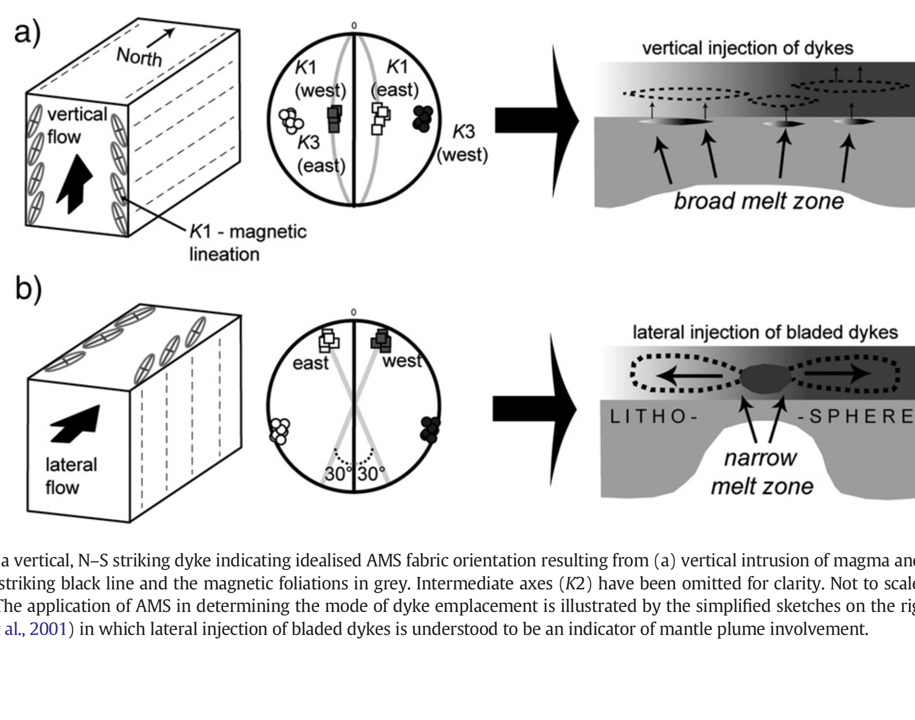

Fig. 5. Schematic diagrams of a vertical, N–S striking dyke indicating idealised AMS fabric orientation resulting from (a) vertical intrusion of magma and (b) lateral intrusion of magma. Dyke plane is shown as N–S striking black line and the magnetic foliations in grey. Intermediate axes ($K2$) have been omitted for clarity. Not to scale (modified from Geoffroy et al., 2002; Gil-Imaz et al., 2006). The application of AMS in determining the mode of dyke emplacement is illustrated by the simplified sketches on the right of each end-member of flow type (re-drawn from Callot et al., 2001) in which lateral injection of bladed dykes is understood to be an indicator of mantle plume involvement.

foliations relative to the dyke plane in predominantly oblate shaped fabric (Aubourg et al., 2002; Geoffroy et al., 2002) while also factoring in the orientation of $K1$ axes (in prolate fabrics) and any field evidence which may assist in flow determinations.

This approach, at the scale of a single dyke, works by plotting the principal AMS axes for each margin (thus two stereoplots per dyke) which when compared may reveal mirrored imbrication of foliations and/or lineations (Fig. 5). The finding of vertical flow (Fig. 5a) and lateral flow (Fig. 5b) across a suite of dykes or a dyke swarm is, however, generally interpreted in a regional context.

This is because the flow history is strongly controlled by the disposition of the melt source (e.g. broad vs. narrow melt zone, see Fig. 5), the characteristics of the crust (e.g. nature and orientation of stress) and the magma (e.g. overpressure, viscosity, neutral buoyancy level and volatile content). If a mantle plume is involved in the injection of dykes from a local, narrow melt zone, it is expected that flow will be lateral (Ernst and Baragar, 1992; Fialko and Rubin, 1999) (Fig. 6).

It is important to note that AMS fabric may not be directly related to the original flow-fabric. For example, late-stage settling can cause compaction fabric, and even back-flow (downward flow) can occur in dykes (Aubourg et al., 2002; Philpotts and Philpotts, 2007).

Indeed late-stage flow may result in more complex or mixed fabrics resulting from significant grain interaction, as has been found in the Rooi Rand dyke swarm (Hastie et al., 2011a). It is thus imperative to view the bulk AMS (and mineral SPO) data to provide a sense of the regional flow-fabric. This is done by plotting all the data together in order to observe and interpret the AMS, or other petrofabric, on a regional scale (Rochette et al., 1991; Tamrat and Ernesto, 1999; Callot et al., 2001; Aubourg et al., 2002; Archanjo and Launeau, 2004).

This can only be achieved, however, if a relatively large number of dykes (>20) are studied across a region (Ernst and Baragar, 1992). From the Northern Lebombo, Rooi Rand and Okavango dyke swarms we discuss the regional significance of AMS data determined from a total of 60 dykes. We show AMS data plotted in a regional manner from a further 30 dykes from west Dronning Maud Land.

This regional AMS fabric is shown in dyke co-ordinates, where data have been rotated in accordance with the orientations of the dykes from which data was collected. This essentially "normalizes" the data for ease of viewing and interpretation (Rochette et al., 1991; Callot et al., 2001; Aubourg et al., 2008) and in this paper is referred to as N–S dyke co-ordinates. In such a framework it is easiest to observe and interpret the imbrication of the fabric (the bulk foliation as determined by the grouping of $K3$ axes), should it occur. The "normalized" regional data presented in the following sections is contoured on an equal area stereonet and type-B fabric has been omitted.

