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Doctor of Philosophy in the University of Oxford

THE STRUCTURAL AND MAGMATIC EVOLUTION OF THE
LEBOMBO MONOCLINE, SOUTHERN AFRICA, WITH PARTICULAR
REFERENCE TO SWAZILAND



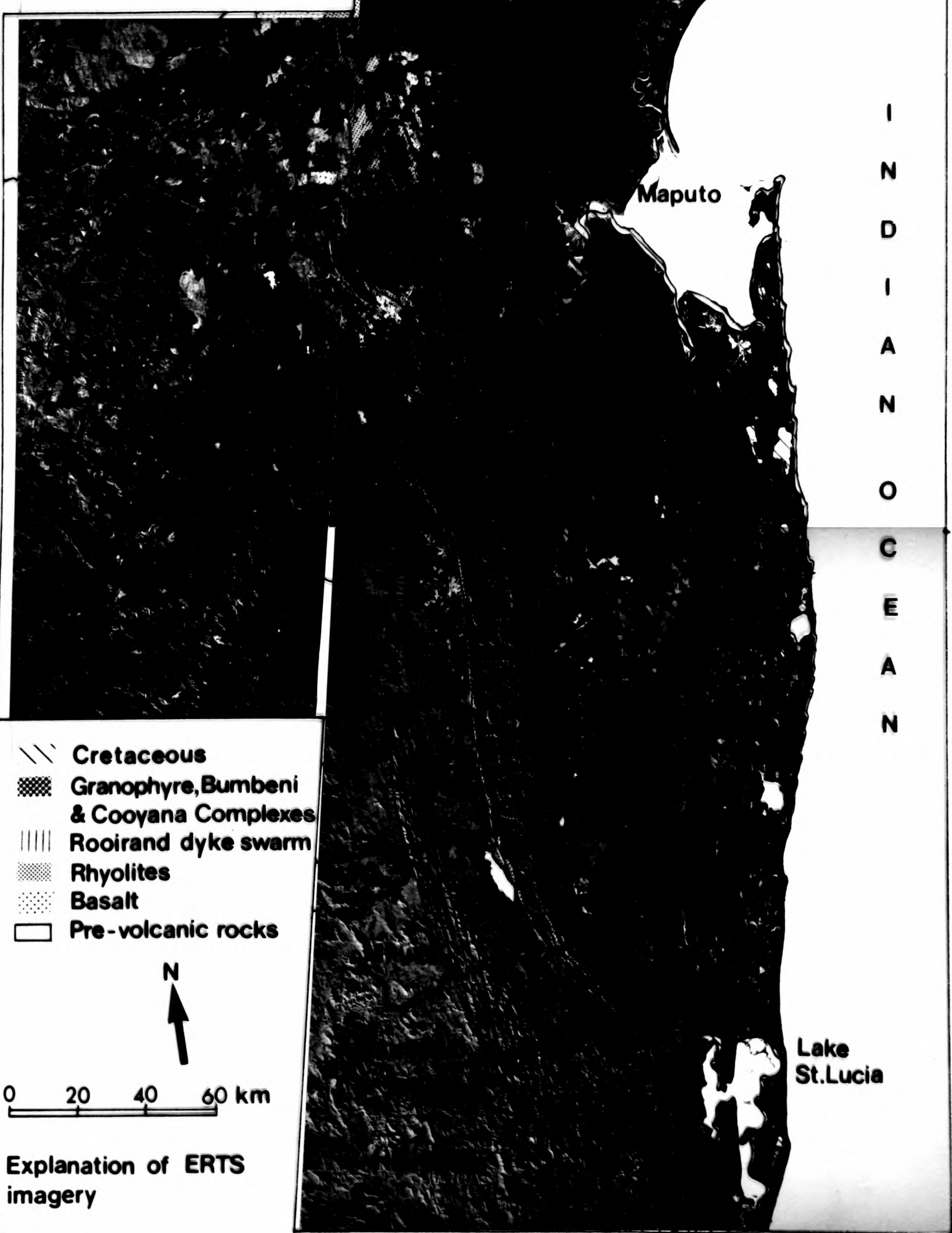
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Geological map of the Southern Lebombo





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ABSTRACT

The Lebombo Monocline is a large flexure of Karroo age extending from Southern Rhodesia to Natal. It consists mainly of a thick sequence of Karroo basalts and rhyolites overlying the Archaean crust. The Karroo strata dip eastwards beneath Cretaceous cover at angles of up to 50° . Four 1:50,000 detailed maps have been produced of the volcanics in Swaziland.

The basalts form a monotonous and poorly exposed succession. They are about 5 km thick but a large dyke swarm and strike-faulting serve to increase the apparent thickness. A few acidic interbeds occur near the base. The rhyolites overlie the basalts with little interbedding. They are also about 5 km thick and can be divided into 30 flows which are interpreted as rheoignimbritic in origin because of the presence of characteristic ignimbrite features and flow-folded tops to the flows.

A dense dyke swarm intrudes the basalts and is younger than most of the basaltic succession; five granophyre bodies intrude the basalts close to the base of the rhyolites.

The monocline is block-faulted in the south but becomes progressively more flexural in the north. The jointing in the rhyolites can be divided into groups related to basement structures, monoclinal flexuring and post-Karroo tectonic activity respectively.

Major and trace element analyses for seventy rocks show that the chemical variation in the rhyolites can be explained by Rayleigh fractionation of a dacitic magma produced by fractional crystallisation of a basaltic precursor. The more basic and most acid rhyolites have compositions more compatible with partial melting of a solid basaltic source. The rhyolite magma was nearly dry but the vapour phase was rich in CO_2 .

The acidic interbeds within the basalts have depleted trace element abundances and relatively radiogenic isotopic chemistry and were produced by crustal anatexis.

The tectonics and magmatism of the Lebombo are discussed in relation to the break-up of Gondwanaland.

LONG ABSTRACT

The Lebombo Monocline is a large flexure of Karroo age which runs along the eastern border of South Africa from Natal in the south to Rhodesia in the north. It consists of a thick sequence of Karroo basalts and rhyolites overlying a thin succession of Karroo continental sediments, which in turn rests on Precambrian crust of the Kaapvaal craton. The Karroo strata dip eastwards beneath a Cretaceous cover at angles of up to 50° .

The Karroo volcanics have been mapped in detail, with emphasis on the acid volcanics, as part of a project concerned with the structural and petrological evolution of the Lebombo and its position in a wider tectonic context. Four detailed 1:50,000 maps have been produced.

The basalts, belonging to the Drakensberg Stage of the Stormberg Series, were the lowest strata studied. They form a thick and monotonous succession and are deeply eroded, forming the flat-lying lowveld. They generally only crop out where baked by dolerite or granophyre intrusions. Individual flows are on average 15 m thick and are usually quartz-tholeiites, fine-grained and poorly porphyritic. Sandstone and tuffaceous intercalations occur throughout the succession and a few rhyolitic interbeds are found towards the base. The thickness of the basalt succession is difficult to estimate because of the presence of a large dolerite dyke swarm and repetition of strata due to strike faulting, but is probably about 5 km.

The basalts are overlain by the rhyolites of the Lebombo Stage with the basal flows of the rhyolites interbedded with the basalts. The rhyolites can be divided into 30 flows, averaging 200 m in thickness and each extending some 50 km along strike. The total thickness of the rhyolite succession calculated from measurements of the width of outcrop and estimates of dip, is

also about 5 km.

Each rhyolite flow can be divided into a number of textural zones which are not all present in all the flows, individual flows generally showing two or more zones. Most rhyolites are underlain by the basal tuffaceous zone which consists of fine, banded tuffs, often with a considerable proportion of aeolian sand. The eutaxitic zone marks the base of most flows showing a gradation from coarse fiamme to fine, streaky banding. With the increase of welding, compaction and devitrification the eutaxitic zone grades into the massive zone which is structureless and accounts for 95 per cent of most flows. At the top of the massive zone contorted flow folding appears, becoming autobrecciated towards the top. The breccia zone occurs at the top of most flows, consisting of extrusive polymict breccias as well as the flow autobreccias. Intrusive tuffisitic breccias also occur.

The general features of the rhyolites: large lateral extent and thickness, and the presence of fiamme are suggestive of an ignimbritic origin. Physical modes for density currents and fluidization are discussed and applied to the eruption of ash-flows. The presence of contorted folding is evidence for movement by flow and an origin similar to that of a rheoignimbrite is proposed for these rhyolites. Thermodynamic aspects of the eruption and emplacement of ash-flows are discussed and the implications for viscosity and thermal convection considered.

The rhyolites have been divided into basic acid, and intermediate, groups on the basis of the abundance of quartz phenocrysts. This mineralogical subdivision and the features diagnostic of the tops and bottoms of the flows have enabled the flows to be mapped out in detail. The rhyolites have in general a uniform north-south strike, parallel to the monoclinial axis but they are warped in two large east-west synforms separated by an antiform.

There are many overlaps and underlaps within the rhyolite succession so that the warps are gradually smoothed out and the uppermost flows have a north-south strike. The flows in the southern synform are in general younger than those in the northern one. There is a slight unconformity at the base of the rhyolites and the interbedded flows are overlapped by the base of the main rhyolite succession at the centre of each of the synforms. This overlap is due to erosion of the basalts and to overstep of the rhyolites westwards across them.

A number of strike faults cut the basalts and underlying sediments and appear to be related to the dyke swarm. They do not cut the main rhyolite succession but do affect the interbedded rhyolites in the north of Swaziland. Small faults oblique to strike cut the interbedded rhyolites.

The rhyolites are all very jointed; some joints have been produced by the formation of the monocline, a number can be related to basement structures, and a prominent north-northeast trend can be attributed to post-Karoo tectonic activity.

The dyke swarm occupies a narrow zone trending obliquely across the basalt outcrop. The dykes have steeply to the west and account for over fifty per cent of the outcrop in places. They are considered to be younger than most of the basalt succession but do not intrude the rhyolites.

The axis of the monocline runs close to the rhyolite-basalt contact, with the steepest dips being developed in the north of the country. Faults and joints with a north-south trend are less pronounced in the north of the country and it seems that the monocline, from being faulted in the south, becomes more flexural in the north.

The structural features of the Lebombo can be interpreted as having formed in an east-west tensional regime related to the rupture of Gondwanaland. The Lebombo may therefore represent the newly-formed continental edge. It is transverse to the major Kaapvaal Precambrian lineaments but parallels the Mozambique belt and seems to have been constrained in its northerly trend by the tectonic grain of the Mozambique belt.

The petrography of the rhyolites is described in detail. Plagioclase (some with sanidine overgrowths), quartz, clinopyroxene, magnetite, pseudomorphs after fayalite, apatite and zircon are present as phenocrysts. The crystals often form lomeroporphyritic aggregates, interpreted as having formed by synneusis, and are often broken and resorbed. The groundmass is usually devitrified showing felsitic, microgranophyric and snowflake textures.

New Rb-Sr and K-Ar data are presented which show that the ages of the rhyolites and dolerites are indistinguishable at 188 ± 5 my. This is in agreement with published K-Ar data and unpublished zircon fission-track work. The dolerites and rhyolites both have an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.704.

Fifty-three samples have been analysed for major and trace elements and a further fifteen for trace elements only. The 'rhyolites' range from dacitic to rhyolitic in composition with one being trachyandesitic. The major elements show regular variations generally attributable to fractional crystallisation of a clinopyroxene plus plagioclase assemblage from a parental dacitic melt. The granophyres fall on the same trends and are considered to be comagmatic. A least-squares petrographic mixing model has been used to quantify the variation, and an extract of 60% plagioclase and 40% clinopyroxene can account for most of the observed variation. Several of the more basic interbedded rhyolites are too potassic and aluminous to belong to the same trend.

The rhyolites are enriched in the elements Nb, Zr and Y relative to most rhyolites described in the literature. The trace element variation is considered in terms of logarithmic plots and shown to be compatible with Rayleigh fractionation. The variation of Rb, Ba and Sr is consistent with the fractionation of a clinopyroxene/plagioclase mixture using published distribution coefficients. When coefficients are calculated relative to Sr for the elements La, Ce, Nd, Y and Nb however, they are greatly in excess of those predicted. La also shows coherent but anomalous behaviour relative to Ce and Nd and is enriched during fractionation while the others are depleted. This abnormal behaviour of trace elements is explained in terms of anomalously high concentrations of them in zircon and apatite. Y and Nb show a similar variation to Zr and their variation is also considered to be mainly controlled by the fractionation of zircon. The granophyres fall on the same trend as the main group of rhyolites but the most acid rhyolites, at the same silica level, are enriched in Y, La, Ce, and Nd and do not seem to be related to the rest of the rhyolites by simple crystal fractionation. They are interpreted as having formed from an enriched source region.

The rhyolitic interbeds which crop out near the base of the basalt succession have depleted trace element abundances and very radiogenic isotopic chemistry. They are interpreted as having formed by anatexis of acid crustal materials. The rest of the rhyolites are interpreted as being derived from a basaltic source by a process of partial melting or fractional crystallisation. The two processes are indistinguishable with the uncertainties in distribution coefficient data; the acid and relatively potassic interbedded rhyolites have compositions more compatible with an origin by partial melting. Trace element considerations indicate that the source cannot have contained residual amphibole or garnet and hence must have been fairly dry. That the rhyolite magma was also dry is supported by the trend towards the dry granite minimum and by the absence of hydrous phenocrysts. The rhyolite

magma required a volatile phase for its eruption as ignimbrites; a CO₂-rich fluid phase is compatible with the observed chemical and petrographical features.

The geochemistry of the Lebombo is reviewed in the overall context of Karroo magmatism and in relation to the potassic northern and sodic southern geochemical provinces. There are several features of the chemistry of the rhyolites from the northern and southern province that are not simply explicable in terms of a regular variation in source chemistry.

A cyclical model relating the volcanism to the break-up of Gondwanaland is proposed to account for the variation within and between the basalts and rhyolites.

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CHAPTER I

INTRODUCTION

I INTRODUCTION

The Lebombo Monocline is a large flexure of Karroo age stretching 600 km from Natal in the south to southern Rhodesia in the north. It runs along the western border of Mozambique and the eastern side of Swaziland (Figure 1.1) forming the eastern margin of the Karroo (U. Carboniferous - L. Jurassic) province. The Lebombo consists of a thick sequence of basalts and rhyolites overlying a thin succession of Karroo sediments resting unconformably on the Archaean granitic basement which constitutes most of Swaziland. The rocks of the monocline have a pronounced easterly dip and are overstepped on the east of the Lebombo by Lower Cretaceous marine sediments (Plates 1.1-3).

The volcanics can be divided into two main groups, each consisting of a basic and an acid member. The lower group is late-Karroo in age and contains a 5 km thickness of basalt, which forms the flat plain of the lowveld, overlain by a 5 km succession of rhyolitic ignimbrite flows which give rise to the Lebombo Mountains. About 30 rhyolite flows can be recognised, averaging 200 m in thickness; they form a peneplained plateau separated from the underlying basalts by a steep escarpment. The upper group is probably disconformable on the lower group and is Lower Cretaceous in age. It crops out mainly in Mozambique with the upper rhyolitic group forming the low hill of the Little Lebombo (Plate 1.4). Rocks similar in age to the Little Lebombo are found in the Bumbeni Complex, Zululand, and in the Lupata Series of the Zambesi Valley.

