

STRONTIUM ISOTOPE CHARACTERIZATION OF KAROO VOLCANIC ROCKS

by

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ABSTRACT

Karoo volcanics of both mafic and felsic composition cover large areas of southern Africa, and are characterized by considerable ranges in their initial $^{87}\text{Sr}/^{86}\text{Sr}$ (i.e. R_0) ratios. Mafic volcanics from the eastern and western continental margins show the greatest range in R_0 ratios, i.e. Lebombo–Nuanetsi area: 0.7035–0.7134 and Etendeka: 0.7030–0.7135 respectively. In the Central Karoo area the mafic rocks show a smaller spread of ratios: 0.7046–0.7094. In both the Lebombo and Etendeka areas R_0 ratios of ≤ 0.704 are characteristic of late stage dykes. In the Lebombo monocline basalts from the southern region show the greatest range in R_0 ratios whereas mafic rocks from the northern region and Nuanetsi exhibit more restricted ranges in R_0 ratios (0.7042–0.7092). The viability of crustal contamination as a process causing the high and variable R_0 ratios of Karoo mafic rocks is evaluated using isotopic, major and trace element and geological data. Apart from the Effingham dolerite suite in northern Natal, contamination is rejected as the major process capable of consistently explaining the R_0 ratios of the mafic rocks. The R_0 ratios essentially reflect those of the mantle source areas of the mafic rocks. Support for this conclusion comes from complementary studies of mantle xenoliths in southern African kimberlites. These indicate a complex history of metasomatic enrichment of the subcontinental lithosphere such that high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios can evolve in quite short periods of time. Metasomatism also leads to enhanced concentrations of incompatible elements. Most Karoo mafic rocks have source areas located in the old enriched subcontinental lithosphere. Voluminous silicic volcanics are confined to the Etendeka and Lebombo regions. In the Etendeka, quartz latites generally have R_0 ratios of >0.718 , are unrelated to the mafic rocks and are considered to be partial melts of the Pan-African basement rocks. In the Lebombo, the main rhyolites have R_0 ratios = 0.7044 ± 2 (southern Lebombo), 0.7048 ± 1 (northern Lebombo) and 0.7085 ± 6 (Nuanetsi). Only in the northern Lebombo is there close agreement between the R_0 ratios of the rhyolites and the restricted range in R_0 ratios for the associated mafic rocks. On the basis of R_0 ratios along the whole length of the Lebombo, the rhyolites appear to be genetically unrelated to the mafic volcanics. Although the origin of the Lebombo rhyolites is not fully resolved, the available isotopic evidence suggests they are probably of lower crustal derivation. The association of high R_0 in mafic volcanics with rifted continental margins may suggest that the pattern of continental rifting is influenced by the distribution of zones of enriched lithospheric mantle, in that the latter is hotter, less viscous and therefore more easily penetrated by material from the asthenosphere. The occurrence of voluminous rhyolites in the rifted margin areas may be a result of the increased heat supply to the crust from these zones of geochemically and thermally anomalous lithospheric mantle.

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I. INTRODUCTION

Radiogenic isotopes, initially Sr and Pb, and more recently Nd, have been widely employed in studies of igneous petrogenesis. Results of these isotope studies have commonly been used in conjunction with major and trace element data, in attempts to identify processes (such as contamination) which may have modified primary magma compositions, to characterize different magma types and their sources (e.g. mid-ocean ridge basalts, ocean island basalts), and to constrain petrogenetic models for different magmatic suites. One of the main results of these investigations is the recognition that mafic rocks of continental areas generally have higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (hereafter referred to as R_o) than their counterparts of similar age in the oceans. In particular, the higher and more variable R_o ratios exhibited by continental flood tholeiites in comparison to those of mid-ocean ridge basalts have been the subject of considerable debate (e.g. Allègre *et al.*, 1982a). The high R_o ratios for continental basalts have frequently been interpreted in terms of contamination by crustal materials (e.g. Compston *et al.*, 1968; Faure *et al.*, 1974; Moorbath and Thompson, 1980; Carlson *et al.*, 1981; Mahoney *et al.*, 1982; Thompson *et al.*, 1982). However, a growing body of data on mantle inclusions in alkaline lavas, especially those from kimberlites in southern Africa, indicate that some parts of the subcontinental mantle have been metasomatically enriched and that these mantle materials may have high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (e.g. Barrett, 1975; Kramers, 1977, 1979; Erlank and Shimizu, 1977; Menzies and Murthy, 1980; Erlank *et al.*, 1982).

Thus an alternative view to crustal contamination proposes that the high and variable R_o ratios in continental basalts are inherited from their source regions in the old enriched subcontinental lithosphere (e.g. Brooks *et al.*, 1976; Erlank *et al.*, 1980; Kyle, 1980). Implicit in this is the interpretation, used in this paper, that the term crustal contamination does not include recycling of crust into the mantle. The resolution of these alternative interpretations of the R_o ratio data of continental basic rocks is of considerable importance in view of the implications for models of Earth structure and mantle convection (see Allègre *et al.*, 1982b; Hofmann *et al.*, 1978). In this paper Sr-isotope data of Karoo volcanic and intrusive rocks of southern Africa are evaluated in the context of this debate. In addition the data are used to examine relationships between the different mafic and felsic magma types, and to constrain models for the petrogenesis of the latter.

Karoo volcanics occur as remnants of a volcanic carapace which probably covered much of southern Africa at the end of the Jurassic. The volcanics include a diverse assemblage of mafic rocks (e.g. nephelinites, picrite basalts, basalts, tholeiitic andesites and shoshonites) and several groups of felsic rocks (e.g. rhyolites, rhyodacites, dacites, quartz latites and syenites). Major remnants of lavas are preserved in the Lebombo–Nuanetsi region, Lesotho and the north-eastern Cape, Botswana, and the Etendeka region though important outcrops of lavas are also found beyond these areas (Fig. 1). In addition, older strata covering very large areas of southern Africa host a network of dolerite dykes, sills, sheets and other intrusive bodies (Walker and Poldervaart, 1949). Many new data are now available for these volcanics and intrusive rocks, and readers may refer to the following papers for information relating to the volcano-stratigraphy, geochronology and geochemistry of the Karoo igneous rocks of the three main areas: *Lebombo*—Cleverly and Bristow (1979), Bristow (1982, 1984a, b), Armstrong *et al.* (1984), Cox and Bristow (1984),

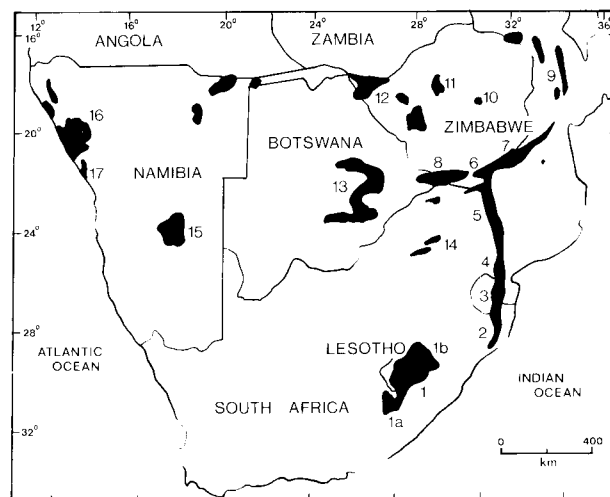


Figure 1.

Distribution of Karoo volcanics in southern Africa and location of various sub-areas discussed in this paper. (1—Central Karoo; 1a—Stormberg; 1b—Lesotho; 2—Southern Lebombo; 3—Swaziland; 4—Central Lebombo; 5—Northern Lebombo; 6—Nuanetsi; 7—Sabi; 8—Tuli; 9—Lupata; 10—Featherstone; 11—Nyamandhlovu; 12—Victoria Falls—Wankie; 13—Botswana; 14—Springbok Flats; 15—Mariental; 16—Etendeka; 17—Cape Cross).

Cox *et al.* (1984); *Central Karoo area*—Marsh and Eales (1984); *Etendeka*—Erlank *et al.* (1984). Studies of a more regional nature have been described by Allsopp *et al.* (1984b), Eales *et al.* (1984), Duncan *et al.* (1984) and Fitch and Miller (1984).

Most of the Sr-isotope data discussed here have been determined during the current study and, together with analyses drawn from other sources, are listed in Appendix A. The listing in Appendix A has been subdivided by region, and further divisions are made within each region on the basis of volcano-stratigraphy or magma-type. A summary of these data is presented in Table I from which it is immediately apparent that the mafic rocks show large variations in R_o ratios from 0.7030 to 0.7135. In addition, considerable variation exists in R_o ratios for samples from a

TABLE I
Summary of R_o Ratios of Karoo Volcanic and Intrusive Rocks

Region	Rock Type	R_o
Southern Lebombo and Swaziland	Basalts and dolerites	0.7038–0.7134 (25)
	Rooi Rand dolerites	0.7035–0.7042 (14)
	Effingham dolerites	0.7117–0.7223 (6)
	Jozini rhyolites	0.7044 ± 2 (54)
	Kuleni rhyolites	0.7114 ± 103 (6)
	Bumbeni felsic rocks	0.7055 ± 2 (4)
Central & northern Lebombo	Mafic rocks	0.7042–0.7069 (19)
	Jozini rhyolites	0.7048 ± 1 (16)
Nuanetsi	Mafic rocks	0.7046–0.7092 (26)
	Rhyolites	0.7085 ± 6 (6)
Central area	Basalts and dolerites	0.7046–0.7094 (67)
	Andesites and dacites	0.7094–0.7130 (5)
Northern Namibia	Basalts and dolerites	0.7065–0.7135 (23)
	Horingbaai dolerites	0.7030–0.7034 (3)
	Latites	0.7153 ± 1 (2)
	Quartz latites	0.7186–0.7218 (16) (one value at 0.7137)

Number of samples is indicated in parentheses. Central area includes western and north-western Cape, Springbok Flats and southern Namibia.

given geographic area, e.g. southern Lebombo. The felsic rocks (except for those from Etendeka) do not show such wide variations in R_o ratios for particular suites within individual areas, although variation between rock suites within a given geographical area does occur. Data for the

mafic and some felsic rocks from the four main areas above have been plotted on diagrams of R_o ratios vs Rb/Sr (Fig. 2a-d). The marked variation in R_o ratios for mafic rocks is clear from these diagrams, particularly for the southern Lebombo and Etendeka regions. In addition the data do

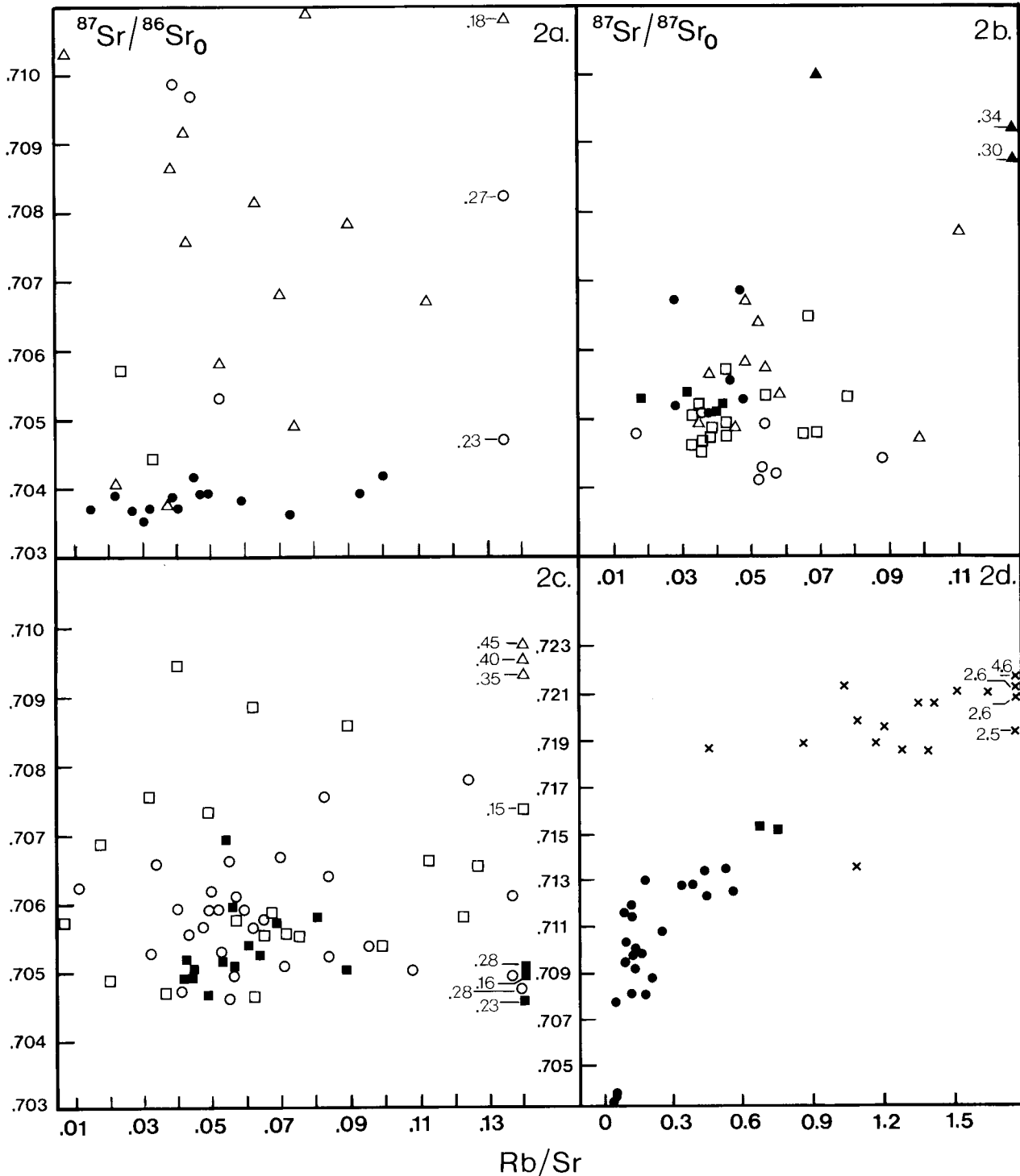


Figure 2

R_o ratio vs Rb/Sr diagrams for the Karoo volcanics. Numbers within diagrams refer to Rb/Sr ratios for samples plotting outside the scale of the diagrams.

- a. Southern Lebombo (including Swaziland): triangles — basaltic lavas of the Sabie River Formation; circles — dolerite dykes; dots — Rooi Rand dolerites; squares — basalts of the Movene Formation.
- b. Central-northern Lebombo and Nuanetsi: dots — nephelinites of the Mashikiri Formation; open squares — picrite basalts of the Letaba Formation; solid squares — picrites (intrusive); open triangles — basalts of the Sabie River Formation; solid triangles — Nuanetsi Interbedded Basalts; open circles — dolerites.
- c. Central Karoo (includes western and north-western Cape dolerites); solid squares — basalts of the Lesotho Formation; open squares — basalts of various other Formations recognized in Stormberg region; open triangles — andesites and dacites; circles — dolerites. Not shown are two Pronsberg dacites with high R_o ratios (0.7124 and 0.7130).
- d. SWA/Namibia (Etendeka Formation volcanics): dots — basalts and dolerites; squares — latites; crosses — quartz latites.

not define Rb-Sr isochrons of Karoo age except for the Jozini and Kuleni rhyolites and the Bumbeni felsic rocks from the southern Lebombo and the rhyolites from Nuanetsi (see Allsopp *et al.*, 1984b), nor do they lie on a 1.6 b.y. linear trend as suggested previously by Brooks and Hart (1978).

II. ANALYTICAL METHODS

Sr-isotope ratios were determined at the Bernard Price Institute of Geophysical Research (BPI), University of Texas at Dallas (UTD), the University of Oxford (OXF) and the Open University (OU). Most Rb/Sr ratios were determined by X-ray fluorescence (Bristow, 1980) at the University of Cape Town (UCT), Rhodes University (RU) and Oxford University except in those cases where Rb or Sr concentrations were determined at the BPI by isotope dilution. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were normally measured on unspiked samples after dissolution and separation by conventional methods. No blank corrections were necessary.

$^{87}\text{Sr}/^{86}\text{Sr}$ ratios quoted in this paper are relative to the following values for either the Eimer and Amend or NBS-987 SrCO_3 standards: OXF — E & A = 0.70800; UTD — E & A = 0.7080; OU — NBS-987 = 0.71015 and BPI — NBS-987 = 0.71023. Inter-laboratory checks were also carried out for both $^{87}\text{Sr}/^{86}\text{Sr}$ and Rb/Sr ratios, and although small differences in these ratios were obtained in some instances these are insignificant in relation to the wide range exhibited by Karoo rocks.

R_0 ratios were calculated, assuming closed system behaviour of Rb and Sr (see discussion in following section), on the basis of $\lambda^{87}\text{Rb} = 1.42 \times 10^{-11}\text{y}^{-1}$ and using either measured ages (e.g. K-Ar) or, in some cases, assumed geological correlations. Where feasible, however, both ages and R_0 ratios were calculated by the isochron method. Further details of the analytical and computational procedures adopted are included as footnotes to Appendix A.

III. RADIOGENIC Sr CORRECTIONS AND SECONDARY PROCESSES

A. The Age Factor

The age of some of the volcanics is not accurately known, and the use of assumed ages to calculate R_0 ratios could contribute to some of the observed scatter. In particular, an age of 190 m.y. has been used for all the Central area and Lebombo mafic rocks although at least some of the dolerites may be as young as 160 m.y. (Fitch and Miller, 1984). However, for these mafic rocks the magnitude of the radiogenic Sr correction is rarely greater than 0.001. Uncertainty in the true ages of the Karoo volcanics cannot, therefore, account for the large variations observed in their calculated R_0 ratios.

B. Secondary Processes

Hydrothermal processes and secondary alteration, unless essentially coincident with emplacement, may have caused redistribution of Rb and/or Sr thereby causing incorrect radiogenic corrections to be made on measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. Likewise, these processes may have affected the actual $^{87}\text{Sr}/^{86}\text{Sr}$ ratios such that calculated R_0 ratios would be incorrect. The effects of hydrothermal and alteration processes have thus been examined to assess what effect they may have had on both Rb/Sr and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios.

1. Hydrothermal Activity

Zeolites and related minerals such as calcite are commonly found in amygdaloids and veins of the basalts, indicating syn- or post-crystallization hydrothermal activity. At one locality from the southern Lebombo Rb and Sr data were obtained for zeolite and calcite from a

hydrothermal vein (Table II) as well as from the freshest basalt found close to the vein. The R_0 ratios of the zeolite and the basalt overlap within error, suggesting that the zeolitization was essentially contemporaneous with basalt crystallization, and the small difference (0.0004) between the calcite and basalt ratios, although statistically significant, is more than an order of magnitude smaller than the variation observed among the basalts in this area. In addition, total Sr in the calcite and zeolite is low such that large amounts of zeolite and calcite would have to be incorporated in the whole rocks to change the Sr-isotope ratios of the basalts significantly.

TABLE II
Rb, Sr and Sr-isotope Data for Mafic and Felsic Lavas and Associated Hydrothermal Calcite and Zeolite from the Lebombo (190 m.y.) and Etendeka (121 m.y.)

Sample	Type	Rb	Sr	$^{87}\text{Sr}/^{86}\text{Sr}$	R_0
<i>Lebombo</i>					
L505c	Basalt	19.4	172	0.70752 ± 12	0.70664
L505c	Calcite	0.126	33.1	0.70703 ± 7	0.70702
L505c	Zeolite	3.657	80.2	0.70705 ± 7	0.70669
<i>Etendeka</i>					
KLS18	Quartz latite	198	80.2	0.73182 ± 5	0.71951
KLS18	Zeolite	19.2	3154	0.71678 ± 4	0.71675
KLS103	Quartz latite	165	119	0.72547 ± 4	0.71856
KLS103	Zeolite	17.1	6268	0.71652 ± 4	0.71651
KLS103	Calcite	1.15	43.4	0.71675 ± 6	0.71662
KLS66	Basalt	18.3	189	0.71208 ± 6	0.71160
KLS66	Zeolite	1.90	75.3	0.71126 ± 6	0.71113

Conversely, zeolites from Etendeka quartz latites have very high Sr abundances which, together with their lower R_0 ratios relative to the host-rock at 121 m.y., could cause the more altered quartz latites to have lower apparent R_0 ratios than would otherwise be the case; see, however, Allsopp *et al.* (1984b) for a discussion of the Etendeka ages. Similarly, a calcite separate from a quartz latite has a lower R_0 ratio than its host, but its lower Sr content suggests that the presence of calcite would have a smaller influence than would the presence of zeolites on the R_0 ratios of the quartz latites. In contrast, zeolite from an Etendeka basalt has only a slightly lower R_0 ratio than the whole rock and together with a lower Sr content cause insignificant variation of R_0 ratios in the basalts compared to the spread of observed ratios.

It should be noted that the above samples are highly amygdaloidal and were chosen to assess the possible effects of zeolitization and calcitization which is common in Karoo lavas. In the case of the other samples, rocks were chosen which were as far as possible amygdale-free or, when amygdaloids were present, they were hand-picked from the sample before final crushing.