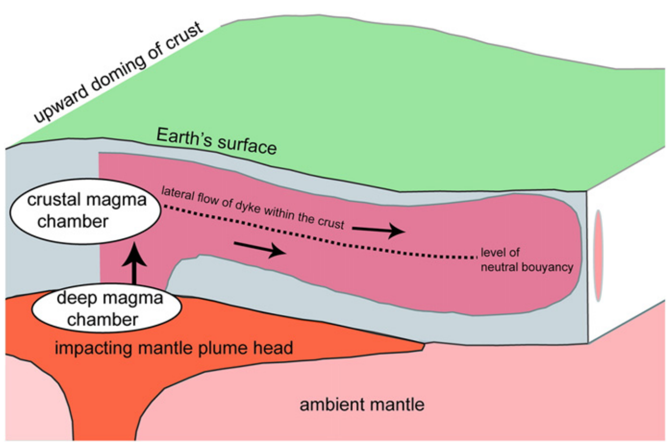

Fig. 6. Schematic figure showing impact of hot (relative to ambient mantle) mantle plume head beneath the crust; proposed lateral flow (arrows) of a dyke within the crust along the level of neutral magma buoyancy (not to scale); and doming of the Earth's surface caused by plume-related thermo-mechanical uplift.
Re-drawn from Fialko and Rubin, 1999.

5.3. Mineral shape preferred orientation

The method of quantifying mineral SPO, as applied to the Northern Lebombo and Rooi Rand dyke swarms, begins by cutting three orthogonal thin sections per AMS sample. From these, between 300 and 2800 plagioclase grains per section are digitally imaged and analysed to determine an inertia tensor from the 2-D SPO inertia ellipses (Launeau and Robin, 2005 and references therein). This ellipsoid, similarly to AMS, is defined by three orthogonal principal axes ($L1 > L2 > L3$). We infer a sense of magma flow in the same way as AMS—imbrication of the foliation defined by the principal ellipsoid axes with respect to the dyke plane. The methodology of SPO study applied to the Rooi Rand and Northern Lebombo dyke swarms is covered in more detail by Hastie et al. (2011a).

6. Results

6.1. Magnetic fabric of the Northern Lebombo dykes

Magnetic fabric has been studied in 14 dykes (224 samples) of the Northern Lebombo dyke swarm (Hastie et al., 2011b) and it is evident that the magnetic fabric is carried by stoichiometric magnetite, and tends to be normal (type-A) to intermediate. Values of P' range from ~1.0 to 1.054 (mean: 1.027). The mean value of T in the Northern Lebombo dyke swarm is 0.027 with oblate fabric in 64% of the data.

Magnetic fabric in the youngest (D4) dykes is typically type-A and oblate in shape. Vertical to sub-vertical $K1$ axes are found only in the prolate fabrics and these prolate fabrics tend to be intermediate, although some are neutral to oblate in shape. Four sites display scattered fabric, and another four sites display type-B fabric. Thus six dykes provide a measure of magma flow, including evidence of vertical and lateral flow. The grouped data (Fig. 7a) reflects these trends and, overall, is consistent with lateral magma flow from the north. Note that the bulk data are split according to western and eastern margins in order to easily see any imbrication of the fabric elements.

6.2. Magnetic fabric of the Okavango dyke swarm

Ernst and Duncan (1995) were first to measure the magnetic fabric in the Okavango dyke swarm using AMS, focussing their attention on the orientation of the magnetic lineation. These authors interpreted steeply plunging magnetic lineations close to the triple junction (<300 km) and shallowly plunging lineations 400 km from the triple junction to indicate steep (vertical) flow and lateral (sub-horizontal) flow respectively.

The study of Aubourg et al. (2008) focussed on two regions of the Okavango dyke swarm, the Thune and Shashe sections in which dykes have intrude basement gneisses. The Thune and Shashe sections are 300 km and 400 km west of the Karoo triple junction respectively (Fig. 7). The overall magnetic fabric determined by AMS of 23 dykes (386 samples) is type-A. The fabric is carried by ferromagnetic grains. Thermomagnetic measurements indicate that paramagnetic susceptibility is negligible. However, Aubourg et al. (2008) showed that the type-B fabric most likely results from strong magnetization of some samples and the development of planar preferred orientation of ferromagnetic grains orthogonal to the dyke plane.

At the dyke scale in the Thune section (Fig. 7b) there is evidence of sub-vertical magma flow in the Okavango dyke swarm, as well as lateral flow to the east and west. The magnetic fabric is less well defined in the Shashe section, although the imbrication of the magnetic foliations, in particular, is consistent with magma flow to the west (Aubourg et al., 2008). On the regional scale, the magnetic fabric in the Okavango

