

Stable isotope study of the Archaean rocks of the Vredefort impact structure, central Kaapvaal Craton, South Africa

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Abstract The Vredefort dome in the Kaapvaal Craton was formed as a result of the impact of a large meteorite at 2.02 Ga. The central core of Archaean granitic basement rocks is surrounded by a collar of uplifted and overturned strata of the Witwatersrand Supergroup, exposing a substantial depth section of the Archaean crust. Orthogneisses of the core show little variation in whole-rock $\delta^{18}\text{O}$ value, with the majority being between 8 and 10‰, with a mean of 9.2‰ ($n = 35$). Quartz and feldspar have per mil differences that are consistent with O-isotope equilibrium at high temperatures, suggesting minimal interaction with fluids during subsequent cooling. These data refute previous suggestions that the Outer Granite Gneiss (OGG) and Inlandsee Leucogranofels (ILG) of the core represent middle and lower crust, respectively. Granulite-facies greenstone remnants from the ILG have $\delta^{18}\text{O}$ values that are on average 1.5‰ higher than the ILG host rocks and are unlikely, therefore, to represent the residuum from the

partial melting event that formed the host rock. Witwatersrand Supergroup sedimentary rocks of the collar, which were metamorphosed at greenschist-to amphibolite-facies conditions, generally have lower $\delta^{18}\text{O}$ values than the core rocks with a mean value for metapelites of 7.7‰ ($n = 45$). Overall, through an ~20 km thick section of crust, there is a general increase in whole-rock $\delta^{18}\text{O}$ value with increasing depth. This is the reverse of what is normal in the crust, largely because the collar rocks have $\delta^{18}\text{O}$ values that are unusually low in comparison with metamorphosed sedimentary rocks worldwide. The collar rocks have δD values ranging from -35 to -115‰ (average -62‰, $n = 29$), which are consistent with interaction with water of meteoric origin, having a δD of about -25 to -45‰. We suggest that fluid movement through the collar rocks was enhanced by impact-induced secondary permeability in the dome structure.

Keywords Vredefort impact structure · Stable isotopes · Fluid-rock interaction · Witwatersrand Supergroup · Archaean basement

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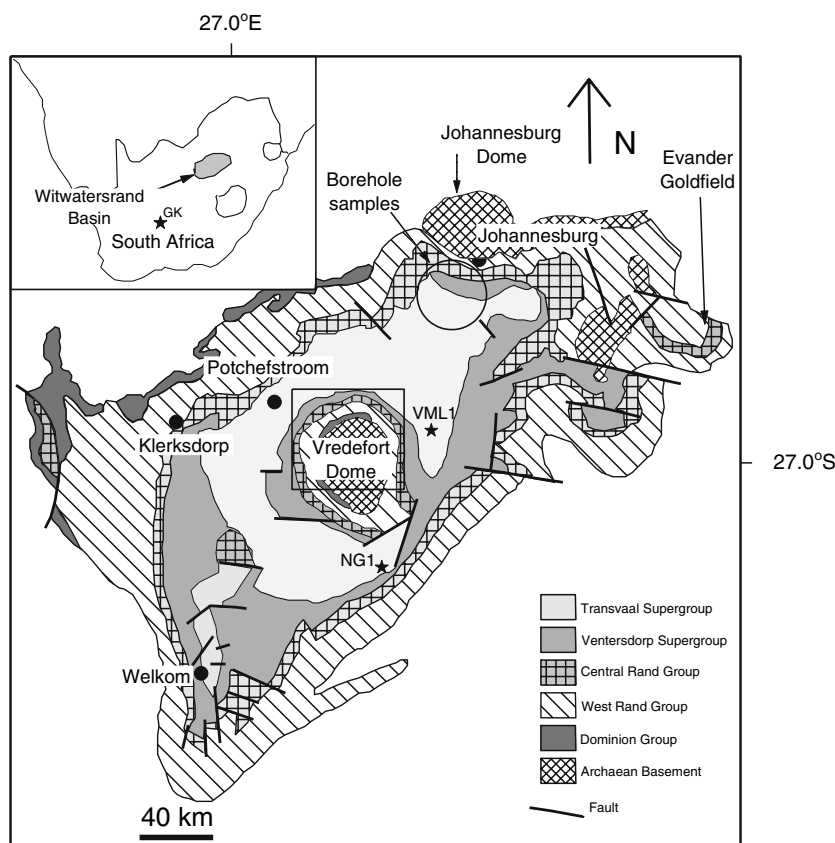
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Introduction

The Vredefort dome is situated in the Witwatersrand basin of the central Kaapvaal Craton (Fig. 1), adjacent to the world's largest gold province, and is the largest and oldest meteorite impact structure on earth (Gibson and Reimold 2001). The 90 km wide structure is deeply eroded and consists of an uplifted core of Archaean basement rocks, exposing a section of the middle and possibly lower crust, surrounded by a collar of upturned Archaean and Proterozoic strata. The structure, therefore, provides a unique opportunity to study a crustal section from the surface down

Fig. 1 Map showing the location of the Vredefort dome within the Witwatersrand basin. The areas covered in Fig. 2, in the vicinity of the Vredefort dome, are indicated. The Witwatersrand Supergroup samples that were analysed, in addition to the Vredefort samples, were collected from just south of the Johannesburg dome (marked “borehole samples”). GK the Goedehoop kimberlite pipe



at least to the middle crust where the metamorphic grade increases from greenschist facies in the outer collar to granulite facies in the centre (Bisschoff 1982, 1988). Previous studies of the metamorphism in the Vredefort collar have documented hydrous mineral assemblages in pseudotachylite veins produced by the impact event (Gibson and Wallmach 1995). It is, therefore to be expected that the rocks of the Vredefort dome have a history of fluid–rock interaction, and Frimmel et al. (1999) suggested that there was a connection among percolation of meteoric waters, the Vredefort meteorite impact and gold mobilisation in the Witwatersrand basin. The aims of this study are as follows:

1. To determine the oxygen isotope composition of the core rocks and their constituent minerals.
2. To determine the $\delta^{18}\text{O}$ and δD values of the collar rocks and thereby evaluate the evidence for fluid–rock interaction in the dome.
3. To establish a stable isotope profile through the Vredefort structure from the centre of the core to the outer edge of the collar.
4. To determine the source of infiltrating fluids and estimate the degree of fluid/rock interaction experienced by various rock types and to establish the relationship (if any) between the impact structure and fluid–rock interaction.

Geological setting

The origin of the Vredefort structure was for long a matter of debate. Numerous authors suggested non-impact origins for the structure (e.g. Slawson 1976; Colliston 1990; Nicolayson and Ferguson 1990), but it is now generally accepted that the structure was formed by a meteorite impact, as first suggested by Boon and Albritton (1936), at 2.02 Ga (Kamo et al. 1996).

The dome consists of a 40–50 km wide core of amphibolite- to granulite-facies Archaean granite-gneisses and a 15–20 km wide collar of upright to overturned, late Archaean to early Proterozoic metasedimentary rocks (Armstrong et al. 1991). Previous studies have indicated that the uplift led to exposure of a cross-section through the mid-crust of the Kaapvaal Craton, exposing progressively older and deeper rocks towards the centre of the uplift (Slawson 1976; D. Stepto 1979, unpublished data; Hart et al. 1981; Lana et al. 2004). The core rocks experienced mid-crustal, upper amphibolite- to granulite-facies metamorphism at ~3,090–3,080 Ma (Hart et al. 1999; Lana et al. 2003, 2004). The metamorphic grade increases towards the centre of the core, consistent with increasing levels of rebound-induced uplift associated with the impact event. Metamorphism within the collar is slightly more

enigmatic. Sediments of the lower Witwatersrand Supergroup have been metamorphosed to a maximum of mid-amphibolite-facies grades. This metamorphism is focused on the northwestern quadrant of the dome (e.g. Stevens and Preston 1999). The origin of this metamorphism has been ascribed to contact effects by alkali granite bodies that intrude this portion of the structure (Bisschoff 1982, 1988). However, Gibson and Wallmach (1995) and Gibson et al. (1998) argued that peak metamorphism was a function of crustal heating during the Bushveld magmatic event, and that this was overprinted by post-shock thermal metamorphism related to the impact process.