2. Low-temperature Alteration

Although the freshest available material was selected for analytical work, low-temperature alteration, e.g. sericitization of feldspars, is fairly common in Karoo igneous rocks. Much of this alteration, particularly in intrusive rocks, is probably deuteric but interaction with groundwater may also be important and result in significant changes in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and Rb and Sr contents. It is obviously essential to demonstrate that such changes are not responsible for the variation in observed R_0 ratios. Using the loss on ignition (LOI) of analysed samples as a crude measure of the extent of low-temperature alteration (i.e. addition of H_2O , CO_2 , etc.) we have searched for any evidence suggestive of positive correlations between the extent of alteration and $^{87}\text{Sr}/^{86}\text{Sr}$, Rb, Sr, and Rb/Sr in the lavas from different areas. The interpretation of any correlations is equivocal, however, because of the complicated chemical effects of the alteration process and

the intrinsic variation of R_o ratios and Rb and Sr between lava flows. Useful information on the effects of alteration on isotopic data could be obtained from studies of single lava flows. Although data for such lava flows are unavailable, several small Karoo mafic intrusions have been studied in detail and the results of these are summarized below.

Seven samples collected along a vertical traverse through the undifferentiated 10 m thick Amherst sill (Marsh and Eales, 1984) exhibit ranges in Rb/Sr (0.010–0.061) and R_o ratios (0.70565–0.70621). Disregarding the two hypocrystalline, chilled contact samples, which are clearly more altered, these ranges reduce to 0.040–0.061 and 0.70566–0.70609 respectively, suggesting that the total observed ranges in these ratios can be ascribed to the effects of low-temperature alteration. The range in R_o ratios is small compared to that exhibited by Karoo mafic rocks. In a more detailed study, Richardson (1984) has shown that the range in Rb/Sr in the differentiated Tandjiesberg dolerite sill cannot all be attributed to magmatic differentiation and that this ratio, and the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, have been changed by alteration although the range in R_o ratios is small (0.0004). Of particular importance is the absence of significant variations in initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios measured on ten samples from the sill and this is attributed to the insensitivity of the Sm-Nd system to alteration. The Nd-isotope results indicate that there are no primary isotopic heterogeneities in the sill and imply that the range in R_o ratios is due entirely to secondary processes.

As with the Amherst sill this range is small relative to the variation in R_o ratios observed for Karoo mafic rocks. We therefore consider that although individual samples may have had either their Rb/Sr or $^{87}\text{Sr}/^{86}\text{Sr}$ ratios slightly changed by alteration processes, the measured Rb/Sr and calculated R_o ratios are essentially those of the magmas at the time of emplacement. Consequently the large ranges in R_o ratios reported in Table I are taken to be a primary feature of Karoo magmas.

IV. PRIMARY PROCESSES AND ISOTOPIC VARIATION

Assuming that hydrothermal activity and alteration have not caused significantly large changes in the Rb/Sr and R_o ratios of the Karoo rocks, the principal problem raised by the data relates to whether the high and variable R_o ratios are the result of crustal origin or involvement, or whether they reflect variations in the mantle sources. These causes of R_o ratio variation in Karoo magmas are considered to be of a primary nature and will be discussed in detail in this and subsequent sections.

A. Crust-related Processes

Given that volcanic and intrusive rocks must upon leaving their source areas commonly travel through substantial thickness of crustal material, it seems almost inevitable that hot magmas will interact in some way with the country rock as they move through conduits to their site of emplacement (see Walker and Poldervaart, 1949). The nature and extent of interaction are difficult to predict and are influenced by a number of variables. Some of these variables must be the size and shape of the magma transport conduits and storage chambers, the rate at which magma moves through these conduits, and residence times if they are stored in crustal magma chambers. Also of importance are the compositions of the rocks in the crustal section traversed by the magmas and the thermal properties of these rocks and magmas. A feature limiting interaction of magma with its wall rocks is the lining of conduit walls with frozen magma. Two recent discussions on physical models of contamination are those of Patchett (1980) and McBirney (1979). In addition Allègre and Minster (1978) and DePaolo (1981) have described chemical models of

contamination for trace elements and some experimental data are available from the work of Watson (1982). In general, crustal contamination of basalts is usually discussed in terms of two dominant processes, bulk and selective contamination, and these are reviewed below.

1. Bulk Contamination

This is the wholesale incorporation of crustal material, usually acid in composition and either liquid or solid in form, into mantle-derived magma. The composition of the contaminated magma is determined by mass balance mixing of the two components involved. The heat required for bulk contamination has traditionally been regarded as being latent heat supplied by fractional crystallization of the magma and the composition of the contaminated magma must therefore be determined by a combined fractional crystallization–contamination process (Bowen, 1928).

The significance of bulk contamination for $^{87}\text{Sr}/^{86}\text{Sr}$ ratios is that the felsic crustal material usually invoked as a contaminant has high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and a suite of variably contaminated basalts should exhibit good positive correlations between R_o ratios and elements like SiO_2 , K_2O , Rb, etc. Conversely negative correlations should be apparent between R_o ratios and elements with low abundances in felsic rocks, e.g. Mg, Cr, Ni and Co. If, as is probable, fractional crystallization accompanies contamination these correlations should be enhanced as most mafic magmas differentiate towards felsic compositions similar in composition to likely contaminants. Such correlations have frequently been used as evidence for crustal contamination of mafic magmas (e.g. Faure *et al.*, 1972, 1974, 1979). On the other hand, if contamination by mafic crustal material is invoked only minor variations in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios will occur and correlations between R_o ratios and geochemical parameters will not exist. Only felsic contaminants are considered in this paper although Allsopp *et al.* (1984a) have noted that mafic contaminants (especially plagioclase xenocrysts; see also Morse, 1983), may have effected some of the small-scale variations in R_o ratios, including lowering of this ratio from initial magma values.

Bulk contamination therefore results in changes in both the isotopic composition of Sr (and other radiogenic elements) and major and trace element concentrations in the contaminated magma. This type of contamination can therefore be tested by using the equations of DePaolo (1981) or Allègre and Minster (1978) to model contamination trends and comparing the results with measured data. Such tests are extensively applied in the sections that follow. In addition, evidence for this type of contamination may be found in petrographic features of the rock (e.g. xenocrysts, reaction coronas) and in geological relationships between the igneous rock and its wall rocks. In order to substantiate bulk contamination an integrated elemental, isotope, and petrographic approach should be employed in evaluating contamination models.

2. Selective Contamination

Selective contamination is a means whereby variable R_o ratios in the magma may be generated without greatly modifying the bulk composition of the magma. References to this process as a cause of high R_o ratios in mafic magmas have been common in the literature for close on 20 years but it remains a vague concept with no general consensus on the actual mechanism involved. For example, Moorbath and Thompson (1980) envisage selective contamination occurring through the medium of a fluid phase which is generated in the crust and transports selected elements, including radiogenic Sr, towards relatively anhydrous mafic magmas. On the other hand Watson (1982) views the process as one of diffusion between mafic melt and felsic rocks or melt in response to activity gradients between the

two. Other proposed mechanisms are zone refining/wall rock reaction (Green and Ringwood, 1967) and crustal disequilibrium melting to produce small-volume melts, perhaps derived by preferential melting of a specific phase, e.g. mica (Pushkar and Stoesser, 1975). Some of these have been discussed in detail by Pankhurst (1969).

Selective contamination is clearly a complex and little-understood process lacking a comprehensive theoretical and experimental basis that allows incorporation of isotopic and major and trace element data into models which can be compared to measured data. However, there are some factors which should be considered when assessing the applicability of this type of contamination. If the contaminating medium is a fluid or melt, the element concentrations in the medium will be controlled by mineral-medium partition coefficients and therefore are predictable, unless generation of the medium is characterized by gross disequilibrium. For a diffusion process, Watson's (1982) results suggest that the activity gradient for Sr between mafic melt and felsic material must be such that Sr will diffuse towards the mafic melt. This means that the Sr concentration in the felsic component must be about equal to or greater than that in the mafic melt. These aspects will be considered in more detail in a later section.

B. Mantle-related Processes

1. Mantle Depletion and Enrichment

It is now generally accepted that the upper mantle beneath both oceanic and continental regions is heterogeneous in composition (Bailey *et al.*, 1980). The depletion and enrichment processes which have led to this, and their intensity, scale and timing are clearly relevant to radiogenic isotope variations in continental basaltic rocks (e.g. Erlank *et al.*, 1980; Kyle, 1980; Weaver and Tarney, 1981). Mantle depletion is commonly ascribed to the removal of a melt, leading to a lowering of Rb/Sr in the residue and subsequent subdued increase of $^{87}\text{Sr}/^{86}\text{Sr}$ with time. Enrichment of the mantle is believed to result from veining by small-volume partial melts or fluids (Wood, 1979) and from metasomatism (Erlank, 1976; Erlank *et al.*, 1980; Bailey, 1982); the attendant increase in Rb/Sr leading with time to increased $^{87}\text{Sr}/^{86}\text{Sr}$ ratios.

Enrichment and depletion processes are clearly complementary in their chemical nature but need not be so in space and time. That such processes have operated in the mantle sources of basalts has been inferred from the studies of basalts themselves in oceanic areas (e.g. Richardson *et al.*, 1982) and continental areas (e.g. Kyle, 1980). Direct evidence for the existence of both enriched and depleted mantle beneath continental areas is found in mantle nodules contained in alkaline mafic volcanics, in particular kimberlites from southern Africa (Erlank and Rickard, 1977; Gurney and Harte, 1980). The presence of enriched mantle can also be inferred from Rb-Sr isotopic measurements on kimberlite micas (Allsopp *et al.*, 1979). These observations are of special importance to the present study and will be discussed further in Section VII.

2. Underplating

Brooks *et al.* (1976) suggested that high and variable Sr-isotope ratios of mafic magmas may be inherited from the subcontinental lithosphere. Their model envisages a continental lithosphere which thickens with time through being underplated by buoyant material rising through the mesosphere. The enriched and variable antiquity of the lithosphere ensures elevated and variable $^{87}\text{Sr}/^{86}\text{Sr}$ ratios within it. Mafic magmas rising through the lithosphere may become contaminated by the old underplated material and hence change the Sr-isotope characteristics. Thus underplating is a process which can lead to *mantle contamination*. Of course if magmas are generated within

the lithosphere they will reflect directly the isotopic character of their source rocks.

In a different model Cox (1980) and Ewart *et al.* (1980) have suggested that underplating at or near the base of the crust may play an important role in the evolution of continental lava suites. In the case of the Karoo, young subcrustal underplated material has been considered to be important in the genesis of the Lebombo rhyolites (Betton and Cox, 1979); this is discussed in a later section. In consideration of the role of young subcrustal underplated material in continental lava petrogenesis it becomes a matter of semantics as to whether such material is "crust" or "mantle". In the sections that follow we shall regard crustal influences to involve upper crust or *old* lower continental crust. *Young* material underplated on to the base of the lower crust will be regarded as "mantle" material as its chemical and especially isotopic character is essentially "mantle" in nature.

3. Subduction

In view of the suggestions of Cox (1978) that subduction may have initiated Karoo volcanism, mention should briefly be made of the possible role of subduction as a mantle process. Subduction is a means of recycling material with relatively high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios back into the mantle. Subducted material may be added to the mantle wedge by fluids or melts rising from the slab and the chemical and isotopic effects resulting from this process have been reviewed by Pearce (1982) and Hawkesworth (1982). These reviews indicate that subduction is a process leading to enrichment of the mantle, albeit in selected elements, a process referred to as "source contamination" by James (1981).

C. Summary

The high and variable Sr-isotope ratios in continental tholeiites may reflect the interaction of the magmas with continental crust or processes affecting the source rocks of the tholeiites at some substantial time before magma generation. The possibility of obtaining information about the chemical evolution of the mantle from the trace element and isotopic characteristics of mantle-derived igneous rocks has been the *raison d'être* of many basalt studies over the last 15 years. From these studies it is clear that crustal contamination may obscure the source rock characteristics in continental mafic magmas and it is important that this possibility be carefully assessed before inferences regarding source rock character are made. The results of such assessments may not be conclusive as they are invariably model-dependent. In addition, the dangers of working with one isotope system alone are well known. Ideally, evidence from additional isotope systems, e.g. U-Pb, Sm-Nd, O, Ar and He, as well as refined interpretative techniques are needed in order to identify the contributions of the different crustal and mantle processes to the compositions of basalts. Though a full range of isotopic data are not yet available on the Karoo rocks discussed in this paper, a detailed assessment of strontium isotope data coupled with petrographic data, major and trace element data, and Nd isotopic data to a limited extent, have been used to assess the effects of contamination and other processes on the Karoo magmas and thereby place constraints on the petrogenesis of these rocks.

V. REGIONAL ASSESSMENT OF CONTAMINATION MODELS FOR KAROO MAFIC ROCKS

A. Bulk Contamination

For the Lebombo and Central Karoo areas contamination trends have been modelled for combined fractional crystallization-assimilation and simple two component mixing using equations of De Paolo (1981) and Langmuir *et al.* (1978). Composition trends between what is regarded as

uncontaminated or least contaminated mafic rocks and a variety of contaminants have been calculated and Tables III, IV and V summarize the appropriate compositional parameters.

Putative uncontaminated mafic compositions used in the calculations have been carefully chosen by inspection of the data for a much larger number of samples than were analysed for $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. Pristine R_0 ratios for these compositions lie between 0.7040 and 0.7050, within the range acceptable for mantle materials and close to the so-called "Bulk Earth" value at 190 m.y. (e.g. Zindler *et al.*, 1982). Some recent contamination models for continental basalts (e.g. Carlson *et al.*, 1981) have assumed $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for the pristine magmas to be less than 0.7040 but in such instances rocks with these ratios are common in the igneous suites being investigated; in the Karoo province the bulk of exposed volcanic and intrusive rocks have R_0 ratios >0.7040 . Nevertheless, our model calculations have been repeated assuming an R_0 ratio of 0.7035 for pristine magma but the results do not alter our conclusions. The compositions of the contaminants vary depending on the area

under examination and are discussed more fully in the appropriate section.

Contamination trends for the variation of R_0 with both major and trace elements have been calculated. Amongst the major elements contamination has been modelled using SiO_2 and Mg-number because of the compositional contrast in these two parameters between mafic magma and contaminant and because of their sensitivity to concomitant fractional crystallization. Amongst the trace elements, Cr and Ni are likewise sensitive to contamination, decreasing rapidly in contaminated magmas because of dilution by the added silicic melt and removal by fractionating mafic phases for which they have high distribution coefficients (a feature that is enhanced by the increased polymerization of the contaminated melt). In contrast, Rb will be enriched in contaminated liquids by both the added silicic melt and fractional crystallization. Rb is also highly concentrated in micas and should also be a sensitive indicator of contamination by melts derived selectively from mica-rich rocks.

For models involving assimilation-fractional crystalliz-

TABLE III
Contamination Modelling for Effingham Dolerites

	1 A	2 B	3 60% A	4 50% A	5 40% A	6 35% A	7 (3)	8 (2)	9 (6)
SiO_2 %	53.00	74.6	61.64	63.80	65.96	67.04	60.74	63.84	66.39
TiO_2	1.31	0.4	1.03	0.86	0.76	0.72	1.85	1.96	1.62
Al_2O_3	14.30	13.6	14.02	13.95	13.88	13.85	15.01	13.59	13.18
Fe_2O_3^*	12.35	3.6	8.85	7.98	7.10	6.66	10.11	9.03	9.18
MnO	0.17	0.02	0.11	0.10	0.10	0.07	0.22	0.11	0.15
MgO	6.22	0.06	3.76	3.14	2.52	2.22	3.18	1.44	1.60
CaO	9.34	1.40	6.16	5.37	4.58	4.18	3.39	4.23	2.21
Na_2O	2.31	2.70	2.47	2.51	2.54	2.56	2.48	2.48	2.64
K_2O	0.76	3.80	1.93	2.28	2.58	2.74	2.63	3.04	2.76
P_2O_5	0.24	0.10	0.20	0.17	0.16	0.15	0.40	0.30	0.27
$^{87}\text{Sr}/^{86}\text{Sr}$	0.7080	0.740	0.7112	0.7125	0.7143	0.7156			0.7172
Sr ppm	366	60	244	213	182	167			

Columns: 1. Average southern Lebombo basaltic lava.

2. Representative Precambrian granite from eastern Natal (Frankel, 1969). The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio is an assumed value but is within the range of values reported from Swaziland and north-eastern Natal (Condie and Hunter, 1976; Glikson, 1979; Allsopp, unpublished data) as is Sr; in the above case the Sr value represents the average of the relatively evolved Sicunusa-type shield forming batholiths of Swaziland (Condie and Hunter, 1976).

3, 4, 5, 6. Compositions derived by mass-balance addition of different proportions of granite (B) to the average basaltic lava (A).

7, 8, 9. Effingham dolerites, normalized to 100% volatile free (from Frankel, 1969). Numbers in parentheses above refer to Frankel's (1969) original sample numbers.

* Total Fe as Fe_2O_3

TABLE IV
Variables used for Contamination and Contamination-Crystallization Trends for Lebombo and Nuanetsi Mafic Rocks

Model Compositions	SiO_2 %	Mg number	Rb ppm	Cr ppm	Ni ppm	Sr ppm	$^{87}\text{Sr}/^{86}\text{Sr}$
A. Pristine magma compositions							
1. S. Lebombo basalt	50.0	59	8	350	120	220	0.7040
2. N. Lebombo basalt	50.0	59	8	200	120	1000	0.7040
3. Picrite basalt	46.0	78	30	1000	1000	1000	0.7050
4. Nephelinite	43.0	57	40	160	90	1000	0.7050
B. Contaminants							
1. Granite	75.0	5	200	10	5	60	0.7400
2. Tonalite	68.4	5	59	10	10	767	0.7400
3. 5% Partial melt of tonalite	75.0	5	104	5	5	283	0.7400
4. Witkoppen mica	—	—	727	—	—	19.1	5.589
C. Bulk Distribution Coefficients							
1. Basalt			0.01	2, 6*	3, 8*	1.0	
2. Picrite basalt			0.01	2, 8*	4, 6*	0.1	
3. Nephelinite			0.01	8, 12*	6, 8*	0.1	

The above parameters are based on data from Allsopp (1961), Condie and Hunter (1976), Glikson (1979), Barton *et al.* (1980), Bristow (1984a, b), Cox and Bristow (1984) and Allsopp (unpublished data). The 5% partial melt has been calculated assuming equilibrium batch melting of a source with 15% hornblende, 15% biotite, 45% plagioclase, 5% magnetite, 10% K-feldspar and 10% quartz (see Reid and Barton, 1983). Distribution coefficients, from Arth (1976), for Rb and Sr respectively are: Hornblende 0.1, 0.4; Biotite 3, 0.1; Plagioclase 0.1, 5; K-feldspar 0.3, 5; Magnetite 0, 0; and Quartz 0, 0.

* Note that 2,6; 2,8; 8,12 refers to distribution coefficients of 2 and 6, 2 and 8, 8 and 12 etc.

TABLE V
Variables used for Calculating Contamination and Combined Contamination—Crystallization Trends for the Central Area Lavas and Dykes

Model Compositions	SiO ₂ %	Mg number	Rb ppm	Cr ppm	Ni ppm	Zr ppm	Sr ppm	⁸⁷ Sr/ ⁸⁶ Sr
A. Pristine basaltic magma	50.00	59	8	350	120	80	220	0.7050
B. Contaminants								
1. Pronksberg dacite	65.40	42.5	146	54	28	203	320	0.7125
2. Hypothetical granite A	75.00	5	200	10	5	200	60	0.7400
3. Hypothetical granite B	75.00	5	200	10	5	200	60	0.800
4. Hypothetical granite C	75.00	5	200	10	5	200	30	0.800
5. Halfway House mica			1176				66.5	2.664
6. Witkoppen mica			727				19.1	5.489
7. Limpopo Belt mica			430				22.7	0.8874
Bulk distribution coefficients			0.01	2.6*	3.8*	0.10	1.0	

The compositional parameters listed above are based on data from Allsopp (1961), Barton (1979) and Marsh and Eales (1984).

* Note that 2,6 and 3,8 refers to distribution coefficients of 2 and 6 and 3 and 8.

ation the mass ratio : contaminant/crystals formed (the R-parameter of DePaolo, 1981) has been set at 1.0. This is probably a conservative estimate as Nicholls and Stout (1982) have recently emphasized that the heat for assimilation is greater than that released by crystallization and this ratio is probably nearer 0.5. The effect of setting $R < 1.0$ is to shift the calculated contamination/crystallization trend lines to more evolved trace or major element compositions for a given ⁸⁷Sr/⁸⁶Sr ratio.

