Originally published as:

Backeberg, N. R., Reid, D. L., Trumbull, R. B., Romer, R. L. (2011): Petrogenesis of the False Bay dyke swarm, Cape Peninsula: Evidence for Basement Assimilation. - South African Journal of Geology, 114, 3-4, 335-352

DOI: [10.2113/gssajg.114.3-4.335](https://doi.org/10.2113/gssajg.114.3-4.335)

Petrogenesis of the False Bay Dyke Swarm, Cape Peninsula: Evidence for Basement Assimilation

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Abstract

The False Bay Dyke Swarm represents igneous activity associated with the opening of the South Atlantic during the early Cretaceous, and can be considered a distal component of the Etendeka – Paraná Large Igneous Province. In contrast to the intense magmatism in Namibia and Brazil, with the Paraná-Etendeka continental flood basalts and huge dyke swarms, the False Bay swarm is considered to be the product of a relatively low magma-flux in a passive rift setting. Previous work suggested that the False Bay dykes consist of a single magma type, with a marked differentiation from olivine-tholeiite basalt to ferro-tholeiitic andesite, accompanied by crustal contamination. This study uses new trace element and radiogenic Sr and Nd isotope analyses to better constrain the processes of magma evolution in the dykes. Combined assimilation – crystal fractionation (AFC) modelling suggests a first stage of nearly closed system fractionation of a gabbroic assemblage (olivine + clinopyroxene + plagioclase + magnetite), accompanied by progressively larger amounts

of crustal assimilation at intermediate and late stage. The AFC models show that the exposed country rocks, including Cape Granites and Malmesbury Group metasediments, are compositionally unfavorable for producing the observed assimilation trends in the dykes. Instead, a more suitable crustal assimilant would be Mesoproterozoic granitic gneisses similar to those exposed in the neighbouring Namaqua Province, which may underly the Cape Peninsula.

Key words: dolerite dykes, South Atlantic rifting, differentiation, AFC, Sr and Nd isotopes, Cape Granite, Malmesbury Group

Introduction

The South Atlantic margin is the product of Cretaceous rifting during Gondwana breakup, which resulted in extensive igneous activity including the Paraná-Etendeka large igneous province and a network of dolerite dyke swarms. The most extensive of these on the African margin is the Henties Bay-Outjo Dyke swarm (HOD) in the Etendeka province of NW Namibia (Trumbull et al., 2004; 2007). The False Bay dolerite dyke swarm (FBDS) described in this paper is the southern-most extension of the coastal dykes and it crops out as a narrow ~70 km wide band along the south-west corner of South Africa around False Bay and the Cape Peninsula (Fig. 1). The first records of the dark intrusive rocks of the Cape Peninsula are summarised by Nell and Brink (1944), Walker (1956), Day (1987) and Reid (1990) and in the literature the dyke swarm has been referred to as the Cape Peninsula dolerites (Walker, 1956; Reid, 1990) and the False Bay dolerites (Day, 1987; Reid et al., 1991). Based on magnetic anomalies, Day (1987) suggested the presence of numerous dykes in the region, but only 24 are actually exposed around the Cape Peninsula and on Robben Island (Fig. 1). Exposures are usually confined to shoreline locations, where dykes are seen to intrude all of the pre-Cretaceous rocks found on the Peninsula (Fig. 1). The dykes are steeply-dipping to subvertical, with local variations between 75° - 89° dip (Backeberg, 2011). The regional NW-SE strike of approximately 300° based on magnetic anomalies from Day (1987) agrees well with field measurements (Backeberg, 2011). The regional trend approximates that of the continental margin offshore between Cape Columbine and Cape Agulhas (Day, 1987).

The most reliable published ages for the False Bay dyke swarm include a mean emplacement age of 132 ± 6 Ma with a 15 Ma range (126 - 141 Ma) from whole rock K-Ar dating (Reid et al., 1991), and an ^{40}Ar - ^{39}Ar date of 131.3 ± 1.3 Ma measured on plagioclase (Stewart et al., 1996). It is likely that the 15 Ma range of K-Ar ages exaggerates the true duration of igneous activity, but this needs confirmation by further work. Similarly, K-Ar ages for dykes in the HOD dyke swarm of Namibia span a wide range of ages between 116 - 143 Ma (Trumbull et al., 2007 and references therein), whereas $^{40}\text{Ar}/^{39}\text{Ar}$ dates from Etendeka basalts and intrusive complexes cluster between 128 and

134 Ma (Renne et al., 1996; Wigand et al., 2004). In any case, the Early Cretaceous age of False Bay dykes links their emplacement with the tectonic evolution of the South Atlantic Ocean during Gondwana rifting, which at around 130 Ma had experienced initial continental stretching (Nürnberg and Müller, 1991).

The mineralogy of the dolerites is dominated by plagioclase and clinopyroxene, with altered olivine pseudomorphs and skeletal grains of Fe-Ti oxides (Walker, 1956; Reid, 1990). Dykes are typically holocrystalline, aphyric and show subophitic relations between plagioclase and clinopyroxene. Many samples show a high density of quartz and calcite amygdales, suggesting a shallow depth of intrusion (Walker, 1956; Reid, 1990).

Previous geochemical work by Reid (1990) and Trumbull et al. (2007) documented a wide range of dyke compositions from the False Bay swarm. The geochemical and Sr-Nd-Pb isotope study of Trumbull et al. (2007) focused on the least contaminated dyke compositions as a basis for comparing mantle source characteristics along the volcanic rifted African margin. They noted contrasts in basaltic magma types between north and south and related them to different tectonic settings and magma volumes along the margin, whereby the northern part is classified as active rifting driven by the Tristan Plume, whereas the southern dykes are characterised by passive rifting. The earlier work mentioned the wide range in initial Sr isotopic values in the False Bay dykes as evidence for crustal contamination, but did not take the subject further.

This paper focuses on the compositional diversity and isotopic range shown by the False Bay dyke swarm and on possible processes of differentiation and contamination during magma evolution. Compositional variations due to different mantle sources can be confidently excluded based on earlier work (Reid, 1990; Trumbull et al., 2007) which demonstrated the monogenetic nature of the swarm. We build on existing whole-rock geochemical analyses from the dykes, mostly based on X-ray fluorescence data, and add new analyses of trace elements and Sr-Nd radiogenic isotopes of key samples chosen to cover the full compositional range. Furthermore, in order to assess the geochemical effects of crustal assimilation on the False Bay magma, we obtained new analyses from potential crustal components cropping out in the region (Cape Granite Suite and Malmesbury Group metasediments).

Geochemistry

Analytical Data and Methods

Most of the data presented here were obtained on dolerite samples originally collected by Reid (1990) and analysed by X-ray fluorescence (XRF) for major and trace elements. Samples from Robben Island were obtained during this study, and all samples were analysed for additional trace elements by quadrupole ICP-MS and for radiogenic Sr and Nd isotopes by multi-collector ICP-MS.