Previous mapping by D. Stepto (1979, unpublished data) suggested that the core of the dome comprised two concentrically arranged units, the Outer Granite Gneiss (OGG) and the Inlandsee Leucogranofels (ILG). Based on major and trace element data, Hart et al. (1990) suggested that the ILG and OGG rocks represent, respectively, the lower and upper crust. Hart et al. (1990) also suggested, on the basis of geochemical and isotope data, that harzburgites exposed in the south-central parts of the core represented uplifted Archaean upper mantle. Subsequent U–Pb single zircon geochronology has led to the suggestion (Moser et al. 2001) that the high-grade metamorphism at 3.1 Ga was linked to significant reworking of 3.3 Ga old lower crustal rocks and formation of a 10 km thick granitic layer that corresponded to the rocks presently exposed in the outer part of the core of the dome. The transition zone would then represent a pre-2.0 Ga intracrustal thrust zone or a mid-crustal tectonic discontinuity, which was reactivated during the 2.02 Ga impact event. However, Lana et al. (2003) found no evidence for a tectonic boundary between the OGG and the ILG, and proposed that the two rock types in the core shared a common polyphase tectonic history, where the Vredefort discontinuity represents the transition between melt-depleted granulites and melt-rich amphibolite facies gneisses formed during the 3.1 Ga event. The southern part of the core is largely covered by Karoo Supergroup sedimentary rocks and dolerite sills. The few outcrops of crystalline basement rocks are exposed in a small window into the sedimentary rocks of the Karoo Supergroup in the southeastern sector. These comprise greenschist- to amphibolite-facies gneisses and metavolcanic rocks of the Greenlands Formation greenstones.

The collar rocks are of the same strata as the rest of the Witwatersrand basin (Fig. 1) and can be stratigraphically divided, from base to top, into the Dominion group, which is overlain first by the Witwatersrand Supergroup then the Ventersdorp and Transvaal Supergroups. The lavas of the Dominion group are mainly mafic, unconformably overlie the Archaean basement rocks, and have been dated at 3.07 Ga (Armstrong et al. 1991). The Witwatersrand Supergroup was deposited between 2.99 and 2.78 Ga

(Frimmel 2005) and consists of fluvial and shallow-marine siliciclastic metasedimentary rocks. These rocks are generally transgressive and upward coarsening, with gold mineralisation in conglomerates, particularly in the upper Central Rand group. The Witwatersrand Supergroup is overlain by the sub-aerial basaltic lavas of the Ventersdorp Supergroup (Gibson and Reimold 2001), which have an age of 2.71 Ga (Armstrong et al. 1991). This implies a hiatus of at least 60 million years at the unconformity (Frimmel 2005). The base of the Transvaal Supergroup comprises dolomites and iron formations of the Chuniespoort group, deposited in a shallow marine environment. This was followed by the siliciclastic Pretoria group, forming the top 3 km of the succession (Walraven and Martini 1995).

The crystalline basement rocks and the overlying Witwatersrand Supergroup strata have been intruded by a number of pre- and post-impact igneous rocks. These include nepheline syenite, alkali granite, gabbro, dunite and wehrlite, which are generally found in the northwest quadrant of the dome (Fig. 1). The largest intrusions in the study area are the Schurwedraai and Baviaanskrans plutons, which have been dated using the single zircon method at 2.078 Ga (Moser 1997). More recently, the Schurwedraai alkali granite has been dated using single zircon methods at $2,052 \pm 14$ Ma (Graham et al. 2005), which is identical, within error, to the accepted age of the emplacement of the Bushveld Complex (2,059 Ma, Buick et al. 2001) and significantly older than the date of the impact.

Of special interest are the occurrences of impact origin melts. Pseudotachylite is present on a scale of tens to hundreds of metres (Thompson and Spray 1996), but does not have any obvious geometric relationship with shear zones; hence the formation mechanism is unclear. One likely explanation is formation by explosive brecciation and shock melting related to the impact event (Dressler and Reimold 2001). Pseudotachylite is most voluminous within exposures in the core, but also common in rocks of the collar, and throughout the Witwatersrand Basin (e.g. Gartz and Frimmel 1999). The Vredefort granophyre forms a series of dykes, which cut the core. On the basis of a Re–Os isotope study, Koeberl et al. (1996) confirmed that the granophyre is of impact origin and contains up to 0.2% of a chondrite meteorite component.

Fluid flow, alteration and gold mobilisation are well documented in the Witwatersrand basin and have been shown to have operated on a variety of scales in the area of the goldfields along the northern and northwestern margins of the basin (e.g. Phillips and Myers 1989; Robb and Meyer 1991, 1995; Frimmel et al. 1993; Fox and Winkler 1997; Barnicoat et al. 1997; Phillips et al. 1997; Stevens et al. 1997; Frimmel et al. 1999; Parnell 2001; Vollbrecht et al. 2002). In combination, these studies have shown that

the gold-bearing ore parageneses record a complex history that can be correlated with several opportunities for widespread fluid–rock interaction in the history of the Witwatersrand basin (Robb and Meyer 1991, 1995; Frimmel et al. 1993).

The only previous study of stable isotopes in Vredefort rocks (O’Neil et al. 1987) determined the whole-rock $\delta^{18}\text{O}$ values of 24 samples taken from the core and the collar. Their principal conclusions were that the OGG and ILG rocks had very similar $\delta^{18}\text{O}$ values that were typical of I-type granites. The small number of pseudotachylite samples they analysed had similar values as that of their host rocks, and the collar rocks of the lower Witwatersrand Supergroup showed a similar range in $\delta^{18}\text{O}$ values as the core samples. Preliminary interpretations based on the present work were reported by La Grange et al. (1999).

Sampling and petrography

Sample collection

In the Archaean basement core, samples were collected along a NW–SE trending traverse (Figs. 1, 2). Samples were taken approximately every 500 m along the traverse, except in the amphibolite to granulite-facies transition, where samples were taken approximately every 50–100 m (Fig. 2). The HM suite of samples was collected from the Helpmekaar farm and included granulite-facies metapelites from within the ILG.

In the collar, samples 412–426 were collected as a continuation of the NW–SE trending traverse taken in the core (Fig. 2). Most of the remainder of the collar samples came from the NW quadrant of the structure, where exposures were most complete. A significant number of samples (VC39–53) were collected from the Kommandonek farm, where a complete section through the collar-core contact zone was exposed. Samples were taken from intrusions that cut the stratigraphy (VC22–24), and a traverse of collar rocks collected at 50 m intervals was made through the Booyens Formation shale (VC27–31), a locality known for the presence of shatter cones. Samples were also collected of pseudotachylites and their host rocks. Samples from a borehole through the West Rand group in the SE part of the structure have been described by Stevens and Preston (1999). Samples of rocks from drill-core through the Witwatersrand Supergroup in areas away from the Vredefort dome were also analysed (Fig. 1) for comparative purposes.

Petrography of studied samples

The core rocks comprise granitic, trondhjemitic and tonalitic gneisses (Gibson and Reimold 2001), which may

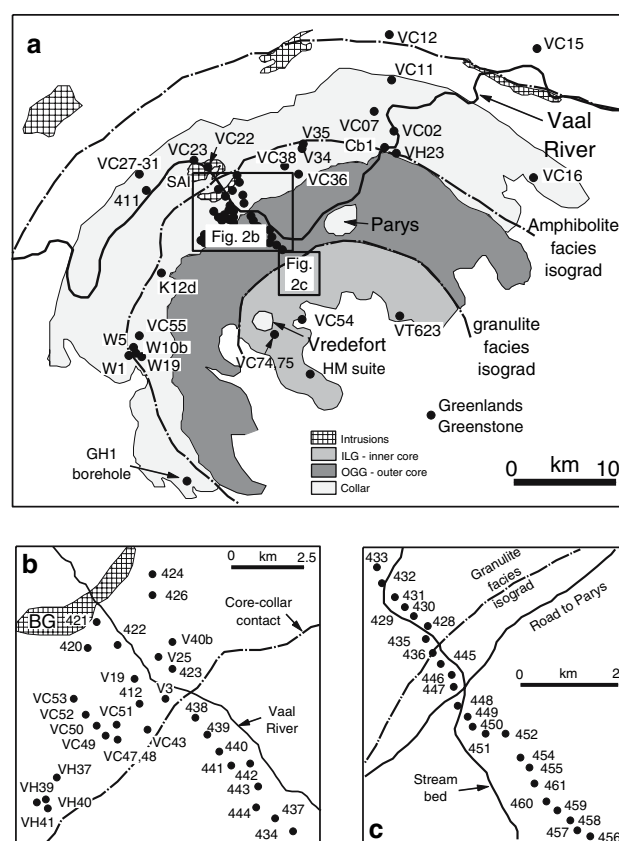


Fig. 2 **a** Geological map of the Vredefort dome, showing the locations of analysed samples, and the towns of Parys and Vredefort. Most samples were collected along a transect through the NW quadrant of the structure. *SAI* Schuurdrain alkaline complex (granites and nepheline syenites). **b** Location of samples taken close to the Vaal river in the NW quadrant. *BG* Baviaanskrans granite pluton **c** shows the location of samples along the transect through the OGG/ILG boundary. Metamorphic grades are from Gibson and Wallmach (1995)