1. Northern Natal, Lebombo and Nuanetsi Mafic Rocks

A continuous belt of monoclinally warped volcanic rocks crop out along the length of the Lebombo mountains extending from Nuanetsi in south-east Zimbabwe southwards for nearly 800 km to northern Natal (Fig. 1). In this paper the term *southern Lebombo* is used to describe that part of the Lebombo range between the northern Swaziland border and the southernmost outcrops of volcanic rocks in Natal. The *central and northern Lebombo* is that area lying between the northern Swaziland border and the Limpopo River. This subdivision has some basis in that there are lateral compositional variations in the mafic volcanic rocks along the Lebombo monocline and rocks from the northern Lebombo are geochemically distinct from those to the south (Cox and Bristow, 1984). The *Nuanetsi* area lies to the north of the Limpopo River.

The volcanic succession in the Lebombo belt consists of a bimodal sequence of mafic rocks (basalts of the Sabie River Formation, picrite basalts of the Letaba River Formation and minor nephelinites of the Mashikiri Formation) and rhyolites (most of which form the Jozini and Mbuluzi formations) and an associated suite of intrusive rocks (Cleverly and Bristow, 1979). A major dolerite dyke swarm, the 200 km long Rooi Rand swarm, intrudes the western portion of the basalt outcrop in the southern Lebombo region (Cleverly and Bristow, 1979; Saggerson *et al.*, 1983). A further group of dykes and sills, referred to as the Effingham dolerites (Frankel, 1969; Absalom, 1970), crops out in the region between Durban and the southern end of the Lebombo belt and are also discussed in this section. A succession of basalts (the Movene basalts) overlie the rhyolites, but these rocks mostly crop out in Mozambique and only two samples were analysed in this work. A broadly similar succession of mafic rocks, rhyolites and dyke rocks occur in the Nuanetsi and adjacent areas of south-east Zimbabwe.

(a) *Effingham Dolerites*. — The Effingham dolerites are considered to be slightly younger than the normal Karoo dolerites and Lebombo lavas (Bristow, 1976). They are characterized by a distinctly evolved chemistry in relation to the mafic lavas and dykes of the Lebombo (Table III), and on the basis of petrographic, major element and Sr-isotope data, Frankel (1969) concluded that these rocks had been produced by the assimilation of sialic rocks into

differentiated iron-rich basic magmas at depth. Field observations support this conclusion since the intrusives are commonly found to carry abundant crustal xenoliths which range in size from small xenocrysts to fragments several centimetres in diameter.

In this section we review and re-examine Frankel's (1969) data for these rocks as we regard them as the most unequivocal example of bulk contamination of Karoo mafic magma. The available isotopic data are presented in Appendix A but it should be borne in mind that these data were obtained prior to the advent of high-precision mass spectrometry and that Rb and Sr were determined by atomic absorption spectrophotometry. Also, important elements such as SiO₂ are not available for the rocks which were analysed for ⁸⁷Sr/⁸⁶Sr, and no trace element analyses other than for Rb and Sr are available.

Although available data do not provide a tight constraint on model calculation, we have attempted to fit a simple two-component mixing model to the data and our results are presented in Fig. 3 and Table III. The best fit mixing curve to the ⁸⁷Sr/⁸⁶Sr data is obtained by assuming a granitic contaminant with ⁸⁷Sr/⁸⁶Sr = 0.740 and Sr = 60 ppm and a basaltic precursor with ⁸⁷Sr/⁸⁶Sr = 0.7080 and Sr = 366 ppm. The characteristics of the granitic component are within the range reported for such rocks from the Archaean terranes of Swaziland and northern Natal. With regard to the basalt precursor the ⁸⁷Sr/⁸⁶Sr ratio is higher than the R₀ ratios reported for the Rooi Rand and other Karoo dolerites in northern Natal but is close to the average for southern Lebombo basalts (Fig. 3; Table III) as is the Sr content.

An additional test of this model is provided by the results of adding varying proportions of the granite contaminant to the average composition of the southern Lebombo basalts (Table III). The calculated compositions are compared with analyses of Effingham dolerites reported by Frankel (1969). There is close agreement between the model and actual data for SiO₂, MgO, Al₂O₃, Na₂O and K₂O but the match for the oxides TiO₂, Fe₂O₃, CaO and P₂O₅ is poor probably because concomitant fractional crystallization has not been considered. Overall, however, both the isotope data and the major element mass balance calculations yield similar results in terms of the proportion of contaminant (65–70%) necessary to produce the most evolved rocks (in terms of SiO₂ and ⁸⁷Sr/⁸⁶Sr ratio).

Some doubt may exist that a R₀ ratio of 0.7080 could be that of an uncontaminated basalt magma. In later sections we demonstrate that such ratios are characteristic of some uncontaminated Karoo basalts. Furthermore, Hoefs *et al.* (1980) have, on the basis of a combined ⁸⁷Sr/⁸⁶Sr and δ¹⁸O study of the Kirkpatrick basalts of Antarctica, identified a similarly high R₀ ratio (0.7090) as being representative of uncontaminated Kirkpatrick basalts. Although more precise isotope and trace element data are needed to refine the models discussed here, they do serve to indicate that

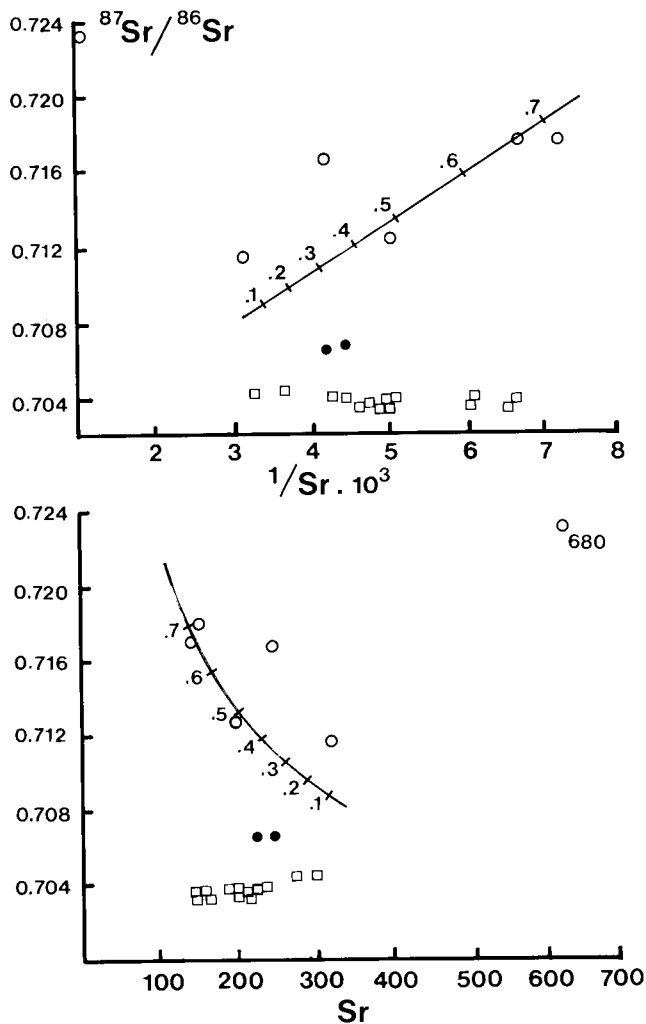


Figure 3

Plots of the R_0 ratio vs $1/\text{Sr} \cdot 10^3$ and Sr (ppm) for the Effingham dolerites (open circles), Rooi Rand dolerites (open squares) and Karoo dolerites (solid circles). Mixing curves have been calculated assuming the contaminant was a granite with $\text{SiO}_2 = 75\%$, Sr = 60 ppm and $^{87}\text{Sr}/^{86}\text{Sr} = 0.740$ and the parental basaltic magma had an R_0 ratio of 0.7080, and Sr = 366 ppm. Numbers on mixing curve represent fraction of granite component in mixture.

significant compositional variation in addition to R_0 ratio variations will accompany bulk contamination of mafic magmas.

(b) *Southern Lebombo Basalts and Dolerites.*—The substantial range in R_0 ratios coupled with high silica contents (up to 58%; see Cox and Bristow, 1984) in some of the southern Lebombo basalts and dolerites (excluding the Rooi Rand), suggests that these rocks may have crystallized from magmas contaminated by crustal material. As outlined previously, this proposal has been tested using the parameters listed in Table IV and the results shown graphically in Fig. 4. The various contaminants are identified according to the numbers indicated in Table IV, and in the Cr and Ni plots curves for different values of D^{bulk} are identified. The assumed R_0 ratio of the uncontaminated magma is 0.7040, which is marginally higher than the mean value for the Rooi Rand dolerites which have the lowest R_0 ratios for all the mafic rocks in the Lebombo.

The exposed granitoid rocks (granites, trondhjemites, tonalites, etc.) of the Archaean terranes of northern Natal and Swaziland are the most likely candidates for contaminants and estimated compositions and the sources for these data are given in Table IV. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of these contaminants at 190 m.y. have been calculated from typical whole rock Rb/Sr and initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios—that for the

tonalite being at the upper end of the range. Contamination by a partial melt of the tonalites and by Sr from an Archaean mica are also evaluated. Contaminants not considered are the considerable volumes of acid volcanics overlying or interbedded with the basic rocks. Their crustal origin is debatable and in any case, except for the late-stage Kuleni rhyolites and the volumetrically unimportant Mkutshane felsic rocks, their R_0 ratios are < 0.7050 (see Section VII).

Examination of Fig. 4 indicates that model bulk contamination trends fail to account consistently for the high R_0 ratios and the major and trace element chemistry of the basalts. Specifically the contaminant should have, or, if fractional crystallization is involved, should allow the contaminated magma to evolve to compositions with high $^{87}\text{Sr}/^{86}\text{Sr}$, Sr, Mg-number and Cr, and low SiO_2 and Rb. These compositional characteristics are unlike any crustal rock or partial melt of crustal rocks or any liquid produced by fractional crystallization of basaltic magmas. In particular the plot of Sr vs R_0 ratio illustrates clearly that contamination cannot produce variable whole rock Sr contents and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in basaltic rocks crystallizing plagioclase unless large degrees of contamination are involved and the contaminant is also Sr-rich. High Mg-numbers are also difficult to sustain with fractional crystallization accompanying contamination.

Even allowing for some degree of contamination to account for the more evolved southern Lebombo basalts, there are samples with high and low R_0 ratios which have similarly primitive major and trace element compositions. These primitive characteristics (low SiO_2 , Rb; high Mg-number, Cr) could not have survived the large degrees of assimilation (up to 40% depending on the contaminant; Fig. 4) required to explain their $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. This is strongly suggestive that the range of R_0 ratios (up to 0.713) exhibited by the southern Lebombo mafic rocks is not a consequence of bulk contamination.

As illustrated in Fig. 4, the Rooi Rand dolerites all have similarly low R_0 ratios (0.7035–0.7042) despite exhibiting a considerable range in major and trace element compositions indicative of fractional crystallization (Armstrong, 1978; Armstrong *et al.*, 1984). Their R_0 ratios are amongst the lowest measured for any rock type in the Lebombo and contrast sharply with the high R_0 ratios of the basalts. These Sr-isotopic characteristics are consistent with the overall "depleted" compositional features of the dykes as exhibited by Nd-isotopic composition and low incompatible elements abundances (Hawkesworth *et al.*, 1984a). It is clear that the Rooi Rand dykes represent relatively uncontaminated magmas from a depleted (possibly asthenospheric?) mantle source (Saggerson *et al.*, 1983).

(c) *Central and Northern Lebombo and Nuanetsi Basalts and Dolerites.*—Relatively few $^{87}\text{Sr}/^{86}\text{Sr}$ data are available for these dolerites and basalts (Sabie River Formation basalts and the Nuanetsi Interbedded basalts). Samples were collected from a wide area and not all have been analysed for both major and trace elements. Relevant data are presented in Fig. 5 together with model contamination trends calculated from parameters in Table IV. Apart from the Nuanetsi Interbedded basalts, these mafic rocks display a more restricted but significant range in R_0 ratios (most lie between 0.704 and 0.706), higher Rb and especially Sr, and lower Cr concentrations than those of the southern Lebombo. As a group they also exhibit a positive correlation of R_0 ratio with Rb, Ni, Sr and Mg-number and this and other aspects of their isotope and whole rock compositions cannot be accounted for by a consistent model of bulk contamination.

The Interbedded basalts from Nuanetsi have much higher R_0 ratios and superficially appear likely candidates for being contaminated. However, more data on all these

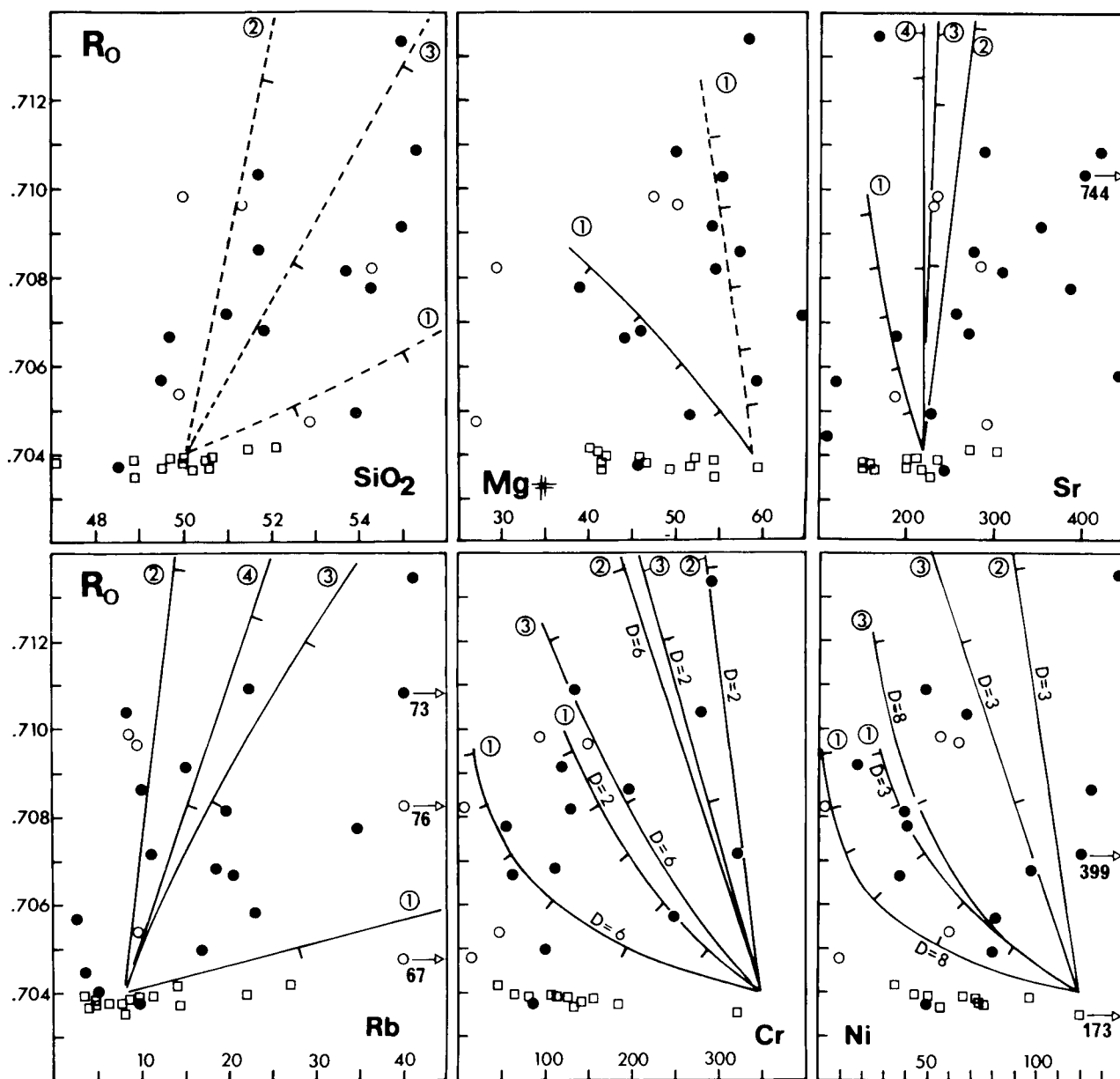


Figure 4

Variation of the R_o ratio with SiO_2 , Mg-number, Sr, Rb, Cr and Ni for southern Lebombo lavas, dykes and Rooi Rand dolerites compared to calculated contamination trends. Dots—basalts; open circles—dolerites; squares—Rooi Rand dolerites. Dashed curves indicate simple mixing between contaminants and basic magma. Continuous curves indicate combined fractional crystallization-contamination trends calculated from the data in Table IV. For trace elements the equations of DePaolo (1981) were used assuming $R = 1.0$. For SiO_2 and Mg-number contamination crystallization trends were estimated from mass balance calculations assuming removal of plagioclase, clinopyroxene and olivine in the proportions 50:40:10 and mass of contaminant = mass of crystals removed. Ticks on curves mark each 10% contaminant added to the magma except for trend 4 where ticks are for each 1% contaminant.

basalts are needed before any unequivocal conclusions can be drawn.

(d) *Nephelinites and Picrite Basalts of the Northern Lebombo-Nuanetsi.*—The Mashikiri nephelinites crop out as a thin (70 m) succession in the northernmost portion of the Lebombo and in the Sabie River area of south-east Zimbabwe (Bristow, 1984a). They form the base of the volcanic succession and are overlain by picrite basalts of the Letaba River Formation which, in contrast, are voluminous, attaining a thickness of some 4 km in the northern part of the Lebombo monocline (Bristow, 1984b). The Letaba Formation is in turn overlain by basalts of the Sabie River Formation discussed above. The nephelinites are unexceptional in their compositional character but the picrite basalts are unusual in being variably but generally highly enriched in incompatible elements and having strongly fractionated REE patterns (Cox *et al.*, 1984).

The R_o ratios for both picrites and nephelinites are presented in Fig. 6. Most samples have R_o ratios in the range 0.7045 to 0.7060 which is very similar to that of the overlying basalts (see previous section). The picrite basalts exhibit a considerable range in composition and in the element vs R_o ratio plots in Fig. 6 are disposed along generally horizontal trends. By comparison with the modelled trends in Fig. 5 it is clear that such compositional arrays could arise through bulk contamination-fractional crystallization provided that the contaminant was granitic in character with low Sr of the order indicated in Table IV. However, because of the high Sr content of the picrites such contamination (with concomitant fractional crystallization) has to be substantial (40–50%) in order to account for the range of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios displayed by these rocks. It is therefore unlikely that the variable R_o ratios of the picrite basalts reflect the influence of crustal contamination.

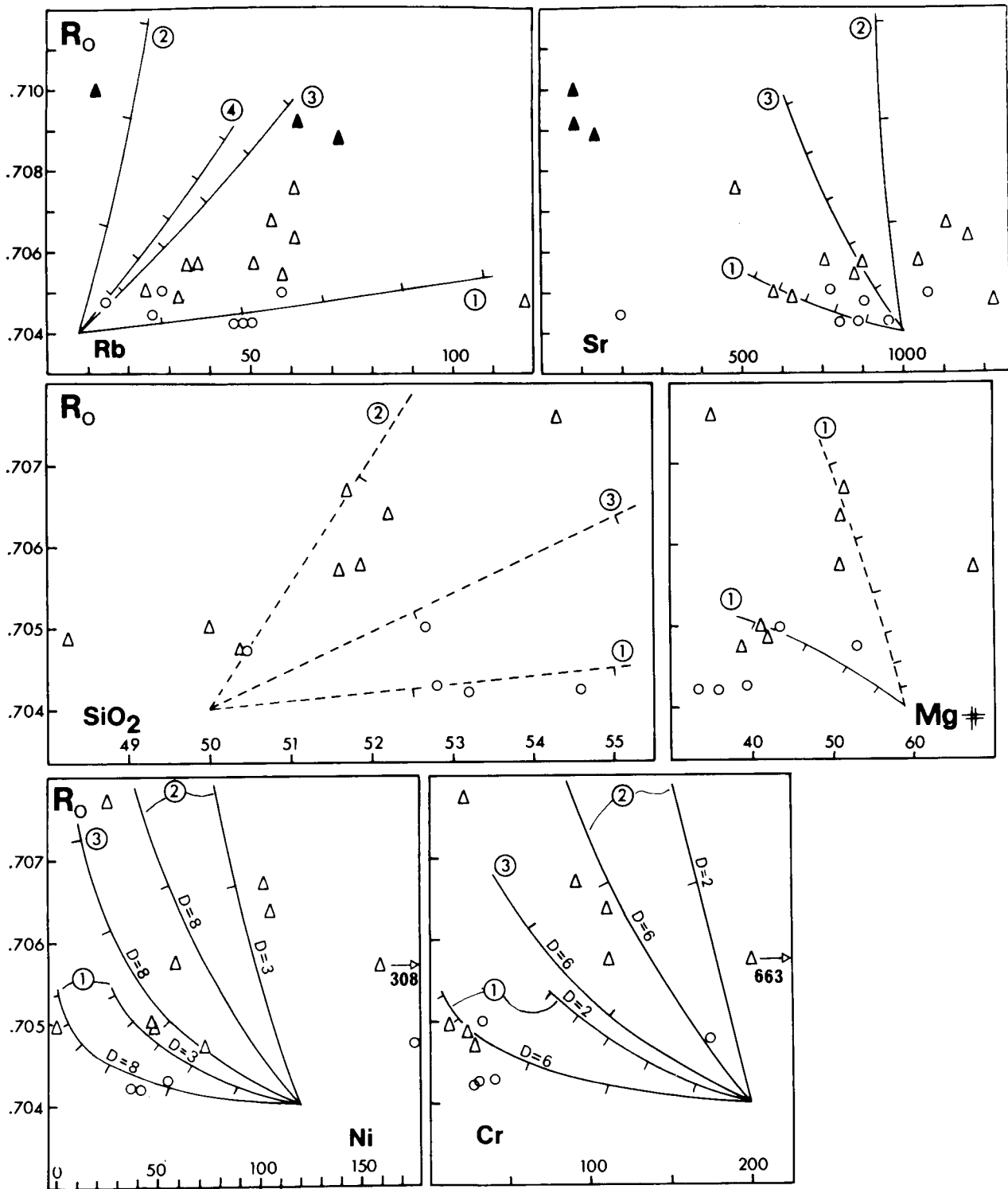


Figure 5

Variation of the R_o ratio with SiO_2 , Mg-number, Sr, Rb, Cr and Ni for northern Lebombo and Nuanetsi basalts and dolerites. Open triangles — basalts of the Sabie River Formation; solid triangles — Interbedded Basalts of Nuanetsi; circles — dolerites. Composition trends for contamination have been calculated as for Fig. 4.