A large dyke swarm follows the axis of the monocline and forms a low range of hills, the Rooirand, in the south of Swaziland (Frontispiece). A number of granophyre bodies crop out along the basalt-rhyolite contact.

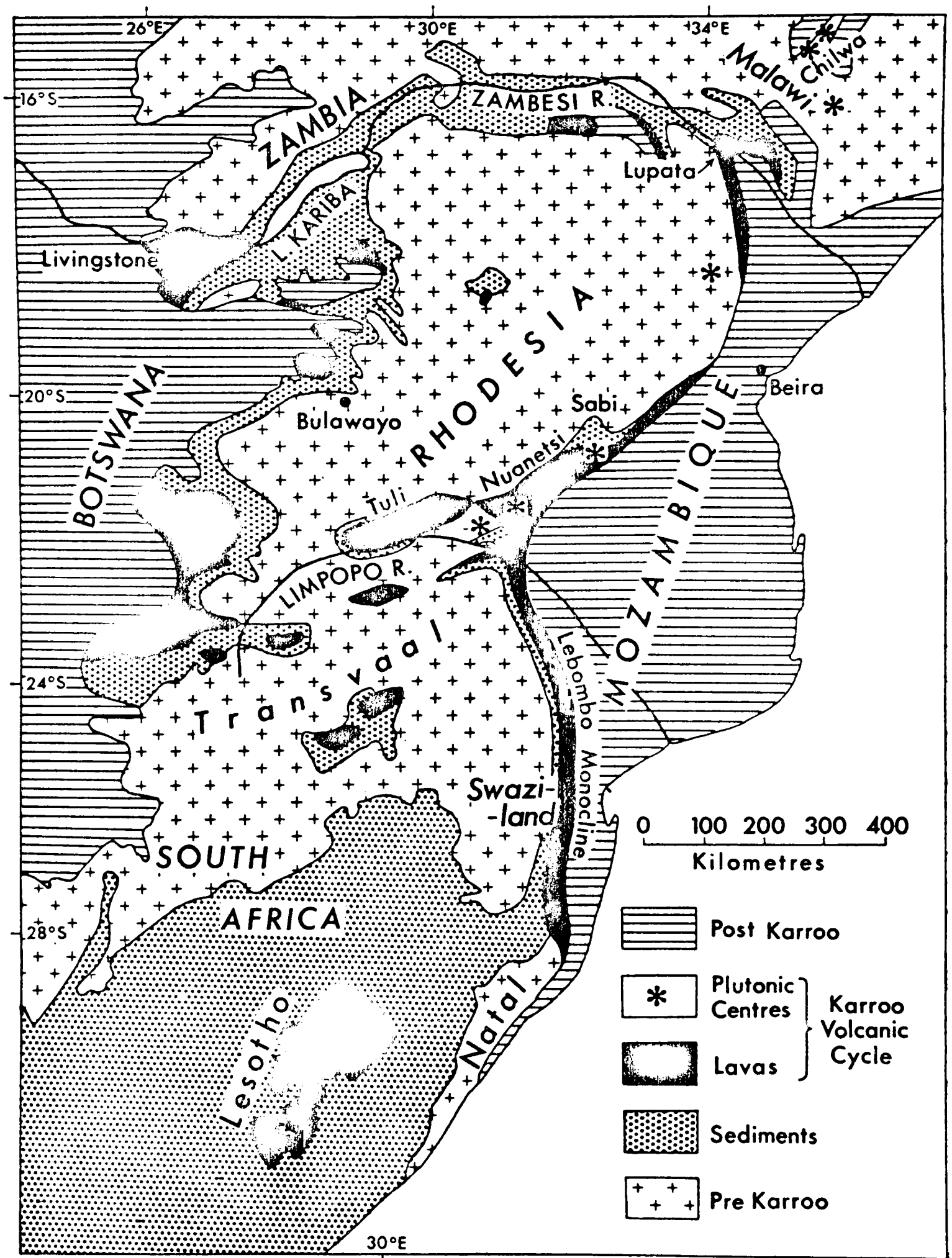


Fig. I.1. Outline geological map of Southern Africa showing the main features of the Karroo province (from Cox, 1972).



PLATE I.1 Lebombo escarpment viewed from the Magwanyana area looking south. Small hills in middleground are formed by the interbedded rhyolites.



PLATE I.2 Escarpment viewed from the Isateki beacon looking towards Big Bend. Parallel rows of hills on the right are formed by the interbedded rhyolites.



PLATE I.3 Aerial view of the Lebombo from 10,000 feet showing the escarpment and inclined rhyolite flows. Coastal plain of Cretaceous sediments at top of photograph. Mkuze river in foreground. (Photograph : Dr. R. Mand)



PLATE I.4 View of the Little Lebombo from the Kruger National Park. Fence in foreground is the South Africa/Mozambique border. (View looking east)

In the south the Lebombo terminates in a series of fault-blocks and arcuate faults and is overlapped by the Cretaceous sediments. North of the Limpopo the Lebombo swings round into the northeasterly trending Sabi monocline which can be followed into the Lupata area of the East African Rift System (Figure 1.1). The Nuanetsi and Tuli synclines are formed at the 'elbow' along the line of the Limpopo belt.

To the west of the Lebombo, the Karroo rocks are relatively flat-lying and are present in great thicknesses in the Karroo basin of northeast Cape Province and, under a thick basalt capping, in Lesotho. Apart from a few outliers the Karroo lavas are generally absent from the central areas of southern Africa due to erosion; only where they are preserved in the Lebombo downwarping is there a continuity of outcrop from one end of the province to the other. An extensive system of dolerite sills cuts the Karroo sediments and, in places, the basement beneath the lavas.

The great majority of the Karroo basalts are olivine-poor tholeiites. Nephelinites, olivine-rich basalts and glassy olivine-bearing lavas (so-called limburgites) are found at the base of the succession in Nuanetsi and olivine-phyric basalts are common in Lesotho and at the southern end of the Lebombo. The upper basalts also largely consist of olivine-poor tholeiites though more alkaline varieties are also present (Assuncao et al., 1962). The Lebombo acid rocks range from andesitic to rhyolitic in composition but the vast majority are rhyodacites with a very uniform composition. More alkaline varieties are reported from Mozambique, the Little Lebombo, the Lupata Series and from the Bumbeni Complex, Zululand.

Cox et al. (1967) have shown that the Karroo basalts of southern Africa can be divided into two geochemical provinces. A northern province includes the Rhodesian localities and is characterised by abnormally high values for

tholeiites of the incompatible elements K, Ti, P, Ba, etc. The southern province covers the outcrops in Swaziland, Lesotho and Natal. These trends may also be reflected in the rhyolites with those in the north being more potassic than those in the south. To define the boundary or transition between these provinces more data are required from the north and central Lebombo as these are the only areas with continuity of outcrop.

Manton (1968) has analysed rhyolites and basalts from Nuanetsi and Swaziland for strontium isotopes. Rhyolites from both areas define isochrons giving ages of about 200 my with initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.708 and 0.704 respectively. These initial ratios are similar to those obtained from basaltic rocks in the two areas and Manton's conclusion was that the rhyolites were evolved from the mantle. An origin from crustal materials depleted in radiogenic strontium cannot, however, be ruled out on the basis of these data.

At the outset of the project the main problem was one of geochemistry: how the Swaziland rocks fit into the geochemical province hypothesis and how such apparently large volumes of acid magma can be generated. It was generally accepted that the basalts were derived from the mantle but no consensus had been reached as to the origin of the rhyolites. The large volumes of rhyolite suggested some form of crustal melting process with heat supplied by the basic magma but this would be incompatible with the isotopic evidence unless a depleted lower crust was involved. It was necessary to examine these hypotheses in more detail by studying the geochemistry of the rhyolites to see if it followed any coherent pattern which could be attributed to either crustal fusion or to some process involving the mantle. If the origin of the rhyolites could be determined and some information on source composition obtained, it should be possible to look at the province problem with the geochemical differences pushed a stage

further back into the crust or mantle. The intrinsic differences between the provinces would then be a function of crust or mantle inhomogeneity or a difference in process perhaps linked with tectonic situation.

Before much weight can be placed on geochemical data, the distribution, volume and field relationships of the various rock-types need to be known. The presence of large-scale interbedding between rhyolites and basalts, for example, would make a significant difference to gross petrogenetic models. The individual flows should be mapped to work out their extent and volume so that samples can be correlated over large distances. Any faulting present must also be assessed as in this type of linear tectonics there is often considerable faulting parallel to strike which can repeat the succession. Strike-faulting is apparent in the sediments beneath the basalts and its importance in affecting the apparent thickness of the volcanics needs to be known. Dilation due to intensive dyke intrusion may also increase the apparent thickness of the basalts.

The linear nature of the monocline poses structural problems as the rocks are only exposed in linear strips and no three-dimensional picture of the flows can be built up. The lateral extent of the flows is largely a function of the mode and type of eruption. Most workers have concluded that the acid rocks are ignimbritic in nature on the basis of their large lateral extent but Wachendorf, working in Mozambique, decided that the rocks originated as lava flows on the basis of fluidal textures and that the outcrop pattern is caused by stacking of elongate 'ribbons' of lava formed by viscous extrusion out of parallel fissures. If the rhyolites are truly ignimbrites, however, then the east-west extent can be assumed to be of the same order of magnitude as the north-south outcrop. The rhyolites are unusual in having such large apparent thicknesses and, taking into account the large lateral extent, volumes. A mechanism needs to be

proposed to account for this and the fluidal structures noted by Wachendorf. The outcrops themselves should be able to furnish information on the physical conditions of the magma at the time of eruption.

All these problems are connected in some way with the regional tectonic situation. Further constraints are placed by the geology of the neighbouring areas that formed Gondwanaland and any conclusions concerning structural implications reached in the Lebombo need to be reviewed in the overall context of Gondwanaland and its break-up.

Note on stratigraphic nomenclature

The stratigraphical nomenclature of South Africa is at present in considerable confusion awaiting the report of the South African Commission on Stratigraphy. Several preliminary reports have been published (e.g., Int. Subcomm. on Strat. classification report 7b, 1972). The use of the hierarchical lithostratigraphical terms Group, Formation, Member, Bed is recommended and these terms have been used by Bristow (1976). Difficulties arise from the presence of geochemically distinct units within the rhyolite sequence and the geochemical variation along the length of the Lebombo in petrographically very similar rocks. The continuation of the rhyolite outcrop for 600 km renders a classification based on the type sections in the southern part of limited use.

To avoid confusion the old terminology has been followed throughout this thesis. Table I compares this older nomenclature with that used by Bristow (1976).

The term *rhyolite* has often been used to denote an acid lava. In this thesis it is used in its strict sense to mean *acid extrusive* and carries no connotations as to the mode of eruption. The terms *lava*, *ignimbrite* are

used where the mode of emplacement is discussed. Similarly *rhyolite*, as used here, has no geochemical connotations and embraces the rock-types *dacite*, *rhyodacite* and *rhyolite*.

TABLE I.1

OLD NOMENCLATURE (used in this thesis)		BRISTOW (1976)		
S T O R M B E R G S E R I E S	LITTLE LEBOMBO RHYOLITES	L E B O M B O · G R O U P	MSUNDUZI RIVER FORMATION	BUMBENI + COOYANA COMPLEXES
	UPPER BASALTS			MPILO BASALTS MEMBER
	LEBOMBO STAGE (Rhyolites)		JOZINI FORMATION (Rhyolites)	
	DRAKENSBERG STAGE (Basalts)		MFOLOZI RIVER FORMATION (Basalts)	
	CAVE SANDSTONE STAGE	CAVE SANDSTONE FORMATION		

CHAPTER II

GEOGRAPHICAL SUMMARY

II GEOGRAPHICAL SUMMARY

II.1 Introduction

The area mapped lies along the eastern border of Swaziland and extends from the Ngwavuma River in the south to the Mbuluzi River in the north. Geological mapping has been extended to the northern border of Swaziland by photo-interpretation. At the time of the fieldwork the political situation in Mozambique made it impossible to obtain entry permits. The international borders are well defined by double foot-and-mouth control fences and are patrolled continuously.

II.2 Access

Access to the area is good with the main gravel road from Mbabane climbing the Lebombo escarpment to Stegi. Dirt roads lead from Stegi north to Mhlumeni and the Du Preez farms near Buchanan and south to the Usutu Poort. Further rough tracks lead to the border in the Maphungwane area and many of the farms have bulldozed tracks which can be used to reach otherwise inaccessible areas. The recently built Lourenço Marques railway runs along the banks of the Mbuluzi and the parallel service road proved to be just passable and furnished access to the railway cuttings. A crude double-fenced track, the 'Red Line', was built along the foot of the escarpment the length of the country to act as an insulation zone against foot-and-mouth disease. In places this track is in excellent condition and proved invaluable in the area north of Stegi. Sledge tracks within the native areas could be adapted to allow a car to pass but would prove impassable when wet.

II.3 Administration

The area falls entirely within the Lebombo administrative district and comes under the jurisdiction of the Stegi District Commissioner. The area contains tracts of European farmland and large areas of Swazi Nation

Land. The European farmers proved hospitable and willing to grant permission to walk over their land. The Swazi Nation Land in this area is composed of eight separate tribal areas each ruled by a Swazi chief. Permission to enter their land needed to be obtained well in advance so their local headmen (indunas) could be informed and alert the local people. The Swazi is naturally inquisitive and friendly, but suspects strangers of trying to steal his cattle and an interpreter was found to save a great deal of embarrassment.

II.4 Population and industry

The population is sparsely distributed in the lowveld but the Lebombo plateau is relatively densely populated, especially adjacent to major roads. The Swazis live in typical Bantu mud-and-wattle huts, thatched with elephant grass, grouped to form villages or kraals each representing a family unit. Local employment is offered on the cattle farms, sugar-cane plantations and cotton fields but the majority of Swazis are unemployed and self-sufficient. Mealies form the staple diet and are grown extensively, along with pumpkins, on every cultivatable patch of land. Goats and cattle are kept as a capital investment and status symbol.

The European farms are almost wholly for beef, though cotton, mealies and several species of nut are grown. There are extensive sugar-cane and citrus plantations in the lowveld around Big Bend and Mhlume.

The game in the area has largely been eaten but some of the farms still have small herds of impala, kudu, zebra and smaller antelopes. The only predators are leopards, hyenas and jackals. Baboons are common and a menace to mealie fields. Snakes, puff-adders and black mambas being the commonest, crocodiles and hippos are found in the lowveld.

II.5 Relief and climate

Swaziland can be divided into four physiographic zones which run from north to south in more or less parallel belts. There are, from the west, the highveld, middleveld, lowveld and the Lebombo. The highveld and middleveld are formed entirely of Precambrian rocks and are incidental to the area under discussion.