contain xenoliths of metapelitic supracrustal rocks. The migmatized outer amphibolites consist of a grey trondhjemitic component and pink, coarse-grained granite gneiss with alkali feldspar megacrysts. The trondhjemitic is medium grained and typically contains plagioclase + quartz + chlorite + biotite + opaques + K-feldspar \pm hornblende, while the coarser grained granite-gneiss usually comprises quartz + K-feldspar + plagioclase + biotite \pm opaques \pm chlorite \pm muscovite. Several minerals in these rocks show impact related recrystallisation features such as undulose extinction and planar deformation features in quartz and kink-banding in biotite. Sericitisation and recrystallisation, as well as perthitic and myrmekitic textures are observed in feldspars. The inner granulites are mainly coarse-grained granitoid gneisses, which are similar to those of the OGG, although more leucocratic, and containing orthopyroxene. Some charnockites are also observed; “xenoliths” (eTable 1) in the trondhjemitic and tonalites of the ILG may be up

to several kilometres in diameter (Gibson and Reimold 2001) and include mafic granulites and metapelites.

A limited number of pseudotachylites were analysed and these samples were fine grained apart from xenoliths of the host rock. Samples VC74 and VC75 were from the Vredefort granophyre (average 67 wt.% SiO₂, Gibson and Reimold 2001) and were fresh orthopyroxene-phyric rocks, with quench textures and a matrix of feldspar and quartz. The analysed samples contained a low proportion of xenolith material.

The quartzites in the collar contain both fine-grained recrystallised quartz and coarser angular grains. Besides quartz, the quartzite samples generally contain some sericite, along quartz grain boundaries and replacing occasional feldspar grains, and micro-scale iron oxide-rich veins. Micro-structures show evidence of both diffusional mass transfer and dislocation creep. Planar deformation features, formed by the extreme shock induced pressure of the impact event, are also observed. The metapelites are characterised by typical greenschist and amphibolite facies mineral assemblages (depending on their position within the collar), although some consist of recrystallised quartz grains separated by sericite. Sample VC47 is richer in sub-hedral, coarser muscovite than the other samples, whereas VC38, 52 and 53 are dominated by biotite, but also contain garnet, cordierite, chlorite, muscovite and quartz. The ironstone samples are predominantly composed of iron-rich orthoamphiboles, quartz and magnetite. In agreement with previous studies (e.g. Bisschoff 1982), the major amphibole phase in these samples was identified as grunerite, sometimes accompanied by cummingtonite.

The collar metabasites (eTable 2) typically have a matrix consisting of fine-grained plagioclase, epidote, actinolite, chlorite and minor quartz and calcite, reflecting the greenschist-facies conditions of metamorphism in the portion of the collar where these rocks were sampled. The felsic intrusions analysed represent a variety of compositions, including granite, alkali granite and nepheline syenite. In addition to some preserved igneous minerals, they contain metamorphic minerals such as chlorite, amphibole and epidote. In accordance with the pre-impact age of these rocks, they are also observed to contain shock metamorphic features such as pseudotachylite veins and planar deformation features in quartz.

Analytical methods

Both whole-rock and separated minerals were analysed from the core samples. The collar rocks are fine grained with intergrown minerals, and it was not possible to obtain mineral separates for analysis. Both hydrogen and oxygen isotope ratios were measured in the stable isotope labora-

tory at the Department of Geological Sciences, University of Cape Town. Isotope ratios were measured using either a Finnegan MAT252 or a DeltaXP mass spectrometer. Results are reported in standard δ -notation relative to SMOW, where δD or $\delta^{18}O = (R_{\text{sample}}/R_{\text{standard}} - 1) \times 1,000$, $R = {}^{18}O/{}^{16}O$ and $R = D/H$ for oxygen and hydrogen, respectively. Where mineral separates were analysed, these were prepared by hand-picking under a binocular microscope from ~2 mm sieved fractions. Quartz separates were cleaned using warm 25% HF in order to remove any feldspar contaminant.

For the oxygen-isotope determination, ~10 mg of material was used for each sample. Ten samples were loaded in each run, two of which were quartz standard. For the VC suite of samples, the standard used was NBS28; for the other samples, an internal standard of Murchison Quartz (MQ) was used. Before loading, the samples were dried at 50°C overnight. After loading into Nickel tubes, they were degassed under vacuum at 200°C then reacted with ClF₃ at 550°C for at least 3 h, and at 600°C overnight for garnet (Vennemann and Smith 1990). This reaction releases oxygen from the silicate in the samples. The following day the free oxygen was passed over a hot carbon rod, converting the O₂ to CO₂ gas, which was collected in “break seal” tubes. The average value obtained for NBS28 or MQ was used to normalise the raw data in each run to the SMOW scale. The average values obtained for MQ and NBS28 in a given run did not differ by more than 0.5‰ from the accepted value. The average difference between duplicate NBS28 or MQ analyses in the same run was 0.34‰, equivalent to a 1 σ of 0.2‰. The method for oxygen extraction and the facilities at UCT are explained in further detail by Harris and Ashwal (2002).

The method of Vennemann and O’Neil (1993) was used for the determination of hydrogen isotope ratios. Water was extracted from between 50 and 100 mg of whole-rock powder with the aim of extracting 1–2 mg H₂O. Samples were dried at 110°C overnight in an oven at 200°C before extraction of water by pyrolysis at ~1,200°C. The water collected was reduced to H₂ using low blank ‘Indiana Zinc’ as a reagent. In each set of 10–15 samples, two internal standards were analysed to monitor the analytical precision. The raw data was normalised to SMOW using an internal water standard (CTMP; $\delta D = -9\text{‰}$). A silicate standard Cape Granite biotite (CGBi; $\delta D = -57\text{‰}$) was analysed nine times and gave an average value of -54‰ ($\pm 2.5\text{‰}$). The water content of the samples was also determined, using either the pressure or the voltage on the mass 2 collector, measured during introduction of the H₂ to the mass spectrometer with the inlet section set to a fixed volume. The relationship between volume and voltage, and the amount of water was determined using accurately weighed amounts of a water standard. Repeated measure-

ments of the water standards indicate that the relative analytical error for water content is $\pm 1.25\%$ for a sample yielding 2 mg of water. However, in samples of lower water yield, common in this study, the analytical error is likely to be higher. The CGBi standard gave an average water yield of 3.72 ± 0.15 wt.% (1σ , $n = 9$).