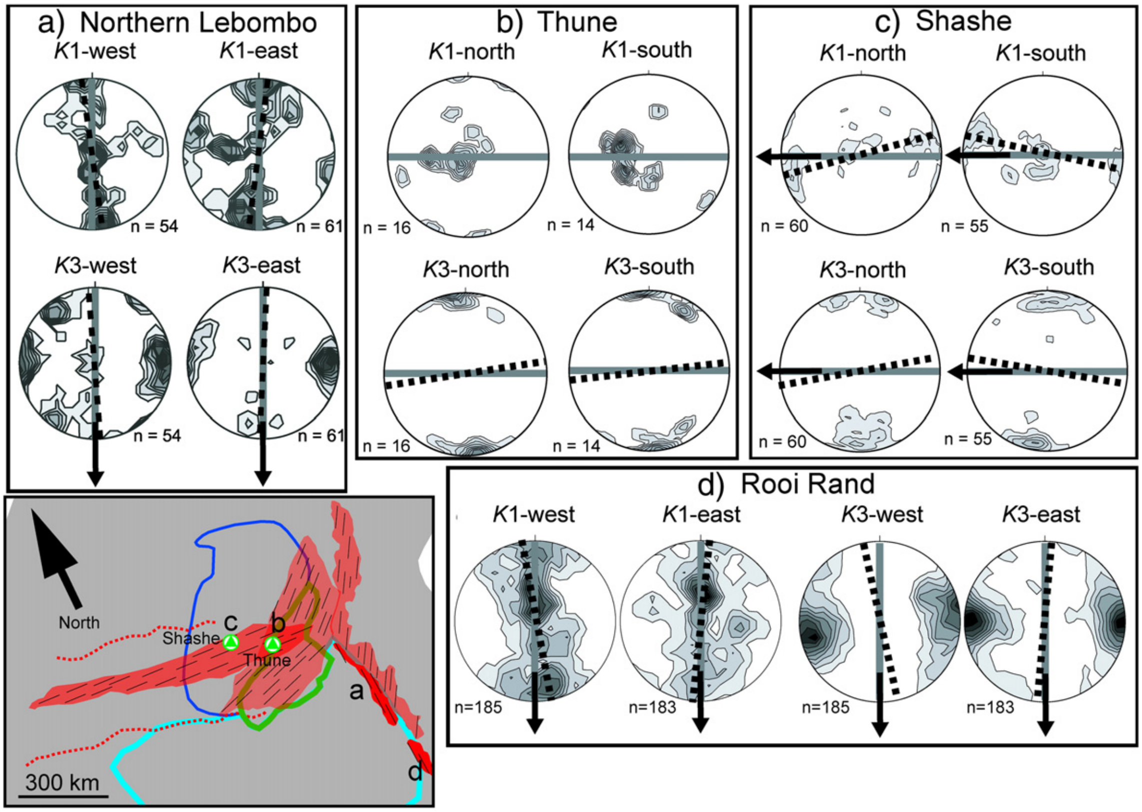

Fig. 7. Grouped AMS data (type-A fabric only) for (a) the Northern Lebombo dyke swarm (Hastie et al., 2011b), (b) the Thune section of the Okavango dyke swarm (Aubourg et al., 2008), (c) the Shashe section of the Okavango dyke swarm (Aubourg et al., 2008) and (d) the Rooi Rand dyke swarm (Hastie et al., 2011a). Inset of the dyke swarms to which the data correlate are shown; a = Northern Lebombo dyke swarm, b and c = Thune and Shashe sections of the Okavango dyke swarm respectively and d = Rooi Rand dyke swarm. The dyke planes in the stereoplots are shown in grey and the foliations as dashed black lines. Note that the Okavango dyke swarm data have been rotated into a common E–W orientation and the Northern Lebombo and Rooi Rand dyke swarm data into a common N–S orientation. The data are split according to the west and east margins (for the Northern Lebombo and Rooi Rand dyke swarm) and north and south (for the Okavango dyke swarm) in order to find imbrication of the foliation or $K1$ axes. The number of samples is denoted by "n". Arrows indicate the magma flow direction inferred from the data. Note that the $K1$ axes for the Thune region, and the Rooi Rand dyke swarm, show a degree of sub-vertical flow.

dyke swarm is well constrained and imbrication is most consistent with lateral magma flow from east to west (Fig. 7c).

6.3. Magnetic fabric of the Rooi Rand dyke swarm

Dykes of the Rooi Rand dyke swarm are plagioclase and augite bearing dolerites. Magnetic fabric measured in 23 dykes (368 samples) is predominantly type-A, carried by fine-grained, low-Ti magnetite. The fabric is generally weakly anisotropic (mean $P' = 1.030$) and neutral to oblate in shape (mean $T = 0.073$). In ~30% of the magnetic fabric data we found that the magnetic foliation or lineation was orthogonal to the dyke plane; this is type-B fabric (Hastie et al., 2011a).

The magnetic foliations are generally steeply dipping, and two dykes have shallowly plunging $K1$ axes, consistent with lateral magma flow. Field data tend to support steep (60°–80°) emplacement, although in 12 of the 23 dykes, inclined, sub-lateral and lateral magma flow is supported by AMS. The density plots of $K1$ and $K3$ axes for the Rooi Rand dyke swarm are well constrained, comprising type-A fabric and the $K3$ axes in particular are very tightly grouped and are weakly imbricated (Fig. 7d). The imbrication is suggestive of inclined to sub-lateral magma flow from the north.

6.4. Dykes of west Dronning Maud Land

As discussed already, Curtis et al. (2008) propose two distinct phases of mafic dyke emplacement (the Jutulrøra and then the Straumsvola dyke swarm at 170.9 ± 1.7 Ma) in the H.U. Sverdrupfjella of west Dronning Maud Land. Curtis et al. (2008) conducted AMS studies on 30 dykes, finding that 42% of the magnetic fabric is type-A. Furthermore, the fabric was found to be triaxial to oblate in shape with 1–4% anisotropy. This is very typical of mafic dykes. It is noted that the fabric becomes less well defined from north (48% type-A) to south (32% type-A). Intermediate fabric was found in 8% of the entire sample suite. The magnetic fabric appears to be carried by low-Ti magnetite, with type-A fabric arising from multi- and pseudo-single domain magnetite.

These authors use the $K1$ axis as a proxy for the magma flow direction, and find that the magnetic fabric is consistent with steep (sub-vertical) magma flow in the Straumsvola area (Fig. 8a), which is consistent with the orientation of stretched amygdales in one dyke of the Straumsvola dyke swarm. When plotted together in N–S dyke coordinates, it is found that 61 of the averaged $K3$ axes (~51% of the data) are inverse (type-B fabric) (Fig. 8a). It has been found by numerous workers in AMS that mafic dykes tend to contain ~50% type-B fabric (Baer, 1995; Callot et al., 2001; Aubourg et al., 2008).

If this type-B component is removed and the margins are plotted separately we find $K1$ consistent with vertical and lateral flow (Fig. 8c). There is a very slight imbrication of the foliations that suggests lateral flow to the south (Fig. 8b). This is consistent with an increasing component of lateral flow ~25 km south of Straumsvola in the Jutulrøra region (Curtis et al., 2008). The lack of shallowly plunging $K1$ axes from the eastern margins is, however, inconsistent with this, as is the arguably weak imbrication, which likely reflects the vertical flow component of the Straumsvola region.