The lowveld is a low-lying undulating plain between the middleveld and Lebombo. It averages 15 km wide and lies 150 to 500 m above sea-level, and is continuous from the Lebombo flats of Zululand through Swaziland and northern Transvaal to Rhodesia. It is near tropical in climate with a low annual rainfall of 50 to 90 cm. Malaria is still endemic in a few places and bilharzia is ubiquitous. It is covered in Karroo acacia with larger acacias and fever trees along the river courses. Aloes are common, especially around Big Bend.

The lowveld is largely underlain by eroded basalt flows with only the interbedded rhyolites, granophyre bodies and, in the south, the Rooirand dyke swarm forming areas of any relief. The presence of strike faulting in the Karroo sediments in the south of Swaziland and in the area north of Stegi also produces diversification of the landscape.

The lowveld is representative of an erosional peneplain, probably Quaternary-Recent in age (King, 1951) and can be considered the hinterland equivalent of the coastal plain lying to the east of the Lebombo.

The Lebombo forms an elevated plateau on the east of the lowveld and separated from it by a steep escarpment. This escarpment is up to 300 m high in the south of Swaziland, but becomes less prominent in the north due to complex interbedding and faulting of rhyolite within the basalt. West of Stegi

the escarpment is developed on the west of the granophyre pluton in the basalts and in the extreme north it is developed on the west of the Mananga granophyre.

The almost geometrical straightness of the escarpment in the south (frontispiece) led Hunter and Urie (1959) and de Blij (1960) to propose a rift-valley origin for the lowveld. There is little evidence for this and, although the escarpment is straight in the south, in the area to the north and immediately to the south of Stegi the arcuate nature of the escarpment and the obvious control on the topography by the interbedded flows and the granophyre bodies militates against an origin for the lowveld by rifting or large scale faulting. Long strips of rhyolite crop out between Stegi and the Pongola River forming low ridges (frontispiece and Plate I.1). These seem to be interbedded flows and do not appear to be controlled or caused by faulting on a large-scale. King (in discussion, de Blij (1960)) supports a purely erosional origin for the lowveld on geomorphological grounds.

The Lebombo Mountain Range is bevelled by an erosional peneplain with a gentle eastward slope (see geological cross-sections (Diag. 1) and Plate II.1). This 650 m surface also forms the middleveld and has been called the Late Cenozoic bevel (King, 1962, p. 298) or the African surface (King, 1951, p. 297) and is Late Tertiary in age. Patches of gravel, ferricrete and occasional limestone (King, 1951, p. 297) are found on the surface. The height of the Lebombo is fairly constant in Swaziland with hills reaching 650 m along the length of the range but falls rapidly to the north till the Lebombo is present as a low ridge to the north of the Komati River. South of Swaziland the elevation decreases to 300 m in the Msunduzi River area, beyond which the range dwindles to a series of small hills with an elevation of 200 m (Bristow, 1976).



PLATE II.1 View of the Lebombo looking west from near the Mozambique border, east-northeast of Stegi. Deeply eroded valley of the Manzimnyame river forming steep rhyolite scarp faces. The peneplained nature of the Lebombo can be clearly seen.

The Lebombo is sub-tropical with an annual rainfall of 75 to 115 cm, falling largely in the summer. It has very little soil on top of the rhyolites and except for the escarpment and deep river valleys, is sparsely vegetated. Small clumps of acacia bush grow on the ridge crests and larger acacias, ferns and various creepers grow in the valleys. There is a large ironwood forest in the Mbuluzi Valley. Many of the kranzes in the east are covered in cycads and aloes. Lines of blue gums and wattles mark the major farms.

II.6 Drainage

The drainage pattern of the Lebombo is very unusual in that a total of thirteen major rivers cut through the Lebombo Range at right angles to the strike. The southern part contains six: (from the north) Mbuluzi, Usutu, Ngwavuma, Pongola, Mkuze and Msunduzi, the first three being in Swaziland (frontispiece). Most of these rivers display incised meanders and are obviously superposed on the present topography.

The lowveld is drained by north-south tributaries of these east-west rivers; the escarpment being formed by scarp retreat (King, 1951). The Lebombo range itself is drained by a number of consequent streams flowing down the easterly dipping peneplain. They show strong structural control (Pl. II.2) and a pronounced trellis pattern due to the interplay of regional joints and the bedding in the rhyolites. On the eastern side these minor streams also become incised (Pl. II.1) and some, tributing to the major rivers, may also be superposed.

The Lebombo therefore has a history of almost continuous uplift since Tertiary times. The Tertiary rivers eroded the basalt flows and incised through the Lebombo rhyolites. Consequent streams developed on the rhyolites and these have also become incised in the east, indicating continuing rejuvenation associated with further uplift.



PLATE II.2 E.R.T.S. photograph (Band 5) of the northeastern corner of Swaziland. Northwest-trending lineament on the left of the picture is the northern Swaziland border. A prominent north-north-east joint set is apparent, controlling the minor streams of the Lebombo. The nature of the escarpment is also clear; straight and due north in the south of the country, becoming warped by the structure of the rhyolite and dividing into three on the northern border of Swaziland. Length of photograph approximately 270 km. (E.R.T.S. E-1047-07184-5-01, negative print)

CHAPTER III

PRE-VOLCANIC ROCKS

Geological Sketch Map of Swaziland

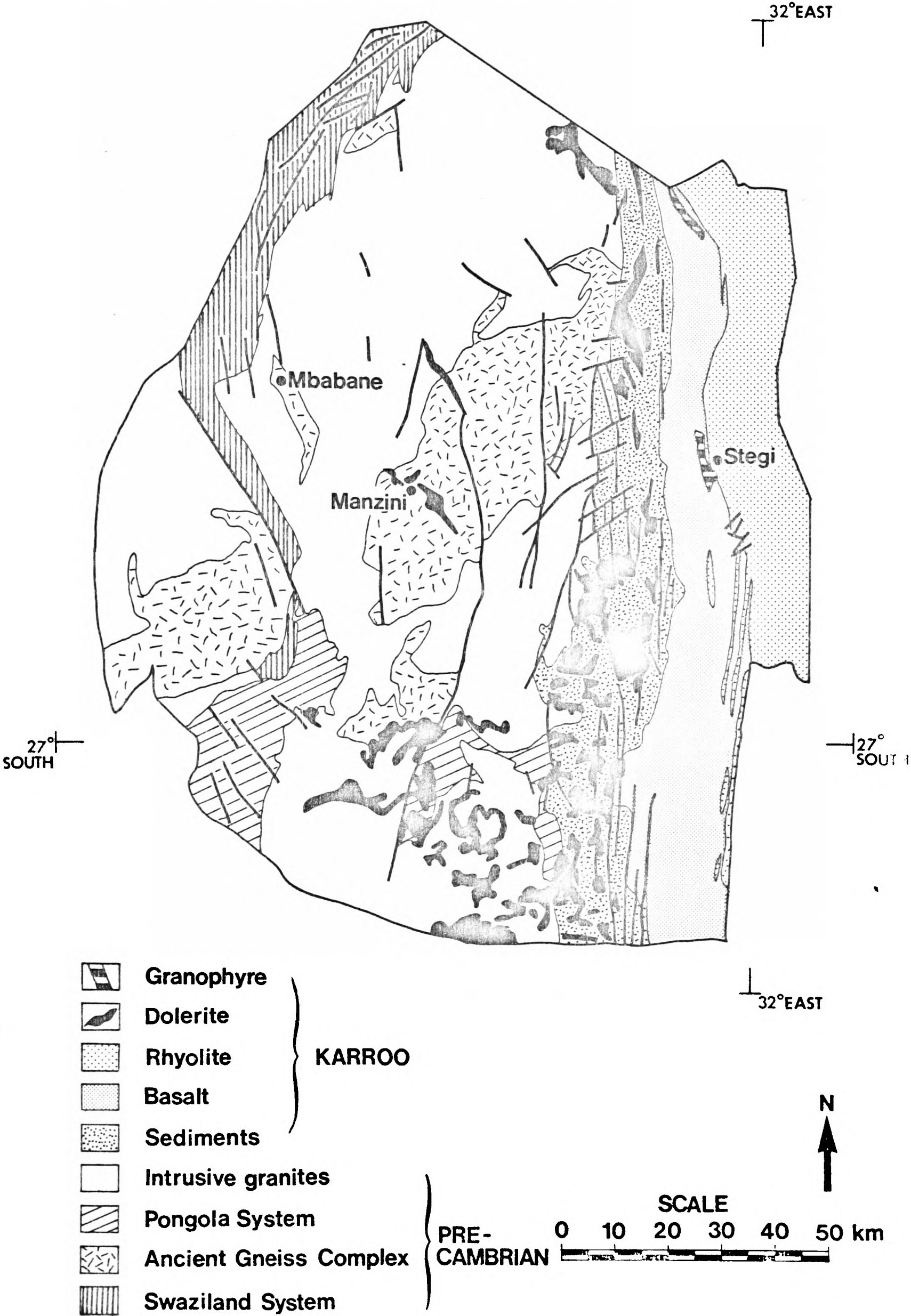


Fig III.1

III PRE-VOLCANIC ROCKS

The fieldwork was entirely concerned with the Stormberg Volcanics which crop out in the east of Swaziland. No rock unit, however, is wholly independent of those beneath it and the older units are discussed to present an overall view of Swaziland geology and place the Karroo volcanics in their regional context. A generalised stratigraphy of Swaziland is given in Table III.1, and a geological sketch map in Fig. III.1.

III.1 Archaean Basement

Archaean rocks form the highveld of Swaziland and extend through the middle-veld till they are overlain by the basal Karroo sediments. They can be divided into three main divisions:

- (1) The granite-greenstone belt of the Swaziland Supergroup* and Ancient Gneiss complex.
- (2) The Pongola Supergroup*.
- (3) Intrusive granites, gabbros, etc.

Hunter (1961, 1974a, b) presents a concise account of the Archaean and Proterozoic rocks and only a brief summary is given here.

(i) Swaziland Supergroup and Ancient Gneiss Complex

The Swaziland Supergroup comprises three major subdivisions: the Onverwacht, Fig Tree and Moodies Groups. These form the greenstone belt of the Barber-

* The terms Supergroup and Group are used in order to conform with the usage by Hunter (1974a) in spite of his reservations and those expressed in Section I.1.

TABLE III.1
Generalised Stratigraphy of Swaziland

(after Hunter, 1961)

AGE		
Tertiary to Recent		Alluvium, ferricrete
Karoo System	188 m.y.	Stormberg Series
		Beaufort Series
		Ecca Series
	± 400 m.y.	Dwyka Series
		<i>unconformity</i>
Pongola System		Mozaan Series
	< 3.06 b.y.	Insuzi Series
		<i>unconformity</i>
Swaziland System		Moodies Series
	3.4 - 3.0 b.y.	Fig Tree Series
		Onverwacht Series
<u>Intrusive Rocks</u>		
Post-Karoo		Dolerites, microgranite and granophyre
Post-Archaeon		Diabase dykes
Archaeon		Post-Mozaan
		Post-Jamestown
		Granites

ton Highlands in northwest Swaziland. The ultramafic and mafic lavas of the Onverwacht Group are characterised by high MgO and $\text{CaO}/\text{Al}_2\text{O}_3$ ratios and low alkali contents and are classed as komatiites (Viljoen and Viljoen, 1969). The majority of the lavas are subaqueous in origin. Layered mafic-ultramafic intrusions forming sill-like bodies have also been recognised in the Onverwacht Group.

The sediments of the Fig Tree and Moodies Groups are mainly immature graywackes showing progressive compositional changes which may indicate the progressive unroofing of a metamorphic terrain (Condie et al., 1970).

The structure of the Swaziland Supergroup is that of a broad synform with a northeasterly trend. Larger faults divide the area into a number of fault blocks. The Supergroup displays greenschist facies metamorphism with superimposed contact metamorphism of amphibolite and pyroxene hornfels grade due to later intrusion of granite.

The Ancient Gneiss Complex underlies central Swaziland to the southeast of the greenstone belt and consists of a bimodal association of interlayered leuco-tonalitic gneisses and amphibolites and a complex suite of metamorphites including quartzites, ironstones, gneisses and granulites. The contact between the Ancient Gneiss Complex and the Swaziland Supergroup is either faulted or obscured by later granite intrusion. The Ancient Gneiss Complex may be either the deep crustal equivalent of the Swaziland Supergroup, implying a large difference in tectonic height across the contact, or may represent an earlier crustal event. The two groups may be distinguished on the basis of their metamorphic and structural styles.

Rb/Sr ages from the Ancient Gneiss Complex have a spread from 3.4 ± 0.09 to $3.14 \pm 0.006 \times 10^9$ y. The low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (0.7006 ± 0.012) is

indicative of no significant crustal contamination (e.g. discussion by Moorbath and Pankhurst, 1976). Ages from the Swaziland Supergroup range from 3.36 ± 0.07 to $2.54 \pm 0.05 \times 10^9$ y. The younger ages obtained from the Onverwacht lavas may have been reset by later intrusive events and probably do not reflect the age of extrusion. The geochronology does not add to the field data and there is as yet no evidence bearing on the relative ages of the Swaziland Supergroup and the Ancient Gneiss Complex.

(ii) Pongola Supergroup

The lavas and sediments of the Insuzi and Mozaan Groups were formed in a sedimentary basin on the Kaapvaal Craton. The volcanics of the Insuzi Group form two small areas in southern Swaziland and are overlain in places by the Karroo. They are mainly andesitic in nature with occasional acid lavas and sedimentary intercalations. The Insuzi Group thins out northwards and is overstepped by the Mozaan Group. The thickness of the Insuzi Group can not be estimated because of the presence of extensive faulting, complex structure and the lack of marker horizons but is unlikely to exceed 3,000 m.

The Insuzi Volcanics are succeeded by a series of alternating shales and quartzites forming the Mozaan Group. These form rugged stepped country in the south of Swaziland due to the harder quartzite bands. A thin layer of amygdaloidal lavas lies at the top of the succession. The whole is structurally complex, being repeatedly intruded by granite, but displays a general lack of metamorphism though garnets are developed in some of the aureoles. The total thickness is about 3000 m though in places the whole of the Pongola Supergroup may reach 10 km in thickness.

The Pongola Supergroup has not been dated though it rests with a sedimentary contact on granite dated at 3.06×10^9 y (Allsop et al., 1962).

(iii) Intrusive Granites

A series of medium- to coarse-grained tonalitic diapiric domes crop out round the southwestern edge of the Barberton greenstone belt. They are intrusive into the Onverwacht Group and cause amphibolite facies metamorphism. Banded gneiss similar to that of the Ancient Gneiss Complex occurs between these diapirs and the lowest members of the Onverwacht Group. The banding of the gneisses becomes progressively more contorted as the contact with the tonalite domes approaches. Partial melting of the tonalite accompanies the deformation and veins and lenses of coarse-grained tonalite are produced. This deformation and partial melting may mark the beginning of the production of the potassic granites forming the Lochiel Plateau to the south. There are no chronological data for these tonalites.

Coarse-grained plutonic rocks ranging in composition from ultramafic to acid are intrusive into the Ancient Gneiss Complex. This Granodiorite Suite shows intrusive contacts with the gneiss and a wider range of composition than the diapirs. It forms elongate northeasterly trending bodies which closely parallel the main synform of the Swaziland Supergroup. It is cut by granites of the 3.07×10^9 y event.