Similar arguments probably also apply to the nephelinites as they are, in general, even more enriched in Sr than the picrite basalts. Two samples from the Pafuri area of the Lebombo have much higher R_o ratios than four other analysed samples. If these high R_o ratios indicate bulk involvement of crust then Fig. 6 suggests that this involvement must be considerable. Such a proposal is not consistent with the relatively high Cr content and Mg-number (compared to other nephelinites with low R_o ratios) and their highly silica undersaturated character. As with the picrite basalts we conclude that the range of R_o

ratios exhibited by the nephelinites do not reflect bulk contamination, but more data are needed for both these rock types to fully substantiate these arguments.

2. Central Area Basalts and Dolerites

Samples from the Central area of the Karoo Province were collected from intrusive and volcanic suites scattered through an area 750,000 km² which covers much of the central part of southern Africa. The main volcanic remnant forms the highlands of Lesotho where the basalts attain a thickness of about 1.4 km. Much of this thickness is

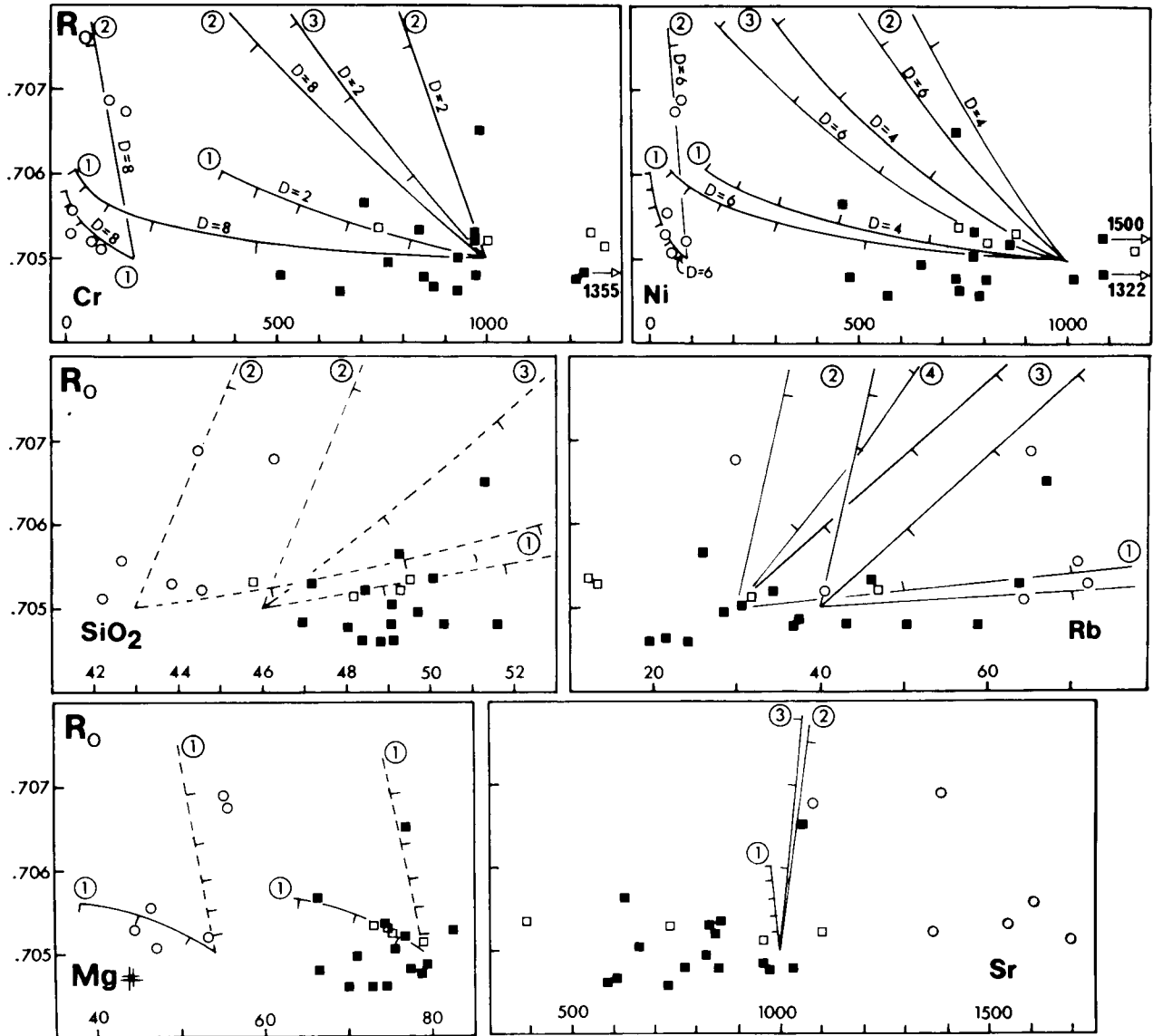


Figure 6

Variation of the R_0 ratio with SiO_2 , Mg-number, Rb, Sr, Cr and Ni for northern Lebombo and Nuanetsi nephelinites (circles), picrite basalts (solid squares) and picrite intrusives (open squares). Contamination trends calculated as outlined in Fig. 4, except that crystallization-contamination trends for SiO_2 and Mg-number were estimated by mass balance removal of 40% orthopyroxene: 60% olivine (picrite basalts) and 80% clinopyroxene: 20% olivine (nephelinites).

dominated by flows of the Lesotho basalt type but smaller volumes of other basalt types form the base of the succession in the north-east Cape where they are interbedded with minor flows of andesite-dacite types (Marsh and Eales, 1984). Other major volcanic remnants are in the Springbok Flats and near Mariental in south-central Namibia. Large parts of the intervening area are covered by thick sequences of Karoo sedimentary rocks and their intrusive suites of dolerite dykes and sills. The major part of this huge volume of intrusive dolerite is compositionally identical to the Lesotho basalt type.

R_0 ratios for the Central area mafic rocks vary from 0.7046 to 0.7094, with the majority in the range 0.7046 to 0.7070. There appears to be no correlation between R_0 ratio and the age of the continental crust through which the magmas were emplaced (Fig. 7) bearing in mind that the Central area is underlain by basement complexes of three distinct ages, the Archaean Kaapvaal Craton, the 1.2–2.0 b.y. Kalahari and Richtersveld provinces, and the Namaqua–Natal Mobile Belt which is 1.0–1.2 b.y. in age (Tankard *et al.*, 1982). Brooks and Hart (1978) previously suggested that higher R_0 ratios in Gondwana volcanics were

characteristic of mafic rocks emplaced through younger mobile belts but the data presented here do not support this proposal.

Since the R_0 ratios are higher and more variable than those considered by many as being typical of mantle-derived magmas, we have tested bulk contamination models adopting an approach similar to that employed with the Lebombo volcanics. The compositional parameters used to model contamination trends are summarized in Table V and the results compared graphically with measured data in Fig. 8. For the choice of contaminants there are two considerations. The only preserved silicic lavas or intrusives emplaced contemporaneously with the basalts are volumetrically trivial andesites and dacites of the north-east Cape which have high R_0 ratios (0.7094–0.7130), corundum in the CIPW norm and are undoubtedly crustal melts. They are prime contenders for the role of contaminants although they are not widespread. Alternatively it is possible that the contaminating melts are not represented in the volcanic suite because they have been mixed in with the basic magmas. In the calculations these contaminants are assumed to be granitic in

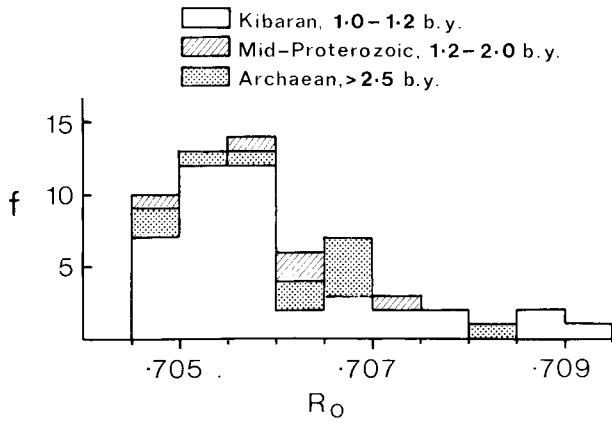


Figure 7

Histogram of R_0 ratios for Central area basalts and dolerites in relation to age of the underlying basement.

composition and the same compositional parameters used for granites in the Lebombo contamination calculations have been used.

Evaluation of Fig. 8 indicates that the contamination trends for the granite and dacite overlap considerably except for model 3 which combines very high $^{87}\text{Sr}/^{86}\text{Sr}$ with low bulk distribution coefficients for Cr and Ni. However, it is clear that a number of samples with high Mg-numbers (>55.0) and low SiO_2 (<51.5%) exhibit R_0 ratios in the range 0.7050 to 0.7080 which cannot be explained by contamination with dacite or granitic rocks. Similar relationships, i.e. samples with low Zr and Rb and high Cr and Ni coupled with high R_0 ratios, are evident in the trace element plots. Note also that the data for an individual sample do not have the same relationships to a given model trend in all plots thereby emphasising the necessity of using several elements to test for contamination. Collectively, the plots indicate that a substantial number of basalts and

dolerites have high R_0 ratios coupled with primitive chemical characteristics and the elevated R_0 ratios cannot be explained by contamination with dacitic or granitic crustal melts. For the samples whose compositions do appear to be accommodated by the dacite and granite contamination trends, contamination has varied up to 30% accompanied by similar amounts of fractional crystallization. Yet these samples have identical major and trace element compositions compared to samples with both higher and lower R_0 ratios. We regard this as compelling evidence against bulk contamination of these basalts. Furthermore, if contamination has affected the Sr-isotope and bulk compositions of all Central area basic rocks, then their uncontaminated parental magmas must have had a wide spread of R_0 ratios possibly ranging up to 0.707 or greater.

The effect of preferential contamination by melts of mica-rich assemblages is indicated on the Rb plot where it is evident that contamination by less than 2% Sr from old micas leads to rapid elevation of R_0 ratios without significant change in the major and trace element composition in the contaminated magma. Although such a process operating with or without attendant or subsequent contamination by granitic melts can explain high R_0 ratios in specific mafic rocks, it is unlikely to have caused the general elevation of R_0 ratios is some 10^6 km^3 of basalt erupted over the vast areas in the Central area.

Marsh and Eales (1984) have summarized the geological evidence against bulk crustal contamination having a significant role in generating compositional variation in Central area basalts and dolerites. Taken together with the isotopic data examined here, we conclude that although we cannot discount bulk contamination playing some role in the generation of basalts with high R_0 ratios, the weight of evidence is against it.

3. Etendeka Basalts and Dolerites

The Etendeka Formation volcanics of north-western

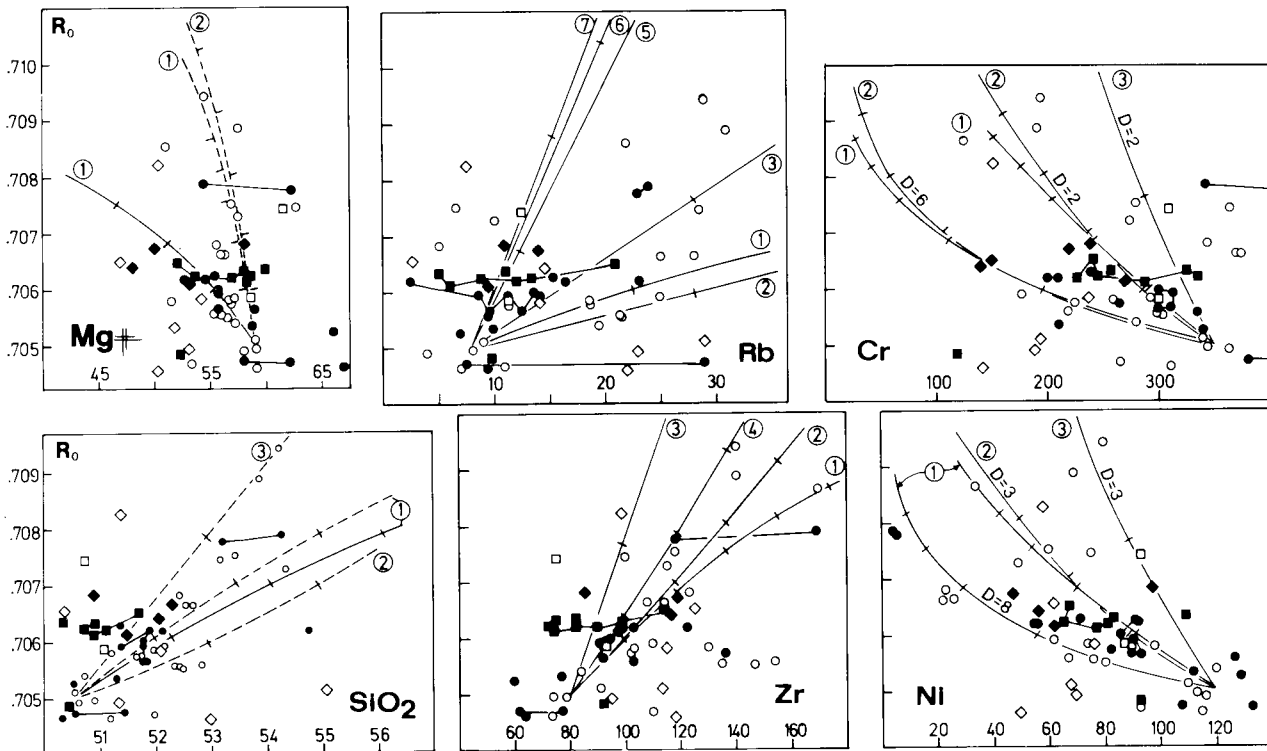


Figure 8

Variation of the R_0 ratio with SiO_2 , Mg-number, Cr, Ni, Rb and Zr for Central area basalts and dolerites. Squares—samples underlain by 1.2-2.0 b.y. basement; diamonds—samples underlain by Archaean basement; circles—samples underlain by 1.0-1.2 b.y. basement. Open symbols—lavas; solid symbols—intrusives. Tie lines join samples from the same intrusion. Contamination trends calculated as indicated in Fig. 4 using data in Table V. Ticks on trends 5, 6 and 7 are for each 1% contamination.

South West Africa/Namibia differ from all other Karoo volcanics in terms of their age, mineralogy and chemical and isotopic relationships (Erlank *et al.*, 1984). They comprise early Cretaceous lavas and associated minor intrusives, and swarms of regional dolerite dykes, possibly of late Jurassic age. The bulk of the lavas comprise a suite of aphyric basalts and evolved basalts (termed the Tafelberg type) interbedded with two important quartz latite units and a thin latite unit of limited extent. The relationship is best shown at the Tafelberg beacon, on the eastern edge of the largest lava remnant where the eroded lava pile reaches a thickness of 900 m; many of the samples analysed are from the Tafelberg locality. Along the coast, plagioclase-phyric basalts (termed the Albin type) underlie and are partly interbedded with the Tafelberg type basalts, and although related to the Tafelberg rocks they tend to be less evolved. The Albin basalts are intruded by thin, aphyric dykes and sills, the Horingbaai dolerites, which along with the Rooi Rand dolerites exhibit a distinctive "primitive" chemistry in comparison to all other Karoo basic rocks. Dolerite dykes from the regional swarms which cut basement and Karoo sediments, but not the lavas, are more evolved than the Horingbaai intrusives, but their precise geochemical relationship to the Albin and Tafelberg lavas, and their age, is not clear at this stage (Erlank *et al.*, 1984).

Data for the Etendeka volcanics have earlier been presented in summary form in Table I and Fig. 2d but are now shown in more detail in Fig. 9. Note that all R_0 ratios are calculated to 121 m.y. (Erlank *et al.*, 1984) and not 190 m.y. as in the case of the samples from other Karoo areas although there are indications that the true age of the Etendeka volcanics is of the order of 130 m.y. (Allsopp *et al.*, 1984a). Inspection of Fig. 9 shows a distinct grouping with respect to R_0 ratios for the individual types recognized, viz. Horingbaai intrusives (0.7030–0.7034), regional intrusives (0.7065–0.7068), Tafelberg and Albin basaltic rocks (0.7081–0.7135), latites (0.7152–0.7154) and quartz latites (0.7186–0.7218, with one low value of 0.7137). Thus the variations in R_0 ratio are unlike those previously discussed for the Lebombo volcanics, where the R_0 ratios of the felsic volcanics are relatively low (0.704–0.705) and

within the lower part of the range encountered for the basic rocks.

The quartz latites are the only Etendeka rocks (taking Rb/Sr variations into consideration) which in any way show a reasonable Rb-Sr age relationship, although regression treatment yields an errorchron with an older apparent age of ~154 m.y. (Allsopp *et al.*, 1984b). This does suggest that some of the samples may have had their $^{87}\text{Sr}/^{86}\text{Sr}$ ratios lowered by introduction of zeolitic material (see discussion in Section III) but the extent of possible change (even assuming a slightly older age for the Etendeka lavas) will not alter the arguments presented below.

(a) *Inter-relationships Between Basaltic Rocks, Latites, and Quartz Latites.* One of the most important features of the Etendeka Formation volcanics requiring interpretation concerns their possible inter-relationships. This is the subject of a separate paper in this volume (Erlank *et al.*, 1984) where the rock types similar to those occurring at the Tafelberg locality, viz. the Tafelberg basalts and evolved basalts and the interbedded latite and quartz latite units, are discussed in detail. Because of the high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the basaltic rocks and their intimate relationship with silicic rocks characterized by high R_0 ratios indicative of probable crustal origin, their inter-relationships bear directly on the question of whether the high R_0 ratios of the basaltic rocks reflect bulk crustal contamination. The compositional and isotopic data available for these rocks is probably more comprehensive and representative than those for any other suite in the Karoo Province and the approach adopted in the sections that follow differs from that used in the Central and Lebombo areas for examining the contamination question.

The three rock types mentioned above exhibit coherent between-type trends, a feature re-inforced by the data shown in Fig. 9 where a possible mixing relationship between the three types is depicted. Accepting the quartz latites as being crustal melts (see the following section) then the dashed mixing line in Fig. 9 could be indicative of bulk crustal contamination such that the more evolved basalts are most contaminated. In this context it is pertinent to note that the majority of the quartz latites (many not analysed for $^{87}\text{Sr}/^{86}\text{Sr}$) have Rb/Sr ratios within the range of 1.0–1.4. They are also demonstrably the most likely crustal contaminant in view of their presence *per se*, and especially in view of their interbedded nature. Support for such a contamination process is provided by a general increase in SiO_2 content when passing from the least evolved to the most evolved basalts, e.g. samples 42 and 98 in Fig. 9 have the highest SiO_2 and incompatible element contents of all the Tafelberg basaltic rocks. The relationship between SiO_2 and R_0 ratio in the basaltic rocks is similar to that observed by Faure *et al.* (1972, 1974) who ascribed such a relationship to simple mixing with crustal material.

However, there are severe difficulties which argue against this possibility. These have been documented in detail by Erlank *et al.* (1984) where an integrated elemental and isotopic approach has been used to evaluate the possible petrogenetic inter-relationship between these rocks. Erlank *et al.* (1984) concluded that the basalts, latites and quartz latites are not genetically related to each other, and that bulk contamination by the quartz latites, or any other average type of crustal material, cannot be responsible for the range of elemental and isotopic variation in the basaltic Tafelberg lavas. Their arguments supporting these conclusions can be briefly summarized as follows.