Country rock samples from the Malmesbury Group meta-sediments and Cape Peninsula Granite collected during the unpublished studies of Mbangula (2004), Mbeje (2005) and Creighton-Jones (1982), respectively, were analysed for radiogenic isotopes. Table 2 reports the new trace element and isotopic results together with the corresponding major oxide and XRF trace element data from Reid (1990), Mbangula (2004), Mbeje (2005) and Creighton-Jones (1982).

The new analytical data were obtained at the University of Cape Town using a newly-installed Thermo Fisher Xseries2 quadrupole ICP-MS, following procedures for sample preparation described by Le Roex et al. (2001). The excellent agreement between the ICP-MS results (present study) and existing XRF data from the same samples (Reid, 1990) gives confidence in our dataset. For 13 trace elements (Sc - Ba in Table 2) determined on 26 samples, this comparison yielded an average correlation coefficient of 0.97. The only obvious discrepancy was the Zr concentrations measured in the Cape Granites, where the ICP-MS results were consistently lower than XRF, likely due to incomplete dissolution of zircon. Other elements such as Hf and perhaps the HREE, which are usually concentrated in zircon, may also be underestimated in the granite samples. The problem of Zr discrepancy between XRF and ICP-MS is not prevalent in the dolerite samples.

A subset of samples covering the compositional range was selected for Nd- and Sr-isotope analysis using the Nuplasma NP51 MC-ICP-MS at the University of Cape Town. Sample preparation followed Miková and Denková (2007) and long term analysis of reference material BHVO-2 (Weis et al., 2006), produced $^{87}\text{Sr}/^{86}\text{Sr}$ values of 0.703468 ± 0.000013 and $^{143}\text{Nd}/^{144}\text{Nd}$ values of 0.512994 ± 0.000005 . Replicate analysis of some samples of Cape Granite and Malmesbury Group rocks were carried out on a Triton MS (for Sr isotopes) and a Finnigan MAT262 MS (for Nd isotopes), at the GFZ German Research Centre for Geosciences in Potsdam, following techniques described by Romer et al. (2005).

Summary of geochemical variations

Previous studies by Reid (1990) and Trumbull et al. (2007), to which the reader is referred for details, noted the tholeiitic affinity of the dolerite dykes from the False Bay swarm and their marked compositional range, from basalt to andesite (Fig. 2). The range of composition in samples used in this study is from 50 - 57 wt.% SiO_2 and 2.7 - 7.5 wt.% MgO . These components were utilised by Reid (1990) and Trumbull et al. (2007) as differentiation indices to demonstrate the systematic and coherent behaviour of several incompatible and fluid-immobile trace elements (e.g. Zr, Nb and Y), and compatible trace elements such as Ni, Cr, V and Sr. Interelement variations of selected incompatible and low-level trace elements from our ICP-MS study (Ta, Hf, U, Th, Pb and Ce) are illustrated in Figure 3. To show dependencies of trace element contents with differentiation, the samples are coded by different symbols to distinguish a high-MgO range (>5 wt.%) from a low-

MgO range. Also shown for comparison is the compositional field of dolerite dykes from the Henties Bay-Outjo (HOD) swarm in NW Namibia (Trumbull et al., 2007). The more extreme differentiation of the False Bay dyke swarm compared to the HOD dykes is apparent in this comparison.

Trace element abundances and the degree of compositional scatter differ with the degree of differentiation expressed by MgO content. For example, the high-MgO samples show rather tightly conserved inter-element ratios (e.g., Ta vs. Hf in Fig. 3a) whereas the low- MgO group has both greater abundances of incompatible elements and greater scatter. An exception is noted for Th and U (Fig. 3b), where the correlation of the two elements is excellent in all samples and the scatter in U/Th ratios is in fact slightly larger in the high-MgO group. This is shown in more detail in the plot of Th/U ratio vs. Pb (Fig. 3c), which also shows higher abundance and greater range in Pb contents in the low-MgO group compared to the high-MgO group. The elements Ce and Pb have similar bulk D-values in basaltic magmas (Hoffman, 1989), and the False Bay dykes show the expected good correlation between the two elements (Fig. 3d), but the Ce/Pb ratios in the low-MgO group are systematically lower than in the high-MgO group, with average values around 8 vs. 5, respectively. A closed-system differentiation process would not cause a change in Ce/Pb ratio, so there must be another factor involved. Hofmann (1989) pointed out that mantle and continental crust have average Ce/Pb ratios of about 10 and 4, respectively. Thus, it is reasonable to relate the shift to lower Ce/Pb in the low-MgO samples, and their higher Pb abundances, to effects of crustal assimilation (see below).

Chondrite-normalised REE patterns (Fig. 4), show a trend of increasing REE abundance and stronger development of a negative Eu anomaly ($\text{Eu}/\text{Eu}^* = 0.98$ to 0.68) with decreasing MgO. Rather than plot all samples on this diagram, we display for clarity only the least and the most-differentiated extremes (LB1 and CP2, respectively) and one intermediate representative, sample SM. The compositions of Namibian dykes (HOD, gray field) overlap with the least-evolved False Bay samples. Also shown for comparison are REE compositions of Karoo (ca. 180 Ma) dolerite sills in the nearby Calvinia district (le Roex and Reid, 1978), which have similarly low REE abundances and flat slope with no Eu anomaly as the high-MgO False Bay sample LB1. More subtle features of the REE compositions are shown in plots of element ratios vs. MgO in Figure 4c-e. An important feature in these plots is the change in slope of compositional trends, which is highlighted by the gray pattern. The high-MgO samples have lower normalised La/Lu and La/Sm ratios, than the others, and these ratios increase with differentiation until about 5 wt.% MgO, beyond which the ratios are unchanged. The heavy REE ratio (Gd/Lu) does not show much variation with differentiation at all (chondrite normalised $\text{Gd}/\text{Lu} = 1.2 - 1.6$).

It is worth noting that most samples from the False Bay swarm, representing many localities

across the peninsula (Fig. 1), appear to form a coherent magma series with only a few exceptions identified by letter codes on Figure 3 and 4. These samples represent particular localities in the region as explained in the figure captions. The Robben Island samples (RI) are particularly distinct, especially in their REE features (Fig. 4). There are no petrographic features of the outlier samples or their host rocks that set them apart from other dyke localities.