The Lochiel Granite (formerly called the Homogeneous Hood Granite by Hunter (1968)) forms a large plateau to the southeast of the outcrop of the Swaziland Supergroup. It is a potassic granite and sheet-like, forming a hood over the Ancient Gneiss Complex. The contact between the two is a complex of granite and pegmatite dykes and sheets forming an apparently migmatitic terrain.

Similar granites are found in eastern Transvaal (Dalmein-type). The Lochiel granite can be traced into the Transvaal where it is overlain with a sedimentary contact by the Pongola Supergroup. It has been dated at 3.07×10^9 y

and provides a maximum age for the Pongola.

Potassic granites intrude the Pongola Supergroup in Swaziland and eastern Transvaal and have been dated at 2.7×10^9 y. They form sheet-like bodies (Pongola Granite) or discordant plutons (Mpageni and Mooihoek Granites). The Pongola Granite is indistinguishable from the Lochiel Granite as regards lithology and mode of occurrence. A number of granites (including the Mpageni) are intrusive into the Pongola Granite.

Granites of this age (around 2.6×10^9 y) are very widespread within the Kaapvaal Craton and it is apparent that this age represents a large thermal event within the crust, perhaps indicating a peak in the rate of granitic additions to the crust.

III.2 Karroo System

(i) General

In spite of its relatively large thickness the Karroo System is confined to a strip along the eastern lowveld and Lebombo Mountains because of the monoclinical flexuring. The generalised stratigraphic column is given in Table III.2. The succession is known in greater detail in the south of Swaziland due to the superior outcrop. Recent detailed drilling in the Eccra coalfield has greatly added to the structural and stratigraphic information available. The thickness of the Karroo sediments is of the order of 1000 m, that of the volcanics in the region of 6 km. The thickness of the basalts is difficult to estimate due to the poor exposure and lack of both marker horizons and structural information.

(ii) Dwyka Series

Tillites and shales of the Dwyka Series are found as isolated patches at

TABLE III.2
Stratigraphy of the Karroo System, Swaziland
 (after Hunter, 1961)

<u>SERIES</u>	<u>BEDS</u>	<u>LITHOLOGY</u>	<u>THICKNESS IN METRES</u>
Stormberg	Lebombo	Rhyolitic tuffs, breccias and welded tuffs	>5000
	Drakensberg	Basalt	~2000
	Cave Sandstone	Massive sandstone, fine-medium grained	60-100
	Red Beds	Mudstones, grey sandstones	30-40
	Molteno Beds	Quartzites	50-65
Beaufort		Grey arenaceous shales	30-40
Ecca	Upper Ecca	Carbonaceous shales, coals, sandstones	100-150
	Middle Ecca	Sandstones, shales, carbonaceous shales, coals	500-650
	Lower Ecca	Grey-black shales	50-100
Dwyka		Tillites, shales, varved shales	0.50
TOTAL (without volcanics)			820-1195

Sicunusa, in the Ngwavuma Valley and in the Mhlosheni area. They are best developed east and southeast of Hluti and irregular outcrops are found overlying the granites at the base of the Karroo outcrop. The erratic distribution of this Series indicates that the pre-Dwyka topography was fairly rugged although it is not known whether the absence of tillite in certain areas is due to faulting or non-deposition.

The tillite, when fresh, has a hard greenish flinty matrix containing angular pebbles and boulders of granite, pegmatite, gneiss, quartzite, quartz and carbonaceous material. Bands of sandstone and varved clays are found interbedded in the tillite. Small areas of glaciated pavement occur in the Manziwayo Valley but no striation directions can be determined. There is no doubt that the Dwyka is glacial in origin and is part of the extensive Upper Carboniferous to Lower Triassic Gondwana glaciation.

(iii) Ecca Series

The Lower Ecca shales are best developed east and southeast of Hluti where they are preserved in dolerite-capped kopjes. They have not been recognised in the north of Swaziland. They rest on the Dwyka or directly on Insuzi lavas or Archaean granites and gneisses. They are fine-grained, thinly bedded and have been interpreted by Davies (1961) as lacustrine.

The Middle Ecca, consisting of sandstones, shaly sandstones and shales with interbedded coal seams, makes up the major part of the Karroo sediments. The base is marked by a succession of feldspathic, often cross-bedded sandstones with grit bands but no carbonaceous matter. This is succeeded by a sequence of alternating shales and sandstones and then by the main coal zone. This latter consists of a succession of coal seams and arkosic sandstones. Three of the seams are of economic significance and are known as the main, No. 1 upper and No. 2 upper. The main seam averages 3 m in thickness but

can reach as much as 9 m. The Nos. 1 and 2 upper average 110 cm and 80 m respectively. The coal zone is succeeded by a sequence of sandstones and shales.

The Upper Eccra shales have been identified in southern Swaziland but no outcrops have been found north of the Usutu. Recent drilling in the coal measures confirmed the existence of a 100 - 150 m sequence of carbonaceous shales and coals correlated with the Upper Eccra. The sequence consists of shales, carbonaceous shales, sandstones and coals. The Upper Eccra coals are rarely of economic importance being thin, shaly and pyritic.

Ryan (in Haughton, 1969) has divided the Eccra into four main facies of which the Northern (or coal-measure) facies crops out in Swaziland. He stated that "the Northern facies consists of sediments mainly deposited under fluvial-deltaic conditions on a stable shelf. Their major source lay to the east and north-east of the existing east coast and the rocks from which they were derived were predominantly granitic in composition".

(iv) Beaufort Series

Overlying the Upper Eccra is a micaceous shale unit 40 m thick which is overlain by the Molteno sandstones. The contact with the Upper Eccra is sharp and a mud-flake conglomerate is developed. The shale does not crop out and is found only in borehole cores. It has been correlated with the Beaufort Series of South Africa on the basis of stratigraphic position and lithological similarity. They are considered to be lacustrine in origin (Davies, 1961).

(v) Stormberg Series

This Series consists of three sedimentary units overlain by volcanics.

(a) Molteno Beds

The Molteno Beds at the base of the Stormberg Series are well exposed and consist of white, coarse-grained, slightly ferruginous sandstones with interlayered beds of shaly sandstone. They have commonly suffered secondary silicification. The maximum recorded thickness is 65 m but the outcrop width on the map is considerably broadened due to migration of boulders to the west.

(b) Red Beds

The Red Beds crop out only in the extreme south of the country but their presence has been confirmed in boreholes as far north as Stegi. They are reddish or maroon siltstones and mudstones and are massive except for a central grey bedded sandstone. They contain small calcareous nodules and thin pebble-conglomerate bands. Davies (1961) interpreted them as formed in a lacustrine environment.

(c) Cave Sandstone

The Cave Sandstone forms a nearly continuous line of outcrops at the base of the basalt. It is a distinctive white fine-grained massive sandstone some 60 to 100 m thick. Large scale cross-bedding is indicative of an aeolian origin for this bed though minor fluvial bands occur at the top and bottom (Beukes, 1970). Basalt flows are interbedded at the top of the unit and sandstone intercalations are found throughout the basalt succession. The Cave Sandstone is present almost everywhere beneath the Karroo lavas in Southern Africa, reaching thicknesses of 300 m in the central Karroo basin, and represent a widespread transition to desert conditions.

(vi) Effects of Karroo dolerites

The sediments are extensively pervaded by dolerite sills and, near the monoclinical axis, dykes. Low grade contact metamorphism is widespread with the production of flinty zones and, in areas of dense dyke intrusion, granular hornfelses. Anthracites produced by the metamorphism of coal seams are mined commercially. Hydrocarbons were released during metamorphism and condensed in overlying beds.

(vii) Structure

The structure of the Karroo sediments is governed largely by the influence of the later developed monoclinical flexure. All the beds have shallow dips to the east, being lower in the west but increasing to 15° in the Stormberg. Major strike faults occur in the Karroo sediments causing repetition of strata and outcrops of Stormberg rocks within the Eccca. The Eccca coal measures are affected by block and trough faulting (unpublished company reports held by the Swaziland Geological Survey and Mines Department) related to the general tension and rifting associated with the formation of the monocline.

CHAPTER IV

STORMBERG VOLCANICS

IV STORMBERG VOLCANICS

IV.1 Introduction

The Stormberg Volcanics form the thickest unit within the Karroo System. They reach their maximum development in the Lebombo Region where a thickness of 10 km is probable. In Swaziland they have an easterly dip due to the monocline and form a broad north-south strip on the eastern side of the country. The basal unit of basalt rests conformably on the Cave Sandstone and, with minor acid flows, forms the lowveld. A thick sequence of rhyolitic ignimbrites forms the Lebombo region with a steep escarpment on the west. The top of the rhyolites, the Upper Basalts and the rhyolites of the Little Lebombo crop out within Mozambique and access was not possible during the present study.

The fieldwork embodied in this thesis was undertaken in two field seasons: August-October 1974 and May-October 1975. Most of the time was spent in Swaziland and four 1:50,000 maps covering the entire Lebombo region in Swaziland have been produced (Map Sheets 1 - 4). Mapping was done on 1:30,000 aerial photographs (mainly 1961 photography, with some 1972 photography in the areas of the railway) and the information transferred to the 1:50,000 topo-cadastral sheets which served as base maps. One unpublished and four published maps relating to the basalt outcrop and underlying sediments were made available by the Director of the Swaziland Geological Survey and information from them has been used to supplement the field-mapping and photointerpretation.

Emphasis has been placed on the acid volcanics, and individual rhyolite flows were traced out to define the stratigraphy and structure within the sequence. The Swaziland localities were sampled in detail though areas

outside Swaziland have been left to workers from the Universities of Natal, Durban and Cape Town. Trips to examine rocks of the Lebombo in South Africa were made with members of the Department of Geology, Durban, to the south of the Kruger Park and to the Mkuze area of Zululand, Natal. In addition, a week was spent measuring detailed sections along some recently excavated canals at the J.G. Strydom Dam at Jozini.

IV.2 Previous work on Lebombo and Karroo volcanics

The early accounts of Lebombo geology (Cohen (1874), Wilson-Moore (1897), Molengraaff (1898), Garrard (1914), Kynaston (1906), Henderson (1909), Young (1920)) were largely descriptions of journeys to and from Lourenço Marques with the geology as a diversion. Prior (1910) reported some petrographic data on dolerite and rhyolite collected in the Lebombo. Henderson (op. cit.) mentioned that the rhyolites are 'in places decomposing, weathering and exhibiting a marked tendency to form pisolitic limonite {ferricrete}'. He also reported that 'the dip of the flowage lines appears to become steeper as one proceeds westward towards the basin rocks which underlie the rhyolites'. Young (op. cit.), also describing a wagon journey, commented on the great thickness of the volcanics and failed to find any evidence for repetition by faulting. He described rocks from the Little Lebombo and some of the interbedded breccias and sandstones. Anderson (1901, 1904, 1907) published a series of reports on the geology of Zululand. He also produced some good sketch-maps showing the relationship between the acid and basic rocks. He included chemical analyses of some rhyolite samples and commented on their high percentage of silica.

The first detailed account was by du Toit (1930) where he described the Lebombo as a region of tension with the monocline formed along a line of weakness. He added to the details reported by previous workers, referring to the petrography of the volcanics, and agreed with them in stating that the

acid lavas were differentiation products of the magma from which the basalts were derived. Du Toit paid particular attention to the intrusive rocks to work out the volcanic cycle and concluded that the larger intrusive granophyre bodies represented consolidated feeders for the acid lavas, and the dyke swarm feeders for the basalts. He interpreted the dyke swarm as representing the axis of the monocline and, assuming that dykes deviate little from the vertical upon injection, stated (p. 209):

"The fact that throughout the Lebombo region the dykes are but rarely vertical, and dip westwards at angles of from 85-70 degrees, proves of utmost moment in signifying that the Lebombo Monocline came into being during the eruption of the lavas. Whereas the dip of the sediments or volcanics is low, e.g. in the extreme east and west, there is a tendency for the dykes to cut the latter almost at right angles, but where the tilting of the strata is considerable, that is to say towards the axis of the monocline, the intrusions are no longer roughly perpendicular to the flows, but intersect the latter at angles so small at times as only 60 degrees, e.g. to the east of Ressano Garcia

..... In interpreting these phenomena it is clear that

- (a) all the warping could not have preceded the dyking, seeing that the effusions themselves must have ascended by means of the dyke planes
- (b) all the warping could not have followed the dyking otherwise in the axial part of the monocline the intrusions would have shared equally in that tilting, which is not the case
- (c) the warping could hence have occurred only during the eruptive period.

The crustal bending began during about the middle of the Lower and ended before the middle of the Upper Basalt phase; its optimum could therefore have fallen within the Middle or Rhyolitic period of vulcanicity."

These statements relating to the relation between monocline formation and magma extrusion have been quoted extensively by other authors.

Lombaard (1952), Walker and Poldervaart (1949), and Stockley (1947) have all described Karroo dolerites and lavas but more detailed studies of the

Lebombo were not made until Cox et al. (1965) and Urie and Hunter (1963) published reports on Nuanetsi and Swaziland respectively.

Cox et al. (op. cit.) described in detail the structure and petrography of the Nuanetsi Province emphasizing the importance of basement control on the Karroo Structures. The basic, acid and intrusive rocks were discussed in detail and an ignimbritic origin was proposed (after Monkman, 1961) for the rhyolites on the basis of textural studies and their large lateral extent. A smaller group of rhyolite lavas were characterised by contorted flow banding. A regional tectonic synthesis was presented drawing analogies with the E. Greenland monocline (Wager, 1947). The rhyolites were considered to represent primary acid magmas and not differentiation products of Karroo basaltic magma. The concept of a Karroo Volcanic Cycle similar to that of du Toit (op. cit.) was proposed to explain the temporal variation of magmatic activity.

Urie and Hunter (1963) working in Swaziland produced a reconnaissance map of the Stormberg Volcanics. They divided the basalts into four groups on petrographic criteria and the rhyolites into four partly by mineralogy and partly by texture and colour. They distinguished a flow rich in quartz phenocrysts and stated that only one intermediate or basic flow had been found interbedded within the main body of rhyolite. They described certain sections in detail, commenting on beds of pyroclastic origin. They followed du Toit in assuming the basalts were derived from fissures now filled by dykes and, on the basis of distribution and general uniformity, interpreted the rhyolites as having an ignimbritic origin. Du Toit's view that the Lebombo volcanics occupy a region of tension was substantiated and a similar origin proposed.

Stratten (1965, 1970) working on the rhyolites in Zululand proposed a stra-

tigraphy of the volcanics dividing the Lebombo rhyolites into an upper and lower Lava Phase with an intermediate Pyroclastic Phase (Table IV.1).