A variety of interelement relationships (e.g. TiO_2 vs Zr; Mg-number vs SiO_2) do not allow for simple mixing between the three magma types but do allow for mixing followed by fractional crystallization. However, in a number of plots involving R_0 vs various incompatible elements such a model fails to account consistently for

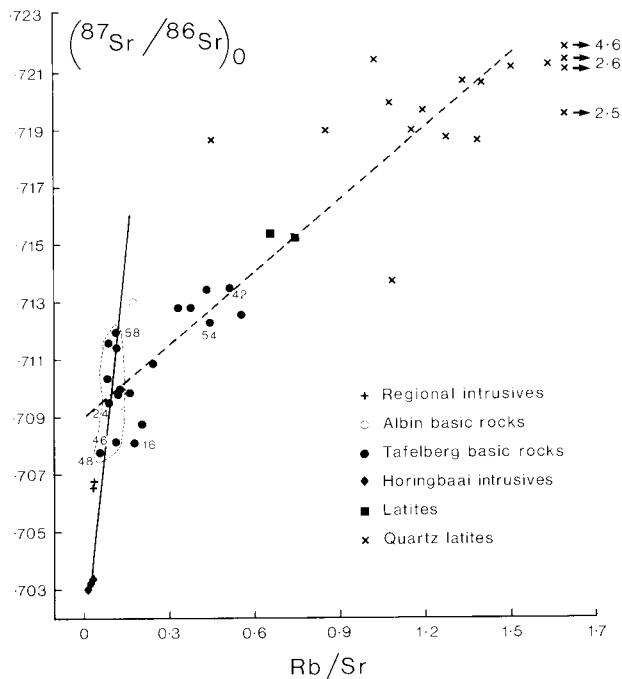


Figure 9

Plot of the R_0 ratio vs Rb/Sr for the Etendeka volcanics. Note that all samples numbered on the diagram are prefixed KLS in Appendix A.

the observed compositional variations. Specifically, the fractional crystallization process is required to produce maximum enrichment of Ti and V in the evolved basalts but almost no enrichment of Ba and Rb. We do not know of any fractional crystallization process that can operate in this way in basaltic rocks.

These arguments are strongly supported by other isotopic and geological data. For example, the latites have the lowest $^{143}\text{Nd}/^{144}\text{Nd}$ initial ratios (Hawkesworth *et al.*, 1984a) and cannot have the same relationship to the basalts and quartz latites as that suggested by the $^{87}\text{Sr}/^{86}\text{Sr}$ data. Model calculations suggest that the most evolved basalts would contain about 55% of the Nd and Sr of the quartz latites. Considering the regional distribution and abundance of the evolved basalts the scale of contamination suggested by these data is enormous and, moreover, it has to operate in such a way as to produce systematic contamination trends of a highly unusual nature.

For all of the above reasons Erlank *et al.* (1984) have concluded that no genetic or mixing relationship exists between the Tafelberg (and Albin) basalts, and the latites and quartz latites.

(b) *Inter-relationships Between the Tafelberg Basalts, Regional Dolerites and Horingbaai Dolerites.* If the question of the derivation of the Tafelberg basaltic rocks is divorced from that of the latites and quartz latites then we are still left with the problem of accounting for the high SiO_2 and incompatible element contents of the evolved basalts and, more importantly, their high Fe_2O_3 and V contents (Erlank *et al.*, 1984). It might be argued that the compositional trends in the basalts and evolved basalts are indicative of a simple mixing relationship, but if this is the case then the implied contaminant must not only have high SiO_2 and incompatible element concentrations, but Fe_2O_3 and V contents substantially in excess of the concentrations in sample 42 (13.9% Fe_2O_3 and 370 ppm V). Such hypothetical contaminants cannot be seriously considered. Moreover, the operation of a simple mixing process, without concomitant fractional crystallization is also unlikely. The most obvious explanation is to consider that the mineralogical and chemical trends in the Tafelberg (and Albin) basaltic rocks are due to fractional crystallization. However, in order to accommodate the isotopic variations in these rocks it is necessary to postulate that the evolved basalts have been derived from basic parental magmas which had a range of isotopic compositions (Erlank *et al.*, 1984). These authors have further shown that quantitative major and trace element calculations for such a fractional crystallization model are in agreement with the derivation of the more evolved from the less evolved Tafelberg basic lavas. These conclusions can be conceptually understood by reference to Fig. 9, where the more primitive basalts 46 and 58 can be envisaged as being the type of parental magmas from which the evolved basalts 16 and 54 are respectively derived.

Having accounted for the derivation of the evolved basalts as indicated above, it is now necessary to provide an explanation for both the range and high values of R_o ratios in the more primitive Tafelberg lavas. These have been arbitrarily chosen as those which have Mg-numbers >55 , and are enclosed by the dotted line in Fig. 9. At this stage it is also necessary to consider the two regional dolerites and the three Horingbaai minor intrusives all of which have lower R_o ratios (Fig. 9) than the Tafelberg and Albin basic rocks. The solid line joining the Horingbaai intrusives, the regional dolerites and the more primitive Tafelberg lavas is again suggestive of a mixing relationship.

Even though the quartz latites are demonstrably the most likely possible crustal contaminants, as argued earlier, it is nevertheless appropriate to consider alternative crustal contaminants in more detail than Erlank *et al.* (1984) have done. A large amount of Rb-Sr isotopic data are available

for crustal materials in north-west Namibia and these have been used in compiling Fig. 10, which is an expanded version of Fig. 9. The outlined fields shown are for Pan-African Damaran metasediments and igneous rocks,

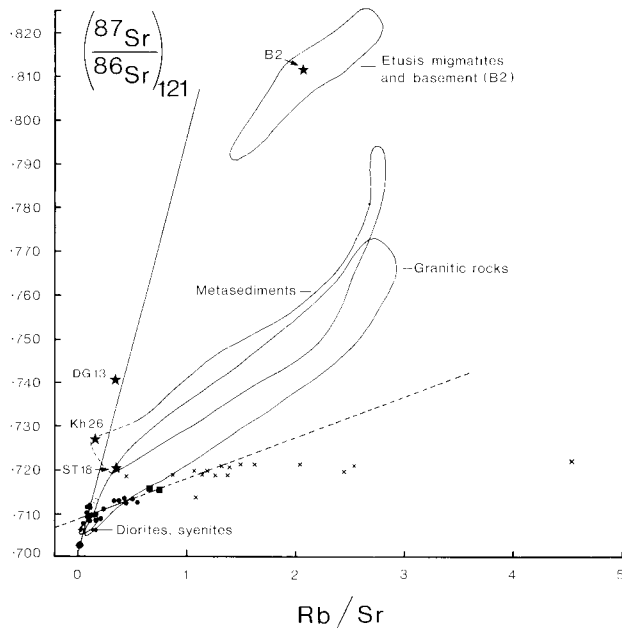


Figure 10

Condensed version of Fig. 9 to allow inclusion of data for Damaran diorites, syenites, granitic rocks and metasediments, Etusis migmatites and one sample of 2.0 b.y. basement (B2—Abbabis granite). All R_o ratios calculated to 121 m.y. Legend for Etendeka volcanics as in Fig. 9. Samples DG13, Kh26 and ST18 are Damaran metasediments with lowest $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, the latter two samples being the only ones which plot within the dashed area of the envelope for the metasediments. Data for pre-Etendeka rocks from Blaxland *et al.* (1979), Downing and Coward (1981), Hawkesworth *et al.* (1981), Haack *et al.* (1982), Hawkesworth *et al.* (1984b) and Hawkesworth (pers. comm.).

mainly granitic in composition, together with one sample of the ca. 2.0 b.y. basement (Abbabis granite), and rocks which are considered to be melts of basement rocks (Etusis migmatites) on the basis of Nd-isotope compositions (C.J. Hawkesworth, pers. comm.), but which occur within the lowermost Damaran metasediments. Further details of the Damaran and basement rocks may be found in Miller (1984).

It is clear from Fig. 10 that the solid trend line linking the primitive Tafelberg basalts and Horingbaai dolerites does not readily identify possible contaminants amongst the crustal rocks. The Damaran intermediate igneous rocks (diorites, syenites) have R_o ratios which are too low, while the more abundant granitic igneous rocks have R_b/S_r ratios which are too high, as are the R_b/S_r ratios of the basement rock and presumed basement melts. Apart from three samples (DG13, Kh26, ST18) the great majority of Damaran metasediments also have R_b/S_r ratios which are too high. It should be stressed that the implied mixing line in Fig. 10 demands a contaminant with a relatively low R_b/S_r ratio. The majority of the crustal materials in Fig. 11 would thus appear to be unsuitable contaminants, and we do not favour special pleading such that only the less abundant metasediments with low R_b/S_r have been assimilated into the basic magmas. Resorting to contamination by partial melts of these rocks worsens the situation, since partial melts can be expected to have even higher R_b/S_r ratios. Of course it is possible that lower crustal materials might have suitably low R_b/S_r ratios but these would also be expected to have relatively low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and therefore will also not be appropriate

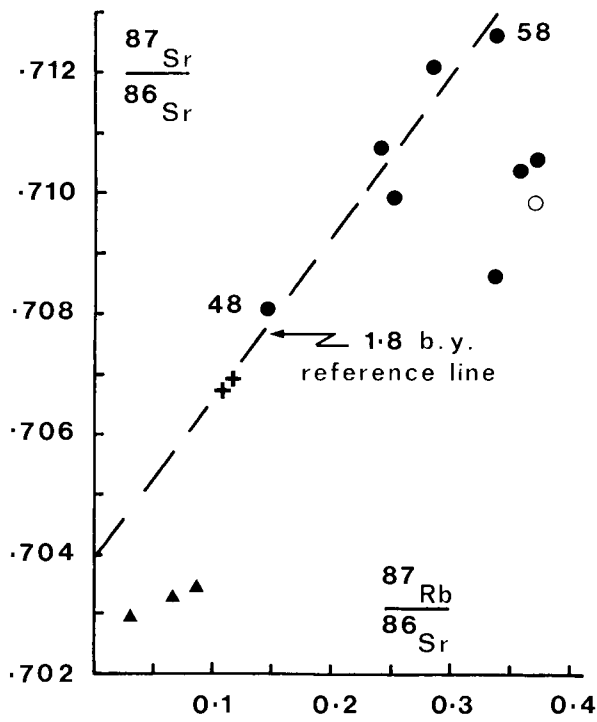


Figure 11

Rb-Sr isochron plot for Etendeka mafic volcanics. Only those samples with Mg-number >55 are depicted. Legend as in Fig. 9.

contaminants. Consequently it does not seem likely that the Horingbaai-regional intrusive-Tafelberg basalt trend in Fig. 10 represents a contamination trend.

B. Selective Contamination

Having evaluated bulk crustal contamination processes in the Karoo Province and found them wanting, we now consider selective contamination, as this process is frequently invoked to explain high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in mafic rocks. As noted in Section IV this process is difficult to test and its applicability to the Karoo rocks is here argued on general grounds.

One of the major constraints on the effectiveness of selective contamination is the very large volume and widespread nature of the mafic rocks that are possibly contaminated. For example, Marsh and Eales (1984) have estimated that in the Central area alone, in excess of $1.0 \times 10^6 \text{ km}^3$ of basaltic magma, the bulk of which appears to have R_o ratios in the range 0.705–0.708, was erupted or intruded as dykes and sills in Karoo sediments. We find it difficult to accept that the diffusion or fluid interaction mechanisms outlined previously could have so thoroughly processed such very large volumes of mafic magmas so as to cause substantial elevation of their R_o ratios to levels observed in the Lebombo, Central and Etendeka areas. It is widely held that magma conduits in the crust will generally have thick linings of frozen basalt which can act as barriers to contamination (Patchett, 1980). Moreover, crustal rocks adjacent to the conduits must soon become depleted in contaminating agents and materials. These probabilities place severe constraints on selective (and bulk) contamination processes. They also suggest that if contamination has operated it should be manifested in the earliest erupted lavas, perhaps as decreasing R_o ratios with height. No such feature has been observed in the Karoo lava suites; in particular, from the southern Lebombo, the uppermost Movene basalts (although only two samples were available for analysis) are not characterized by significantly lower R_o ratios.

Patchett's (1980) thermal modelling indicates that

substantial generation of fluids or partial melts in crustal wall rocks adjacent to magma conduits occurs under conditions prevailing only at lower crustal levels. Because of the large volumes of Karoo magmas which would need to be contaminated, as outlined above, it is pertinent to enquire whether the most favourable site for this, viz. the rocks of the lower crust of southern Africa, have the appropriate isotopic composition. Rogers and Hawkesworth (1982) have shown that lower crustal nodules from Lesotho kimberlites have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios that are mostly lower than the basalts and dolerites we have analysed and could not have contributed to their high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. If this is general for the lower crust beneath southern Africa, this suggests that generation of an effective contaminant would have to be confined to the upper crust where lower ambient temperatures serve to inhibit production of melts or fluids.

Considering now the compositions of selective fluids or melts it is clear that unless gross disequilibrium prevails during their generation and movement through the crust, their compositions must be controlled by the appropriate mineral-fluid or mineral-melt distribution coefficient (D) for the various elements. Data for the former are few but Flynn and Burnham (1978) and Cullers *et al.* (1973) have determined fluid-melt D values for some of the REE. Assuming that the behaviour of Sr is very similar to Eu these data can be used together with feldspar-melt D values to allow a crude estimate of the feldspar-fluid D values for Sr. The results indicate that Sr is strongly partitioned into feldspar, a major phase in crustal rocks, in preference to both melt and fluid phases. Thus selective melts and especially fluids are likely to be rather Sr-poor. Returning to the models for bulk contamination depicted in Figs. 4, 5 and 6 and especially the plots of R_o ratios vs Sr, it is clear that large amounts of crustal Sr have to be exchanged with that in the magmas to produce the necessary changes in $^{87}\text{Sr}/^{86}\text{Sr}$ especially in the Sr-rich mafic rocks of the northern Lebombo-Nuanetsi region. The calculated models are equally applicable to selective contamination as they are independent of the agent of contamination. We consider that the selective fluids are quantitatively incapable of providing the necessary Sr to contaminate such large volumes of mafic magma.

The limitations of diffusion-controlled selective contamination have been outlined by Watson (1982). We wish to emphasize the severe constraints his experimental results place on selective contamination of Sr-rich mafic magmas such as those from the Lebombo by this particular process. The Sr contents of crustal melts must be about 1 to 0.7 times those in the mafic magma before Sr will diffuse into mafic melt. For the Lebombo rocks this means crustal contaminants must have very high Sr contents before this mechanism of selective contamination becomes effective, a feature which clearly inhibits its viability as a major process.

Finally, in both the Lebombo and Etendeka regions, we have demonstrated that the large volumes of silicic volcanics of probable crustal origin have not been involved in any significant contamination interactions with the mafic magmas. It seems most unlikely that the mafic magmas having avoided interaction with these acid melts, should become selectively contaminated by large-scale interaction with fluids of similar selective agents that in all likelihood have been derived from the same crustal segment as the felsic volcanics. We conclude that the variable R_o ratios of the Karoo mafic rocks are unlikely to be accounted for by processes of selective contamination, although individual rocks may obviously show the effects of such processes.

C. Summary

Karoo mafic volcanics and intrusives are characterized by high and variable initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios mostly with values greater than that of Bulk Earth. The largest range is

exhibited by the mafic rock suite of the southern Lebombo and Etendeka regions on the eastern and western continental margins respectively. The range in R_o ratios for rocks from the Central area and the northern Lebombo–Nuanetsi region are more restricted. Significantly rocks with the lowest R_o ratios (<0.7040) are late dykes—the Rooi Rand and Horingbaai dolerites—from the marginal areas. These dolerites also exhibit other isotopic and compositional features indicative of having a depleted source, possibly located in the asthenosphere. In contrast, the vast bulk of Karoo mafic rocks have much higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios which we have attempted to account for by bulk crustal contamination accompanied by fractional crystallization. Contamination models have been tested using both isotopic and major and trace element data. Although contamination can account for some of the chemical character of a number of our samples it fails to explain consistently both the elemental concentrations and isotopic composition of a substantial proportion of the samples from the different areas. Specifically, in all regions there are samples with very similar primitive compositions exhibiting a wide range of R_o ratios; 0.704–0.709 in the Central area, 0.704–0.713 in the Lebombo, and 0.706–0.712 in the Etendeka. Selective contamination is probably also incapable of explaining these data.

We conclude that the above ratios essentially reflect those in the source areas for these rocks and that the Karoo mafic volcanics are derived from the subcontinental lithosphere which has been heterogeneously enriched on both a vertical and horizontal scale. In the Etendeka, although this can readily be accepted for rocks such as the Horingbaai and regional intrusives (see discussion on Central area with respect to lavas of similar composition to the latter), many workers will doubt whether the upper mantle is enriched to the extent implied by the Tafelberg basic rocks. Specifically, a range in R_o ratio of 0.7081–0.7120 is required, as shown by samples 48 and 58 in Fig. 9.

If these conclusions are to be accepted then it is necessary to show that the increase in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios is feasible in terms of realistic Rb/Sr ratios in the mantle sources and time. We use here the Tafelberg basalts as an example because of the high and variable R_o ratios of these primitive basalts which adhere closely to the steep trend line in Fig. 9. This trend line effectively passes through two samples—58 and 48—and allows calculation of the implied age of enrichment in the source on the assumption that both the Rb/Sr and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of these samples are those of the source. A two-point age calculation yields a figure of about 1.8 b.y. with a R_o ratio of 0.704 (Fig. 11). The 1.8 b.y. reference line also passes through the plotted positions of the two regional dolerites. Model age calculations using Bulk Earth parameters for the Rb–Sr system yield slightly older ages of about 2.0 b.y. Thus it is theoretically possible to achieve the required enrichment in the source regions of the Tafelberg basic lavas in some 1.8–2.0 b.y. provided that the Rb/Sr ratio in the source was increased by enrichment to a value of about 0.12 (i.e. that of sample 58). In Section VI we demonstrate that Rb/Sr ratios of 0.1–0.2 are common in metasomatized garnet peridotite xenoliths from southern African kimberlites (Fig. 13) while separated diopsides from such xenoliths have yielded $^{87}\text{Sr}/^{86}\text{Sr}$ ratios as high as 0.713 (Fig. 14). We note that Nd-isotope data for some of these Tafelberg basalts support the existence of enriched mantle in this area (Hawkesworth *et al.*, 1984a). However, since they yield model (T_{CHUR}) Nd ages of around 1.0 b.y., i.e. younger than that inferred from the Sr-isotopic data, it is likely that the pre-history of sub-continental lithosphere in this region is more complex (multi-stage) than portrayed here. For example, lack of isotopic homogeneity in the mantle source regions at the time of enrichment would imply that the reference line in Fig. 11 has no age significance.

VI. KAROO FELSIC VOLCANICS

Substantial volumes of felsic volcanics of Karoo age are restricted almost entirely to the eastern (Lebombo–Nuanetsi) and western (Etendeka) margins of southern Africa. In the Central area only trivial volumes of felsic rocks occur interbedded with the lowermost basalts in the north-eastern Cape and southern Lesotho and they are not considered further in this paper. In the Etendeka and the Lebombo regions two important features characterize the Sr-isotope data of the felsic rocks: they have different R_o ratios than most of the associated mafic rocks and they exhibit isochronous or near isochronous Sr-isotope systematics. There are also important differences between the felsic rocks of Etendeka and the Lebombo and they are therefore discussed separately below.

A. Lebombo Felsic Volcanics

With the exception of a few thin interbedded units (the Mkutshane, Twin Ridge Beds and Olifants Beds, Cleverly, 1977; Bristow, 1982) the voluminous felsic rocks of the Lebombo overlie mafic volcanics. They are referred to collectively as rhyolites and are subdivided into a number of types (Jozini and Mbuluzi rhyolites, the Oribi Beds and Kuleni rhyolites) on the basis of volcano-stratigraphy and petrography (Cleverly and Bristow, 1979). In northern Natal rhyolites and syenites also form a volcanic complex, the Bumbeni complex, which occurs near the top of the main Jozini rhyolite sequence.

Rocks from the Jozini–Mbuluzi, Kuleni and Bumbeni suites define Rb–Sr isochrons with ages of 179 ± 4 , 145 ± 3 and 133 ± 4 m.y. and R_o ratios of about 0.7045, 0.71 and 0.7055 respectively (Allsopp *et al.*, 1984b). For the main Jozini rhyolites which extend along the length of the Lebombo there are small variations in R_o ratios if the isochron regression is calculated on subsets of the data subdivided on the basis of area ($R_o = 0.7044 \pm 2$ for southern Lebombo; $R_o = 0.7048 \pm 1$ for northern Lebombo). The Twin Ridge rhyolites have identical R_o ratios but those of the Mkutshane Beds are much higher ($R_o > 0.715$; see also Betton, 1978, 1979). Rhyolites from Nuanetsi also define an isochron but the R_o ratio (0.7085 ± 6) is again higher than those of most of the Lebombo felsic volcanics.