Results

The core

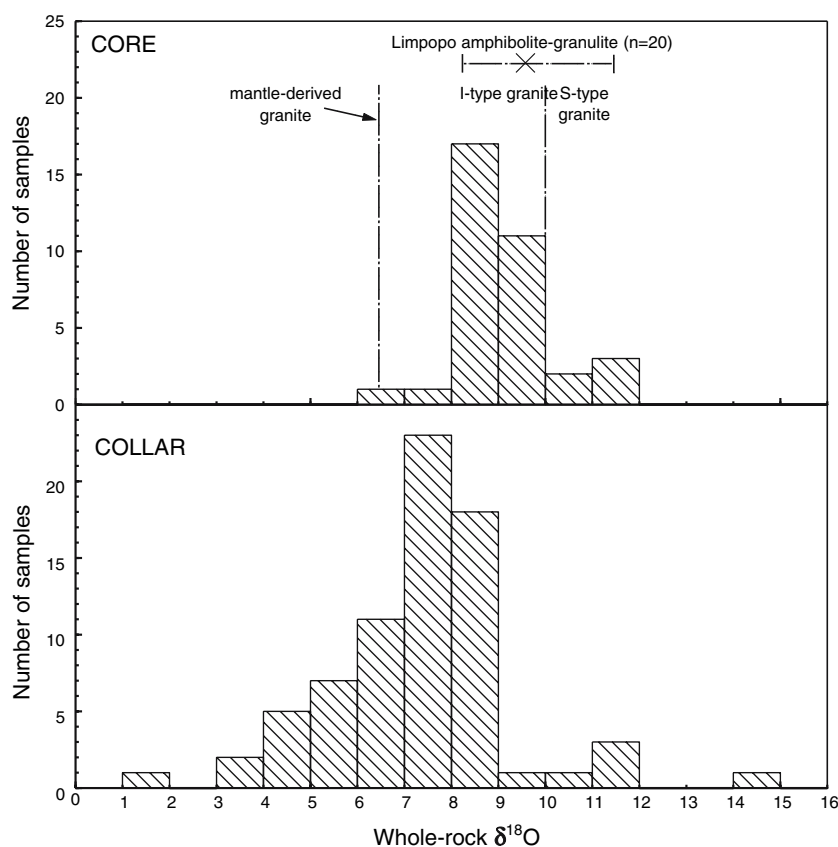
The core rocks have a relatively restricted range in oxygen isotope ratios (eTable 1, Fig. 3). The $\delta^{18}\text{O}$ values for the outer amphibolite facies granitic rocks (OGG) range from 8.4 to 11.5‰, with a mean value of $9.32 \pm 0.90\text{‰}$ (1σ , $n = 14$). The single sample of dolerite analysed from the outer part of the core has a $\delta^{18}\text{O}$ value of 6.8‰, which was omitted from this average. The ILG felsic rocks (granites, leucogranites and charnockites) have $\delta^{18}\text{O}$ values, which range from 8.2 to 10.8‰ with a mean of $9.02 \pm 0.76\text{‰}$ (1σ , $n = 19$). Granulite xenoliths within the ILG range in $\delta^{18}\text{O}$ value from 6.4 to 12.1‰ with a higher mean value of $10.49 \pm 1.53\text{‰}$ (1σ , $n = 14$) than the host rocks. The lowest xenolith $\delta^{18}\text{O}$ value of 6.4‰ is a mafic xenolith

(454a), and expected due to its higher mafic mineral content relative to the granitoids of the OGG and ILG. The key feature of these data are, therefore, the comparative uniformity of the OGG and ILG data and the fact that the garnet-bearing restite xenoliths have significantly higher $\delta^{18}\text{O}$ values (on average $>1\text{‰}$) than the more felsic rocks that enclose them.

Major and trace element analyses of the core samples were presented by M. S. La Grange (2004, unpublished data). Apart from the xenolith sample, there is no significant difference in composition between the OGG and ILG samples. In terms of major elements, SiO_2 ranges from 60 to 76 wt.%, with one mafic xenolith (454a) having a value of 49.4 wt.%. However, those samples with <67 wt.% SiO_2 are granulite xenoliths, so the range in composition of the host rocks is relatively restricted. The majority of samples (and all the OGG samples) have alumina saturation indices (ASI of Zen 1986) between 1 and 1.1 showing that they are mildly peraluminous. Four of the granulite xenoliths have ASI values that are significantly higher (1.7–3.2), consistent with them being garnet-rich. There is no correlation between chemical composition and whole-rock $\delta^{18}\text{O}$ values for the granitoid gneisses.

Separated quartz and feldspar from the core samples have $\delta^{18}\text{O}$ values that average 9.94‰ (± 1.07) and 8.76‰

Fig. 3 Histograms of core and collar $\delta^{18}\text{O}$ values. Also shown are the range and mean $\delta^{18}\text{O}$ values for amphibolite-granulite facies rocks of the southern marginal zone of the Limpopo belt (Vennemann and Smith 1992)



(± 0.61). The average difference between quartz and feldspar is 0.81‰ , which is consistent with O-isotope equilibrium at high temperature (e.g. Chiba et al. 1989). The $\delta^{18}\text{O}$ value of feldspar and whole-rock is plotted against that of quartz from the same rock in Fig. 4. The feldspar represents a mixture of plagioclase and alkali feldspar and if it is assumed that the average An content is 30%, the expected per mil difference (Δ) between quartz and feldspar at 1,000 and 550°C is -0.77 and -1.85‰ , respectively. The former represents the probable upper limit for the crystallisation temperature of the igneous protolith; the latter is the approximate temperature of closure to oxygen diffusion of quartz and feldspar, which make up most of the rock (e.g. Giletti 1986). About half the samples plot between the two isotherms in Fig. 4 and the remaining samples have $\Delta_{\text{feldspar-quartz}}$ values greater than -0.77‰ , which are indicative of low temperature alteration (Fig. 4). Whole-rock quartz pairs are also plotted in Fig. 4 and these are more variable than the feldspar-quartz pairs. Another important feature of the mineral data (eTable 1) is that the granulite xenoliths with high whole-rock $\delta^{18}\text{O}$ values (456a and b) also have high quartz and garnet $\delta^{18}\text{O}$ values. This shows that the higher whole-rock $\delta^{18}\text{O}$ values of the xenoliths relative to the host rock cannot be explained by low-temperature alteration, because both these minerals are resistant to O-isotope exchange at low temperatures.

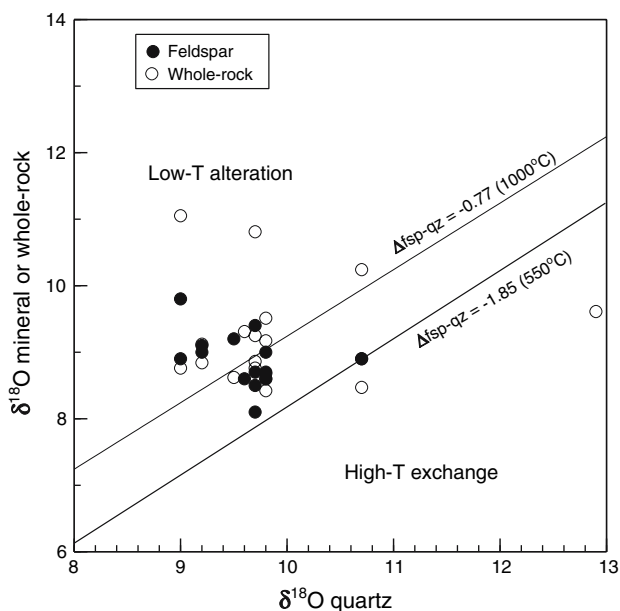


Fig. 4 Plot of feldspar and whole-rock $\delta^{18}\text{O}$ value versus quartz $\delta^{18}\text{O}$ value in the same sample. Isotherms for $\Delta_{\text{feldspar-quartz}}$ for 1,000 and 550°C were calculated, assuming that the feldspar was plagioclase (An_{30}), using the fractionation equations of Chiba et al. (1989). The expected domain of rocks whose feldspars were affected by low- and high-temperature alteration is indicated

The variation of quartz and feldspar $\delta^{18}\text{O}$ values with distance from the contact with the collar is shown in eFig. 1. Although there is no strong correlation between the $\delta^{18}\text{O}$ value of either mineral and distance, the quartz $\delta^{18}\text{O}$ values generally increase away from the collar-core contact, whereas the feldspar $\delta^{18}\text{O}$ values generally decrease away from the collar-core contact. Overall, this has resulted in a general increase in $\Delta_{\text{feldspar-quartz}}$ with increasing distance from the collar.