6.5. Shape preferred orientation fabric

The results of the mineral SPO study are restricted to the bulk plagioclase fabric for the Northern Lebombo and Rooi Rand dyke swarms. In

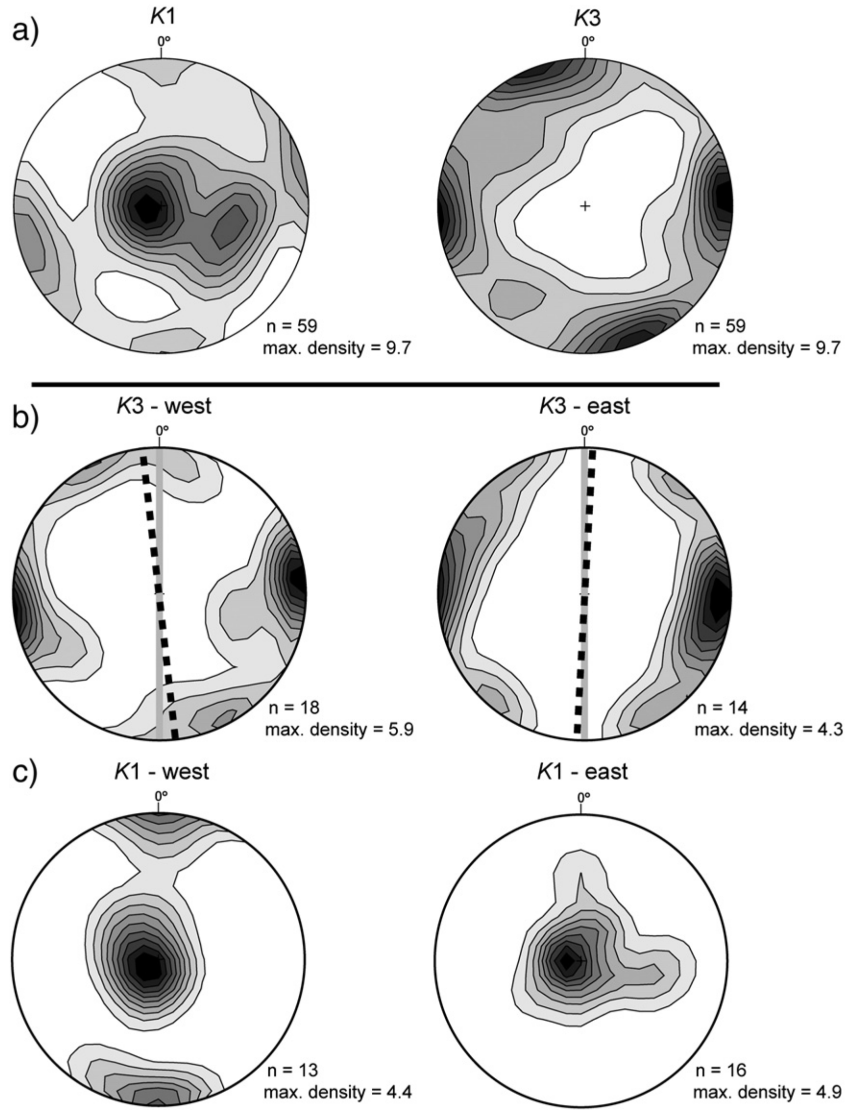

Fig. 8. Bulk AMS data of the Straumsvola and Jutulrøra dyke swarms from Curtis et al. (2008) plotted in (a) N–S dyke co-ordinates and separated into (b) *K*3 axes (N–S dyke co-ordinates) and (c) *K*1 axes (N–S dyke co-ordinates). The N–S dyke plane is shown in grey and the foliation as a dashed black line. Note the slight imbrication of the foliations, which is consistent with the shallow plunging *K*1 axes from the western dyke margins in (c). The (sub-) vertical *K*1 axes in c) are consistent with a component of steep, upward flow. The axes plotted are the averages from Table 2 of Curtis et al. (2008) (i.e. n = 59). The total number of samples measured = 480.

the Okavango dyke swarm Aubourg et al. (2008) does indicate that there is generally good agreement between AMS and the orientation of plagioclase grains. There are no mineral SPO data available for the dykes of Dronning Maud Land.

The plagioclase fabric in the Northern Lebombo dyke swarm (7 dykes, 28 ellipsoids) is generally type-A, consistent with the magnetic fabric. The plagioclase fabric has a mean P' value of 1.208 and is predominantly triaxial to oblate (mean $T = 0.042$). The highest P' values for plagioclase are found in phenocryst-rich samples. Imbrication of the plagioclase SPO fabric is found in four dykes and southward verging imbrication of foliations in the bulk fabric occurs (Fig. 9a).

Although the *L*1 axes would be consistent with vertical magma flow, it has been shown that the foliations in this case are more reliable because of the coincidence between intersection lineations and the *L*1 axes (Hastie et al., 2011b). Furthermore, the geological significance of the lineation in a predominantly oblate fabric is debatable.

A total of 41 ellipsoids representative of the plagioclase fabric have been determined for 10 dykes of the Rooi Rand dyke swarm. The fabric is predominantly neutral to oblate in shape ($T = 0.04$) with a mean P' value of 1.19. Approximately 30% of the plagioclase SPO fabric is type-B and prolate in shape (Hastie et al., 2011a). The remaining type-A fabric of 10 dykes is shown in N–S dyke co-ordinates in Fig. 9b. Similarly to the Northern Lebombo dyke swarm we rely upon the orientation of the foliations to infer magma flow direction.

The foliation defined by the *L*3 axes for the western margins is well imbricated with respect to the average dyke orientation, and verges southward. The foliation from the eastern margins appears similar, although the imbrication is arguably not as obvious. Overall, this imbrication is consistent with magma flow directed from north to south. The *L*1 axes are consistent with flow inclined ~20° from the vertical.