Radiogenic Isotopes

Neodymium and strontium initial isotope ratios (Table 2) were age-corrected assuming emplacement at 130 Ma, and the initial ratios are plotted in Figure 5 together with fields of the HOD dykes from NW Namibia, as well as relevant crustal components (see discussion). Initial Sr ratios of the False Bay dykes vary from 0.7063 to 0.7203, the corresponding values of ϵNd are -1.6 to -6.6, and there is an excellent negative correlation between the two. The high-MgO dykes plot in a rather tight cluster around $^{87}\text{Sr}/^{86}\text{Sr} = 0.706$ and $\epsilon\text{Nd} = -2$, whereas samples with < 5 wt.% MgO account for most of the isotopic variation and show a strong correlation of initial ratio and MgO content (inset Fig. 5). The Sr-Nd isotope trend for the False Bay dykes is shallower than that shown by the HOD dykes from Namibia, i.e., the False Bay samples reach higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios at a given range of ϵNd . Another important difference between the two dyke swarms is that some HOD dykes overlap with the MORB field whereas all of False Bay samples have negative values of ϵNd .

Variations in initial Sr - Nd isotope ratios are compared with differentiation indices (Rb/Sr and MgO) in Figure 6. The upper panel shows that the trend of $^{87}\text{Sr}/^{86}\text{Sr}$ vs. Rb/Sr for all dyke samples "all FBDS trend" has a similar slope to that of basement rocks from the Cape Peninsula (Malmesbury Group and Cape Granite). If one interprets the all-FBDS line as a pseudo-isochron, the apparent age is 518 ± 150 Ma, similar to the range reported by Allsopp and Kolbe (1965) for the basement. The trend line defined by the MgO-rich dykes only is much shallower and it does not project towards any of the crustal components shown. The relationships between initial Sr and Nd ratios and differentiation are perhaps best seen by their relation with MgO (Figs. 6b and c), where a significant break in slope of the data array is apparent at about 5 wt.% MgO.

Discussion of magma evolution

Crystal Fractionation

Evidence was cited in earlier studies of the False Bay dyke swarm for fractional crystallisation as a dominant process of magma differentiation (Reid, 1990; Trumbull et al., 2007). This evidence includes systematic trends in major element variations consistent with fractionation of a gabbroic assemblage (olivine-clinopyroxene-plagioclase). Also, there is an approximately three-fold increase

in the abundance of incompatible trace elements Zr, Nb and Y, consistent with 60-70% fractionation, whereas interelement ratios remain constant (Reid, 1990). The conservation of trace element ratios also holds true for many of the incompatible elements determined in this study (e.g. U, Th, Ta, Hf, Ce; see Fig. 3). Fractional crystallisation of plagioclase is suggested by the progressive development of a negative Eu anomaly with increasing differentiation (Fig. 4). Finally, the isotope ratios $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ are nearly constant for samples with 7.4 to about 5 wt.% MgO, which is consistent with closed-system fractional crystallisation during early differentiation. In the more evolved samples, isotope ratios vary substantially and systematically with MgO, and this isotopic shift is one of the key arguments for the increasing role of crustal assimilation in the later stages of differentiation (see below).

The chilled margins of several dykes contain microphenocrysts of plagioclase and rare clinopyroxene, as well as altered olivine. Crystallisation models using the MELTS program (Ghiorso and Sack, 1995) applied to a dyke composition with the highest MgO content (LB1 with 7.4 wt.% MgO) reproduce the observed microphenocryst phases as liquidus minerals. Olivine is predicted as a liquidus mineral for bulk compositions with greater than about 4 wt.% MgO. The modelled range in olivine compositions, Fo₇₀ for LB1 to about Fo₄₀₋₅₀ for samples with around 6 wt.% MgO (e.g. MP), are in reasonable agreement with the observed olivine range of Fo₅₆₋₇₂ for the more primitive dykes (Reid, 1990). No olivine grains were found in the most differentiated dykes, which is consistent with the prediction of the MELTS crystallisation model. Least-squares modelling (Bryan et al., 1969) was used to simulate the effect of fractional crystallisation on the major element composition. Crystallisation was modelled for separate intervals of MgO content to allow for changes in the crystallising assemblage. The results are compared against specific samples for each range, and agreement between computed and observed compositions is assessed by the sum of the squares of residuals. The values of less than 0.36 for the high-MgO group indicate a satisfactory match (Table 3). The fit is poorer for the more evolved compositions (< 5 wt.% MgO), which is due at least in part to the increasing influence of assimilation as described in the next section. The predicted modes of the fractionating assemblages remain gabbroic throughout, with olivine decreasing in importance from early to late crystallisation, and with Fe-Ti oxides joining the assemblage in the late stages, as predicted by the MELTS model. An independent test of fractional crystallization (Fig. 7) compares observed Rb, Sr and Zr variations with computed fractionation trends for different plagioclase:clinopyroxene ratios based on partition coefficients from McKenzie and O'Nions (1991) and McCallum and Charette (1978). The best agreement to the observed differentiation trends for high-MgO samples (dark symbols) is with a ratio of approximately 60% plagioclase and 40% clinopyroxene, which is in reasonable agreement with the plagioclase and clinopyroxene ratios predicted from least-squares modelling (Table 3).

The geochemical trends and modelling therefore suggest that initial differentiation of the False Bay dolerite magma, from about 8 to 5 wt.% MgO, was governed predominantly by fractional crystallisation. This implies that the least-differentiated dykes retain compositional features of their mantle source. The ratios of Th/U for the dykes have a mean value of ~ 3.8 , which corresponds well to mantle values (O’Nions and McKenzie, 1993). Figure 8 shows variations in element ratios of Nb/Yb vs. Th/Yb of the False Bay dykes compared with the enriched and depleted mantle (the mantle array) and with crustal components in the selected local crustal basement. The high-MgO dykes plot within the mantle array and have apparently retained a mantle signature. The variations of Nb/Yb and Th/Yb within these samples may reflect different contributions of lithospheric vs. asthenospheric mantle in the primary magmas as suggested by Trumbull et al. (2007). In contrast, the low-MgO samples plot outside the mantle array in the direction toward the composition of local continental crust.

Assimilation and Fractional Crystallisation

The best evidence for crustal assimilation during magma evolution in the False Bay dyke swarm is the correlation between Sr and Nd isotope ratios and MgO for samples with less than about 5 wt.% MgO (Fig. 6). Also important is the fact that the high-MgO dykes plot within the mantle array in terms of Nb-Yb-Th relationships (Fig. 8) whereas the low-MgO samples plot between the mantle array and continental crust. The ca. 5 wt.% MgO level therefore marks a consistent change in the trends of many geochemical components, which indicate the increasing importance of open-system behavior in the later stages of differentiation. The following discussion addresses the amount of assimilation and the nature of the crustal component involved based on combined assimilation and fractional crystallization (AFC) modelling.