TABLE IV.1
Stratigraphy of the Lebombo Stage (Stratten, 1970)

Post-Upper Lebombo Suite		dolerite (basalt?)
Upper Lebombo Suite	Upper Bumbeni Phase	igneous quartz viens soft pyroclastic rocks rhyolite granite and microgranite syenite and microsyenite
	Middle Bumbeni Phase	laminated tuff massive ignimbrite laminated ignimbrite
	Lower Bumbeni Phase	Bumbeni conglomerate and arkose, pheno-trachytic lava
Middle Lebombo Suite		andesite, basalt
Lower Lebombo Suite	Upper Lava Phase	rhyodacite, dacite, andesite
	Pyroclastic Phase	red sedimentary beds pyroclastic flow breccia massive, vesicular, jointed lava tuff
	Lower Lava Phase	rhyodacite, dacite andesite, red tuff

The lava phases consisted of massive lavas varying in composition from andesite to rhyodacite; the pyroclastic phase was characterised by thick flow breccias which alternated with massive lava. Stratten (1965) also recognised a number of pyroclastic breccia dykes which he supposed have acted as feeders for the acid magmas and maintained they were the major scarp-forming rock-type. His map was used in the field, without success, to locate breccias and tuff.

Saggerson and Logan (1970) provided a summary of Lebombo geology and a

description of layered basic intrusions near Komatipoort and in the southern Kruger Park. Cox (1970) discussed the tectonic setting of Karroo vulcanism.

In Mozambique Assuncao et al. (1962) published a map of the upper half of the Lebombo Volcanics with some petrographic descriptions of the basalts and rhyolites. Wachendorf (1971, 1973) added detail to Assuncao's map to the northeast of Swaziland and described the type-section through a rhyolite flow. On the evidence of contorted flow-banding he concluded that the Lebombo rhyolites were emitted as lava flows and not as ash-flows or ignimbrites. Only the rhyolitic tuff of the basal part was considered to be always a product of an initial more explosive phase of the same acid eruption. He thought the rhyolites flowed out of elongate fissures and formed ribbon-like outcrops. He concluded that the acid magmas were differentiates from a basic magma, a conclusion reached by Manton (1968) who reported strontium isotopic data showing the acid rocks to have low initial Sr-isotope ratios.

Bristow (1976) has recently completed a thesis on the geology and geochemistry of the southern Lebombo (south of Swaziland). He divided the volcanics into the Mfolozi River Formation of basalts (equivalent to the Drakensberg Stage of the Stormberg Series), the Jozini Formation of rhyolites (equivalent to the Lebombo Stage of the Stormberg) and a later Msunduzi River Formation consisting of the Bumbeni and Cooyana complexes and a thin basalt sequence. Intrusive rocks included an extensive dolerite dyke swarm (the Rooirand), later dolerites, acid quartz-porphyrics, breccia dykes and syenites and granites associated with the Bumbeni Complex. The petrography and field relations of the rocks are described in detail with emphasis on the basaltic Mfolozi River Formation. The rhyolites are compared with those of Mozambique (Wachendorf, 1971, 1973) and Nuanetsi (Monkman, 1961) as well as those elsewhere in the world. An origin as a partially deflated ash-flow tuff was proposed consistent with the observed features apparently

transitional between ash-flows and lavas. The geochemistry was discussed in an outline fashion and further data will soon be forthcoming.

Geophysicists have applied themselves to the problem of determining the thickness and structure of the Lebombo. Darracott (1974), Darracott and Kleywegt (1974), Smit (1962) and Burley et al. (1970) have all published detailed gravity data. The consensus is poor and a variety of models seem compatible with the available geophysical data. These geophysical data and the geochemical data pertaining to geochemical provinces are discussed in later chapters.

IV.3 Drakensberg Stage Basalts

The basalts of the Drakensberg Stage are generally poorly exposed throughout the lowveld. North of the Usutu River they seldom crop out and are seen only in stream beds at the foot of the escarpment, and where baked by intrusions. South of the Usutu the outcrop is better because of the dense dyke swarm although very little basalt is exposed. Extensive road cuttings through the Rooirand in northern Natal show the abundance of dolerite dykes with weathered basalt forming screens. The basalts are moderately well exposed adjacent to the large granophyre bodies west of Stegi, and especially on the west of the Lusoti granophyre where recent blasting for the railway has left large blocks of basalt. The various acid flows interbedded with the basalts are discussed along with the overlying rhyolites in Chapter V.

A number of basalt samples were collected during the present study but no effort was made to subdivide the basalts into mineralogical groups. Hunter (Urie and Hunter, 1963) working in the better exposed areas in the south has divided them into four groups based on textural and lithological differences:

1. Nsoko type

The Nsoko-type is confined to the area between the Rooirand and Lebombo, south of the Usutu River. It is normally dark in colour but at the top of the succession, just beneath the rhyolites, it weathers to a reddish bole. At the top it is also highly vesicular with amygdal lined with or infilled by quartz, zeolites or epidote.

2. Lubuli type

The Lubuli-type occurs throughout Swaziland and is the commonest basalt-type. It is dark, fine-grained, relatively aphyric and commonly contains white amygdales of quartz, carbonate or zeolites, or pea-sized black amygdales of chlorite and magnetite.

3. Mgwawini type

Interbedded with the Lubuli-type is the Porphyritic Mgwawini-type. This is distinguished by the presence of large feldspar phenocrysts up to 3 cm in length.

4. Sinyamantulu type

The Sinyamantulu-type is restricted in outcrop to the hill of the same name 8 km west of Lubuli. It is characterised by abundant small white amygdales and feldspar phenocrysts.

Urie and Hunter (1963) report one olivine-basalt from a point near Lubuli but attempts to relocate it failed. The olivine-rich basalts and 'limburgites' reported from Nuanetsi (Cox et al., 1965) do not appear in Swaziland though olivine-bearing basalts occur at the base of the succession in southern Zululand (Nondwane-type of Bristow 1976) and in the Maluti Mountains of Lesotho (Cox and Hornung, 1966).

Bristow (1976) describes Lubuli-type basalts from Zululand (called Ngqwatayi-type) and also the feldsparphyric Mgwawini-type (called Msunduzi River-type). The similar stratigraphic positions of the Msunduzi River and Mgwawini basalt types may support a correlation of the two types.

One flow of the Mgwawini-type, coarsely porphyritic, forms a low ridge west of Lubuli and can be traced for 30 km with an apparent thickness of 500 m. A flow exposed in the cuttings below the Nyetane (or Magwanyana) Dam has a base very enriched in feldspar phenocrysts which grades up into an aphyric upper portion. The Lubuli and Mgwawini basalt-types may, therefore, represent differing degrees of crystallisation from a common magma type.

Eleven boreholes were drilled for a Ubombo Ranches' dam-site near Big Band. Four penetrate mainly basalt, the rest are entirely in dolerite. Boreholes 3 and 8 are 24 m (80 feet) deep, the rest only 10 m (30 feet). Borehole 3 contains only basalt. The top 12 m (40 feet) are very amygdaloidal with chlorite, zeolite, and epidote amygdales with chlorite pseudomorphs after plagioclase phenocrysts in a pale zeolitised matrix. Below this the rock is black though chloritised and still amygdaloidal. Two flow contacts were observed at 9 m (30 feet) and 10.5 m (35 feet). The former shows a contact between a pale amygdaloidal basalt overlain by a darker less amygdaloidal basalt. The dip of the contact is 50° but this may bear little relation to the true dip. The contact is sharp with a 1-2 mm chloritised layer between the flows. The second contact shows slight veining and gas brecciation of the overlying flow. A 2 mm tuffisite vein cuts the core 5 m from the top.

No flow contacts were seen on outcrop in Swaziland because of the poor exposure. Exposures in the Kruger Park, along the Sabi. River, show highly vesicular flow tops overlain by massive basalt often with pipe amygdales at the base. Rare pipe amygdales were seen in Swaziland but the outcrop was

never sufficient to be able to determine their relationship to flow contacts. Bristow (1976) recognised a zonal scheme similar to that of Walker (1959) involving a central massive entablature with amygdaloidal zones at top and bottom. He also stated that flow tops were rarely weathered and infrequently autobrecciated. Stockley (1947) reported an almost complete lack of weathering of flow tops in Lesotho and that the base of nearly every flow had pipe amygdales.

From exposures in Zululand, the Kruger Park and in the Drakensberg the thickness of individual flows is seen to vary from a few metres to fifty metres. Bristow (op. cit.) quoted an average for fifteen flows of 5.5 m. Stockley (op. cit.) estimated an average thickness of 24 m (80 feet) to 27.4 m (90 feet) for the Drakensberg and Cox et al. (1965) noted that the Nuanetsi (Chikomedzi) basalts were generally less than 15 m.

Tuffaceous horizons are rare within the basalt succession. Bristow (op. cit.) reported a complete absence in the Southern Lebombo though Stockley (1947) and Lock et al. (1974) described ash deposits in the Lesotho and Barclay East areas respectively. Urie and Hunter (1963) noted several narrow tuffaceous horizons within the basalts south of the Usutu River. More extensive deposits were observed at the base of the basalts, on the south bank of the Usutu at Big Bend and in the Sabana Pass, 11 km south of the Ngwavuma Poort. Sandstone interbeds are common at the base of the succession where the lavas interdigitate with the Cave Sandstone but also occur higher up indicating the continuance of Cave Sandstone conditions.

These Drakensberg Stage lavas of the Lebombo are generally accepted as being sub-aerial in origin. The presence of aeolian sandstone interbeds and the absence of pillow lavas and marine sedimentary horizons support a theory of extrusion on to an arid land-surface. McCarthy (1970) and Lock et al. (1974)

reported pillow lavas from the Drakensberg Mountains but these appear to be localised occurrences representing extrusion into inland lakes. A number of sedimentary horizons contain plant remains (Urie and Hunter, 1963).

The absence of volcanic necks and the abundance of dolerite dykes suggest that the basalt flows originated as fissure eruptions. The dolerites are similar chemically to the basalts and have been shown to be contemporaneous by palaeomagnetic data (Graham et al., 1962). Apart from one locality mentioned by Bristow (op. cit.), however, there is no field evidence supporting the direct connection of the lavas and feeder-dykes. In Zululand the main dyke-swarm is west of the basalt outcrop but, in Swaziland, the axis of the dyke-swarm trends obliquely across the basalt outcrop and cuts progressively higher horizons of basalt till the main swarm peters out around Stegi (Map 1 - 4). Bristow (op. cit.) suggested that the dyke swarm represented a localised source of eruption for the basalts to the east and that the locus of intrusion and extrusion migrated eastward with time. This idea, though tenable in Zululand, fails to explain the cross-cutting dyke swarm and the difference in dyke density between the north and south of Swaziland. The problems of the dyke swarm and its relationship to basalt extrusion and monoclinial formation are discussed further in sections IV.5 and V.6.

Nowhere in the Lebombo are the basalts sufficiently well-exposed to enable dips to be measured though a dip of 45° has been recorded at Big Bend (Cox, pers. comm.). The dip must flatten out considerably at the western edge of the outcrop as the underlying sediments have dips generally less than 15° and outliers of basalt are found in small synclines and downfaulted blocks within the Karroo sediments. The dip of the rhyolite-basalt contact is usually in the range $30 - 35^{\circ}$, becoming steeper in the north. The thickness of the basalt is difficult to measure and field estimates are based on measurements of the width of outcrop and an assumption of dip. The presence

of strike-faulting and the dyke swarm serve to increase the apparent thickness but although sedimentary marker-horizons show the base of the basalts to be extensively faulted, the amount of faulting within the basalts is impossible to measure. The outcrop width varies considerably along the length of the Lebombo and is only 6 km in the north of Swaziland compared with 17 km in the south. Increased dips and density of dolerite dykes, frequency of strike faults or simply decrease in thickness may be responsible for this change of outcrop width. Taking an average width of 10 km for the basalt outcrop and assuming an average dip of 25° the thickness can be calculated to be 4.2 km. Estimates based on geometrical construction give a slightly larger value. Strike-faulting and the presence of dykes may reduce this value by as much as 50 per cent. The detailed structure of the monocline is discussed at greater length in section V.7. Previous thickness estimates are given in Table IV.2,

TABLE IV.2

Thickness estimates of the Drakensberg Stage Basalts, Lebombo

<u>Source</u>	<u>Thickness</u>	<u>Evidence</u>
du Toit (1930)	1600 m	Field data
Hales and Gough (1962)	8000 m	Gravity
Urie and Hunter (1963)	2000 m (6000 - 7000 feet)	Field
Burley et al. (1962)	4000 m in South Swaziland	Gravity
	8000 m in North Swaziland	Gravity
Darracott and Kleywegt (1974)	1000 - 2000 m	Gravity
Van Wyk (in Darracott and Kleywegt, 1974)	6000 m	Field
Bristow (1976)	4100 m	Field

IV.4 Lebombo Stage Rhyolites

(i) Introduction

The Lebombo Stage Rhyolites rest with a slight unconformity (see Chapter V) on the Drakensberg Stage basalts. A number of thin rhyolite flows are interbedded at the top of the basalts and form long parallel ridges which pass through Big Bend and are observed north of Stegi and the Mbuluzi. Minor rhyolitic interbeds are also present within the basalt outcrop. There are no basalts interbedded in the main body of the rhyolites. The average dip is about 20° though this increases to near vertical where the monoclinial axis is nearer the rhyolite outcrop. No strike faulting could be detected within the rhyolites and the total thickness must be at least 5 km.

The exposure is very good relative to the basalts and is of the order of one per cent. The rocks are well exposed along the crest of the range and on the escarpment itself. There is very little soil present on the rhyolites and sporadic outcrops are present on most of the ridges running between the rivers. Many of the river beds contain solid exposures but, as the peneplain becomes progressively more dissected in the east, the rivers run in small gorges and access is difficult though the exposures are good in the cliff faces.

Individual flows were mapped out; some thirty flows, averaging 200 m in thickness and traceable over distances along strike of up to 50 km, were recognised. Ninety-five per cent of each flow is completely massive and although at first sight the rhyolites are very similar, there are subtle differences in colour, degree of weathering and type of phenocrysts which serve to distinguish them. The top of each flow is commonly contorted and often breccias are developed. This breccia horizon and the basal tuff unit form a softer bed between flows in contrast to the massive central portion. The resulting slope-break features are often the most useful criteria

for mapping out flow contacts. These observations were greatly helped by study of the well exposed cuttings along the Mbuluzi railway and the irrigation canals beneath the J.G. Strydom Dam at Jozini.

Because of the topographic features produced by weathering of the rhyolites, photo-geological interpretation proved invaluable in interpolating field data and interpreting the areas to the north of the Mbuluzi and the Usutu. Aerial photographs were used extensively in the field for location and also to find the outcrops and features of interest.

Four 1:50,000 maps of the geology of the Lebombo Region (Maps 1 - 4) and a more detailed map of the Mbuluzi Valley (Diagram 1) have been drawn. A detailed sketch of the Mbuluzi railway cuttings is presented in Diagram 2. The individual flows have been traced out as far as possible and are labelled I - XXIX in the main maps and A - T in the Mbuluzi area for purposes of identification and discussion.

(ii) General description of the acid rocks

To illustrate the problems concerning the mode of eruption and emplacement mechanism of the rhyolites a generalised eclectic cross-section is described in detail (Figure IV.1). This is not a typical section as very rarely are all the units present in one flow and, indeed, one flow may contain a variety of textural types dependent on its distance from the origin, the nature of the topography and the overburden at the time of cooling. When observed, however, the textural zones tend to be in the same order, even if certain zones are absent. The excellent exposures in the Mbuluzi illustrate most of the observed textural variations (Diagram 2).