7. Discussion

7.1. Magma flow in the dyke swarms

The significance of AMS fabric in 90 dykes related to Karoo magmatism has been presented in the context of the previous and current understanding of the Karoo LIP (Riley et al., 2006; Aubourg et al., 2008; Curtis et al., 2008; Hastie et al., 2011a,b). Data for the Northern Lebombo dyke swarm are consistent with early vertical flow, followed by lateral flow in the later (D4) dykes. The regional pattern of AMS and SPO are consistent with each other, and subtly suggest a lateral

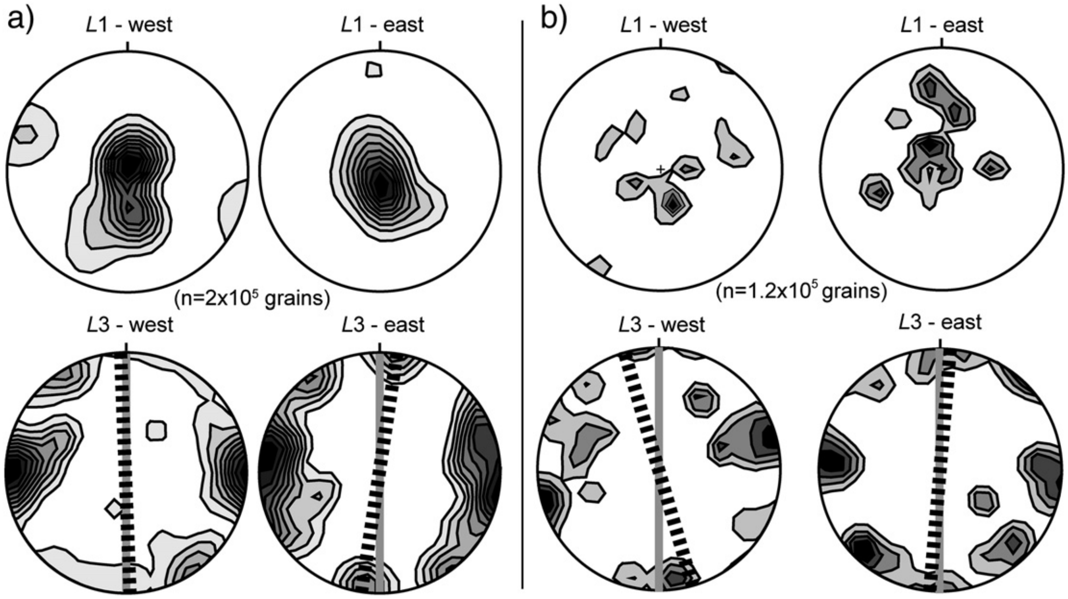

Fig. 9. Mineral SPO data in N–S dyke co-ordinates for (a) the Northern Lebombo (from Fig. 10 of Hastie et al., 2011b) and (b) Rooi Rand dyke swarms. The average dyke planes are shown in grey, the foliations as dashed black lines. The $K1$ axes, of the Northern Lebombo dyke swarm in particular, suggest vertical flow but the imbrication of the foliations, although slight, is suggestive of lateral magma flow.

sense of magma flow from north to south. This would be consistent with the Karoo triple junction being a viable magma source.

This is also reflected in the AMS data of the Okavango dyke swarm; the shallowly plunging $K1$ axes and imbricated foliations of the Shashe region are both consistent with lateral flow from SE to NW, distally to the triple junction. The predominantly (sub-) vertical $K1$ axes from the Thune section (closer to the triple junction) are consistent with steeper (more vertical) magma flow. The lack of mirrored imbrication of the foliations in the Thune data also appears to preclude any significant degree of lateral magma flow. The magnetic and SPO fabric of the Rooi Rand dyke swarm both have imbricated foliations consistent with flow from north to south. Both the $K1$ and $L1$ axes of the eastern dyke margins are consistent with a component of steep, sub-vertical magma flow. Kattenhorn (1994), using the SPO of plagioclase, also found relatively steep flow (23° from the vertical) in a dyke of the Rooi Rand dyke swarm directed from north to south, which is also consistent with field evidence such as steeply plunging broken and rotated bridges along some dyke margins (Nicholson and Pollard, 1985; Bussel, 1989). This suggests that the Rooi Rand dyke swarm was not fed from the Karoo triple junction, as does the fact that the “signature” of the Rooi Rand dyke swarm (MORB-like composition, geochronological results) declines approximately 500 km south of Mwenezi.

If the Rooi Rand dyke swarm indeed post-dates the Lebombo rhyolites and is indicative of lithospheric rupturing, it may be that the Rooi Rand dyke swarm is analogous to a mid-ocean ridge segment. In such segments the dominant flow direction would be lateral in dykes distal to the vertically fed central axis. This type of dyking regime has been previously demonstrated in such segments by Abelson et al. (2001) and Archanjo and Launeau (2004).

The magnetic fabric in the Jutulrøra and Straumsvola dyke swarms has two components. There is data consistent with steep (sub-vertical) magma flow in the north (Straumsvola) and with a lateral flow component further south (Jutulrøra) (Curtis et al., 2008). The steep component is particularly evident from the $K1$ axes. The regional pattern of magma flow determined from the type-A fabric is consistent with this picture of flow from north to south, as can be seen from some of the shallowly plunging $K1$ axes of the western margins and the imbricated foliations. The flow directions for the dyke swarms under consideration are summarised diagrammatically in Fig. 10. The flow direction determined from the Underberg dyke swarm is included, but we do not speculate further on its significance because only three dykes showed reliable flow directions.

Curtis et al. (2008) suggest that the Jutulrøra dyke swarm (irrespective of the vertically or laterally intruded components) and Group 1 dykes should not be considered contiguous with the later Karoo LIP.

Magma flow data from the Alhmannryggen and Vestfjella dyke swarms would assist in shedding light on dyke emplacement in this region, because flow directed from the previously conjugate Africa–Antarctica margin would be consistent with the plume/melting