The two most obvious candidates for crustal assimilation are the exposed host rocks which are cut by the False Bay dykes, namely, the Cape Granite Suite and the Malmesbury Group metasediments (Tygerberg Terrane). A third possible component is the Mesoproterozoic basement (ortho- and paragneisses) which is exposed to the north of the Cape Peninsula. To test the effects of assimilation and fractional crystallization (AFC) in a quantitative way, we used the compositions of samples CJ-12 (Cape Granite) and MM-001 (Malmesbury Group) from this study (Table 2), and data for the Mesoproterozoic Bushmanland gneiss (BML) from Reid et al. (1997). Modelling employed the equations of DePaolo (1981) and bulk distribution coefficients (D-values) given in the caption of Figure 9. The AFC curves were calculated using an intermediate sample OD2 (with 5.2 wt.% MgO) as a starting point, since it was shown above that the high-MgO samples are unaffected by assimilation (Fig. 6). Apart from the mineral-melt D-values, which are reasonably well known, the critical variable in AFC models is the mass ratio of assimilation to crystallization

(r). The model curves are labelled according to the r-value (0.1, 0.2 or 0.4) and the crustal component. All curves extend from 0 to 70% crystallisation ($F=1$ to 0.3), with ticks at 10% intervals.

Model results for the Rb-Sr system (Fig. 9a), show little distinction between the trends for Malmesbury (MM) or Cape Granite (CG) assimilants. Assimilation of granite yields higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, but even for a high proportion of assimilation to crystallisation ($r=0.4$), the AFC models fail to explain the high radiogenic Sr isotope compositions of low-MgO False Bay dykes. Binary mixing curves (dashed gray lines) pass through the low-MgO data array, but mixing of hot basaltic magma with crustal rocks without crystallization of the basalt is unrealistic. A reasonable alternative, in terms of both composition and energy considerations, is assimilation of Bushmanland gneiss (BML), which matches the low-MgO data reasonably well for low r-values of 0.1 to 0.2. The corresponding plot for the Sm-Nd system (Fig. 9b) shows a horizontal trend of the high-MgO dykes (dark squares), underscoring the point that the least-evolved magmas were little affected by crustal assimilation (compare Fig. 6c). The low-MgO samples have lower $^{143}\text{Nd}/^{144}\text{Nd}$ values but variable Sm/Nd ratios. As in Figure 9a, AFC curves are shown for the three crustal components starting at intermediate sample OD2 but in this case it is clear that no AFC model starting from a single point can explain all of the low-MgO data. Possibly the Sm/Nd ratio of the crustal assimilant(s) is variable and not represented by three compositions used in modelling (see below). It is also likely that sample OD-2 is not a parental magma for all of the low-MgO samples. Still, the curves illustrate the point that high amounts of assimilation of Malmesbury metasediments or Cape Granite ($r=0.4$), will produce less isotopic shift than moderate assimilation of Bushmanland gneiss ($r=0.2$). Finally, we point out that only the AFC model for Bushmanland gneiss with $r=0.2$ yields a reasonable fit to both the Sm-Nd and Rb-Sr data (Fig. 9a).

The AFC modeling is oversimplified because it treats crustal assimilation as a bulk mixing process. A more realistic scenario would consider assimilation of partial melts of the crust rather than the bulk rock, meaning that components entering the melt (feldspars, hydrous phases) have more influence than those left in the solid residue. Depending on the details of this process, the composition of assimilant and bulk rock can be different, and the effects of AFC on different isotopic systems may also vary due to the particular mineral hosts for the parent-daughter elements (Nd-Sm, Rb-Sr) and their behaviour during melting (e.g., Zeng et al., 2005; Vásquez et al., 2009). Thus, it can be expected that partial melts will be richer in radiogenic Sr than the bulk rock because feldspars and biotite are major hosts for Rb and they participate early-on in melting reactions. Similarly, if partial melting leaves zircon or other phases rich in HREE in the solid residue (garnet, xenotime), the melts will have lower Sm/Nd ratios and a less radiogenic Nd isotope composition than the bulk rock. These arguments suggest that crustal assimilation can be a complex process which is only partly simulated by standard AFC models like those shown on Figure 9, however

realistic models that take partial melting into account will require more information than is currently available. The results shown in Figure 9 must be interpreted with caution, but we consider it significant that the Mesoproterozoic Bushmanland Gneiss appears to provide a better fit to the data than the exposed basement rocks. Outcrops of Bushmanland Gneiss and related Mesoproterozoic basement units are found in the Namaqua-Natal mobile belt north of the Cape Peninsula, but similar units are thought to underlie the Malmesbury Group metasediments. Supporting geophysical evidence for Namaqua-Natal basement below the Cape Fold Belt is described by Lindeque et al. (2011).

Tectonic setting, intrusion dynamics and differentiation

Two deep-seismic profiles from the region provide insights on the structures and composition of the lower and middle crust that are relevant to this discussion. One profile oriented ENE-WSW crosses the continental margin near Springbok, about 700 km north of Cape Town (Hirsch et al., 2009), and the other is a north-south profile across the southern margin about 400 km east of Cape Town (Stankiewicz et al., 2007; Lindeque et al., 2011). Both profiles indicate a high velocity layer at the base of the 35-45 km thick crust, which was suggested by Trumbull et al. (2007) and Stankiewicz et al. (2007) to represent underplating of breakup-related gabbroic intrusions. In the context of the False Bay dolerites, intrusion and collection of primary mantle magmas in the lower crust could be a favorable mechanism for assimilation of Mesoproterozoic crust similar to that of the Bushmanland Gneisses. The temperature of the False Bay magmas is suggested to be about 1300°C (Trumbull et al., 2007), and intrusion of an equivalent magma at the base of the crust should provide ample heat for partial melting of granitic material at these depths according to experimental evidence (Boettcher and Wyllie, 1968; Huang and Wyllie, 1973). The reason why assimilation of crust becomes effective only after an initial stage of fractional crystallisation can only be speculated upon: (1) more extensive attenuation of the crust was needed to bring basaltic magma into broader contact with the crustal assimilant, (2) early contact with the crust involved refractory, granulitic or mafic material unfavorable for melting and only the more evolved magma came into contact with fertile crustal material similar the Bushmanland Gneisses, or (3) sufficient heating of the crust to begin significant melting and assimilation was only achieved after an initial stage of crystallisation.

Conclusions

The geochemical and Sr-Nd isotopic data and interpretative models presented in this study suggest that the False Bay dolerite magma evolved in two stages. The first stage, from an MgO content of about 8 to 5 wt.%, was dominated by fractional crystallisation of mainly plagioclase and

clinopyroxene in proportion of around 6:4 respectively, with minor contribution of olivine. The fractionation of plagioclase from basaltic melts implies relatively low-pressure conditions of less than approximately 13 kbar (ca. 40km depth). The False Bay dykes that cut the Cape Granite and Malmesbury Group rocks are vesiculated, indicating final emplacement at a shallow depth (Walker, 1956; Reid, 1990). The second stage of differentiation, leading to the most-evolved compositions (andesites) with about 2 wt.% MgO, was more complex and involved assimilation of crustal material in addition to fractional crystallisation. The exposed country rocks of Cape Granite and Malmesbury Group metasediments which are cut by the dykes, were tested as possible components involved in the crustal assimilation, but examination of the Sr and Nd isotope trends and simple AFC models suggests a more likely assimilant would be Mesoproterozoic Bushmanland Gneisses present at deeper levels in the crust. Bushmanland gneisses crop out in the Namaqua-Natal belt north of the study area, but equivalent rocks are presumed to underlie the study area in the middle and/or lower crust. It is suggested that parental basaltic magmas formed in the mantle ponded in the mid to lower crust, where fractional crystallisation and assimilation took place. Variably-evolved magma batches then ascended rapidly through extensional faults and fractures in the overlying Cape Granite and Malmesbury Formation giving rise to the variable compositions of dykes observed in the False Bay swarm.