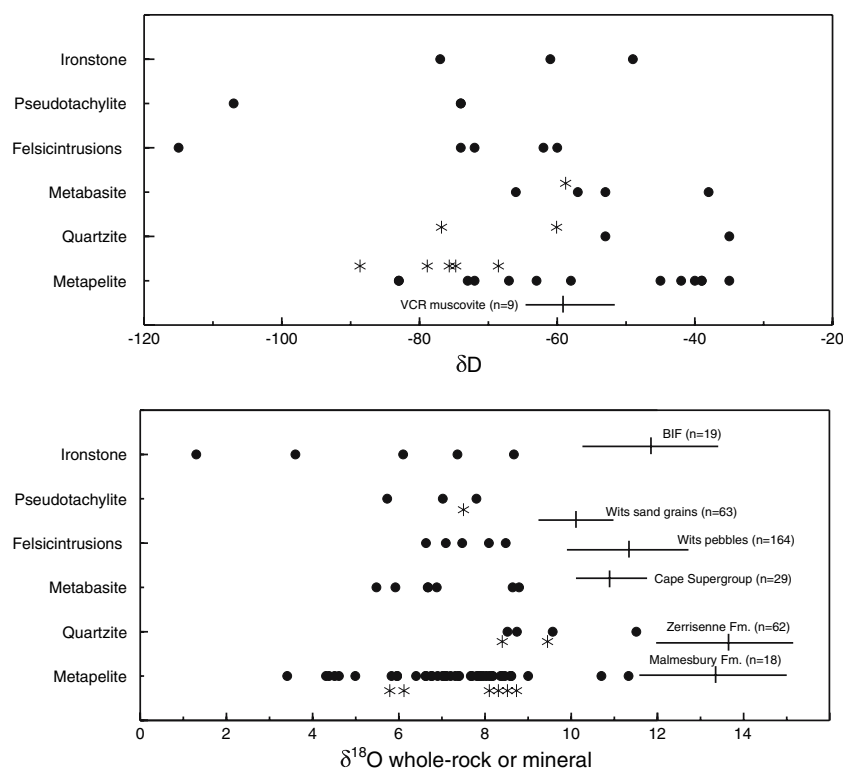
Only three samples from the core were analysed for H-isotopes. A sample of pseudotachylite and host OGG (VC18/19) have identical δD values (-74‰) and almost identical water contents (0.29 and 0.31 wt.%, respectively). One sample from the ILG (VC54) has a δD value of -60‰ and a water content of 0.48 wt.%.

The collar

The collar samples (eTable 2, Figs. 3, 5) have $\delta^{18}\text{O}$ values ranging from 1.3 to 21.2‰ (mean $7.5 \pm 2.7\text{‰}$ (1σ), $n = 61$). Overall, the collar samples have lower $\delta^{18}\text{O}$ values than those from the core (Fig. 3) and whereas the core $\delta^{18}\text{O}$ values approximate a normal distribution, the collar $\delta^{18}\text{O}$ values are negatively skewed. The quartzites (Fig. 5) nearly all have $\delta^{18}\text{O}$ values above the mean value (7.2 – 14.1‰) and the lowest $\delta^{18}\text{O}$ values are seen in metapelites and banded iron formations (BIF) (1.3 – 8.7‰). The metapelites have a range of $\delta^{18}\text{O}$ values from 3.4 to 11.3‰ (mean 7.7‰ , $n = 45$), which is low for a sedimentary rock. Pelitic rocks worldwide (albeit of more recent age) have much higher $\delta^{18}\text{O}$ values with typical values for pelitic schists of 15 – 20‰ (e.g. Eiler 2001, Fig. 7).

The δD values of the collar rocks (eTable 2) range from -115 to -35‰ , with an average of $-62 \pm 20\text{‰}$ (1σ , $n = 29$). The δD values for the Vredefort dome (Fig. 5) are within the usual values found for both igneous and sedimentary rocks, although on the less negative side (Hoefs 1997; Eiler 2001). The range and average for the various rock types are more variable than for $\delta^{18}\text{O}$, but still largely overlap (Fig. 4). The lowest δD values occur in the felsic intrusions (-115 to -60‰), with a very low value in alkali granite (-115‰) and a low value in nepheline syenite (-72‰). The highest δD values were determined for quartzites, but there are only two analyses (-53 and -35‰). A plot of δD against water content (eFig. 2) shows that, although there is no correlation between them, the samples with the most negative δD values have the lowest water content. Apart from one sample, those with >2 wt.% water have δD values of about -40‰ . The low δD values in igneous and metamorphic rocks with low water content are not unusual and are commonly explained by degassing and loss of water where D is partitioned into the exsolved fluid (e.g. Taylor 1986).

Fig. 5 The $\delta^{18}\text{O}$ and δD values determined for each rock type in the collar (solid circles). Also plotted (asterisks) are values of the additional Witwatersrand Supergroup samples analysed (eTable 2). Also shown are the range and mean values of δD for muscovite from the VCR footwall shales (Zhao et al. 2006) and the range and mean $\delta^{18}\text{O}$ of BIF rocks from the Kalahari manganese field (Tsikos et al. 2003), sand grains and pebbles from the Witwatersrand Supergroup (Vennemann et al. 1995), quartzites from the Cape Supergroup, shales from the Malmesbury formation (R. E. Diamond 1997, unpublished data) and the metapelites from the Zerrissene group (P. H. Macey 2003, unpublished data)



A small number of samples from rocks that are the stratigraphic equivalent of the collar rocks from outside of the Vredefort structure were analysed as part of this work (eTable 2) for the purpose of comparison (Fig. 5). These samples have a mean $\delta^{18}\text{O}$ of 7.8‰ ($1\sigma = 1.2\text{‰}$, $n = 8$) and δD of -72.5‰ ($1\sigma = 10.0\text{‰}$, $n = 8$).

Impact generated melts

As described above, an OGG host and a pseudotachylite vein from the core were analysed. The vein has a slightly lower $\delta^{18}\text{O}$ value than the host rock (7.8 and 8.1‰ , respectively), but δD values are identical and water contents are similar and very low. A comparable relationship was observed in a pseudotachylite/host pair from the Schuurdrai granite (eTable 2), where the vein has a $\delta^{18}\text{O}$ value of 7.0 , 0.5‰ lower than the host, δD values are similar (-107 and -115‰) and water contents are very low (vein 0.38 and host 0.22 wt.%). A pseudotachylite vein in metagabbro in the collar (VC43 and 44, eTable 2) has a $\delta^{18}\text{O}$ value of 5.7 , which is identical within error to that of its host. This pseudotachylite is more water-rich (1.02 wt.%) than those of the core, but has water content similar to the host (1.36 wt.%). The δD values are -74 (vein) and -62‰ (host). Two samples from the Vredefort granophyre (VC74 and 75) have identical whole-rock $\delta^{18}\text{O}$ values of 7.6‰ , significantly lower than the average (9.2‰) for the surrounding OGG and ILG rocks.

Discussion

Low $\delta^{18}\text{O}$ values in the collar?

Before trying to explain the generally low $\delta^{18}\text{O}$ values in the collar rocks, it is important to establish just how low these values are compared to ‘normal’ sedimentary and metasedimentary rocks. The metabasites from the collar (eTable 2) have $\delta^{18}\text{O}$ values that are typical of mafic igneous rocks, although one sample has a value $<5.7\text{‰}$, which is lower than expected for a fresh basaltic rock (e.g. Eiler 2001). Because these igneous rocks lack fresh phenocrysts, it is not possible to determine the $\delta^{18}\text{O}$ values of the unaltered rock. The alkali granite that intrudes the collar (VC23, eTable 2) has a $\delta^{18}\text{O}$ value (7.5‰) typical of I-type granites (e.g. Harris et al. 1997), and the nepheline syenite (VC22, eTable 2) has a $\delta^{18}\text{O}$ value (7.1‰) typical of nepheline syenite (e.g. Harris and Grantham 1993). These rocks show no evidence of subsequent sub-solidus changes in $\delta^{18}\text{O}$ value and have the same shock and metamorphic history as the other collar samples.

The quartzites have $\delta^{18}\text{O}$ values that are similar to those of comparable rock types in the Witwatersrand Supergroup (Harris and Watkins 1990; Zhao et al. 2006) and elsewhere (Fig. 5). For comparison, the mean and 1σ for $\delta^{18}\text{O}$ values for a large number of quartz-pebbles from gold-bearing conglomerates and sand grains (Vennemann et al. 1995) are plotted in Fig. 5. Also shown is the mean and 1σ for

$\delta^{18}\text{O}$ values for the much more recent Cape Supergroup quartzites (R. E. Diamond 1997, unpublished data), which have values similar to the Witwatersrand Supergroup quartzites. These comparisons suggest that the Vredefort quartzites have $\delta^{18}\text{O}$ values that are on average slightly, but not significantly, lower than expected for quartzite.