The section is described as objectively as possible; the interpretation and other subjective observations are left to Section IV.iii where the various

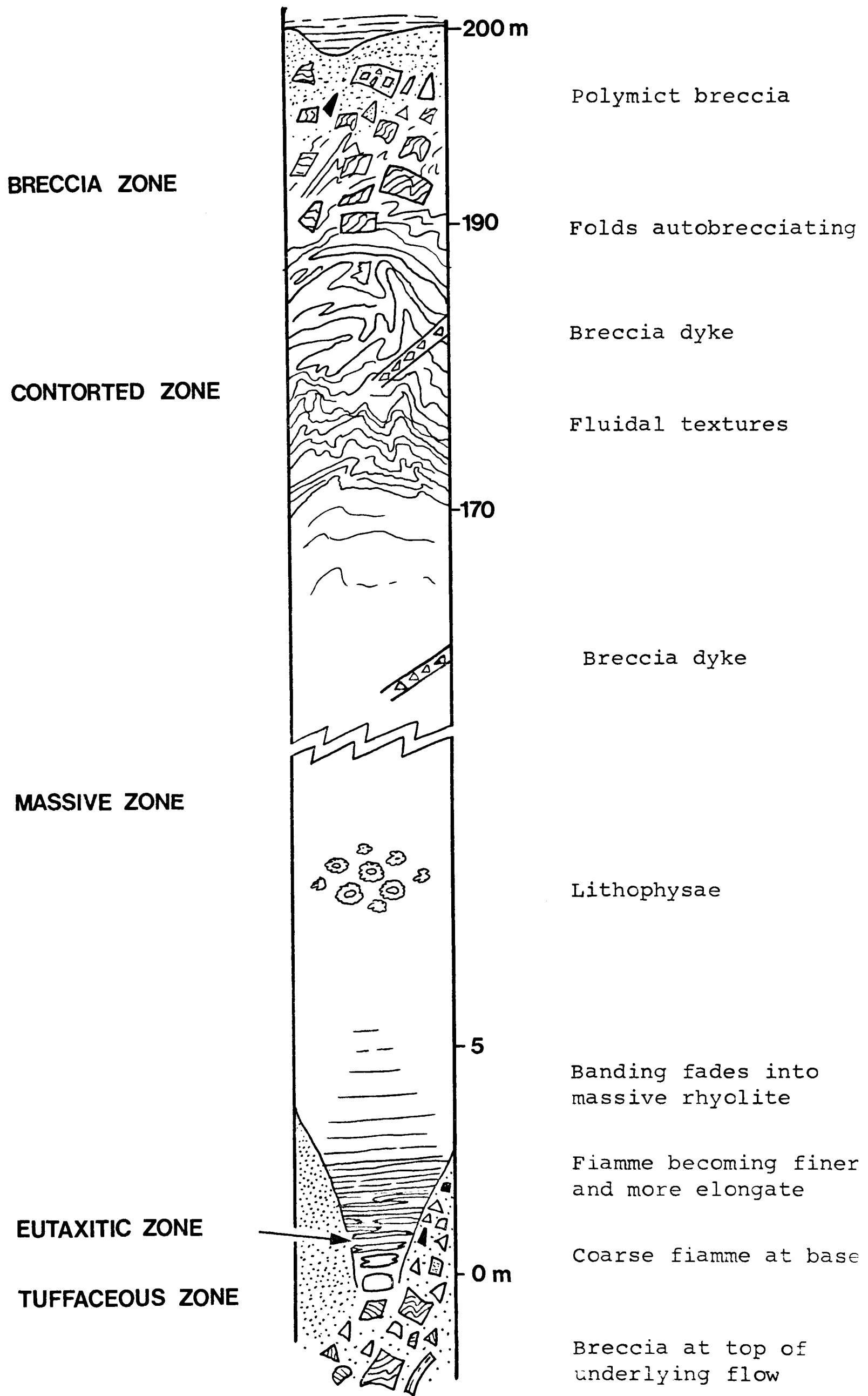


Fig.IV.1. Idealised section through a rhyolite flow.

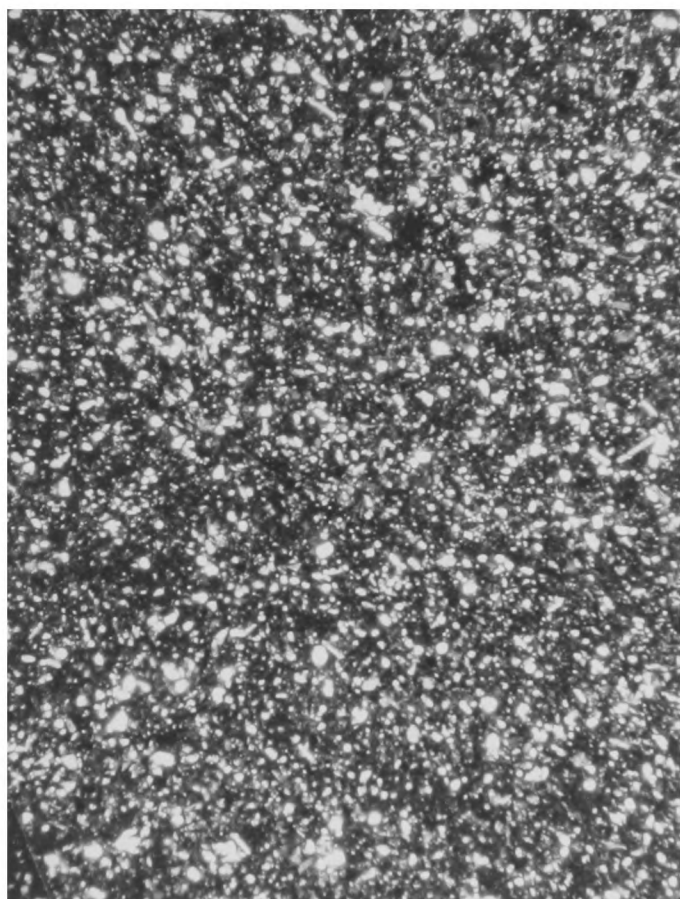
theories for the origin and mode of eruption are discussed. The detailed stratigraphy and discussion of individual flows is left to Section V.3.

(a) TUFFACEOUS ZONE

The base of many flows is marked by a poorly-exposed but often persistent tuffaceous zone which typically has a thickness of a few centimetres to a few metres locally. One tuff unit below flow XVII east-southeast of Stegi is 80 m thick.

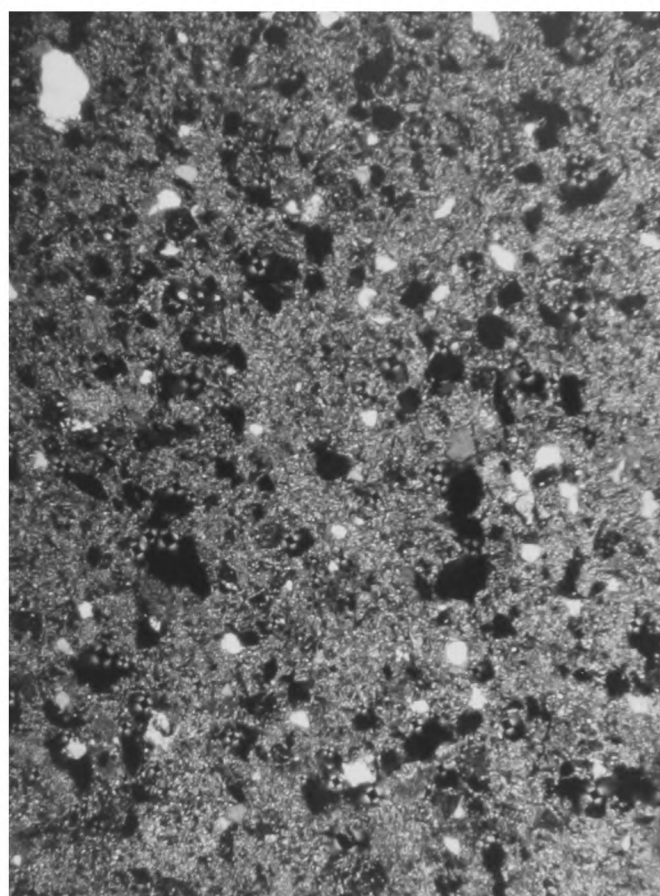
The rocks are often sandy in colour and texture, consisting largely of angular quartz grains with some glassy material and fine-grained interstitial dust. The percentage of quartz is never as high as in the equivalent zones within the basalt where quartzitic sandstones are found. The abundance of very angular quartz and minor feldspar grains with some small mica flakes suggests that this material is derived from outside the volcanic zone and is not entirely related to the formation of the rhyolites (Plate IV.1).

There is a complete gradation from the quartzitic tuffs described above to true tuffs. These latter are white or buff, soft and floury in nature. They are sometimes slightly banded with the banding emphasised by iron-staining and occasionally show secondary Liesegang lamination. Large vesicles are common, and some specimens have 0.5 cm to 2 cm lumps of weathered pumice. Corroded quartz and weathered feldspar phenocrysts are also found. These tuffs contain a much higher proportion of glassy material, usually very weathered, and interstitial dust than the sandy tuffs, and in extreme cases no derived quartz is present at all (Plate IV.2). Individual shards can rarely be recognised though one unit near Jozini, at the top of the basal flow, has the primary structures preserved adjacent to a dolerite dyke, and one sample from the Mbuluzi Gorge (MZ 50) has well preserved shard structures in a tuffaceous layer welded by the overlying rhyolite flow (Plate IV.3).



2 mm

PLATE IV.1 Quartzose tuff. Angular grains of quartz and small flecks of mica (small elongate crystals) in a fine irresolvable matrix (Sk 1).



1 mm

PLATE IV.2 Vitric tuff. A low percentage of quartz crystals set in a very fine matrix. The hand specimen has a floury nature (Sk 61).



2 mm

PLATE IV.3 Glass shards, now devitrified, preserved in a vitric tuff. Haematite staining helps to emphasize the original outlines (MZ 50).

Occasionally there is evidence for reworking of the underlying flow tops. Flow T has a slightly laminated tuff at its base which rests unconformably on a contorted, tuffaceous flow top. The flow top has been eroded and the reworked tuff redeposited as a layer above the unconformity. A tuff at the base of flow XXII east of Stegi has very fine, light and dark laminae represented by concentrations of quartz grains and glassy material respectively. An aqueous origin seems most likely to explain the fine banding (Plate IV.4).

(b) EUTAXITIC ZONE

The base of the flow proper is usually characterised by a sharp contact against the underlying tuffaceous rocks or breccia (see section { f }). The rock at the base of the flow is often glassy with streaky lenses of rhyolite interfingering with a darker glassy matrix to form a fine eutaxitic texture. The lenses and matrix are often complementary and the discontinuous nature of the rhyolitic lenses serves to distinguish them from the matrix. This zone has been called the 'streaky basal zone' by Bristow (1976) and the 'eutaxitic horizon' by Monkman (1961).

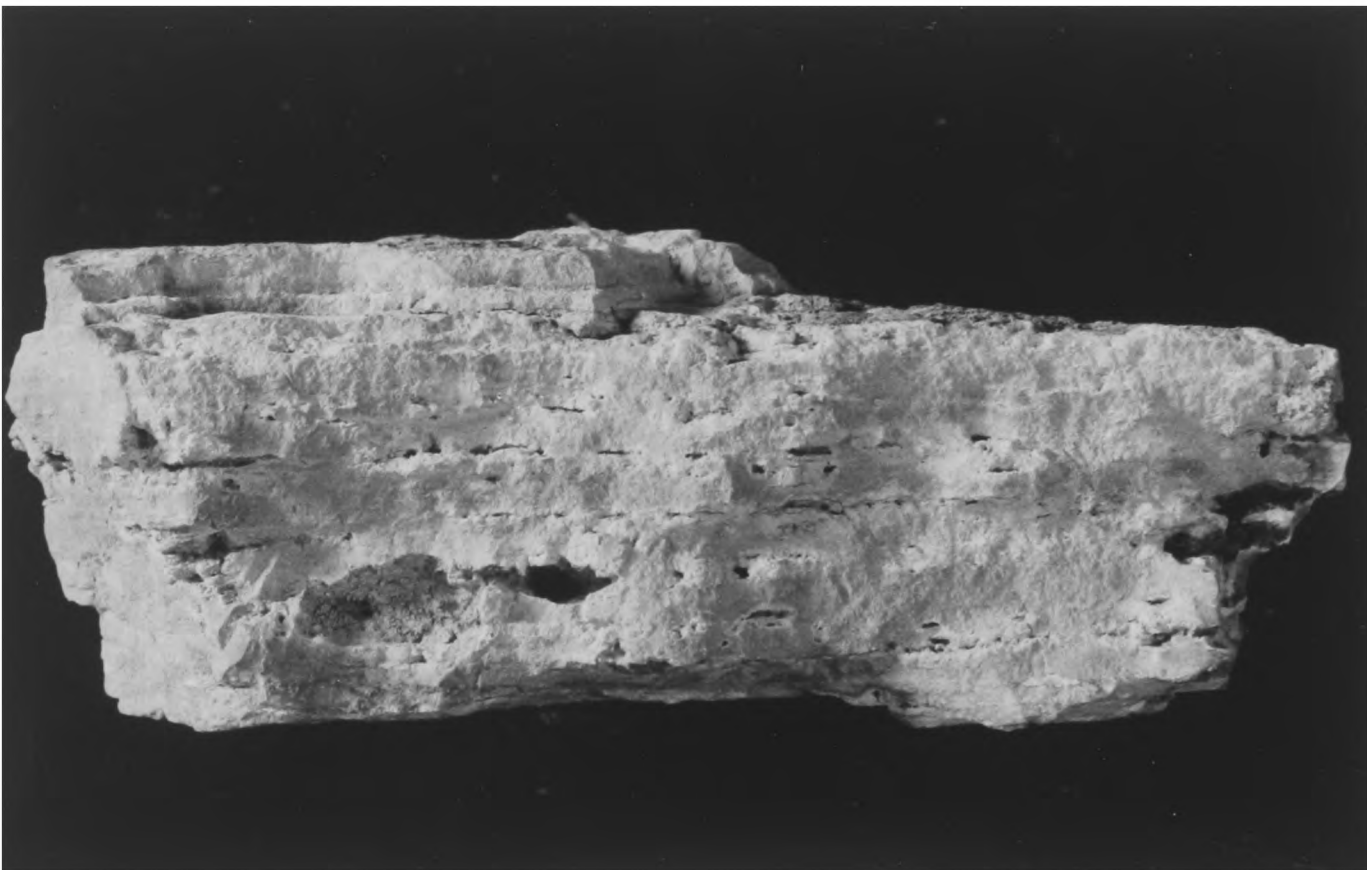
Samples from the base of flow XVIII, 20 km southeast of Stegi, show a well-developed fiamme (or lenticulitic) texture (Plates IV.5, 6). The individual fiamme are commonly 1 cm thick, but one 7 cm thick and at least 15 cm long was seen. The length:thickness ratio is about 10:1 in the specimens collected and seems fairly constant at this horizon. The fiamme are equidimensional in plan view and show no signs of elongation as a result of flowage. More rigid clasts are also found, including some small basic clasts, which show no signs of compaction and were thought to be more solid than the material forming the fiamme at the time of eruption.