Acknowledgements

This research is a product of the Inkaba yeAfrica program in cooperation between the University of Cape Town and the German Research Centre for Geosciences (GFZ) in Potsdam, and was funded by this program. We thank Fayrooza Rawoot, Christel Tinguely, Shireen Govender and Petrus le Roux for processing samples in the analytical laboratories at the Department of Geological Sciences at University of Cape Town. The quality and content of the manuscript was greatly improved by the journal review by Chris Harris, for which we are grateful. This is Inkaba yeAfrica publication number XX.

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Figure Captions

Figure 1: (Left) Magnetic lineament map of the False Bay dyke swarm from Day (1987). Solid and dashed lines denote positive and negative anomalies, resp. Numbers 1-24 show the sampled dyke localities listed in Table 1. The inset map illustrates basement provinces in southern Africa, after Moore et al. (1990) and Eglington (2006). (Right) schematic block diagram of dyke intrusive relations incorporating observations by Walker (1956) and Reid (1990).

Figure 2: Major element discrimination of the False Bay dyke samples studied here, on the total alkali-silica diagram of Le Bas et al. (1986).

Figure 3: Variation diagrams for selected trace elements in the False Bay dolerites analysed in this study. Symbols are shaded to represent high (> 5 wt.%) and low (< 5 wt.%) MgO contents, see text. The shaded field HOD compares data from the Henties Bay - Outjo dykes in NW Namibia (data from Trumbull et al., 2007). Analytical errors of ICP-MS are about 2 % relative or less for all elements shown, and smaller than the symbol size. Letter symbols denote outlier samples from Robben Island (RI), Froggy Pond (FG) and Millers Point (MP).

Figure 4: Chondrite-normalised REE composition of the False Bay dolerites (chondrite values from McDonough and Sun, 1995). (a) REE profiles of samples covering the range of False Bay differentiation (LB1, SM, CP2) and the samples from Robben Island (RI). Shown for comparison are data fields for the Tafelberg type dolerites from the HOD swarm in Namibia (Trumbull et al., 2007) and for the Calvinia sills, Karoo Province (le Roex and Reid, 1978). (b) Eu anomaly (Eu/Eu^*) vs. MgO. (c – e) Variation plots of selected REE ratios vs. MgO, showing a consistent knick at about 5 wt.% MgO (see text).

Figure 5: Plot of initial Sr and Nd isotope ratios of False Bay dolerites and relevant mantle and crustal components. All data were age-corrected to 130Ma. The inset highlights the False Bay data with symbols keyed to MgO contents. Data for the other fields: Bushmanland Gneisses (Reid et al., 1997), Richtersveld Subprovince (Reid et al., 1987b; Reid, 1997), Damara granite and granodiorite (Jung et al., 2003), MORB (Hofmann, 1997), HOD and CFB (Trumbull et al. (2007). Additional Sr isotope data for the Cape Granites and Malmesbury Formation are from Allsopp and Kolbe (1965), Nd isotope data for the Malmesbury Formation from Buggisch et al. (2010).

Figure 6: (A) Variation of initial Sr-isotope ratios (for 130 Ma) vs. Rb/Sr ratio for the False Bay dyke swarm (FBDS) compared with published data from the Cape Granite Suite and, Malmesbury

Group (Allsopp and Kolbe, 1965) and the Bushmanland Gneisses (BML) after Reid et al. (1997). (B and C): Variations of initial Sr- and Nd-isotope ratios against MgO. The correlation of isotope ratios with MgO in samples with less than about 5 wt.% MgO indicates the onset of crustal assimilation in the more evolved magmas, see text for discussion.

Figure 7: Variations of Zr content vs. Rb/Sr ratio in the False Bay dykes (symbols keyed for MgO content) are compared with model trend lines calculated for crystal fractionation of clinopyroxene and plagioclase in the ratios indicated (using mineral-melt partition coefficients from McKenzie and O’Nions (1991) and McCallum and Charette (1978). Data trends for the high-MgO samples (unaffected by assimilation) are best fit by the fractionation trend for a 60:40 proportion of plagioclase to pyroxene.

Figure 8: Th/Yb vs Nb/Yb plot of False Bay dolerites (symbols keyed to MgO) compared with the mantle array (Pearce, 1982); MORB (Hofmann, 1997); Primitive Mantle (Pearce, 1982), and with crustal components from the region: Richtersveld Subprovince (Reid, 1997) and Bushmanland Gneisses (Reid et al., 1987a, 1997); Malmesbury Group and Cape Granite suite (this study; Allsopp and Kolbe, 1965), as well as unpublished student theses from UCT (Mbangula, 2004; Mbeje, 2005). Note that most intermediate to low-MgO dolerites plot above the Mantle Array toward crustal components.

Figure 9: Assimilation and fractional crystallisation models of magma evolution in the False Bay dyke swarm. Model AFC curves start from sample OD2 with 5.2 wt.% MgO, and represent a range of $f = 1$ to 0.3 (30% melt remaining), with ticks at intervals of 10%. Labels indicate the crustal component and r -value used for each curve (CG-Cape Granite, MM-Malmesbury metasediment, BML-Bushmanland gneiss). The arrows labeled FC show trends for fractional crystallization only.

(a) model results for the Rb-Sr system rule out Cape Granite and Malmesbury metasediments as assimilants and favor the Bushmanland gneiss at moderate r -values of 0.1 to 0.2. Binary mixing (dashed lines) would fit the data but is considered unlikely for thermal reasons (see text).

(b) The Sm-Nd data suggest a first stage of magma evolution with no change in Nd isotopic composition (high-MgO samples), after which crustal assimilation progressively lowers $^{143}\text{Nd}/^{144}\text{Nd}$ ratios. AFC models starting from sample OD2 (labeled curves with tic marks) show that Bushmanland gneiss can produce the observed Nd- isotopic range at moderate r -value of 0.2, whereas the Cape Granite and Malmesbury curves fall short even for $r=0.4$. The spread in Sm/Nd may reflect variable Sm/Nd in the crust or different starting points from the FC trend (see text).

Bulk partition coefficients used in calculations are: $D_{\text{Rb}} = 0.04$, $D_{\text{Sr}} = 1.13$, $D_{\text{Sm}} = 0.17$ and $D_{\text{Nd}} = 0.168$, based on D-values from McKenzie and O'Nions (1992) and an assemblage of 60% plagioclase to 40% clinopyroxene from Fig. 7. Compositional data are from Table 2 (samples CJ12 and MM001 for Cape Granite and Malmesbury, respectively), and for Bushmanland Gneiss: Reid et al. (1997).