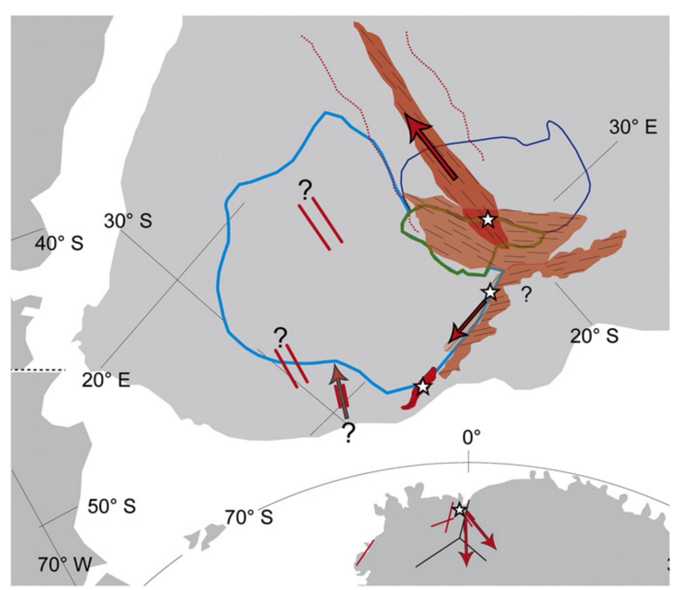

Fig. 10. Regional map of southern Gondwana between 178 and 170 Ma with magma flow directions inferred for the Okavango, Northern Lebombo, Rooi Rand, Straumsvola and Jutulrøra dyke swarms. The red arrows (exaggerated 2.5× for the Antarctic dykes) indicate predominantly lateral flow while stars indicate predominantly vertical flow. The flow directions of the Underberg, Southern Botswana and Southern Lesotho dyke swarms are uncertain (“?”).
Re-drawn from White and McKenzie, 1989; Storey et al., 1992; Encarnación et al., 1996; Storey and Kyle, 1997; Watkeys, 2002; Jourdan et al., 2004; Mekonnen, 2004; Ferraccioli et al., 2005; Riley et al., 2006; Curtis et al., 2008; Veevers, 2012.

anomaly position shown in Fig. 4c (Cox, 1989; White and McKenzie, 1989; Storey et al., 1992; White, 1997; Elliot and Fleming, 2000; Curtis et al., 2008). The north-to-south flow direction found in the Straumsvola and Jutulrøra dyke swarms is not compatible with the flow direction that would be expected from the Weddell Sea triple junction (Fig. 4e).

The overall flow pattern, although somewhat simplified, is consistent with the Karoo triple junction being an important magma source for the Northern Lebombo and Okavango dyke swarms, at least. It furthermore highlights the strong role played by the Limpopo Belt and Kaapvaal Craton in controlling the distribution of the Lebombo and Save–Limpopo monoclines and the Okavango dyke swarm. The locus of flow in the Jutulrøra and Straumsvola dyke swarms is at least 500 km from the Karoo triple junction, as is the apparent locus for the Rooi Rand dyke swarm. Linked with the field evidence and geochronology that argues for the following dyking progression: Jutulrøra >Northern Lebombo > Straumsvola > Okavango > Rooi Rand dyke swarms, it seems unlikely that each melt locus could have been a separate mantle plume. Indeed, these loci may have been "weak spots" exploited by magma impingement under tensional conditions.

7.2. Magma flow and the plume hypothesis

The triple junction geometry, geomorphological features and geochemistry of the rocks of the Karoo LIP have all been used as evidence of a starting mantle plume in the Karoo LIP (Burke and Dewey, 1973; Cox, 1989; White, 1997). Partnered with evidence of lateral magma flow from the Karoo triple junction in the Northern Lebombo and Okavango dyke swarms, and evidence of the earliest (nephelinites) and youngest (Mwenezi intrusions) igneous activity at the triple junction, it would be reasonable to view the Karoo triple junction as a relatively long-lived (~183–174 Ma) magmatic point source within the region—perhaps originating from a mantle plume.

However, inspection of the magma flow data regionally, and other evidence, calls this relatively simple scenario into question. There is a protracted history of lithospheric structural control of the distribution of the Karoo basalts; with low-Ti basalts erupted through Proterozoic basement, and high-Ti basalts erupted through Archaean basement (Hawkesworth et al., 1984; Sweeney and Watkeys, 1990; Sweeney et al., 1994; Hawkesworth et al., 1999). There is also evidence of structural control of the triple junction whereby Proterozoic structures and dyke orientations have been exploited by the later Karoo-age intrusions (Le Gall et al., 2002; Jourdan et al., 2004; Le Gall et al., 2005). Indeed, the structure and geochronology of the Karoo LIP show that the orientation of the Save–Limpopo and Okavango dyke swarms were inherited from Proterozoic intrusive events and flood basalt eruptions appear to have been diachronous from south to north.

Similar to the Karoo LIP in southern Africa, the Ferrar Province and associated magmatism of west Dronning Maud Land display evidence of subduction related processes (Hergt et al., 1991; Encarnación et al., 1996) although there is a potentially significant degree of lateral magma flow that occurred during the emplacement of the Ferrar magmas (Elliot and Fleming, 2000; Leat, 2008).

Only three proponents of the original mantle plume models suggest that the plume impinged directly beneath the triple junction (Burke and Dewey, 1973; Campbell and Griffiths, 1990; Storey, 1995). The other models place the mantle plume "head" between Dronning Maud Land and the southern African continent (Fig. 4c). This position, however, is not consistent with the directions of magma flow found in the dyke swarms of southern Africa.

When the age constraints on the dyke swarms are considered a number of inconsistencies come to light as well (Jourdan et al., 2004). Firstly, the Save–Limpopo dyke swarm was the earliest swarm to intrude and, along with the Okavango dyke swarm, contaINS subparallel Proterozoic dykes which demonstrate a Jurassic-age exploitation of the older dyke direction. Secondly, dykes of the Northern Lebombo dyke swarm mostly pre-date, but also overlap with the Okavango dyke swarm. Thirdly, the Rooi Rand dyke swarm is considerably younger (174 Ma) and of asthenosphere-derived, MORB-like composition in comparison to other Karoo magmas (Saggerson et al., 1983; Armstrong et al., 1984; Meth, 1996). If we consider the principal $K1$ axes, and the finding of Kattenhorn (1994), it is evident that some degree of steep flow occurred in the Rooi Rand dyke swarm. Fourthly, the remaining dyke swarms (Underberg, Southern Lesotho and Southern Botswana dyke swarms) bear no relation to the triple junction at all. There are currently no ages of the Southern Lesotho and Southern Botswana dyke swarms available, although the fact that they have intruded through the basalts, similarly to the Underberg dyke swarms, suggests that they may be ~178–176 Ma in age.