The origin of the ironstones analysed in this study is uncertain, but BIFs are known to exist in the Witwatersrand Supergroup (Frimmel 1995). In Fig. 5, they are compared to the Hotazel BIF rocks of the Kalahari manganese field in the Northern Cape. These rocks are part of the Transvaal Supergroup (e.g. Tsikos et al. 2003) and at 2.2–2.4 Ma are significantly younger than the ironstones exposed in the Vredefort collar. The Hotazel BIFs have much higher $\delta^{18}\text{O}$ values (Tsikos et al. 2003) with a mean of around 12‰, which are much closer to the presumed original sedimentary values (Tsikos et al. 2003). All five ironstone analyses from Vredefort have lower $\delta^{18}\text{O}$ values, some significantly so. The Vredefort metapelites have generally lower $\delta^{18}\text{O}$ values than the Vredefort quartzites and these metapelites have significantly lower $\delta^{18}\text{O}$ values than expected for metapelites in general. Eiler (2001) gives a range of between 15 and 20‰ for shales, and in the Southern African context, the largest data set for metapelites is for the late Pre-cambrian Malmesbury (Harris et al. 1997) and Zerrisene groups (P. H. Macey 2003, unpublished data), both of which have been metamorphosed to similar greenschist facies grades to the Vredefort samples. The Malmesbury and Zerrisene group metapelites have mean $\delta^{18}\text{O}$ values of 13.3‰ ($n = 18$) and 13.6‰ ($n = 62$), respectively, which are significantly higher than the Vredefort samples.

One explanation of the generally low $\delta^{18}\text{O}$ values in the collar rocks is that Archaean sedimentary rocks have inherently low $\delta^{18}\text{O}$ values compared to Proterozoic rocks. Low $\delta^{18}\text{O}$ values would be expected in marine cherts and carbonates if seawater was significantly hotter, as suggested by Knauth and Lowe (2003). However, the pelites analysed are terrestrial rather than marine, and recent work on zircon has shown that high $\delta^{18}\text{O}$ values are found in sedimentary material due to interaction with the hydrosphere before 4 Ga (e.g. Valley et al. 2002). Furthermore, if Archaean sedimentary rocks in general had low $\delta^{18}\text{O}$, then the granitoids of the core would be expected to have even lower $\delta^{18}\text{O}$ values, as they must ultimately have been derived from a mixed sedimentary-igneous source.

The low $\delta^{18}\text{O}$ values in Vredefort metapelites and ironstones could theoretically have been produced by closed-system exchange with quartzites during metamorphism. Before the onset of metamorphism, the $\delta^{18}\text{O}$ values of the pelitic protoliths would have been higher than those of the quartzites. During prolonged metamorphism of a package of mica-rich metapelites and quartzites, the difference between a mica-rich unit and a quartz-rich unit

would depend on the O-isotope fractionation factor between quartz and mica. At 550°C, the value of $\Delta_{\text{quartz-mica}}$ is 1.5‰ (Chiba et al. 1989) and this increases with decreasing temperature. This means that metamorphism could result in a lowering in $\delta^{18}\text{O}$ value of the metapelite layer. Because the fractionation factors between quartz and most minerals are large at low temperatures (e.g. Bottinga and Javoy 1975; Zheng 1993), similar arguments could be applied to quartz and amphibole in iron formations. There are, however, several reasons why closed-system O-isotope re-equilibration is not a viable explanation for low $\delta^{18}\text{O}$ values in the metapelites and ironstones. First, closed system exchange between metapelites and quartzites would raise the $\delta^{18}\text{O}$ in the quartzites. As discussed above, this has not occurred. Second, mass balance dictates that the desired lowering of metapelite $\delta^{18}\text{O}$ values would only be achieved with a relatively high proportion of quartzite to metapelite. The ratio of shale to quartzite in the West Rand group (from which most of the collar samples come), however, is ~1:1 (Stevens and Preston 1999).

Fluid–rock interaction is a more likely mechanism for isotopic resetting of the collar rocks, as fluids typically have a great influence on stable isotope values of rocks and minerals with which they exchange (e.g. Clayton 1981). Interaction with an internally derived fluid as a means of lowering $\delta^{18}\text{O}$ is not feasible because this fluid would have a $\delta^{18}\text{O}$ value too similar to the rock itself, if the system is in isotope equilibrium. However, open system behaviour, where a rock interacts with an externally derived fluid, would cause a lowering of $\delta^{18}\text{O}$ values in the rock at high temperatures, provided the $\delta^{18}\text{O}$ value of the fluid was <5‰. This is because of the small fractionation factors between minerals and fluid (e.g. Chacko et al. 2001). The conditions and extent of fluid–rock interaction will be discussed below.

The data from other areas of the Witwatersrand basin are also plotted in Fig. 5 for comparison. Although the data set is small, there is no indication that the Vredefort collar rocks have significantly lower $\delta^{18}\text{O}$ values than those of stratigraphically equivalent samples from elsewhere in the basin. Five metapelite samples from the collar have significantly lower $\delta^{18}\text{O}$ values (<5‰), but this could simply be a reflection of the small size of the sample set available for comparison. These data indicate that low $\delta^{18}\text{O}$ values are typical of the Witwatersrand Supergroup in general.

Fluid–rock interaction in the collar rocks

There is no correlation between δD and $\delta^{18}\text{O}$ values (eFig. 2), and although the collar rocks as a whole show a large range of compositions, the range is much smaller for rocks with high water content (>2 wt.%). The temperature of metamorphism estimated by Gibson and Stevens (1998)

and Gibson and Wallmach (1995), and the calculated rock-water fractionation factors given in eTable 3, can be used to estimate the isotope composition of the water in equilibrium with these rocks. Fluid in equilibrium with the average metapelite had a δD value of around -27‰ and a $\delta^{18}O$ of about $+5\text{‰}$. This is reasonably consistent with the δD of fluid inclusions from syn-metamorphic quartz veins in the Vredefort collar (-31 and -43‰ , eTable 2). If only the rocks with $H_2O > 2$ wt.% are considered, the δD value of the fluid with which they are in equilibrium at the same temperatures is about -5‰ . Assuming that the fluid was ultimately of meteoric origin, as suggested by fluid chemistry data (Frimmel et al. 1999), the $\delta^{18}O$ of the fluid before the onset of exchange can be estimated using the equation of the meteoric water line at about -2 to -5‰ , depending on which values are taken for the δD of the alteration fluid. These estimates are consistent with palaeomagnetic studies that suggest that Vredefort was near the equator at the time of impact (Salminen et al. 2005), hence ambient meteoric water would not be expected to be particularly negative in respect to its stable isotope composition.

The Witwatersrand basin is dominated by sedimentary rocks, and as these would have had high initial water content before burial, the current low water contents of the rocks indicate the expulsion of large volumes of water during the history of the basin (Phillips et al. 1990). However, it is likely that much of this dehydration occurred at lower temperatures than the peak metamorphism, when rates of isotope exchange were slow. Although some isotope resetting of the rocks would occur as a result of fluid loss, it would not have led to a fluid/rock ratio that was high enough to account for all the lowering of $\delta^{18}O$ values in the area.

The extent of fluid/rock interaction can be calculated using simple equations based on mass balance as defined by Taylor (1977). Although the equations of Taylor (1977) have been superseded by more sophisticated models involving continuum mechanics (e.g. Baumgartner and Valley 2001), the mass balance approach still provides a semi-quantitative estimate of water/rock ratio. For the Vredefort metapelites and ironstones, the pre-metamorphic $\delta^{18}O$ values can be estimated by reference to Fig. 5. It seems unlikely that the initial $\delta^{18}O$ values of these rocks was $< 12\text{‰}$ and may have been significantly higher. The approximate rock-water fractionation factors for the ironstones and metapelites can be estimated if it is assumed that the isotope fractionation factor for each rock equals the sum of the fractionations of the individual minerals multiplied by their mole fraction (Vry et al. 2001; Zhao and Zheng 2002). Mole fractions have been estimated from the volume % determined by thin section study. As the fractionation factors used are defined for equilibrium systems,

it is also assumed that the rocks reached isotopic equilibrium, both with the fluid and internally. The temperature of isotope equilibration is assumed to be that of peak metamorphism, which has been estimated for each rock, based on petrographic observations and previous metamorphic studies (e.g. Gibson and Wallmach 1995). The estimated rock-water fractionations are given in eTable 3, along with the fractionation factors used in the calculation.