Most of these features were first observed by Manton (1968) who also noted that in the southern Lebombo the main volume of felsic rocks have lower R_o ratios than the associated mafic rocks (i.e. basalts of the Sabie River Formation) but in the Nuanetsi region the R_o ratios for the felsic rocks lie within the range exhibited by the mafic volcanics. Our new data essentially confirm these observations except that in the southern Lebombo the Rooi Rand dolerites have been found to have R_o ratios similar to those of the Jozini, Mbuluzi and Twin Ridge rhyolites and at Nuanetsi the rhyolites have higher R_o ratios than the underlying basalts and picritic basalts but similar R_o ratios to the interbedded basalts with which they interdigitate.

On the basis of their low R_o ratios, a crustal origin for the main volume of rhyolites in the southern Lebombo has generally been rejected. Manton (1968) suggested a mantle origin for these rhyolites and the model of Betton and Cox (1979) derives the rhyolites by melting young (early Karoo) basaltic material underplated on to the base of the crust. In addition Cleverly *et al.* (1984) have developed models whereby Jozini and Mbuluzi rhyolites can be produced by partial melting of a gabbroic source represented by a typical Rooi Rand dolerite composition. Nd-isotopic data (Hawkesworth *et al.*, 1984a) are not consistent with this view even though Fig. 12 depicts a linear relationship between the data for the Rooi Rand dolerites, the Jozini rhyolites and the Bumbeni Complex on a Sr-isotope evolution diagram. This linear relationship was initially considered to strongly support the model of Betton and Cox (1979). However, the Rb/Sr ratio of the implied Rooi Rand

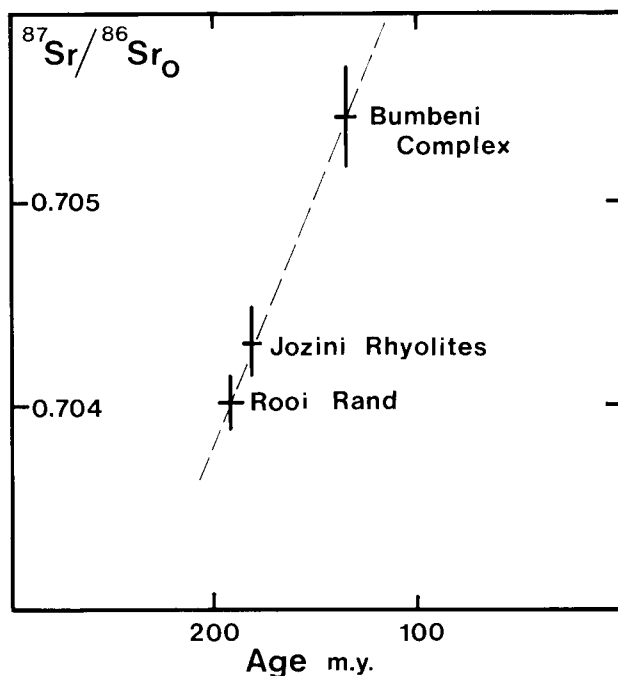


Figure 12

R_o ratio for the Rooi Rand dolerites, Jozini rhyolites and Bumbeni Complex from the southern Lebombo plotted against age.

source would have to be 0.6 whereas actual Rooi Rand dolerites have Rb/Sr ratios in the range 0.015 to 0.100. A gabbroic source with a R_o ratio coincidentally the same as the Rooi Rand could be envisaged, but even if such a gabbroic source derives from an older underplating event it would be considered crustal according to our terminology. Alternatively, the Rooi Rand dolerites might represent a depleted phase of a source that is not exposed but has the same R_o ratio, though such an argument requires special pleading. Equally unsatisfactory would be the necessity for only basalts with $R_o < 0.705$ to be underplated and involved in the rhyolite genesis whereas the bulk of the exposed mafic volcanics have R_o ratios > 0.705 . It should be noted that the Betton and Cox (1979) model was developed on the assumption that the high R_o ratios of the Lebombo mafic volcanics resulted from contamination; if, however, the evidence to the contrary given earlier in this paper is accepted the model becomes difficult to sustain.

Returning to the question of a crustal origin for the southern Lebombo felsic volcanics it is worth noting that Van Niekerk (1968) reported Precambrian U-Pb ages for zircon concentrates from two Jozini rhyolites. This suggests crustal involvement in their generation. Also, the results of Rogers and Hawkesworth (1982) indicate that some lower crustal granulite nodules from kimberlites in Lesotho have similar Sr- and Nd-isotopic compositions when compared to those of the rhyolites (Hawkesworth *et al.*, 1984a). The isotopic evidence for these granulites indicates that they are of Upper Proterozoic age. These nodules appear to have been sampled from an area close to the boundary between the Proterozoic mobile belt and Archaean craton. However, the Lebombo rocks overlie crust of either Archaean or supposed Pan-African age and the applicability of the data of the Lesotho granulite xenoliths to the question of the petrogenesis of the Lebombo rhyolites is therefore debatable.

In the central and northern Lebombo the R_o ratios of the rhyolites (0.7048) and the tholeiitic (low-MgO) rocks (0.7042–0.7067) overlap, a feature which may suggest that they are related—either by differentiation or by derivation from the same source. However, the northern and central Lebombo rhyolites have the same age and very similar R_o

ratios as the rhyolites from the southern Lebombo and this, together with the continuous nature of the outcrop and their similar compositional character, is not suggestive of a separate origin for the rhyolites of these two areas.

Given the above evidence together with the fact that the southern Lebombo felsic rocks are isotopically unrelated to their associated mafic rocks, we regard the mafic and felsic rocks of the northern Lebombo to be unrelated and the overlap in their R_o ratios to be fortuitous. On the basis of available isotopic evidence we therefore conclude that the Jozini–Mbuluzi suite of rhyolites from the Lebombo is probably of crustal origin. However, considering the low R_o ratios of the rhyolites and geochemical modelling results of Cleverly *et al.* (1984) it would appear that mafic lower crustal material (e.g. mafic granulite) was the most likely source of the rhyolites. In view of this, the basic concept of melting of underplated mafic material as proposed by Betton and Cox (1979) and modelled by Cleverly *et al.* (1984) remains an attractive model. However, inherent in the above is that underplating took place in pre-Karoo times giving rise to old material now considered to be crustal (see section IV.B.2). In the broadest sense Nd data on lower crustal rocks from Lesotho (Rogers and Hawkesworth, 1982) support the above.

The Nuanetsi rhyolites, on the basis of their higher R_o ratios, also appear to be unrelated to the underlying mafic rocks and hence probably of crustal origin. Clearly, more data are needed for all the felsic rock types of the Lebombo before unequivocal conclusions may be reached regarding their origin and relationships to the mafic rocks; combined stable and radiogenic isotope measurements would be especially valuable in this regard.

B. Etendeka Latites and Quartz Latites

In the Etendeka region two thick units of quartz latite and one thin unit of latite occur interbedded with Tafelberg-type basalts, and at the Tafelberg locality the thickness proportions of basalt, quartz latite and latite are estimated as 70 : 25 : 5 respectively. In the discussion on the Etendeka basaltic rocks in a previous section we followed Erlank *et al.* (1984) in assuming that the latites and quartz latites are of crustal derivation in view of their overall major, trace element and isotopic compositions. We have also stressed that the R_o ratios for these rocks are much higher than those of the Lebombo rhyolites (Table I), indicative of a different source area for the latites and quartz latites and that their R_o ratios are much higher than the associated mafic rocks. The proposed crustal origin for these rocks can be briefly considered in the light of the information summarized in Fig. 10. If the information portrayed in Fig. 10 is representative of the crustal lithology in this region it is immediately apparent that the latites and quartz latites could not have been derived from the older basement in view of the high ratios of the latter. Conversely, it is quite feasible that the latites and quartz latites could have been derived from those Damaran granitic rocks which had $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.715–0.722 at the time of emplacement of the latites and quartz latites. The latites, and most of the quartz latites, have Rb/Sr ratios slightly higher than the Damaran rocks of appropriate $^{87}\text{Sr}/^{86}\text{Sr}$ ratio. This is what would be expected from melting considerations and partitioning relationships during the melting of granitic rocks. This explanation would, however, probably not account for the few quartz latites with Rb/Sr ratios > 2 , and for these it would be necessary to increase their Rb/Sr ratios by fractional crystallization processes.

The proposal that the latites and quartz latites are derived from Damaran granitic rocks rather than from the older basement, is supported by unpublished $^{143}\text{Nd}/^{144}\text{Nd}$ data (C.J. Hawkesworth, pers. comm.) for these crustal materials. Detailed major and trace element modelling should be carried out to substantiate this conclusion, but

sufficient data are not yet available for this purpose. Nevertheless, it is clear that the latites and quartz latites must have been derived from slightly different crustal sources, and have undergone different evolutionary paths, in view of their different mineralogy, chemistry and Sr-, Nd- and Pb-isotopic characteristics (Erlank *et al.*, 1984).

VII. EVIDENCE FOR MANTLE HETEROGENEITY AND MANTLE ENRICHMENT PROCESSES

There is now general agreement that the mantle source areas of ocean floor and ocean island basalts are heterogeneous in composition (Bailey *et al.*, 1980). To what extent the subcontinental mantle is heterogeneous is less clear in view of the current debate concerning the possible role of crustal contamination in modifying continental basaltic compositions (Allègre *et al.*, 1982a). This section focuses on the evidence for heterogeneous and variably enriched mantle source regions for Karoo basaltic rocks and the mechanisms which could have led to such enrichment. Such attention is necessary since we have concluded that for most of the Karoo basaltic rocks we have studied that their variable and generally high R_0 ratios are essentially those of their source areas, although minor crustal contamination, in so far as Rb-Sr relationships are concerned, cannot be completely dismissed in any of the areas we have discussed above.

Direct evidence for heterogeneous enrichment of the subcontinental lithosphere comes from the mantle xenolith suites in alkaline basalts and kimberlites. The first persuasive evidence of mantle metasomatism as a mechanism for mantle enrichment was documented by Lloyd and Bailey (1975), Harte *et al.* (1975) and Erlank (1976) in their studies of mantle nodules contained in alkali basalts and kimberlites. At the same time, Barrett (1975) and Shimizu (1975) presented the first reliable Rb-Sr whole rock and mineral isotopic data for mantle nodules from southern African kimberlites. Many more results and data are now available, particularly on southern African Cretaceous kimberlites and their nodules which are of direct relevance to this section (Erlank and Rickard, 1977; Erlank and Shimizu, 1977; Kramers, 1977, 1979; Erlank *et al.*, 1980; Menzies and Murthy, 1980; Kramers *et al.*, 1981; Hawkesworth *et al.*, 1982; Haggerty *et al.*, 1983). The textural, mineralogical, major and trace element and isotopic data reported by these authors for peridotite nodules not only convincingly demonstrates the existence of infiltration metasomatism as a mechanism for enriching the upper mantle in K, Rb, Sr and other incompatible elements, but also provides evidence for the existence of earlier enrichment events and suggests that it is possible to obtain information on the timing of these events.

Of particular relevance to this study are the data and conclusions reached by Erlank *et al.* (1982) on peridotite nodules from the Bultfontein dumps in the Kimberley area. These authors have divided the suite into garnet peridotites (GP) which contain no texturally equilibrated or primary phlogopite, garnet phlogopite peridotites (GPP), phlogopite peridotites (PP) and phlogopite K-richertite peridotites (PKP). The latter two groups contain no garnet, while diopside may or may not be present in all four groups. This simple mineralogical classification is borne out by trace element data, since it is possible to separate garnet-bearing from garnet-free nodules on the basis of their HREE patterns and their Rb/Ba and, to a lesser extent, their Rb/Sr ratios.

Rb-Sr isotopic data presented by Erlank *et al.* (1982) are reproduced in Fig. 13 and the high present-day $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the metasomatized peridotite are apparent. Of immediate concern is the question of whether these whole rock nodules have been affected by kimberlite contamination. However, since all but two samples have distinctly higher R_0 and Rb/Sr ratios than the associated

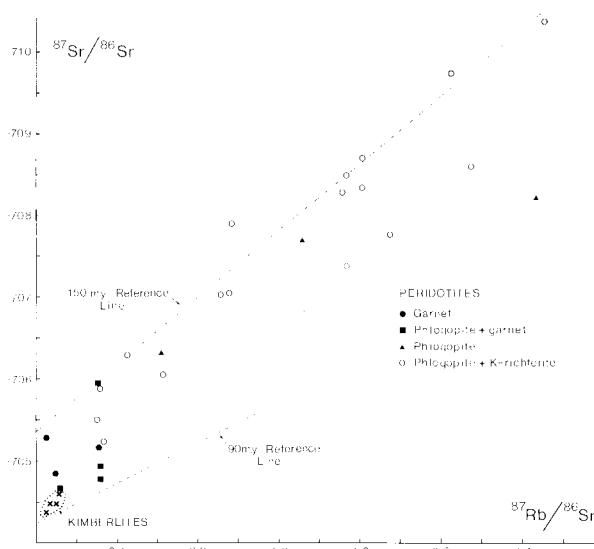


Figure 13

Rb-Sr isochron plot for peridotite whole rocks and kimberlites from Bultfontein, Kimberley. Data from Erlank *et al.* (1982) and unpublished data.

kimberlites at the time of kimberlite emplacement (90 m.y.), it is clear that kimberlite contamination during emplacement cannot be responsible for the high Rb/Sr and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in these nodules. Rb/Sr isotopic data on mineral separates, HCl-leached whole rocks and minerals, and associated leach solutions, suggest that some whole rocks do contain small amounts of unequilibrated strontium which could be of kimberlitic origin, such that their Rb/Sr and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were slightly lowered by kimberlite contamination during pipe emplacement. This would be most serious for those samples which have lowest Rb/Sr ratios — such samples also have lowest Rb and Sr contents and constitute the samples which are least metasomatized. Significantly, separated diopsides and K-richertites have higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios than the kimberlite host rocks and this supports the conclusion reached above. Erlank *et al.* (1982) therefore concluded that the scatter shown in Fig. 13 is essentially due to mantle metasomatism and that the variable R_0 ratios exhibited by the nodules at 90 m.y. (0.7045–0.7075) are those of a variably enriched mantle region at that time.

A further indication of the range of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the upper mantle beneath Kimberley is demonstrated in Fig. 14 which represents a compilation of available analyses on diopsides from nodules and megacrysts (discrete nodules). The measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios shown are effectively initial ratios in view of the very low Rb/Sr ratios in these diopsides. More important, however, is that the range observed is very similar to the overall variation in R_0 ratios found for all Karoo basaltic rocks we have discussed (Table I). The very high Sr content of some of the diopsides (500–900 ppm) is further evidence of metasomatism since, as first noted by Shimizu (1975), on the basis of clinopyroxene/liquid partition coefficients in basaltic systems at high pressures, such diopsides could not have equilibrated with any known silicate liquids. The partition coefficient data, if applicable, imply that these high-Sr diopsides equilibrated with liquids or fluid phases containing of the order of 1% Sr, which can be taken as an indication of the amount of Sr transported by metasomatic fluids in the upper mantle. Two new Fe-Cr titanates found as accessory minerals in K-richertite bearing nodules also contain percentage levels of Sr (Haggerty *et al.*, 1983).

Additional evidence for mantle enrichment has been provided by Smith (1983) who has shown that "micaceous" kimberlites (Dawson, 1967) are isotopically distinct from

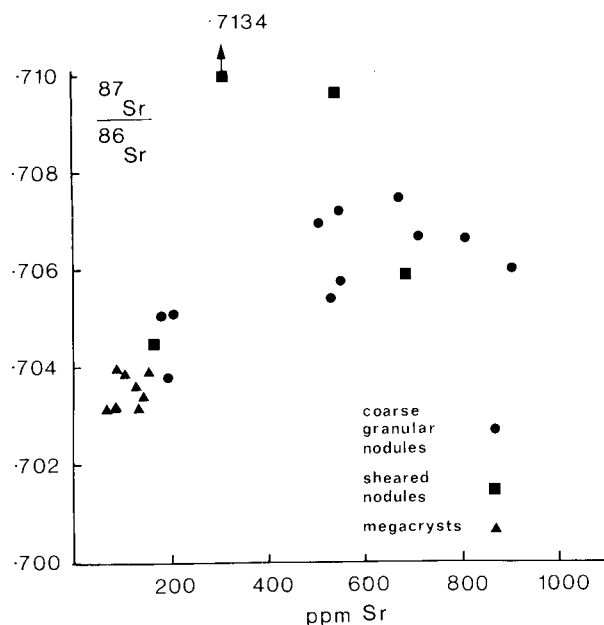


Figure 14

Plot of $^{87}\text{Sr}/^{86}\text{Sr}$ vs Sr for peridotite diopsides and megacryst diopsides from the Kimberley area. Data from Barrett (1975), Shimizu (1975), Kramers (1977, 1979), Kramers *et al.* (1981), Allègre *et al.* (1982b) and this study.

the "basaltic" types, and suggests that the former are derived from an enriched, and the latter from a slightly depleted, mantle source. Cohen *et al.* (1982) have also reported $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of about 0.83 for clinopyroxene from a peridotite nodule from a recent Lashaine volcano (Tanzania). Taken together with the very similar $^{87}\text{Sr}/^{86}\text{Sr}$ ratio for phlogopite from the same sample (Barrett, 1975; see also Ridley and Dawson, 1975), the implication is that portions of the mantle, although in isotopic equilibrium on a mineral-to-mineral scale, can be characterized by very high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, i.e. even higher than for the metasomatized nodules discussed above. Such highly enriched mantle regions would have to be sufficiently large in extent to resist isotopic homogenization with surrounding (less enriched) mantle. The widespread extent and antiquity of such enriched mantle regions is also implied by the results of Allsopp *et al.* (1979) who noted that high-Ti matrix phlogopites from numerous southern African kimberlites yield Rb-Sr model ages (i.e. with R_0 assumed 0.705) older than that of pipe formation; in most cases geologically acceptable ages would be obtained on the assumption of R_0 ratios in the range of 0.71 to 0.75.

The question of the timing of mantle metasomatic events is problematic, since co-existing separates from the metasomatized peridotites from Kimberley for the three dominant host minerals for Rb and Sr, namely phlogopite, K-richterite and diopside are, as might be predicted, in approximate isotopic equilibrium at 90 m.y. (Erlank *et al.*, 1982). Consideration of the whole rock data indicates that the samples analysed are bounded by the 90 m.y. and 150 m.y. reference lines in Fig. 13. The simplest interpretation is that the whole rock data denote a young metasomatism and that an upper limit of 150 m.y. can be assigned to this metasomatism, characterized by the K-richterite peridotites in view of their location on Fig. 13. Since this post-dates the main phase of Karoo volcanicity at 190 m.y. (Fitch and Miller, 1984), this volcanism clearly cannot be the main cause of the enrichment event, though Erlank *et al.* (1980) have suggested a connection between the waning of Karoo volcanicity and this metasomatic event. However, of particular importance is the demonstration that mantle metasomatism is capable of generating $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of as high as about 0.71 in peridotites with pre-

metasomatic $^{87}\text{Sr}/^{86}\text{Sr}$ ratios approximating Bulk Earth in just 150 m.y. This type of high Rb/Sr metasomatism may well have relevance for the Etendeka lavas in view of the high Rb/Sr ratios inferred for the source region of these volcanics in a previous section.

As mentioned previously, Erlank *et al.* (1982) noted that the garnet-bearing nodules could be separated from the PP and PKP nodules using trace element criteria. Thus Fig. 14 indicates lower Rb/Sr ratios in the garnet-bearing nodules but the limited data and variation observed in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios does not allow any "age" distinction to be made when compared with the non-garnet-bearing rocks with higher Rb/Sr ratios. However, a Nd-isotopic study on these same nodules discussed more fully in Hawkesworth *et al.* (1984a) indicates that the Nd-isotope data carry an age signature which is much older than the 150 m.y. age previously inferred from Rb/Sr data for the metasomatic event that led to the formation of the K-richterite peridotites. Model age considerations, based on the lowest $^{143}\text{Nd}/^{144}\text{Nd}$ ratios measured, indicate the existence of an older upper mantle enrichment event with an age of the order of 1 b.y. The chemical and mineralogical signature of this older enrichment event cannot yet be categorized because of the suspected imprint of the younger phlogopite-K richterite metasomatism, nor can this older enrichment be ascribed to metasomatism since other mechanisms of mantle enrichment (e.g. subduction) could have been responsible, but its existence is beyond dispute. An old enriched lithosphere is of particular importance for the derivation of the Karoo basaltic lavas, particularly in view of the fact that model Nd ages obtained on the latter are also of the order of 1 b.y. (Hawkesworth *et al.*, 1984a).

We have shown that it is possible to infer from nodules from a single kimberlite that two separate episodes of mantle enrichment with different trace element enrichment patterns have occurred at different times. It is obvious that future multi-isotopic and trace element work on nodules from other kimberlite localities should be equally rewarding in deciphering and categorizing the fractionation events that have contributed to the evolution and development of the sub-continental lithosphere beneath southern Africa. This is especially pertinent for the source areas of Karoo lavas since the combined evidence now available points strongly to the derivation of many Karoo lavas from old enriched lithospheric mantle material similar to that now contained as nodules in kimberlites.