The agreement between the magma flow determinations and the triple junction does not exclude the triple junction from being an important magmatic source region, but the timing of the various dyke swarms is an important factor to consider, as is the pre-Karoo uplift which may have preferentially allowed lateral magma flow, rather than by a plume-driven mechanism. This is apparent in the dykes of west Dronning Maud Land. For example, the Alhmannryggen, Jutulrøra, Straumsvola and Group 1 dyke swarms appear in a radiating geometry (Fig. 1b) which, along with other compositional and flow characteristics, and the protracted magmatic history of the Straumsvola region, could be considered indicative of a mantle plume. As discussed, however, magma flow from an igneous centre in the vicinity of Straumsvola at ~205 Ma predates Karoo magmatism by ~15 Ma, while the Straumsvola dyke swarm was emplaced closer to 178 Ma. Thus, the Jutulrøra dyke swarm and Group 1 dykes should not be considered to be synchronous with the later Karoo LIP, and the remaining Karoo-aged dykes do not form a radiating pattern that could have resulted from a mantle plume, as first described by Curtis et al. (2008). It also appears highly unlikely that these dykes were fed by laterally intruding melts from the Weddell Sea or Karoo triple junction, but rather from a source coincident with thinned crust between Africa and Antarctica; a position between (b) and (c) in Fig. 4.

Having considered the relevant dyke swarms in relation to the pre-existing crustal architecture and in relation to relative and absolute ages, we can make two basic inferences. Firstly, the Karoo triple junction has been an important magmatic source for the Karoo LIP and, secondly, there is no evidence from the dyke swarms themselves for the influence of a mantle plume. There is a conflict between the magmatic sources amongst the southern African dykes, however. The Northern Lebombo and Okavango dyke swarms certainly seem to have been fed from a magma source related to the triple junction at Mwenezi, but the Underberg dyke swarm may have been fed from the SE and the Rooi Rand dyke swarm intruded at a later stage, possibly after the Jozini rhyolites, and unrelated to a mantle plume. Without magma flow or any other data the origin of the SW-1 dyke swarm remains enigmatic. It is most likely related to extension and crustal thinning between SE Africa and Antarctica (N–S direction in Fig. 4). It is apparent that taking into account the triple junction geometry and the magma flow directions is insufficient evidence for proposing a mantle plume origin for the outwardly radiating dyke swarms and associated magmatism.

7.3. Magma flow and passive melting

If a mantle plume did not give rise to the triple junction and the dyke swarms during the Jurassic, under what conditions did it arise, and how did the distribution of Karoo magmatism occur? Firstly, it is important to note that a triple junction can develop because of tensional stresses in the lithosphere owing to its curved shape, because a curved surface (i.e. the Earth's surface) responds differently under tension to a flat surface. Perhaps more fundamentally, it is likely that a principle of energetic economy is at work in nature, resulting in an intrinsic geometry in the event of crack formation, whether it is fracture patterns, mudcracks, ice cracks, columnar jointing, and indeed tectonic plates (Anderson, 2002).

Therefore, if dykes or dyke swarms exploit regions of lithospheric weakness, which has been demonstrated, it follows that the development of extensional regions on a curved surface would converge in triple junction-type geometry, without necessitating the influence of an upwelling plume from beneath.

In a passive model of magmatism the development of the Karoo LIP may have started because of a lack of cooling rather than active heating by a plume. Anderson et al. (1992) have suggested that crustal thinning above non-isothermal, inhomogeneous mantle is responsible for the formation of volcanic ridges and hotspots, as the homogeneity and thermal state of ambient mantle assumed in the plume hypothesis have been largely ruled out by geophysics. Passive melting may manifest as vertical magma flow in dykes, perhaps with some lateral flow outward from relatively small igneous centres. Evidently the accumulation and/or generation of heat and subsequent magmatism beneath the supercontinent of Gondwana would have weakened the crust, making continental break-up more likely as opposed to necessarily being a direct result of the emplacement of a plume beneath the supercontinent (Storey et al., 1992; White, 1997; Hawkesworth et al., 1999).

There has been considerable research into the potential effects of thermal insulation of the sub-continental mantle during times of supercontinent assembly that precedes flood basalt formation and continental break-up (Anderson, 1994; Lenardic et al., 2005; Coltice et al., 2007; O'Neill et al., 2009; Heron and Lowman, 2011; Rolf et al., 2012). For example, the depleted MORB-like magmas of Dronning Maud Land, generated from an anomalously "hot" source, are interpreted as being consistent with the melting model of Coltice et al. (2009). This potentially implicates internal heating of the upper mantle in Karoo magmatism, as opposed to a mantle plume (Heinonen et al., 2010).

Coltice et al. (2009) have simulated mantle convection beneath a supercontinent in 3-D, and the results show that a supercontinent at the Earth's surface can have a significant impact on convection and temperature within the mantle. Their mantle global warming model predicts that (1) heating occurs over an area comparable in size to the overlying supercontinent, (2) melting occurs at modest ($\leq$100 °C) temperature increases in the mantle, (3) melt is predominantly sourced from the asthenosphere and the continental lithosphere and (4) tectonic processes control the extraction of melt.

For continental cover >10% of the Earth's surface, the authors expect the sub-continental mantle to increase in temperature by >75 °C over an area comparable in size to the "insulating" supercontinent. Rolf et al. (2012) have found that this figure may be as high as 100 °C. This process of mantle warming, alternatively to mantle plumes, is thus a feasible mechanism for the origin of some CFBs and the authors suggest that the mantle global warming model is becoming a more favourable explanation for the origin of the Karoo LIP. Their model is understandably not valid for all CFB provinces, including those that are unrelated to supercontinent dispersion, and for those in which mantle plume signature can be robustly demonstrated.

Modelling which incorporates both continental and oceanic plates in a supercontinent-type formation suggests that this insulating effect is not strong and does not assist in elevating the mantle temperatures (Heron and Lowman, 2011). Instead, the presence of subduction zones surrounding such a supercontinent appears to strongly influence thermal convection beneath the lithosphere (O'Neill et al., 2009) and furthermore, a lack of movement of a supercontinent is a predominant factor in mantle insulation.