The results of the water/rock ratio calculations in atomic proportions for ironstone and metapelite are given in eTable 4. Two values for the initial $\delta^{18}O$ value were assumed, one representing the average for each rock type given by Hoefs (1997) and a lower value based on the local examples shown in Fig. 5. The calculations were also made for two different water $\delta^{18}O$ values, which bracket the range estimated from Fig. 6. The water/rock ratios in eTable 4 are generally between 0.5 and 1.0 and represent the typical water/rock ratio required for the average shifts in $\delta^{18}O$ value. These values increase by around 0.2 if a temperature of 300°C is used for calculating rock-water fractionation factors. Water/rock ratios are significantly higher if lowest observed values are taken for metapelite and ironstone rocks (1.47–3.62, depending on the starting $\delta^{18}O$ of the water, eTable 4).

Fluid–rock interaction in the core basement rocks

Before considering the stable isotope constraints on the origin of the core rocks, it is important to assess the extent of fluid–rock interaction affecting these rocks. Mineral pairs that plot outside the two isotherms in Fig. 4 have probably exchanged oxygen with an external reservoir, most likely a fluid. About half the samples have lower than predicted values of $\Delta_{\text{feldspar-quartz}}$ and plot along a line of $\Delta_{\text{feldspar-quartz}}$ of approximately zero. These samples show no particular geographical distribution. This can most easily be explained by an $\sim 1\text{‰}$ increase in feldspar $\delta^{18}O$ value caused by low-temperature retrograde alteration. The spread of whole-rock–quartz pair data is more variable than the feldspar–quartz pairs (Fig. 4). This is probably due to the additional alteration minerals present in the whole-rock material. Thus, unlike the collar rocks, any O-isotope resetting in the OGG or the ILG rocks was at low rather than high temperatures and was of minor significance.

Nature and source of the core basement rocks

The $\delta^{18}O$ values of the rocks in the core are very close to the boundary between S- and I-type granites, about 9‰ (e.g. Sharp 2007, p. 255), and both the OGG and ILG rocks are slightly peraluminous, which also suggests that they are somewhat intermediate between typical I- and S-type granitoids. Therefore, there must have been a substantial

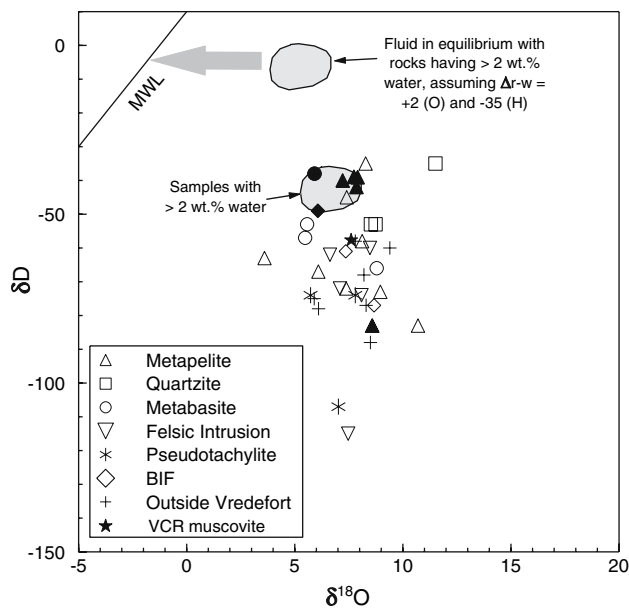


Fig. 6 Plot of whole-rock δD versus $\delta^{18}O$ for the Vredefort samples. Samples with >2 wt.% water are shown as *solid symbols*. These data plot within the shaded field, with one exception (see text). Also plotted are data from the Witwatersrand Supergroup outside Vredefort (eTable 2) and the average value for muscovite from the VCR footwall (Zhao et al. 2006). The composition of water in equilibrium with samples having >2 wt.% water is shown and the composition of possible meteoric water precursor fluid is indicated by an arrow. The meteoric water line (MWL) is from Craig (1961)

sedimentary component in the precursor for both the OGG and the ILG. This is not consistent with the suggestion of Hart et al. (2004) that the ILG represents a melt depleted, relatively mafic lower crust, which would be expected to have a lower $\delta^{18}O$ value. It is also important to note that the OGG and ILG have similar average $\delta^{18}O$ values ($\sim 9.2\text{‰}$), which suggest that they come from similar levels of the crust.

The O-isotope data is consistent with the idea that the Vredefort “discontinuity” represents a thermal boundary (Gibson and Stevens 1998; Lana et al. 2003), rather than a boundary between two significantly different parts of the crust (Hart et al. 2004). Lower crustal granulite xenoliths in general have $\delta^{18}O$ values around 7.0‰ (Kempton and Harmon 1992). Oxygen isotope data for lower crustal xenoliths from southern Africa are rare, but amphibolite, garnet granulite and eclogite xenoliths brought up by the Goedehoop kimberlite (500 km SW of Vredefort, Fig. 1), have average plagioclase $\delta^{18}O$ values of 7.9‰ ($n = 14$) and average pyroxene $\delta^{18}O$ values of 6.3‰ ($n = 4$) suggesting a bulk rock $\delta^{18}O$ value for a gabbroic assemblage (50% plagioclase, 50% pyroxene) of about 7.1‰ (C. Harris, unpublished data). These data suggest that the true lower crust is likely to have significantly lower $\delta^{18}O$ values than those obtained for the Vredefort core rocks, and

it is suggested that none of the Vredefort core is of lower crustal origin.

An extremely interesting aspect of the Vredefort core is the relationship between the ILG rocks and the xenoliths. These xenoliths are tens of metres to kilometres in diameter, and there is no clear petrogenetic relationship with the ILG. As described above, the ILG host has higher $\delta^{18}O$ values than the xenoliths. The stable isotope data therefore demonstrate that these xenoliths are not restite from formation of ILG magma, because this difference is the opposite to what would be predicted if the relationship between ILG and xenoliths is that of melt and residua.

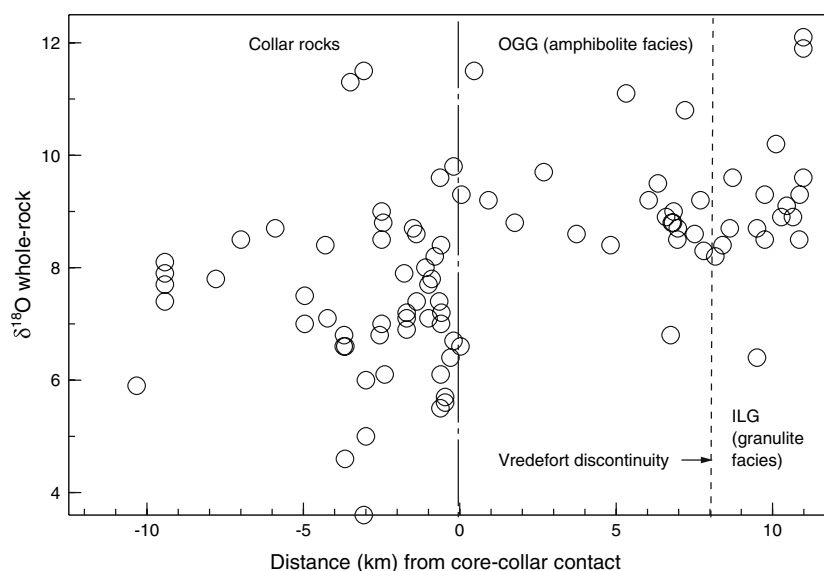
Impact generated melts

In both VC18/19 and VC23, the pseudotachylite melt has a lower $\delta^{18}O$ value than the host rock. The Vredefort granophyre also has $\delta^{18}O$ value (7.6‰) very similar to the pseudotachylites, which is lower than the average $\delta^{18}O$ value (9.6‰) for the core granofels rocks. As discussed above, a partial melt should have a higher $\delta^{18}O$ value than the bulk rock. Similar $\delta^{18}O$ relationships between pseudotachylite and host rock, in general, have been noted by O’Hara and Sharp (2001). As in the case of Vredefort, the low water content of the pseudotachylite ($<0.4\%$) suggests that the $\delta^{18}O$ value is unlikely to post-date the rapid crystallisation of the melt. These data support the suggestion of O’Hara and Sharp (2001) that pseudotachylite formation involves open system conditions, and presence of fluids, during melting. In VC43/44, the $\delta^{18}O$ of the host rock is the same, within error, as that of the pseudotachylite, whereas the δD of the host rock is significantly less negative than that of the pseudotachylite, likely due to hydration and alteration of the host. These samples have higher water contents (1.36 and 1.02 wt.%) than the pseudotachylite/host-rock pairs of felsic rocks from the core, which is another indication of more significant fluid–rock interaction in the collar than in the core (Fig. 6).