VIII. CONCLUSIONS AND GEODYNAMIC IMPLICATIONS

In this paper we have shown that the mantle sources supplying Karoo mafic magmas generally had high and variable $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and that the Karoo mafic rocks have R_0 ratios which reflect this character because, with a few minor exceptions, they have not interacted in any significant way with crustal materials. This important conclusion differs from the interpretations of isotopic data of similar continental tholeiitic suites (e.g. Faure *et al.*, 1974; Carlson *et al.*, 1981; Mahoney *et al.*, 1982). In reaching this conclusion we stress that it is essential to evaluate Sr-isotope data in relation to whole rock geochemical analyses for both major and trace elements, as well as petrographic and other geological information. In addition, our interpretations have been influenced by the results of complementary studies of mantle xenoliths in kimberlites of southern Africa. These demonstrate unequivocally that the subcontinental lithosphere is markedly heterogeneous and has been metasomatically enriched to the extent of being able to produce, at least in the Kimberley area, high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the order of 0.711 in quite short periods of time.

A regional summary of Sr-isotopes in Karoo rocks is presented in Fig. 15. Basalts with the highest R_0 ratios

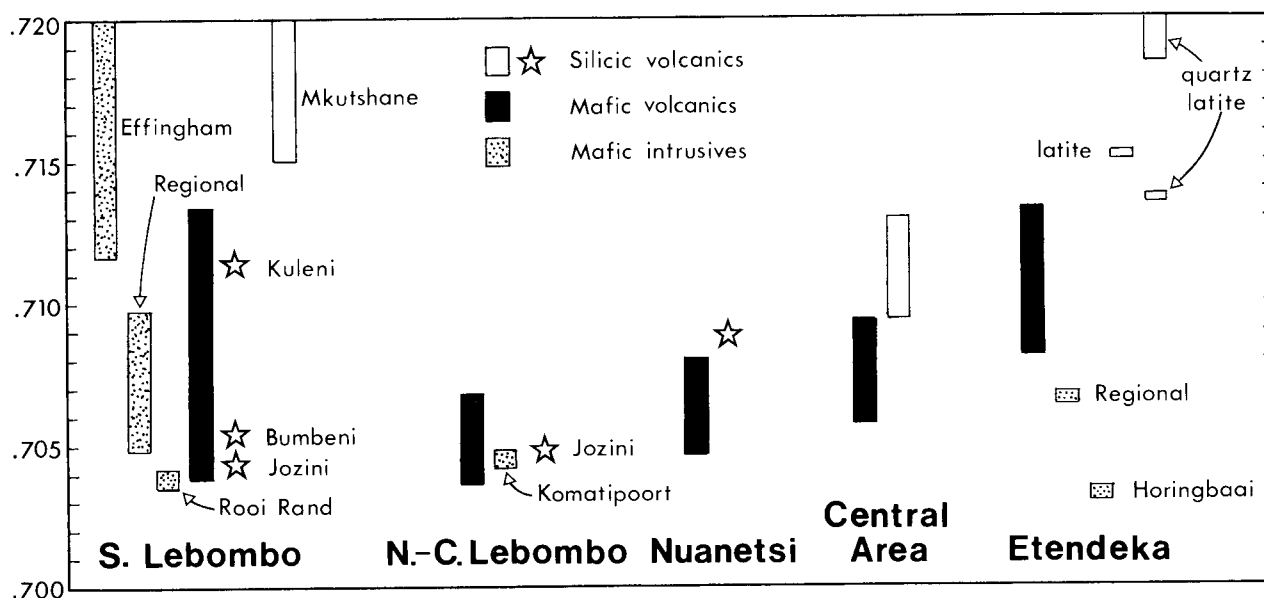


Figure 15

Summary of R_0 ratio data for the Karoo Province. Rock suites yielding isochrons are indicated by stars. Mafic lavas and intrusives not distinguished in northern-central Lebombo, Nuanetsi and Central areas.

(ranging up to 0.713) are found in the Lebombo and Etendeka regions, i.e. close to the sites of continental rifting during the break-up of Gondwanaland. In contrast, the ratios of basalts from the intracratonic Central area are lower (0.7046–0.7094). In both the Lebombo and Etendeka areas a swarm of post-basalt dykes, the Rooi Rand and Horingbaai dolerites respectively, exhibit R_0 ratios between 0.703 and 0.704. These ratios are the lowest measured for any rocks in the Karoo Province and are consistent with other isotopic and elemental compositional data, indicating that these dykes have been derived from "depleted" source areas. Rocks with similar compositions are unknown from the Central area.

Again, in both the Lebombo and Etendeka regions substantial volumes of silicic volcanics (quartz latites, rhyodacites, rhyolites) occur overlying the main thickness of basalts or interbedded with them in their upper layers. In the Etendeka the silicic volcanics have higher R_0 ratios than the associated mafic types, are genetically unrelated to them, and are probably partial melts of the Pan-African basement underlying the Karoo sequence. In the southern Lebombo the main volume of rhyolites (i.e. those of the Twin Ridge, Mbuluzi and Jozini formations) have on average lower R_0 ratios than the mafic volcanics (but not the Rooi Rand dolerites). In the northern Lebombo the range in R_0 ratios of the mafic rocks overlaps that of the rhyolites. Unlike the basalts, the rhyolites are isotopically extremely uniform along the whole length of the Lebombo. These features preclude a simple genetic relationship between the mafic and felsic rocks, based on our interpretation of the derivation of the mafic rocks from enriched mantle. The low R_0 ratios (<0.705) of the rhyolites are inconsistent with their derivation from upper crustal rocks. It is possible that the rhyolites are derived by partial melting of mafic material underplated on to the base of the lower crust. However, the origin of the rhyolites cannot be decided on isotopic evidence alone, but taken together the balance of the evidence, especially their very large volumes, suggests a crustal origin.

In the Nuanetsi region the mafic volcanics (excluding the Interbedded basalts) have R_0 ratios similar to those of the northern and central Lebombo, but the associated rhyolites have distinctly higher but uniform R_0 ratios (0.7085 ± 6). Again, this precludes a simple genetic relationship between the bulk of the mafic rocks and the voluminous rhyolites and the latter are arguably of crustal origin. However, it is

clear that much more data of both an isotopic and whole rock compositional nature are required before some of the uncertainties involving these felsic rocks are resolved.

Taken together, the data for the Karoo mafic rocks and mantle xenoliths contain evidence of a complex pattern of sub-continental lithosphere evolution. In some areas (Nuanetsi and northern Lebombo) pronounced enrichment of incompatible trace elements is not accompanied by high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. In other areas (Etendeka and southern Lebombo) enrichment is reflected in high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios combined with more subdued elevation of incompatible element abundances. The contrast between the southern and northern Lebombo mafic rocks is emphasized in Fig. 16 where R_0 is plotted vs Sr content. The change in isotopic and chemical character of these mafic rocks along the Lebombo occurs, on the basis of many more Sr analyses than shown in Fig. 16 (Microfiche Card 1 attached to this volume), between the Komati and Sabie rivers at approximately latitude $25^\circ 10' \text{S}$. This striking observation remains an unsolved problem in terms of source identities and the apparent older age of enrichment as inferred from Sr-model ages for the presumed mantle source of the

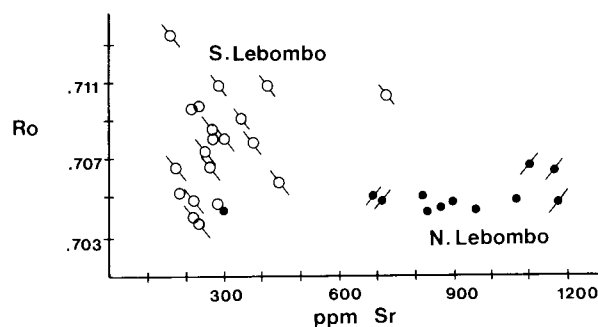


Figure 16

Plot of the R_0 ratio vs Sr for Sabi River Formation basalts and dolerites from the Lebombo (known Rooi Rand dolerites have been excluded).

S. Lebombo: Basalts—open circles with ticks; dolerites—open circles.

N. Lebombo: Basalts—closed circles with ticks; Dolerites—closed circles.

Note that the inclusion of the Mashikiri Formation nephelinites and Letaba Formation picrites from the northern Lebombo in the plot would not change the groupings shown.

southern Lebombo rocks. This suggests that isotopes and trace elements may be decoupled during mantle enrichment. The timing of these enrichment events and their chemical nature, particularly the relative behaviour of Rb and Sr (as well as other parent–daughter pairs) are critical in this regard. We have demonstrated that mantle processes like metasomatism can produce substantial variations in Rb/Sr ratios and generate very high concentrations of incompatible elements in some mantle phases (e.g. Sr in diopside). In addition, mantle nodules contain evidence of multi-stage enrichment events, one at least as old as ~1 b.y.

With the supporting evidence from the mantle nodules the R_0 ratio data for the Karoo mafic magmas are strongly suggestive of their source areas being located in the sub-continental lithosphere. These source areas were isolated from any convective mixing for long periods of time (at least 1 b.y.) and were heterogeneous in terms of isotopic ratios and element concentrations on both a horizontal and a vertical scale. High R_0 ratios are also a feature of Jurassic basalts and dolerites on other Gondwana continents (Antarctica, South America and Tasmania) and we suggest that our conclusions may apply to these suites as well. Indeed, Kyle (1980) has made such proposals regarding the Ferrar Group in Antarctica. However, we stress that such conclusions should be drawn only from detailed isotopic and whole rock compositional studies.

One outstanding feature of the Karoo Province which requires further consideration is the association of mafic rocks with high R_0 ratios (Etendeka and southern Lebombo) or high concentrations of incompatible elements (northern Lebombo and Nuanetsi) with large volumes of silicic volcanics and rift tectonics accompanying the breakup of Gondwanaland. To view the situation in broader perspective, we note from the analyses of Rüegg (1976) that quartz latites are apparently abundant in the southern portion of the Paraná lava field in Brazil, around the point of juxtaposition with their Etendeka equivalents in pre-drift reconstructions. Two Paraná (Serra Geral) quartz latites analysed by us for Rb, Sr and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Appendix A), together with their major and trace element (including REE) contents (Microfiche Card 1 attached to this volume) shows that these two samples are chemically and isotopically indistinguishable from their Etendeka counterparts. The large amount of crustal melting implied, contemporaneous with basalt emplacement, clearly demands a heat source of unusual size and intensity, and the distribution of the little studied Etendeka and Paraná quartz latites requires further investigation.

The association of highly enriched hydrous lithospheric mantle with continental rifting may reflect the ease with which such lithosphere is thinned and penetrated by the rising asthenosphere, possibly because it is hotter and therefore less viscous and weaker. If this suggestion is valid the distribution of highly enriched zones in the sub-continental lithosphere may influence the pattern of continental rifting. It is significant that in the marginal zones asthenospheric-type material becomes involved in producing late-stage “depleted” Karoo magmas as evinced by the Rooi Rand and Horingbaai dolerites. The temporal relationships between Karoo volcanism, Gondwana rifting and the commencement of sea-floor spreading around southern Africa are controversial and, therefore, hampers a better understanding of these suggestions. However, the available evidence, summarized in Duncan *et al.* (1984), suggests that the onset of Karoo vulcanicity preceded rifting and sea-floor spreading on both the western and eastern continental margins.

With regard to the rhyolites and quartz latites we partly concur with the ideas recently emphasized by Hildreth (1981) that crustal (silicic) magmatism is fundamentally basaltic in origin, not in the parent–daughter sense but in

that mantle-derived magmas provide the thermal basis for magma generation in the crust. It should therefore not be unusual that large volumes of silicic volcanics occur in major basalt provinces like the Karoo. However, the scarcity of felsic rocks in the Central area is not in accord with Hildreth's viewpoint. Although it may be argued that evidence for their presence has been lost to erosion, this explanation is hard to sustain. One explanation is that the rift environment of the continental margin areas allowed silicic magmas to erupt freely. Such an argument is not by itself convincing when one considers that there was little hindrance to copious effusion of basaltic magma in the Central area. Another possibility is that the increased thermal input into the crust at the continental margins may have resulted from a greater supply of basaltic magma, but the eroded nature of the Karoo volcanic succession offers only equivocal evidence in support of this proposal. Alternatively the heat for increased crustal melting was largely mantle-derived in that the crust at the rifted margins was sited over distinct geochemical and thermal anomalies in the underlying enriched lithospheric mantle as compared with the Central area where the lithosphere is less enriched. Certainly, this proposal is supported by the Sr-isotope and trace element data of the Karoo mafic rocks discussed in this paper.

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APPENDIX A

Rb-Sr Isotopic Data for Karoo Volcanic and Intrusive Rocks

NOTES:

- $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are normalized to $^{88}\text{Sr}/^{86}\text{Sr} = 8.375$. Numbers in parentheses after the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are in-run uncertainties expressed as two standard errors. Note that no in-run uncertainties are available for some of Manton's data, including Effingham. In this case errors of 0.0002, based on replicate analyses, have been assumed.
- Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, where shown, are calculated either for 190 m.y. or 121 m.y. according to ages quoted by Allsopp *et al.* (1984), and Fitch and Miller (1984). In those cases where initial ratios are not shown, ages are generally younger and should be calculated according to ages cited in Allsopp *et al.* (1984), e.g. Lebombo, Kuleni and Bumbeni rhyolites.
- The following conversion formula has been used to calculate $^{87}\text{Rb}/^{86}\text{Sr}$: $^{87}\text{Rb}/^{86}\text{Sr} = \text{Rb}/\text{Sr} \times (2.693 + 0.2830 \times ^{87}\text{Sr}/^{86}\text{Sr})$.
- In certain cases isotope dilution methods were used to calculate ^{87}Rb and ^{86}Sr values, viz. Kuleni rhyolites and

samples analysed by Manton. In these cases the $^{87}\text{Rb}/^{86}\text{Sr}$ ratios as listed by the original authors are thus correct. Rb (absolute) was in these cases calculated as $\text{Rb} = ^{87}\text{Rb} \times 3.5399$ and Sr (absolute) was calculated from the equation listed in (c) above.

(e) With the exception of data from Manton (1968) and Frankel (1969), the locality details of analysed samples are included in the Karoo Database (Microfiche Cards 1 and 2 attached to this volume). Data presented for the Effingham dolerites are from Frankel (1969) and represent analyses of dykes and sills found along the coastal area of Natal (South Africa) between

Empangeni and Durban. In the case of the unpublished data obtained by Manton, samples with an "MB" prefix are from the Mbuluzi section of the Lebombo in Swaziland, whereas details of the remaining sample localities were published in Manton (1968). Additional details of sample localities will be provided by the authors on request.

(f) Eight additional granophyre analyses from the Komatipoort area are not listed in Appendix A but were used to calculate the Central and Northern Lebombo rhyolite age listed in Table I. These data are presented in Allsopp *et al.* (1984b).

AREA/ SAMPLE	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	R_O (190 m.y.)
ZIMBABWE						
VICTORIA FALLS						
Basalt						
KL23[6]	8.2	345	0.024	0.069	0.70435(4)	0.70416
NUANETSI						
Mashikiri Formation Nephelinites						
NTS7[3]	70	1500	0.047	0.135	0.70566(6)	0.70529
NTS8[8]	69	1566	0.044	0.127	0.70585(5)	0.70551
NTS12[8]	63	1663	0.038	0.110	0.70540(6)	0.70510
Letaba River Formation Picritic Basalts						
NB7[7]	68	1027	0.066	0.191	0.7070	0.7065
N399[8]	19.4	585	0.033	0.095	0.70487(5)	0.70461
N149[8]	21.5	604	0.036	0.104	0.70493(4)	0.70465
N187[8]	25.6	613	0.042	0.121	0.70599(5)	0.70566
N406[8]	24.0	703	0.034	0.098	0.70485(4)	0.70459
N113[8]	48.0	735	0.065	0.188	0.70531(6)	0.70480
N126[8]	33.6	787	0.043	0.124	0.70528(5)	0.70495
N60[8]	36.4	935	0.039	0.113	0.70514(4)	0.70483
N133[8]	36.0	958	0.038	0.109	0.70505(4)	0.70476
N135[8]	33.6	970	0.035	0.100	0.70548(5)	0.70521
N400[9]	42.3	998	0.042	0.123	0.70516(4)	0.70478
Sabie River Formation Basalts						
NB2[7]	34.5	898	0.038	0.110	0.7060	0.7057
NB3[7]	64	575	0.110	0.318	0.7085	0.7076
NB4[7]	58	983	0.059	0.170	0.7059	0.7054
N175[3]	37	703	0.053	0.153	0.70617(22)	0.70576
N386[3]	50	1028	0.049	0.142	0.70621(4)	0.70583
Interbedded Basalts						
NB1[7]	62	183	0.339	0.980	0.7118	0.7092
NB5[7]	12.5	182	0.069	0.199	0.7105	0.7100
NB6[7]	72	237	0.304	0.879	0.7112	0.7088
Picritic Intrusives						
NTS17[8]	13.1	742	0.018	0.051	0.70545(6)	0.70531
N21[8]	45.8	1130	0.041	0.117	0.70554(4)	0.70522
N22[8]	33.2	842	0.039	0.114	0.70544(4)	0.70513
N163[8]	12.2	389	0.031	0.091	0.70563(7)	0.70538
Nuanetsi Formation Rhyolites						
NR1[7]	153	67.6	2.257	6.54	0.7266	
NR3[7]	143	109	1.312	3.80	0.7186	
NR4[7]	142	100	1.412	4.09	0.7194	
NR7[7]	116	96.3	1.202	3.48	0.7179	
NR8[7]	192	204	0.940	2.72	0.7163	
LM466[7]	207	74.2	2.797	8.11	0.7306	

AREA/ SAMPLE	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	R_0 (190 m.y.)
LEBOMBO						
PAFURI AREA						
Mashikiri Formation Nephelinites						
KP82[3]	28.6	1034	0.277	0.080	0.70696(5)	0.70674
KP127[3]	62	1337	0.046	0.134	0.70723(6)	0.70687
Letaba River Formation Picritic Basalts						
KP112[3]	61.0	795	0.077	0.222	0.70590(5)	0.70530
KP121[3]	45.0	831	0.054	0.157	0.70577(4)	0.70535
Dolerites						
KP89[2]	14.2	898	0.016	0.046	0.70488(4)	0.70476
KP129[2]	58	1065	0.054	0.156	0.70541(5)	0.70499
SHINGWEDZI RIVER AREA						
Letaba River Formation Picritic Basalt						
KS3[3]	58.0	842	0.069	0.199	0.70533(4)	0.70479
Dolerite						
KS36[3]	26.0	297	0.088	0.253	0.70513(6)	0.70444
Jozini Formation Rhyolites						
KS16[3]	111	500	0.222	0.642	0.70648(6)	
KS19[3]	133	226	0.589	1.703	0.70914(3)	
OLIFANTS RIVER AREA						
Mashikiri Formation Nephelinite						
KA16[3]	38.7	1320	0.029	0.084	0.70543(6)	0.70520
Letaba River Formation Picritic Basalt						
KA24[3]	29.5	902	0.033	0.095	0.70529(4)	0.70503
Sabie River Formation Basalts						
O169[3]	113	1177	0.096	0.278	0.70549(4)	0.70474
KA112[3]	32	712	0.045	0.130	0.70524(4)	0.70489
Dolerites						
KA31[3]	45	871	0.052	0.149	0.70461(5)	0.70421
KA52[3]	48	836	0.057	0.166	0.70468(5)	0.70423
KA52B[3]	50	953	0.053	0.152	0.70469(4)	0.70428
Jozini Formation Rhyolites						
KA6B[3]	119	189	0.630	1.822	0.70928(2)	
KA78[3]	111	179	0.620	1.794	0.70933(6)	
Acid Intrusive						
KA13[1]	132	284	0.465	1.345	0.70834(10)	