In modelling the assembly of a supercontinent Heron and Lowman (2011) have shown that a mantle plume forms beneath the crust at ~150 Ma after the subduction of oceanic crust between the adjoining continental plates. The return mantle flow, therefore, is predisposed to be beneath the supercontinent as the subduction zones are marginal (O'Neill et al., 2009). In the case of southern Africa, this would have occurred ~150 Ma prior to the Pan-African assembly of Gondwana at ~540 Ma (Rino et al., 2008). However, major expressions of Karoo volcanism did not occur for a further ~360 Ma. From the modelling, it appears that 6 mantle transit times are required to produce continental dispersal (Zhong and Gurnis, 1993; Heron and Lowman, 2011). This is, perhaps fortuitously, equal to 60 Ma for each mantle transition, or 360 Ma. It is after 360 Ma that an unstable thermal anomaly forms beneath the supercontinent in the model, related to continuing subduction at the supercontinent margins (O'Neill et al., 2009).

Similarly, Rolf et al. (2012) found in 3D modelling of mantle thermal structure that the position of oceanic plate boundaries relative to the continental margin affects the temperature of the mantle beneath the continent. Indeed, a convection cell almost completely covered with a continent was found to be efficient at insulating the sub-continental mantle, to the extent that sub-continental mantle became up to 140 K hotter than sub-oceanic mantle. Such a temperature increase could induce partial melting in mantle beneath a supercontinent, leading to enhanced magmatic activity and potentially continental breakup.

Relating the above model directly to Gondwana break-up and the facets of the pre-cursory Karoo LIP is speculative at best. However, it does fit quite well with the time constraints shown above, and the structure and the geochemistry of the Karoo LIP that is best explained by a sub-continental lithospheric mantle melt source (Sweeney et al., 1991, 1994; Hawkesworth et al., 1999; Jourdan et al., 2007a). Jourdan et al. (2009) have shown that only a slight geochemical evolution has occurred in the enriched, shallow lithospheric mantle beneath southern Africa in the 900 Ma prior to the development of the Karoo LIP. These authors conclude that the sub-continental lithospheric mantle model for the Karoo is well explained by enhanced melting resulting from subduction processes prior to, and the insulating effects during, the formation of Pangaea which broke-up ~200 Ma ago. This is consistent with the work of Heron and Lowman (2011), who show that subduction related intra-continental magmatic processes are an important consequence of supercontinent assembly.

8. Conclusions

We have briefly reviewed the geological setting, field relationships, petrology and geochronology of the Karoo LIP, focussing on magma flow determinations in dyke swarms for which there are data. We can conclude that:

1. Evidence for vertical and lateral flow exists in the magnetic and plagioclase SPO fabric of the Northern Lebombo, Okavango and Rooi Rand dyke swarms.
2. Lateral flow predominates in the Okavango dyke swarm and the youngest dykes of the Northern Lebombo dyke swarm. This is consistent with the Karoo triple junction being a viable magma source for these swarms.
3. The only position for a mantle plume that would be consistent with this finding would be one emplaced directly beneath Mwenezi (Fig. 4a), a position inconsistent with the other dyke swarms and the south-to-north younging of magmatism in the Karoo LIP.
4. Evidence of lateral magma flow in the Rooi Rand dyke swarm is consistent with the findings for the Northern Lebombo dyke swarm, but difficult to interpret in light of its restricted occurrence and compositional and age differences. Steep flow coupled with evidence of lateral flow from AMS and SPO is most likely the result of outward flow from a central magmatic source.
5. The positions and orientations of the Okavango dyke swarm, Save–Limpopo dyke swarm and Lebombo monocline are a product of the lithospheric structure, most obviously with respect to the ENE-trending structural fabric of the Limpopo Belt and the eastern edge of the Kaapvaal Craton.
6. Dykes of west Dronning Maud Land show a more protracted history (~205 Ma to ~175 Ma) than those of the Karoo LIP. The age and progression of magma flow characteristics are similar to the Karoo dyke swarms, however. Magma flow in the Straumsvola and Jutulrøra dyke swarms is similar to that in the Northern Lebombo and

Okavango dyke swarms. Dykes closest to Straumsvola have been fed (sub-) vertically and emplaced under high magmatic pressure.

7. Thus, it is highly unlikely that these dykes were fed by lateral intruding melts from the Weddell Sea or Karoo triple junction, but rather from a source coincident with a position between (b) and (c) in Fig. 4. There is, therefore, inconsistency between magma sources of the southern African and Antarctic dykes of the Karoo LIP, and basaltic eruptions appear ~3 Ma after those in southern Africa.
8. The lack of plume-related regional uplift deems it unlikely that the Karoo triple junction developed as the result of a mantle plume.
9. These results, in the framework of the known structure, field relationships and geochronology related to the Karoo triple junction, are significant. The magma flow regime appears superficially consistent with the pattern of radiating dyke emplacement typically associated with mantle plumes. However, in conjunction with high-resolution $^{40}Ar/^{39}Ar$ dating, relative timing of dyking events and geochemical and geodynamic considerations, the results do not necessitate a mantle plume for their origin.

A single, large-scale natural geological event did not give rise to the entire Karoo LIP, other associated magmatism and the variety of dyke swarms over a period of ~10 Ma. There can be little doubt that a melt source in the sub-continental lithospheric mantle, which has undergone only slight geochemical evolution in the last ~1.0 Ga (Jourdan et al., 2009), is the most viable bulk melt source for the magmas of the Karoo LIP erupted and intruded in a region with a protracted tectonic history. Relatively small, independent igneous centres most likely gave rise to the various dyke swarms considered here, as evidenced from flow related, field and geochronological data.

Ideally, more flow-related petrofabric studies and geochronology in the lesser known dyke swarms, and sills, mentioned are needed to further broaden the understanding of the origin of the Karoo LIP. That is, provided they are linked to rigorous field, structural and geodynamic constraints.

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