These fiamme are found at two localities at the base of the flow, 1 km apart. In the southern of these two localities (RC 138, MW 14) the fiamme



1cm

PLATE IV.4a Finely laminated tuff from the base of flow XXII. Light and dark laminae correspond to concentrations of quartz grains and glassy material respectively.

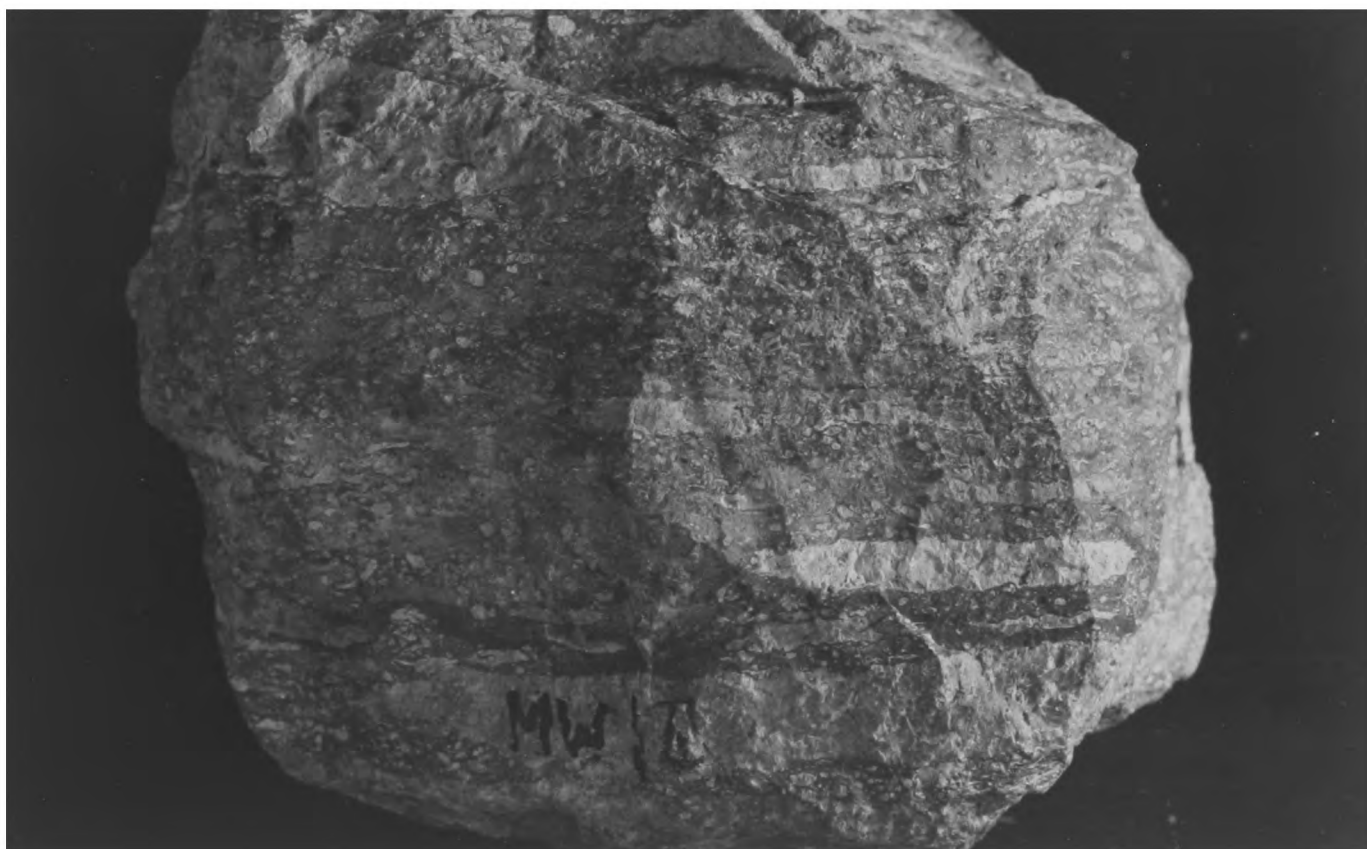


1cm

PLATE IV.4b Coarser laminated tuff. Holes are largely caused by pumice fragments weathering out.



1cm



1cm

PLATES IV.5, 6 Specimens from the base of flow XVIII showing well-developed fiamme. The length:thickness ratio is about 10:1.
(RC 138; MW 14)

seem to grade from an underlying breccia. The exposure was not perfect but the breccia blocks appear to become more elongated and to develop frayed ends. It is not possible to estimate the thickness of this breccia band nor its relationship to the flow beneath. Above the lenticulitic horizon, the fiamme grade into a finer laminated horizon and shortly into massive rhyolite.

A band of well-laminated rhyolite occurs at a locality near the base of flow II north of Stegi (SK 131A). The laminations are formed by extremely elongate, pale fiamme in a dark matrix. These bands reach a maximum 1 cm in thickness and are continuous on an outcrop scale (i.e. probably longer than 5 m). The rare ends to these laminae have flame-like ends. The banding is coarser and more pronounced towards the base of the outcrop and becoming finer and more obliterated by devitrification towards the top. One isoclinal fold closure was seen near the base.

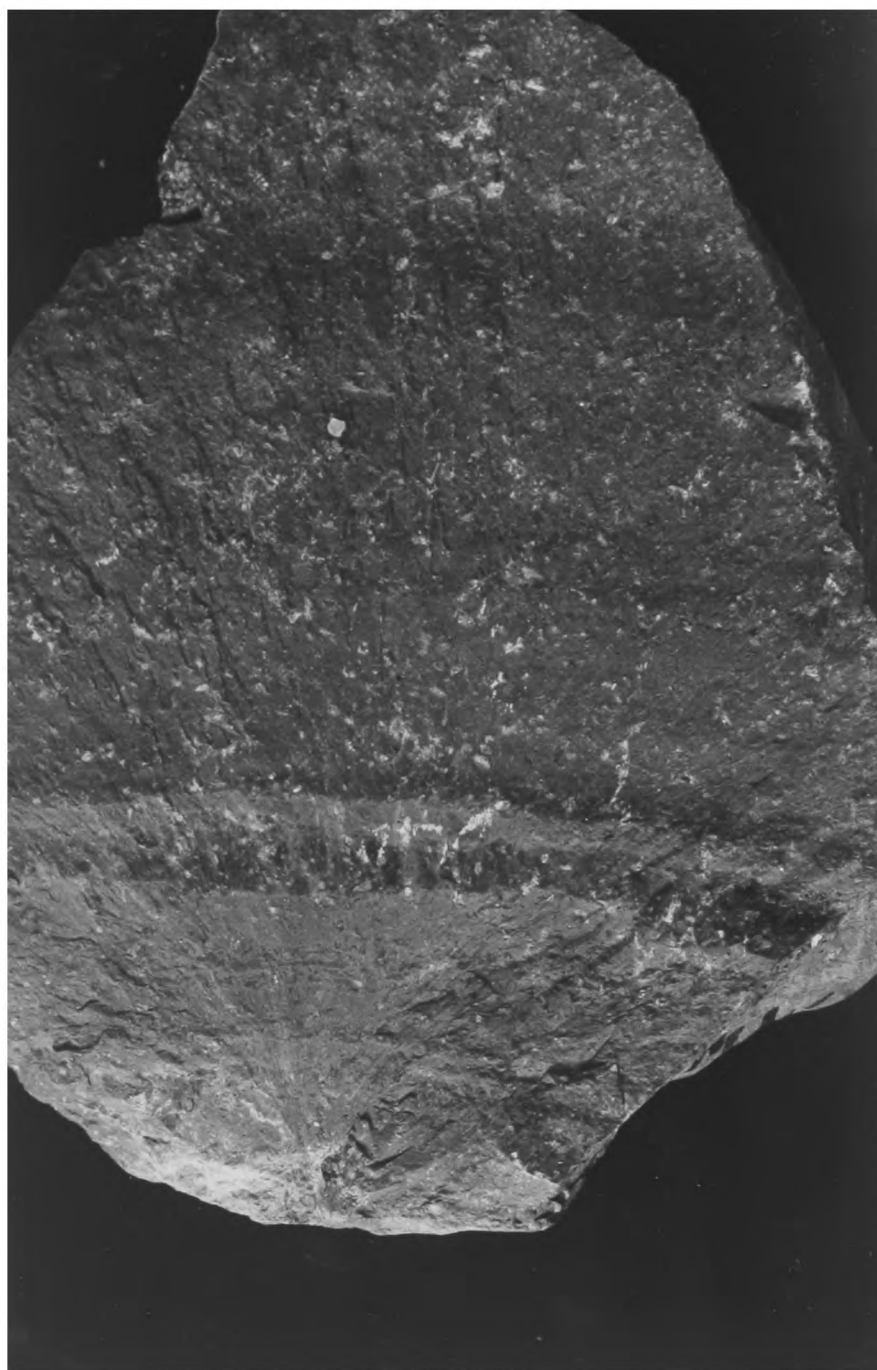
The length-thickness ratios of this unit are hard to measure but must be at least 100:1 and perhaps as much as 1000:1. Ross and Smith (1961) observed that length:thickness ratios of pumice fragments in the range 20-60 have usually suffered stretching in excess of that produced by simple collapse of pore space. The rocks of this unit have probably suffered flowage or secondary remobilisation in excess of that produced by flattening; the presence of a fold closure is evidence for shear caused by flow having taken place. Some of the thicker fiamme are sandwiched between layers of quartz; differential slippage may have caused the release of volatiles along these contact zones to produce cavities later infilled by quartz.

In both these types of lenticulite, the fiamme account for approximately 50% of the rock. The fiamme themselves are massive and display a blocky fracture typical of weathered pitchstone and feathery ends indicative of

flattened pumice fragments (Ross and Smith, 1961). They do not appear to differ in composition from the matrix, in that the phenocryst types and amounts are similar. Devitrification has caused a colour contrast between the fiamme and the matrix and has also obliterated any primary structures within the fiamme, though the matrix shows finely-laminated eutaxitic texture (Plate IV.7).

The commonest rock-type of this eutaxitic zone, as described at the beginning of this section, is very finely laminated and often pitchy in texture. The laminae are similar to flow laminae but their discontinuous nature and, on occasion, their downwards gradation into the lenticulite described above, suggests that they are also formed by a similar process of compaction and subsequent remobilisation. The streaks in these streaky-banded rocks have a finite, albeit large, length: breadth ratio and it seems improbable that they have been formed by extreme extension of the lenticulite horizons. It seems more likely that the quantity and size of pyroclastic fragments and pumice decreases sharply away from the base so that the finer lenses represent smaller elongated clasts. As the quantity of large pumice clasts decreases, and the streaks become finer, it becomes progressively easier for welding and devitrification to obscure the primary textures (Plates IV.8, 9), and the eutaxitic zone grades into the massive zone.

Parataxitic textures (as defined by Beavon et al., 1961) are also seen. They develop as a result of extreme welding and compaction. Thin parataxitic bands are found within the rhyolites, usually near the base of flows (Plate IV.10). These bands, usually less than 2 cm thick, have a texture very different from the surrounding matrix, being relatively aphyric and fairly rich in angular quartz grains. They probably represent horizons of tuff welded and compacted by the surrounding rhyolite. The tuff may have originated as a clast, but as the parataxitic bands are generally parallel to the



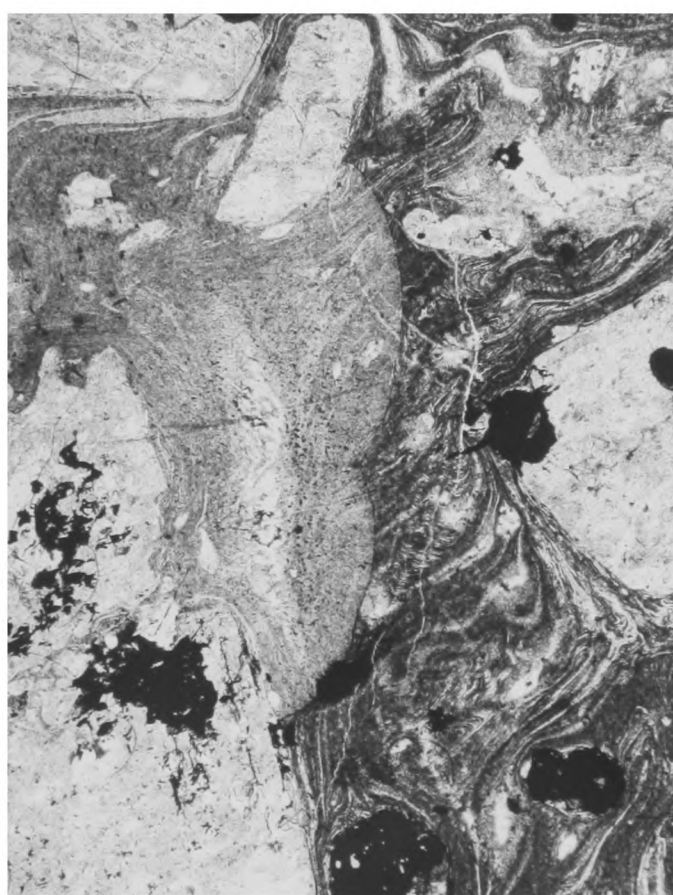
1cm

PLATE IV.8 Eutaxitic or fiamme structures grading into massive pitchstone. The fiamme are emphasized by weathering. (MW 5)



1 mm

PLATE IV.7 Flame-like end to one fiamma.



0.5 mm

PLATE IV.9 Primary eutaxitic banding becoming obscured by spherulite. Matrix within the spherulite is formed by many interfering spherulites of quartz and feldspar. (Sk 134)



1 mm

PLATE IV.10 A thin parataxitic band within a rhyolite flow. It is relatively aphyric and rich in angular quartz grains, and probably represents a thin tuff horizon welded and compacted by the surrounding rhyolite. (MZ 35)



PLATE IV.11 Massive rhyolite exposed in the 'Pot', southwest of Stegi. Figure top left gives scale.

observed foliation, it seems more likely that they were deposited in situ during an interval in the eruption, and hence could act as marker horizons for flow units within a cooling unit. Unfortunately, these horizons are too rare to have anything other than curiosity value.

The banding observed in this zone often appears to be steeper than the dip of the flow as defined by topographic features (e.g. scarps, slope features, dip of columns, etc.). The dip of the flow contact is generally in the range $20 - 25^{\circ}$ whereas the banding may be as much as 60° . It is often difficult to decide which is primary banding as often weathering differences along joints can cause secondary laminations. The volcanological and structural implications of this are discussed in Section IV.iii(d).