AREA/ SAMPLE	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	R_0 (190 m.y.)
SABIE RIVER AREA						
Sabie River Formation Basalts						
CL110[3]	54	1100	0.049	0.142	0.70711(5)	0.70673
CL115[3]	24.4	696	0.035	0.101	0.70527(6)	0.70500
CL120[3]	60	1162	0.052	0.149	0.70678(4)	0.70638
Dolerite						
CL105[3]	28.2	822	0.034	0.099	0.70531(5)	0.70504
Granophyre						
CL99[1]	152	76	2.000	5.794	0.72028(8)	
Jozini Formation Rhyolites						
CL209[3]	122	208	0.587	1.697	0.70925(4)	
CL451[1]	139	153	0.908	2.629	0.71088(9)	
SWAZILAND						
Sabie River Formation Basalts						
LB2[7]	5.0	225	0.022	0.064	0.7042	0.7040
EH8[3]	18.5	265	0.070	0.202	0.70734(4)	0.70679
Rooi Rand Dolerites						
RC110[4]	13.3	298	0.045	0.130	0.70446(4)	0.70411
RC120[4]	15.5	213	0.073	0.211	0.70425(6)	0.70368
RC121[4]	4.1	198	0.021	0.060	0.70409(4)	0.70393
Mkutshane Interbedded Rhyolites						
RC42[9]	95	158	0.601	1.741	0.71968(4)	
RC119[4]	122	129	0.946	2.747	0.74571(6)	
Twin Ridge Interbedded Rhyolites						
RC36[3]	127	193	0.658	1.904	0.70914(4)	
RC40[3]	117	174	0.672	1.946	0.70938(6)	
RC128[3]	88	245	0.359	1.039	0.70705(2)	
Jozini Formation Rhyolite						
RC66[3]	121	168	0.720	2.084	0.70951(4)	
Mbuluzi Formation Rhyolites						
RC74[4]	144	105	1.371	3.971	0.71460(9)	
RC131[4]	144	124	1.161	3.362	0.71276(8)	
SK138[4]	143	191	0.749	2.167	0.70953(6)	
SK153[4]	125	131	0.954	2.762	0.71142(8)	
Acid Volcanics (Undifferentiated)						
NO1[7]	154	71.4	2.151	6.23	0.7203(4)	
MB1[7]	123	127	0.968	2.80	0.7115(4)	
MB2[7]	129	142	0.908	2.63	0.7117(4)	
MB3[7]	120	134	0.892	2.58	0.7114(4)	
MB4[7]	133	138	0.892	2.58	0.7117(4)	
MB5[7]	122	141	0.861	2.49	0.7109(4)	
MB6[7]	126	90.5	1.388	4.02	0.7154(4)	
MB7[7]	118	127	0.926	2.68	0.7111(4)	

AREA/ SAMPLE	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	R_0 (190 m.y.)
MB8[7]	147	124	1.185	3.43	0.7134(4)	
MB9[7]	130	131	0.992	2.87	0.7115(4)	
MB10[7]	108	134	0.802	2.32	0.7102(4)	
MB11[7]	143	98.5	1.447	4.19	0.7154(4)	
MB12[7]	142	79.6	1.778	5.15	0.7176(4)	
MB13[7]	129	135	0.953	2.76	0.7119(4)	
MB14[7]	104	159	0.653	1.89	0.7109(4)	
MB16[7]	143	174	0.819	2.37	0.7104(4)	
MB17[7]	125	117	1.068	3.09	0.7125(4)	
MB18[7]	130	136	0.957	2.77	0.7116(4)	
MB19[7]	123	135	0.908	2.63	0.7113(4)	
MB20[7]	117	145	0.805	2.33	0.7106(4)	
MB21[7]	124	145	0.854	2.47	0.7111(4)	
MB22[7]	127	93.8	1.350	3.91	0.7145(4)	
MB23[7]	154	134	1.147	3.32	0.7131(4)	
MB24[7]	130	131	0.988	2.86	0.7121(4)	
MB25[7]	145	121	1.199	3.47	0.7135(4)	
MB26[7]	144	93.2	1.540	4.46	0.7162(4)	
MB27[7]	143	69.9	2.040	5.91	0.7198(4)	
SR1[7]	126	132	0.954	2.76	0.7115(4)	
SR4[7]	115	141	0.816	2.36	0.7105(4)	
JR4[7]	204	186	1.098	3.18	0.7121(4)	
SG1[7]	117	164	0.712	2.06	0.7095(4)	
SG2[7]	139	117	1.182	3.42	0.7135(4)	
Movene Formation Basalts						
NG25[1]	2.8	121	0.023	0.067	0.70587(5)	0.70569
NG26[1]	3.6	111	0.032	0.094	0.70471(8)	0.70446
SOUTHERN LEBOMBO						
Sabie River Formation Basalts						
L12[1]	6.7	727	0.009	0.027	0.71043(3)	0.71036
L18[1]	10.7	273	0.039	0.113	0.70891(8)	0.70860
L29[1]	19.1	302	0.063	0.183	0.70865(5)	0.70816
L32[1]	10.9	256	0.043	0.123	0.70784(6)	0.70751
L36[1]	14.9	352	0.042	0.123	0.70947(4)	0.70914
L57[1]	34	378	0.090	0.260	0.70849(6)	0.70778
S14[3]	22.4	288	0.078	0.225	0.71150(9)	0.71089
S45[1]	75	417	0.180	0.521	0.71229(6)	0.71088
L285[9]	16.5	222	0.074	0.215	0.70513(4)	0.70491
L348[3]	41.0	166	0.247	0.715	0.71536(4)	0.71343
L505c[1]	19.4	172	0.113	0.326	0.70752(2)	0.70664
J3[1]	8.8	235	0.038	0.108	0.70404(4)	0.70375
LB3[7]	23	446	0.052	0.150	0.7062	0.7058
Zeolite Separate						
L505c[3]	3.657	80.2	0.0456	0.132	0.70705(7)	0.70669
Calcite Separate						
L505c[3]	0.126	33.1	0.0038	0.011	0.70703(7)	0.70702

AREA/ SAMPLE	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	R_0 (190 m.y.)
Dolerites						
L94[1]	9.9	228	0.043	0.127	0.71001(5)	0.70967
L249a[1]	9.7	188	0.052	0.149	0.70574(4)	0.70534
L503[1]	9.1	234	0.039	0.113	0.71011(3)	0.70981
L508[1]	74	276	0.268	0.776	0.71031(3)	0.70821
J55[3]	66	288	0.229	0.663	0.70653(4)	0.70474
Rooi Rand Dolerites						
L496b[1]	21.8	232	0.094	0.272	0.70466(8)	0.70393
A21[2]	8.2	205	0.040	0.116	0.70404(4)	0.70373
A22[2]	9.4	159	0.059	0.171	0.70428(6)	0.70382
A49[2]	3.0	202	0.015	0.043	0.70382(6)	0.70370
A114[2]	10.8	224	0.048	0.140	0.70429(6)	0.70391
A117[2]	26.8	268	0.100	0.289	0.70494(6)	0.70416
A129[2]	5.7	150	0.038	0.110	0.70415(4)	0.70385
A135[2]	6.3	213	0.030	0.086	0.70373(4)	0.70350
A175[2]	9.3	198	0.047	0.136	0.70424(6)	0.70387
A185[2]	4.8	150	0.032	0.093	0.70398(6)	0.70373
A186[2]	4.3	161	0.027	0.078	0.70387(4)	0.70366
Effingham and Karoo Dolerites						
A[10]	50	680	0.074	0.213	0.7229	0.7223
B[10]	110	240	0.458	1.328	0.7199	0.7163
C[10]	100	140	0.714	2.070	0.7228	0.7172
D[10]	80	200	0.400	1.159	0.7153	0.7122
E[10]	60	150	0.400	1.159	0.7206	0.7175
F[10]	30	320	0.094	0.272	0.7124	0.7117
G[10]	20	230	0.087	0.252	0.7075	0.7068
H[10]	20	240	0.083	0.240	0.7071	0.7064
A-F Effingham Dolerites: A - Glassy vein: B - Glassy margin: C - Sill centre: D - Sill: E - Effingham type rock: F - Coarse grained dolerite: G + H - Normal Karoo dolerites. These isotopic data were obtained prior to the advent of high precision mass-spectrometry and Rb and Sr were determined by atomic absorption spectrophotometry.						
Jozini Formation Rhyolites						
L333[1]	119	143	0.832	2.408	0.71033(2)	
L335[1]	121	151	0.801	2.319	0.71004(7)	
L336[1]	122	139	0.878	2.540	0.71042(4)	
L431[1]	128	172	0.744	2.154	0.70984(7)	
L441[1]	103	213	0.484	1.399	0.70810(5)	
L467[1]	111	201	0.552	1.598	0.70834(5)	
J5[1]	131	133	0.985	2.851	0.71165(5)	
J11[1]	131	134	0.978	2.830	0.71151(7)	
J16[1]	113	204	0.554	1.603	0.70844(6)	
J18[1]	116	197	0.589	1.704	0.70850(3)	
J66[1]	129	147	0.878	2.540	0.71066(6)	
JR7[7]	111	207	0.536	1.550	0.7081(4)	
JR9[7]	123	154	0.795	2.300	0.7101(4)	
JR13[7]	119	172	0.691	2.000	0.7097(4)	

AREA/ SAMPLE	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	R_0 (190 m.y.)
Kuleni Rhyolites						
78-5[3]	260	6.16	42.1	125.0	0.9695(25)	
78-40[3]	268	2.37	113.0	350.2	1.4317(15)	
78-40b[1]	282	2.32	121.6	378.9	1.4962(30)	
BK-6[1]	280	2.38	118.1	367.0	1.4672(10)	
BK-7[1]	274	2.59	106.0	327.3	1.3963(30)	
BN-6[1]	237	2.80	84.5	257.10	1.2403(25)	
Bumbeni Complex Rhyolites and Syenites						
78-13[3]	165	158	1.044	3.022	0.71103(4)	
78-36[3]	173	80	2.163	6.263	0.71735(4)	
78-38[3]	113	183	0.618	1.787	0.70892(7)	
L507[3]	145	50	2.900	8.402	0.72145(6)	
CENTRAL KAROO						
SPRINGBOK FLATS						
Basalts						
KL50[5]	7.4	164	0.0451	0.131	0.70861(6)	0.70826
KL48[5]	2.7	176	0.0153	0.044	0.70666(8)	0.70654
NATAL						
Dolerite						
2[5]	10.9	197	0.0553	0.160	0.70724(10)	0.70681
WITSIESHOEK						
Dolerites						
D2[5]	14.1	202	0.0698	0.202	0.70728(6)	0.70673
D34[5]	14.6	176	0.0830	0.240	0.70705(9)	0.70640
D40[5]	9.4	189	0.0497	0.144	0.70655(6)	0.70616
LESOTHO						
Lesotho Formation Basalts, Kao Core						
KLK1[5]	20.9	90	0.232	0.672	0.70639(5)	0.70457
KLK2[5]	27.8	173	0.161	0.465	0.70639(5)	0.70513
KLK7[5]	13.9	176	0.079	0.229	0.70645(5)	0.70583
KLK15[5]	18.7	347	0.0539	0.156	0.70738(6)	0.70696
KLK24[5]	21.9	513	0.0427	0.124	0.70525(6)	0.70492
Lesotho Formation Basalts, Oxbow						
PN2101[8]	10.5	186	0.056	0.163	0.70641(6)	0.70597
PN2105[8]	13.0	191	0.068	0.197	0.70629(6)	0.70576
PN2110[8]	10.0	157	0.064	0.184	0.70578(4)	0.70528
PN2130[8]	7.0	145	0.048	0.140	0.70509(4)	0.70471
PN2139[8]	25.5	182	0.140	0.405	0.70613(5)	0.70503
PN2140[8]	18.0	205	0.088	0.254	0.70577(6)	0.70508
PN2157[8]	10.0	180	0.056	0.161	0.70557(4)	0.70514

AREA/ SAMPLE	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	R_0 (190 m.y.)
PN2164[8]	6.5	156	0.042	0.121	0.70553(8)	0.70520
PN2165[8]	9.0	178	0.051	0.147	0.70563(3)	0.70523
PN2175[8]	12.0	201	0.060	0.173	0.70590(5)	0.70543
STORMBERG AND EASTERN CAPE						
Pronksberg Basalt (Drumbo Type)						
KR28[1]	21	297	0.071	0.205	0.70610(5)	0.70555
Pronksberg High-K Basalts, Blikana						
KL24[9]	29.6	488	0.061	0.178	0.70935(4)	0.70887
KL25[9]	27.8	691	0.040	0.116	0.70972(4)	0.70941
Pronksberg Dacites, Jamestown						
KRP6[9]	117	320	0.366	1.059	0.71522(4)	0.71236
KRP8[9]	95	344	0.276	0.799	0.71512(4)	0.71296
Belmore Andesites, Barkly East						
JR31[5]	111	279	0.398	1.152	0.71267(8)	0.70956
KRB1[9]	135	299	0.452	1.307	0.71325(4)	0.70972
KRB8[9]	88	252	0.349	1.011	0.71216(4)	0.70943
Roodehoek Unit Lavas, Omega type Basalt, Jamestown-Molteno						
AM22[9]	4.0	198	0.020	0.058	0.70506(4)	0.70490
AM36b[9]	11.0	177	0.062	0.180	0.70517(4)	0.70468
AM25[9]	7.0	197	0.036	0.103	0.70498(4)	0.70470
Vaalkop Unit Lavas, Jamestown-Molteno						
AM29[9]	28	222	0.126	0.365	0.70761(4)	0.70662
AM34[9]	25	226	0.111	0.320	0.70750(4)	0.70664
AM13[9]	5.0	307	0.016	0.047	0.70702(4)	0.70689
Moshesh's Ford Formation (Drumbo Basalt Member)						
KL38[5]	18.5	272	0.068	0.197	0.70636(5)	0.70583
CLA-01[5]	21.3	286	0.075	0.215	0.70613(8)	0.70555
Moshesh's Ford Formation (Donnybrook Basalt Member)						
DBM11[3]	19.6	303	0.065	0.188	0.70602(4)	0.70551
Kraai River Formation Basalts						
S3/10[5]	21.0	236	0.089	0.258	0.70934(8)	0.70864
KF2[9]	6.3	201	0.031	0.091	0.70776(4)	0.70752
KF11[9]	26.3	176	0.149	0.432	0.70861(4)	0.70744
KRF13[9]	9.9	210	0.047	0.136	0.70765(4)	0.70728
Lesotho Formation Basalts, Naude's Nek						
JP22[9]	8.7	196	0.044	0.128	0.70545(4)	0.70510
JP27[9]	8.0	192	0.042	0.121	0.70552(4)	0.70495
JP43[2]	18.0	181	0.099	0.288	0.70618(3)	0.70540
JP48[1]	1.1	199	0.006	0.016	0.70578(5)	0.70574
JP54[9]	11.0	194	0.057	0.164	0.70623(4)	0.70579
JP58[1]	24.0	199	0.121	0.349	0.70677(8)	0.70583

AREA/ SAMPLE	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	R_0 (190 m.y.)
Dolerites, Various Localities						
KL37[5]	18.5	288	0.0642	0.186	0.70624(4)	0.70574
JR18[5]	9.6	176	0.0545	0.158	0.70504(5)	0.70461
JR2[5]	6.6	192	0.034	0.099	0.70687(5)	0.70660
JR36[5]	9.6	224	0.043	0.124	0.70590(5)	0.70557
JR70[5]	22.9	168	0.136	0.394	0.70725(4)	0.70618
Amherst Dolerite Sill, Fort Beaufort						
SK14[9]	9.6	203	0.047	0.137	0.70602(4)	0.70565
SK15[9]	8.7	217	0.040	0.116	0.70625(4)	0.70594
SK17[9]	11.2	228	0.049	0.142	0.70628(4)	0.70590
SK18[9]	13.7	244	0.056	0.162	0.70653(4)	0.70609
SK19[9]	13.9	237	0.059	0.170	0.70637(4)	0.70591
SK20[9]	12.7	208	0.061	0.177	0.70614(4)	0.70566
JR29[5]	2.4	244	0.010	0.029	0.70629(6)	0.70621
WESTERN CAPE						
Dolerites						
KL32[5]	10.6	197	0.0538	0.156	0.70576(6)	0.70534
KL33[5]	7.0	220	0.0318	0.092	0.70550(4)	0.70525
NORTH WESTERN CAPE (Calvinia)						
Blaaukrantz Dolerite Sill						
LR001[1]	29.0	167	0.174	0.502	0.70608(8)	0.70472
LR007[1]	7.60	187	0.041	0.118	0.70506(6)	0.70474
Hangnest Dolerite Sill						
LR025[1]	24.1	194	0.124	0.360	0.70885(6)	0.70788
LR035[1]	17.1	210	0.082	0.236	0.70840(7)	0.70776
Hantam Dolerite Sill						
LR045[1]	11.5	204	0.0564	0.163	0.70547(5)	0.70503
SOUTHERN NAMIBIA						
Kalkrand Basalts						
KLS1[5]	10.9	242	0.0450	0.130	0.70619(6)	0.70584
KLS4[1]	12.0	239	0.0502	0.145	0.70781(8)	0.70742
Dolerites						
KL11[5]	15.4	200	0.0770	0.223	0.70687(9)	0.70627
KLS7[5]	11.2	196	0.0571	0.165	0.70681(10)	0.70636
KLS13[1]	9.9	252	0.0393	0.114	0.70514(3)	0.70483

AREA/ SAMPLE	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	R_0 (121 m.y.)
NORTHERN NAMIBIA						
ETENDEKA						
Regional Intrusives						
KLS43[9]	10.2	274	0.0372	0.108	0.70673 (4)	0.70654
KLS44[9]	11.0	275	0.0400	0.116	0.70696 (4)	0.70676
Albin Basalts						
KLS144[9]	41.0	234	0.175	0.507	0.71388 (4)	0.71301
KLS196[9]	33.6	260	0.129	0.373	0.70984 (4)	0.70920
Horingbaai Minor Intrusives						
KLS122[9]	4.3	186	0.0231	0.067	0.70334 (4)	0.70323
KLS126[9]	7.3	241	0.0303	0.088	0.70350 (6)	0.70335
KLS145[9]	2.2	192	0.0115	0.033	0.70305 (4)	0.70299
Tafelberg Volcanics						
Basaltic Lavas						
KL16[9]	44.0	212	0.208	0.601	0.70976 (4)	0.70873
KLS22[5]	31.0	257	0.121	0.349	0.71198 (4)	0.71138
KLS24[5]	18.7	214	0.0874	0.253	0.70990 (6)	0.70947
KLS26[5]	44.0	268	0.164	0.475	0.71063 (10)	0.70981
KLS40[5]	71	213	0.333	0.965	0.71445 (4)	0.71279
KLS41[5]	75	198	0.379	1.097	0.71469 (5)	0.71280
KLS42[5]	96	187	0.513	1.486	0.71605 (5)	0.71349
KLS46[1]	32.9	281	0.117	0.339	0.70867 (4)	0.70809
KLS53[5]	52	214	0.243	0.703	0.71206 (7)	0.71085
KLS54[5]	83	186	0.446	1.292	0.71445 (8)	0.71223
KLS58[9]	32.0	275	0.116	0.336	0.71257 (4)	0.71199
KLS66[1]	18.3	189	0.0968	0.280	0.71208 (6)	0.71160
KLS98[9]	111	257	0.432	1.251	0.71558 (2)	0.71343
KLS197[9]	34.6	270	0.128	0.371	0.71057 (2)	0.70993
Dolerites						
KLS16[9]	52	293	0.177	0.514	0.70895 (4)	0.70807
KLS38[5]	43.0	349	0.123	0.356	0.71039 (6)	0.70978
KLS48[5]	16.1	313	0.0514	0.149	0.70805 (4)	0.70779
KLS100[9]	20.0	241	0.0830	0.240	0.71072 (4)	0.71031
KLS152[1]	93	168	0.554	1.605	0.71526 (4)	0.71250
Latites						
KLS69[1]	142	215	0.660	1.913	0.71866 (5)	0.71537
KLS71[1]	150	201	0.746	2.162	0.71893 (4)	0.71521
Quartz Latites						
KL13[5]	225	87.9	2.560	7.425	0.73422 (5)	0.72145
KL20[5]	161	139	1.158	3.357	0.72472 (14)	0.71895
KLS18[1]	198	80.2	2.469	7.160	0.73182 (5)	0.71951
KLS33[1]	164	152	1.079	3.127	0.72528 (7)	0.71990
KLS34[5]	163	136	1.199	3.474	0.72563 (8)	0.71965
KLS36[5]	139	163	0.853	2.471	0.72320 (8)	0.71895
KLS51[1]	162	158	1.025	2.972	0.72653 (8)	0.72142
KLS52[1]	170	133	1.278	3.705	0.72508 (6)	0.71871

AREA/ SAMPLE	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	R_0 (121 m.y.)
KLS55[5]	164	117	1.402	4.063	0.72758(9)	0.72059
KLS56[1]	134	123	1.089	3.155	0.71916(7)	0.71373
KLS77[1]	225	168	1.339	3.883	0.72734(17)	0.72066
KLS78[1]	169	112	1.509	4.375	0.72872(8)	0.72120
KLS81[9]	95.7	211	0.454	1.314	0.72093(4)	0.71867
KLS83[9]	188	115	1.635	4.740	0.72935(4)	0.72120
KLS84[1]	399	156	2.558	7.419	0.73378(7)	0.72102
KLS103[1]	165	119	1.387	4.019	0.72547(4)	0.71856
AG38-4[5]	245	53.8	4.554	13.223	0.74458(7)	0.72184
Zeolite Separates						
KLS18[5]	19.2	3154	0.00609	0.0176	0.71678(4)	0.71675
KLS66[1]	1.9	75.3	0.0252	0.0730	0.71126(6)	0.71113
KLS103[1]	17.1	6268	0.00273	0.00791	0.71652(4)	0.71651
Calcite Separate						
KLS103[1]	1.15	43.4	0.0264	0.0767	0.71675(6)	0.71662

BRAZIL

PARANA BASIN

Quartz Latites

BRA11[1]	176	121	1.455	4.217	0.72787(6)	0.72062
BRA12[1]	170	136	1.250	3.623	0.72684(6)	0.72061

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