- do** False Bay dolerite dykes (132 Ma)
- Op** Peninsula Formation (Ordovician)
Op* → Massive Quartzite
- Og** Graafwater Formation (Ordovician)
----- unconformity
- Ec** Cape Granite (540 - 500 Ma)
- Nt** Malmesbury Group (750 - 560 Ma)

Table 1. List of dyke sample locations.

Location	Name and sample abbreviation
1	Sea Point (SP)
2	Clifton 2 (CL2)
3	Clifton 1 (CL1)
4	Bakoven (BK)
5	Kasteelpoort (KS)
6	Oudekraal 1 (OD1)
7	Oudekraal 2 (OD2)
8	Llandudno Ravine (LR)
9	Logies Bay (LB)
10	Sandy Bay (SB)
11	Chapman's Peak 1 (CP1)
12	Chapman's Peak 2 (CP2)
13	Chapman's Peak 3 (CP3)
14	Chapman's Peak 4 (CP4)
15	Chapman's Peak 5 (CP5)
16	Oukaapseweg (OK)
17	Nursery Ravine (NS)
18	Tokai 1 (TK1)
19	Tokai 2 (TK2)
20	Froggy Pond (FG1)
21	Oatland Point (FG2)
22	Millers Point (MP)
23	Smitswinkelbaai (SM)
24	Robben Island (RI)

Locations are shown in Figure 1.

Sample abbreviations are in Table 2

Table 2: Geochemical and radiogenic isotope compositions of the False Bay dykes and basement samples

Sample Dyke location*	GCH89-20 TK1	LB-07 LB1	GCH89-22 TK2	GCH89-11 CP1	LB-11 LB2	GCH89-14 CP3	GCH89-03 SB	GCH89-33 MP
SiO ₂	49.6	49.5	49.8	49.6	50.1	52.4	50.1	50.7
TiO ₂	1.49	1.21	1.41	1.46	1.62	0.99	1.99	2.09
Al ₂ O ₃	14.7	16.7	15.9	15.4	14.9	15.7	14.3	14.4
FeO	11.0	9.5	10.6	10.3	11.4	10.4	13.1	12.2
MnO	0.35	0.16	0.18	0.19	0.20	0.21	0.23	0.23
MgO	7.31	7.33	7.26	7.16	6.97	6.29	6.05	5.75
CaO	11.1	11.5	11.2	11.0	10.7	10.3	9.3	9.3
Na ₂ O	2.00	2.17	2.34	2.52	2.37	2.13	2.56	2.77
K ₂ O	0.38	0.49	0.66	0.69	0.65	1.05	0.95	1.26
P ₂ O ₅	0.24	0.18	0.21	0.23	0.28	0.16	0.32	0.46
LOI	1.84	1.20	0.41	1.46	0.84	0.40	1.10	0.93
Sc	42	39	39	43	45	37	40	42
V	293	254	276	298	319	299	368	348
Cr	270	336	285	194	139	144	36	77
Co	46	44	44	44	47	42	51	43
Ni	68	89	79	63	64	69	51	33
Cu	85	80	71	83	83	112	85	40
Zn	97	86	93	101	111	90	142	118
Rb	14.9	11.9	16.4	16.1	14.4	33.1	22.3	31.1
Sr	327	209	214	204	213	163	355	301
Y	25	21	23	28	30	24	26	37
Zr	97	80	85	127	135	99	129	171
Nb	9.5	7.7	9.0	12.0	12.2	7.4	11.8	24.2
Ba	400	138	165	181	181	219	252	324
La	11.2	9.5	11.0	14.5	14.6	12.7	15.7	24.6
Ce	24.8	20.9	24.1	31.7	32.4	27.4	35.3	52.2
Pr	3.28	2.76	3.16	4.10	4.25	3.47	4.67	6.40
Nd	14.9	12.3	14.2	18.2	18.9	14.7	20.8	27.4
Sm	3.84	3.14	3.60	4.49	4.68	3.63	4.93	6.10
Eu	1.29	1.08	1.21	1.41	1.49	1.08	1.68	1.94
Gd	4.36	3.62	4.09	4.98	5.33	4.12	5.35	6.98
Tb	0.70	0.58	0.65	0.77	0.85	0.67	0.80	1.06
Dy	4.64	3.78	4.37	5.12	5.49	4.47	4.99	6.68
Ho	0.95	0.77	0.87	1.04	1.11	0.91	0.99	1.33
Er	2.80	2.30	2.61	3.10	3.30	2.72	2.83	3.85
Tm	0.40	0.33	0.36	0.43	0.46	0.38	0.39	0.52
Yb	2.65	2.15	2.44	2.93	3.12	2.55	2.53	3.41
Lu	0.40	0.32	0.36	0.44	0.46	0.38	0.39	0.50
Hf	2.56	2.14	2.31	3.25	3.45	2.62	3.32	4.28
Ta	0.52	0.46	0.48	0.65	0.70	0.41	0.70	1.23
Pb	3.1	3.7	3.3	4.0	4.4	5.4	4.1	5.6
Th	1.2	1.3	1.4	1.7	1.5	3.2	2.4	3.2
U	0.3	0.4	0.4	0.4	0.4	0.8	0.5	0.8
⁸⁷ Sr/ ⁸⁶ Sr ± 2σ (m)		0.706671 ± 12	0.706850 ± 13	0.707372 ± 12	0.706703 ± 11		0.707569 ± 16	0.707304 ± 11
⁸⁷ Sr/ ⁸⁶ Sr (130Ma)		0.70637	0.70644	0.70695	0.70634		0.70723	0.70675
¹⁴³ Nd/ ¹⁴⁴ Nd ± 2σ		0.512509 ± 9	0.512516 ± 6	0.512498 ± 8	0.512497 ± 9		0.512473 ± 6	0.512511 ± 6
¹⁴³ Nd/ ¹⁴⁴ Nd		0.51238	0.51239	0.51237	0.51237		0.51235	0.51240
εNd (130 Ma)		-1.8	-1.7	-1.9	-2.0		-2.3	-1.5

Dyke locality abbreviations from Table 1.

Major element oxides in weight percent, measured by XRF from Reid (1990), with total Fe as FeO, recalculated to 100%

Trace elements in ppm, measured by ICP-MS except where noted (Zr* by XRF)

Table 2: continued.