O-isotope profile through the crust at Vredefort

The total thickness of the crust examined is over 20 km with the centre of the core representing mid-crustal conditions at *ca.* 3.1 Ga (e.g. Stevens et al. 1997; Hart et al. 1999) and most likely similar depths prior to impact. Figure 7 shows a plot of whole-rock $\delta^{18}O$ versus distance from the core-collar contact. Although there is no strong correlation between distance and $\delta^{18}O$ value ($r = 0.24$), the $\delta^{18}O$ value generally increases with increasing depth. The collar rocks on average have a $\delta^{18}O$ value (7.5‰) that is significantly lower than that of the core rocks (ILG = $9.02 \pm 0.76\text{‰}$, OGG = $9.32 \pm 0.90\text{‰}$). This is the opposite of what would be intuitively expected, based on

Fig. 7 Plot of whole-rock $\delta^{18}\text{O}$ value versus distance towards the centre of the Vredefort structure measured from collar/core contact. Rocks from the collar therefore plot at negative distance. The position of the so-called Vredefort discontinuity is from Hart et al. (1990)



the typical variation of $\delta^{18}\text{O}$ in igneous, metamorphic and sedimentary rocks (e.g. Eiler 2001, Fig. 7), and very different from typical $\delta^{18}\text{O}$ profiles through the oceanic crust (e.g. Gregory and Taylor 1981). Wickham et al. (1991) pointed out that a lowering of $\delta^{18}\text{O}$ values characterises metamorphic rocks at higher grades; but what is seen at Vredefort is the opposite, in that lower $\delta^{18}\text{O}$ values are observed in rocks that represent both lower metamorphic grades and higher crustal levels. Several studies (e.g. Grunder and Wickham 1991; Wickham et al. 1993) have shown that stable isotope discontinuities exist within the crust with abrupt increases in $\delta^{18}\text{O}$ values with depth, but these isotopic shifts are on a much smaller scale than that observed in the present study. Thus, a major conclusion of this work is that the $\delta^{18}\text{O}$ profile through the crust at Vredefort suggests that $\delta^{18}\text{O}$ values increase with increasing depth. This could be an inherent feature of the crust, reflecting primary $\delta^{18}\text{O}$ differences between the fundamentally different sediments of the cover sequence, the amphibolite facies granitoids of the OGG and the more lithologically complex rocks of the granulite facies ILG domain. Alternatively, it may have resulted from essentially closed-system metamorphism, or it could represent the result of extensive fluid–rock interaction in the collar, with the possibility that the latter alternative was facilitated by impact induced deformation and secondary permeability, creating favourable conditions for post-impact hydrothermal circulation, as seen in other impact structures (e.g. Ames et al. 1998; Zurcher et al. 2005).

Relationship between fluid–rock interaction and impact

According to the data presented in this paper, the relatively high degree of fluid–rock interaction affected the Vredefort

collar metapelites, and the low $\delta^{18}\text{O}$ appear to be a general characteristic of the Witwatersrand basin. This suggests either that the fluid–rock interaction is a general characteristic of the Witwatersrand Supergroup, which existed prior to the meteorite impact, or that the fluid–rock interaction is related to the impact and it established a hydrothermal system on a scale much larger than the immediate collar. In the absence of a systematic basin-wide stable isotope study, it is not possible to decide between these hypotheses. However, given that the age of peak of metamorphism, fluid-flow and alteration are similar to that of the impact event (e.g. Phillips et al. 1997; Stevens et al. 1997; Frimmel et al. 1999), it is possible that these events are related.

The idea that there is a relationship between the Vredefort impact event and fluid–rock interaction is not new. Frimmel et al. (1999) proposed extensive fluid flow in the Witwatersrand basin rocks due to hydrothermal infiltration, aided by the creation of impact induced secondary permeability in the form of a dense network of microfractures developed in the already metamorphosed Witwatersrand rock sequence. These authors further argued that after progressive burial metamorphism reached lower greenschist facies conditions, the primary permeability and any connate water would have disappeared. They point out that a tectonic event, like the Vredefort impact, would have been required to increase the secondary permeability, and that the upturning and fracturing of the rocks around the impact site, together with a significant disturbance of the geotherm, would have permitted the surface waters to reach depths as great as 10–15 km.

It can therefore be speculated that the fracturing and increased temperatures induced by the impact event led to an enhanced hydrothermal circulation in the rocks affected

by the impact, and that this would explain the low $\delta^{18}\text{O}$ values in the collar rocks. The core-rocks were not fractured nearly as extensively as the collar, and consequently their permeability remained low, and their $\delta^{18}\text{O}$ values were not significantly affected by post-impact fluid flow. It is generally accepted that the gold mineralisation in the Witwatersrand basin occurred by remobilisation of placer deposits (e.g. Frimmel 2005 and references therein). The hydrothermal circulation responsible for gold remobilisation may have been the same as that which lowered the $\delta^{18}\text{O}$ values in the collar rocks. As previously speculated by McCarthy et al. (1990), it may not be a coincidence that the world's greatest gold province is hosted within the rim synclinorium of the largest impact structure on earth.

The establishment of a hydraulic head in the elevated and upturned rocks of the Vredefort structure, along with the impact-induced fracturing, are specific to Witwatersrand basin rocks adjacent to the impact structure. Similar basins present elsewhere at the same time would not be expected to share this rather unique feature. Basins in southern Africa of equivalent age to the Witwatersrand Supergroup, such as the Pongola Supergroup in KwaZulu-Natal, represent a possible test of the relationship suggested above, because normal rather than low $\delta^{18}\text{O}$ values would be expected.

Conclusions

1. The collar rocks have generally low $\delta^{18}\text{O}$ values, and the metapelitic rocks and the ironstones have much lower than expected $\delta^{18}\text{O}$ values. A small amount of data from elsewhere in the Witwatersrand basin suggests that this feature is common to the Witwatersrand Supergroup.
2. The granitic gneisses of the core of the Vredefort structure have an average $\delta^{18}\text{O}$ value of 9.2‰. This is close to the boundary between S- and I-type granites and suggests an ultimate origin from mixed igneous-sedimentary material, which became largely homogenised. The $\delta^{18}\text{O}$ values of co-existing minerals in a subset of rocks show no evidence for significant fluid-rock interaction.
3. The $\delta^{18}\text{O}$ values of the core rocks are significantly higher than granulite and amphibolite xenoliths brought up by kimberlite pipes, whose origin can be shown from geothermometry and geobarometry as being from the lower crust. This suggests that the Vredefort granite gneisses represent mid-crustal rocks, which is in accordance with published work on the metamorphism (Lana et al. 2003; Gibson and Stevens 1998). There is no difference in $\delta^{18}\text{O}$ between the OGG and ILG rocks of the core, across the “Vredefort

Discontinuity” of Hart et al. (1991). This confirms the suggestion of Lana et al. (2004) that the “discontinuity” represents the transition from granulite to amphibolite and not juxtaposition of blocks from different crustal levels.

4. Granulite xenoliths from the ILG have higher $\delta^{18}\text{O}$ values than their host rocks. This suggests that the xenoliths do not represent entrained restite material from the partial melting that produced either the ILG or OGG granite gneisses.
5. Comparison between the $\delta^{18}\text{O}$ of pseudotachylites and their host rocks show that the pseudotachylites also have lower $\delta^{18}\text{O}$ than their hosts (and presumed restite). This may have been caused by the presence of water during pseudotachylite formation, as the pseudotachylites are unlikely to be influenced by post-impact fluid flow. Similar differences in $\delta^{18}\text{O}$ values are present between the Vredefort granophyre and the OGG/ILG rocks.
6. The $\delta^{18}\text{O}$ profile through the 15 km of crust exposed “edge-on” at Vredefort shows no strong correlation between the $\delta^{18}\text{O}$ value and depth, but in general is the opposite to that expected in that the $\delta^{18}\text{O}$ values are higher in the rocks that originated from greater depths. This is a function of the effects of fluid-rock interaction at greenschist-amphibolite facies conditions in the sedimentary sequence. This is also consistent with the generation of a post-impact hydrothermal system, as also seen in the Sudbury and Chicxulub impact structures (Ames et al. 1998; Zurcher et al. 2005).

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