Sample Dyke Location	GCH89-28 FG1	GCH89-45 OD2	FBD-601 RI	GCH89-31 FG2	GCH89-40 SP	GCH89-39 SL	LB-50 SW
SiO ₂	51.8	53.0	50.2	52.3	49.3	50.6	52.8
TiO ₂	2.10	1.83	2.52	2.20	2.89	2.98	1.83
Al ₂ O ₃	14.7	17.0	12.2	14.2	12.7	13.2	13.4
FeO	11.4	7.6	17.3	11.7	15.0	13.9	14.2
MnO	0.21	0.11	0.26	0.19	0.26	0.68	0.23
MgO	5.24	5.09	5.02	4.97	4.74	4.65	4.61
CaO	8.7	10.5	9.1	8.4	8.5	7.8	8.1
Na ₂ O	2.87	2.23	1.79	2.84	2.26	2.25	2.94
K ₂ O	1.62	0.91	0.83	1.70	1.12	1.68	1.15
P ₂ O ₅	0.49	0.31	0.30	0.54	0.54	0.54	0.23
LOI	0.90	1.42	0.53	0.90	2.68	1.62	0.52
Sc	35	44	49	36	45	40	44
V	300	290	634	329	484	433	535
Cr	69	270	61	49	37	29	43
Co	39	43	51	39	51	63	47
Ni	37	62	51	20	38	21	42
Cu	27	57	420	12	41	47	198
Zn	119	129	179	121	167	140	136
Rb	46.2	21.7	27.8	49.3	37.1	51.0	31.8
Sr	302	252	116	288	326	207	143
Y	37	35	56	38	49	58	40
Zr	220	146	201	218	237	283	154
Nb	28.8	15.3	12.3	28.5	24.0	28.0	11.3
Ba	378	240	195	390	396	408	251
La	30.1	18.9	16.5	30.9	28.2	33.0	18.4
Ce	64.2	35.4	38.3	66.1	61.2	71.0	40.0
Pr	7.91	5.12	5.17	8.17	7.64	8.80	5.05
Nd	32.9	22.4	24.2	33.9	33.6	38.0	21.7
Sm	7.25	5.25	6.72	7.39	7.81	8.30	5.36
Eu	2.00	1.70	2.02	1.99	2.42	2.20	1.61
Gd	7.39	6.16	8.82	7.48	9.22	9.60	6.47
Tb	1.12	0.97	1.46	1.13	1.43	1.50	1.06
Dy	7.06	6.16	9.97	7.07	9.28	9.60	7.14
Ho	1.39	1.25	2.07	1.40	1.88	2.00	1.50
Er	4.09	3.59	6.30	4.08	5.58	5.70	4.53
Tm	0.57	0.50	0.90	0.57	0.79	0.80	0.64
Yb	3.76	3.29	6.03	3.73	5.24	5.30	4.33
Lu	0.56	0.49	0.89	0.55	0.78	0.80	0.65
Hf	5.36	3.74	5.37	5.39	5.95	5.30	4.14
Ta	1.55	0.82	0.69	1.54	1.29	1.40	0.73
Pb	8.0	5.8	4.1	8.7	7.2	11.1	7.8
Th	4.8	2.6	2.8	5.4	3.9	7.7	5.2
U	1.3	1.0	0.5	1.4	1.0	2.1	1.3
⁸⁷ Sr/ ⁸⁶ Sr ± 2σ (m)	0.707628 ± 10	0.707854 ± 14	0.709223 ± 10	0.708965 ± 15		0.717372 ± 12	0.713931 ± 17
⁸⁷ Sr/ ⁸⁶ Sr (130Ma)	0.70681	0.70739	0.70794	0.70805		0.71605	0.71275
¹⁴³ Nd/ ¹⁴⁴ Nd ± 2σ	0.512496 ± 7	0.512469 ± 7	0.512559 ± 7	0.512466 ± 7		0.512362 ± 6	0.512405 ± 5
¹⁴³ Nd/ ¹⁴⁴ Nd	0.51238	0.51235	0.51242	0.51235		0.51225	0.51228
εNd (130 Ma)	-1.7	-2.4	-1.1	-2.3		-4.3	-3.8

Table 2: continued.

Sample	LB-40	LB-43	GCH89-17	GCH89-36	GCH89-43	PD-2	GCH89-41	LB-33
Dyke Location	CL1	CL2	CP5	OK	BK	LR	KS	OD1
Major oxides (wt%)								
SiO ₂	52.1	53.1	54.1	54.8	52.1	55.0	53.3	55.0
TiO ₂	2.21	2.15	2.35	2.21	2.88	2.49	2.63	2.52
Al ₂ O ₃	14.4	14.2	13.6	13.6	12.5	13.3	12.3	13.0
FeO	12.8	12.7	12.5	12.4	15.2	12.9	15.7	12.5
MnO	0.22	0.20	0.29	0.20	0.24	0.41	0.24	0.19
MgO	4.26	4.19	4.10	3.84	3.49	3.23	3.08	2.96
CaO	6.4	8.0	7.3	6.8	7.3	6.1	7.0	5.9
Na ₂ O	3.04	2.46	2.34	2.41	2.29	1.70	2.14	2.91
K ₂ O	2.48	1.69	2.18	2.11	1.80	2.82	1.94	2.64
P ₂ O ₅	0.41	0.40	0.52	0.51	0.75	0.64	0.41	0.66
LOI	1.67	0.93	0.79	1.16	1.47	1.49	1.29	1.72
Sc	39	39	35	39	41	38	42	37
V	444	440	410	349	375	421	401	404
Cr	42	46	23	65	15	13	3	8
Co	41	40	39	35	43	39	45	35
Ni	30	32	25	29	16	16	4	12
Cu	47	46	26	20	16	12	16	7
Zn	140	140	132	136	169	145	169	146
Rb	134.1	67.0	80.3	83.7	62.8	105.3	76.1	133.5
Sr	170	181	182	187	142	198	138	159
Y	44	44	42	45	55	54	52	48
Zr	214	220	225	195	230	239	220	243
Nb	19.8	20.1	21.9	18.3	20.5	23.9	16.8	24.3
Ba	372	337	413	358	375	961	356	408
La	26.7	27.3	31.3	27.6	32.3	34.5	26.5	35.2
Ce	57.7	58.4	66.9	58.6	69.0	73.6	57.3	75.4
Pr	7.17	7.24	8.26	7.09	8.42	8.90	7.02	9.23
Nd	30.4	30.7	34.1	29.6	35.8	37.2	30.1	38.1
Sm	7.01	7.04	7.60	6.72	8.18	8.25	7.20	8.36
Eu	1.93	1.92	1.95	1.95	2.32	2.15	2.22	2.12
Gd	7.81	7.84	7.94	7.79	9.76	9.23	8.98	8.86
Tb	1.22	1.25	1.21	1.22	1.53	1.41	1.44	1.37
Dy	8.05	8.16	7.79	8.06	10.12	9.08	9.49	8.75
Ho	1.62	1.65	1.55	1.64	2.06	1.87	1.96	1.76
Er	4.89	4.95	4.57	4.90	6.14	5.46	5.84	5.27
Tm	0.69	0.70	0.64	0.69	0.86	0.75	0.83	0.73
Yb	4.64	4.70	4.26	4.56	5.75	4.87	5.49	4.89
Lu	0.69	0.69	0.63	0.68	0.84	0.72	0.82	0.74
Hf	5.50	5.63	5.78	5.05	5.94	6.11	5.71	6.25
Ta	1.20	1.21	1.36	1.19	1.19	1.53	0.99	1.64
Pb	9.4	10.1	12.4	12.3	11.7	14.3	10.9	14.6
Th	6.2	6.4	8.3	8.3	8.0	10.0	7.3	10.4
U	1.6	1.7	2.3	2.5	2.1	2.8	1.8	2.7
Radiogenic Isotopes								
⁸⁷ Sr/ ⁸⁶ Sr ± 2σ (m)	0.721096 ± 18		0.717324 ± 17				0.724626 ± 16	
⁸⁷ Sr/ ⁸⁶ Sr (130Ma)	0.71686		0.71496				0.72011	
¹⁴³ Nd/ ¹⁴⁴ Nd ± 2σ	0.512341 ± 7		0.512371 ± 6				0.512337 ± 6	
¹⁴³ Nd/ ¹⁴⁴ Nd	0.51222		0.51226				0.51222	
εNd (130 Ma)	-4.8		-4.2				-4.8	

Table 2: continued, with samples of Cape Granite(*) and Malmesbury Formation (**)

Sample Location	GCH89-24 NS	GCH89-07 CP2	CG-A1 * Cable Way	CJ-08* Sea Point	CJ-12* Sea Point	MM-001** Grotto Bay	NM-012** Robben Island
SiO ₂	54.2	54.6	71.8	71.4	71.5	72.9	60.9
TiO ₂	2.54	2.62	0.48	0.38	0.38	0.88	0.91
Al ₂ O ₃	12.2	12.4	13.7	14.6	14.5	11.8	19.4
FeO	15.2	14.7	3.02	2.57	2.60	4.9	7.3
MnO	0.24	0.27	0.07	0.07	0.06	0.08	0.12
MgO	2.88	2.70	1.11	0.88	0.92	2.34	4.23
CaO	7.4	6.2	1.3	1.1	1.2	1.3	0.3
Na ₂ O	2.02	2.07	2.55	3.00	2.77	2.67	2.04
K ₂ O	2.05	2.46	5.14	4.97	4.95	2.92	4.65
P ₂ O ₅	0.43	0.77	0.21	0.24	0.23	0.23	0.15
LOI	0.79	1.20	0.55	0.83	0.88	-	-
Sc	39	37	< 3	4	5	10	16
V	343	286	31	30	31	87	123
Cr	3	11	15	14	15	68	85
Co	40	32	4	5	5	11	16
Ni	3	9	7	7	7	30	40
Cu	13	1	< 0.5	1	7	23	46
Zn	166	156	41	49	55	-	-
Rb	76.3	98.3	261	238	247	101	153
Sr	427	161	90	75	80	136	57
Y	53	53	24	21	23	30	19
Zr	226	239	*135	*147	*155	145	112
Nb	17.0	21.6	15.8	12.4	13.2	12.2	13.7
Ba	845	614	340	364	357	499	491
La	27.8	35.1	16.2	25.2	25.5	40.1	25.1
Ce	60.3	76.4	42.7	53.7	54.3	84.8	56.3
Pr	7.54	9.51	4.34	6.20	6.36	9.72	6.35
Nd	32.2	39.8	17.0	23.9	24.2	38.2	25.3
Sm	7.81	9.18	3.52	4.99	5.36	7.18	4.97
Eu	2.16	2.23	0.44	0.84	0.87	1.41	0.97
Gd	8.93	9.69	3.02	4.56	5.05	6.53	4.44
Tb	1.44	1.52	0.45	0.71	0.80	0.96	0.67
Dy	9.62	9.75	2.75	4.35	4.68	5.64	4.12
Ho	1.96	1.96	0.49	0.76	0.82	1.12	0.81
Er	5.91	5.83	1.35	2.15	2.13	3.14	2.39
Tm	0.85	0.82	0.19	0.30	0.28	0.47	0.36
Yb	5.64	5.51	1.15	1.89	1.68	2.99	2.49
Lu	0.84	0.81	0.15	0.27	0.22	0.44	0.36
Hf	6.01	6.28	1.88	1.594	1.545	3.99	2.94
Ta	1.04	1.41	0.94	1.554	1.267	0.87	7.28
Pb	11.5	14.0	62.26	23.94	62.26	16.6	31.9
Th	7.8	10.2	11.5	11.07	11.5	10.855	12.2
U	2.0	2.6	4.33	3.068	4.33	3.15	1.51
⁸⁷ Sr/ ⁸⁶ Sr ± 2σ (m)		0.723599 ± 10	0.782517 ± 12	0.779682 ± 11	0.779632 ± 10	0.731501 ± 4	0.778418 ± 4
⁸⁷ Sr/ ⁸⁶ Sr (130Ma)		0.72032	0.76472	0.76401	0.76291	0.72752	0.76411
¹⁴³ Nd/ ¹⁴⁴ Nd ± 2σ		0.512253 ± 7	0.512177 ± 9	0.512171 ± 6	0.512213 ± 6	0.512163 ± 3	0.511868 ± 5
¹⁴³ Nd/ ¹⁴⁴ Nd		0.51213	0.51209	0.51204	0.51210	0.51207	0.51177
εNd (130 Ma)		-6.6	-7.5	-8.4	-7.2	-7.9	-13.7

Unpublished major element data by XRF for Cape Granites (CJ-08 and CJ-12) from Creighton-Jones (1982). Unpublished major and trace element data for Malmesbury Formation samples MM-001 and NM-012 from Mbeje (2005) and Mbangula (2004) respectively.

*Zr concentration of granites shown as measured by XRF.

Table 3. Least square models for evolution of the False Bay dykes.

Intervals	1	2	3	4
MgO range (wt%)	7.4 - 7.0	7.0 - 5.8	5.8 - 4.7	4.7 - 2.7
Olivine (Fo)	75	70	0	0
Plagioclase (An)	79	75	48	45
Mineral modes (vol. %)				
Olivine	13	10	0	0
Plagioclase	67	46	52	47
Clinopyroxene	20	44	43	40
Magnetite	0	0	5	7
Ilmenite	0	0	0	6
error (sum of squares)	0.01	0.36	0.31	5.0*
melt proportion (f)	0.76	0.8	0.66	0.52

Closed-system least-squares crystallisation models for intervals shown.

Mineral compositions: olivine (Fo) and plagioclase (An) as shown on table, clinopyroxene (En48, Fs13, Wo39) and magnetite (Mt88, Usp12) from Reid (1990), Ilmenite was taken as pure ilmenite (FeTiO₃).

The excessive error shown for interval 4 (*) indicates failure of the closed-system assumption.