(c) MASSIVE ZONE

The central massive zone accounts for 95% of the thickness of each flow (Figure IV.1). It is frequently devoid of any obvious textural features and some flows (e.g. flows XI and T) show large thicknesses of very massive rhyolite with no internal structure and virtually no joints (Plate IV.11). Because of its resistance to weathering, the massive zone forms the vast majority of outcrops and, in the larger flows, forms large pavements of bare rock (e.g. flow XI east of Moyeni and at the Isateki Beacon). Similar rocks have been called 'The Massive Central Zone' by Bristow (1976) in Zululand and 'felsites' by Monkman (1961) and Tyndale-Biscoe (1949) in Nuanetsi.

The rocks of the massive zone are grey, fine-grained, evenly porphyritic and finely devitrified or micro-spherulitic. Thicker flows may develop a granophyric texture in the centre. Occasionally devitrification is not complete and rounded, darker glassy patches are found. The lower flows often develop a spotted texture with white 2 mm spherulites set in a greyish or rusty-weathered matrix. Larger spherulites and lithophysae are also found

(Section VI.4). Primary banding is usually absent though layers of spherulites may develop along former eutaxitic bands to give a banded appearance. Secondary banding, formed by devitrification, veining and weathering along microjoint planes, is more common and the effect is similar to eutaxitic banding at first sight. These joint planes are often discordant to primary banding observed elsewhere in the flow.

Columnar jointing is common and is well-displayed in the Lebombo escarpment and in krantzes elsewhere. The columns are roughly perpendicular to the flow contact but irregular in cross-section with a number of randomly-orientated, vertical, curvilinear joint-places (Plate IV.12).

Due to the availability of exposures, the identification of flows was usually based on comparison of samples from the massive portions of flows. Apart from mineralogical criteria, slight differences in the amounts of jointing, veining, weathering and haematite staining were all used to differentiate between flows. Later discoveries of intervening breccia and tuff horizons usually supported the evidence from hand-specimens.

(d) CONTORTED ZONE

The central massive zone grades into the contorted zone with the reappearance of laminations. These laminations are characteristically contorted in a variety of fold styles emphasised on the weathered surface. This zone is present in the majority of the flows in Swaziland; it can vary in thickness from a metre-thick horizon at the top of a flow to a unit tens of metres thick forming most of a flow.

Many of the best flow-features were found in the acid flows IX and J. Flow folding is present throughout these flows (Diagram 2) and also in cliff exposures around the J.G. Strydom Dam where large-scale folding was observed



PLATE IV.12 Irregular columnar joints
exposed in section in the cliff and in plan
view on the bedrock. Hammer shaft is 90 cm
long.

over thicknesses of 30 m. In the majority of flows, however, the contorted zone is confined to a few metres at the top of the flow. It seems probable that flow folding is generally present throughout most flows but that the folded laminae are obscured by the increased welding and devitrification present in the massive zone. The laminations are caused by devitrification differences between layers and not, as in the eutaxitic zone, by primary textural features. These differences in recrystallisation may have a primary origin but only in one instance was complex folding seen in rocks with a eutaxitic texture.

The amplitudes of the folds vary from a few millimetres to several metres with styles ranging from broad open structures to tight isoclinal and kink folds (Plate IV.13-21). Although some of the folds exhibit fairly regular structures similar to those caused by plane shear in solid rocks, in most cases the rhyolites have undergone plastic flow and hence the shear planes and axial planes of the folds do not bear a regular relationship to the stress direction.

In some outcrops (e.g. Plates IV.15, 16) the fold axes are vertical and often parallel over long distances. This verticality and lack of contortion suggests that the fluid rhyolite flow was moving in a laminar fashion producing, in places, the effect of being stirred by a spoon held vertically. Most of the outcrops, however, display fluidal textures with highly irregular fold axes showing large variations over short distances (e.g. Plates IV.16, 20), and it was never possible to calculate the direction of flow except over very small distances.

The degree of contortion frequently increases towards the top of the flow so that it can be used as a qualitative estimate of the distance below the top of the flow. In Swaziland it is apparent that the more silica-rich flows



PLATES IV 13,14 Fold plane exposed in the Mbuluzi Gorge railway cuttings. Small-scale tectonic ripples can be seen on the fold surface. (Flow J)

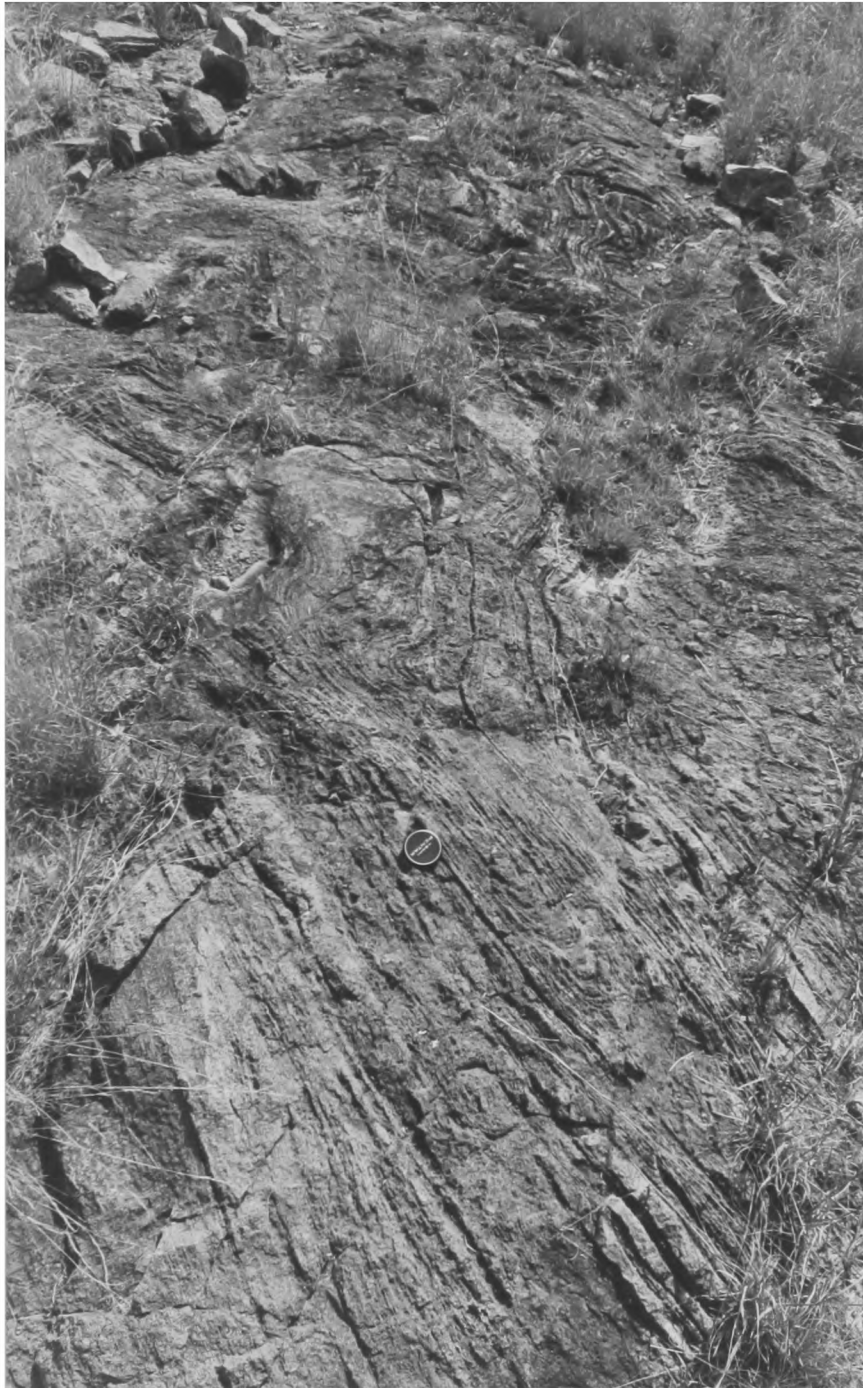


PLATE IV.15 Large scale irregular folding.
Fold axes are generally steep. (Flow XI)



PLATE IV.16 Nearby outcrop to Pl. 15.
Pronounced lamination with fold closures and
'boudin' structures visible in foreground.
Fold axes steep.



PLATE IV.17 'Orange-pip' structure. The formation of a more competent fold core has resulted in the outer laminae flowing past. (Flow XI)



PLATE IV.18 Contorted flow folds showing curved vertical axial plane. (Flow XI)



PLATE IV.19 Ptygmatic folding in rhyolite. (Kruger National Park)



PLATE IV.20 Fluidal textures in exposure west of J.G. Strydom dam wall. Right-hand half shows break-up of more competent bands.



PLATE IV.21 Shear folding from Bumbeni beds. Folds show thinning on the limbs. (Life-size)

show a greater abundance of complex folding; this may reflect the compositional dependence of viscosity or the lack of complete devitrification obscuring the structures. Bristow (1976) commented on the fact that the degree of complexity seems to increase towards the north within Zululand; this may also be composition-related due to the presence of more basic rhyolites in the extreme south (Bristow, pers. comm.). Contorted banding is rare in the Nuanetsi rhyolites and Monkman (1961) distinguished the 'Maose rhyolite lava' on the basis of the presence of 'highly contorted, ropy flow-banding' in contrast to the other massive flows. This Maose flow has a sporadic outcrop along the contact between two massive flows. When its basal contact is seen, however, it appears to grade into the underlying flow with no discontinuity (Norry, pers. comm.). It seems likely that the Maose rhyolite flow represents the contorted upper part of the underlying Bakuji flow.

Evidence of viscosity contrasts can also be seen within a flow. They are often manifest in 'orange-pip' structures (Plate IV.17) or in cases where a rigid band has been broken by the surrounding flow (Plate IV.20). Small elongate boudins (Plate IV.15) are common in the more planar-laminated areas. Towards the top of the flow, as the rhyolite solidified, the folding becomes autobrecciated and the contorted zone grades into the breccia zone.

(e) BRECCIA ZONE

Definition of terms. The following terms are defined in the sense in which they are used in the ensuing discussion on breccias.

Autobreccia. A breccia produced by autoclasis. A flow autobreccia is produced by fragmentation of a moving magma flow.

Autoclasia (adj. Autoclastic). A process of fragmentation of semisolid and/or solid lava during confinement beneath the surface or by relatively slow movement of unconfined lava flows, rather than by explosive disruption (Fisher, 1960). It is taken to include processes resulting in intrusive breccias as well as flow breccias.

Cataclasis. A process of mechanical fracture of rocks.

Epiclastic. Epiclastic volcanic breccias contain fragments produced by any type of rock fragmentation which are transported by surface geomorphological agents or gravity transfer (Fisher, 1960).

Gas breccia. A gas breccia is one formed by streaming gases and perhaps emplaced by a process of fluidization (Reynolds, 1954).

Intrusion breccia (volcanic). Volcanic intrusion breccias form by any brecciation process within a volcanic rock and show crosscutting relationships.

Monomict. Monomict is a term describing breccias where all the clasts are of one type (opp. polymict).

Pervasive breccia. Pervasive breccia is used here to describe a special case of intrusion breccia where a solid rock is veined and apparently brecciated over large areas by large scale hydrothermal or gaseous processes. It is not a breccia in the true sense.

Polymict. Polymict breccias contain a variety of clasts (opp. monomict).

Pyroclasis (adj. pyroclastic). A process of formation of a rock by

explosions and ejections of liquid and/or solid fragments from volcanic sources.

Tuffisite. An intrusive tuff. A term used to distinguish between intrusive tuff (tuffisite) and the similar extrusive varieties and for distinguishing crosscutting pyroclasts from cataclastic rocks.

The tops of the majority of flows are characterised by zones of breccia. These breccias are either monomict (where the clasts and host rock are the same) and autoclastic in origin or polymict and probably pyroclastic in origin. In addition to these breccias at the tops of flows, intrusion breccias are found in the form of narrow veins of autobreccia with a tuffisitic matrix and pervasive breccias, formed by intense veining giving the appearance of brecciation, affect large areas of rock. The term *allobreccia* was used in the field to mean the opposite of *autobreccia* or *polymict* (e.g. Diagram 2).

Autobreccias

The commonest breccia-types are the autobreccias found at the tops of flows. These are formed largely by flow autobrecciation and are gradational downwards into the contorted zone. The clasts are generally small, seldom exceeding 15 cm in diameter. They are angular, unsorted and identical to the host rock which commonly shows contorted banding. The matrix is fine-grained and often weathered, containing smaller fragments of rhyolite with fine glassy material and later chalcedony. The process envisaged for the formation of these breccias is one of break-up of the more solidified upper portion of the flow due to differential movement of the part-solidified magma beneath. As the folding became more intricate towards the top of the flow and the viscosity increased, the folds tended to fracture and then

become more broken up and disorientated with further movement. This produce angular fragments which often have an obvious 'fit', and this, together with the similarity to the host rock, suggests that the brecciation took place in situ and that little subsequent movement occurred. These flow breccias are different from those observed on basalt lavas because the rock was in a partially solidified state when the brecciation happened and primary flow textures are observed in both the clasts and the host rock. Basaltic auto-breccias are formed when the crust breaks up on top of a moving magma and the blocks formed, together with pyroclastic debris, roll down the front of the advancing flow to form a zone both at the top and bottom. The flow auto-breccias observed in the rhyolites are characteristic of contorted flow-tops and were only very rarely observed in the middle of flows, even when highly contorted, and never at the flow base. They vary in thickness from a few centimetres to 20 m but are generally several metres thick.

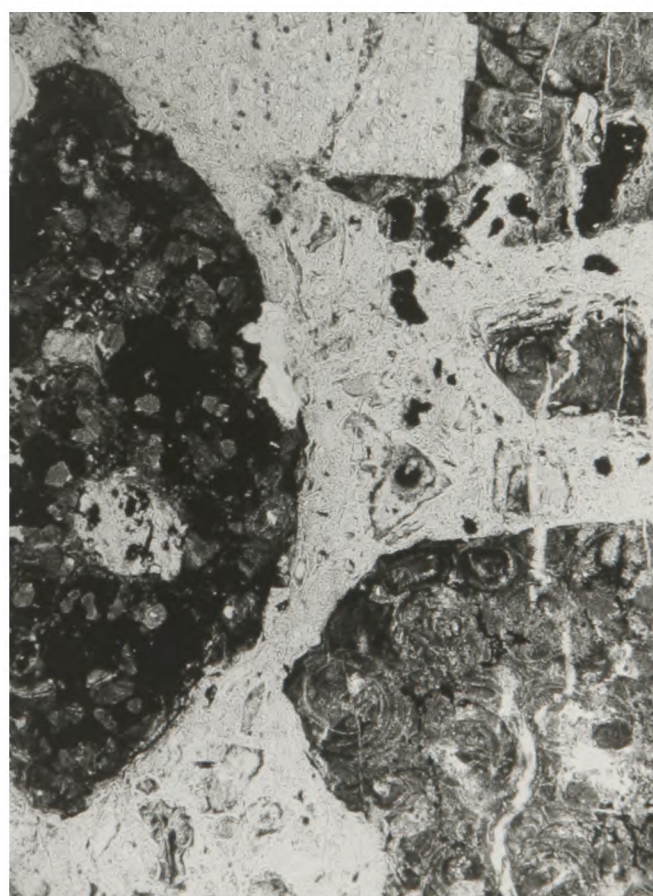
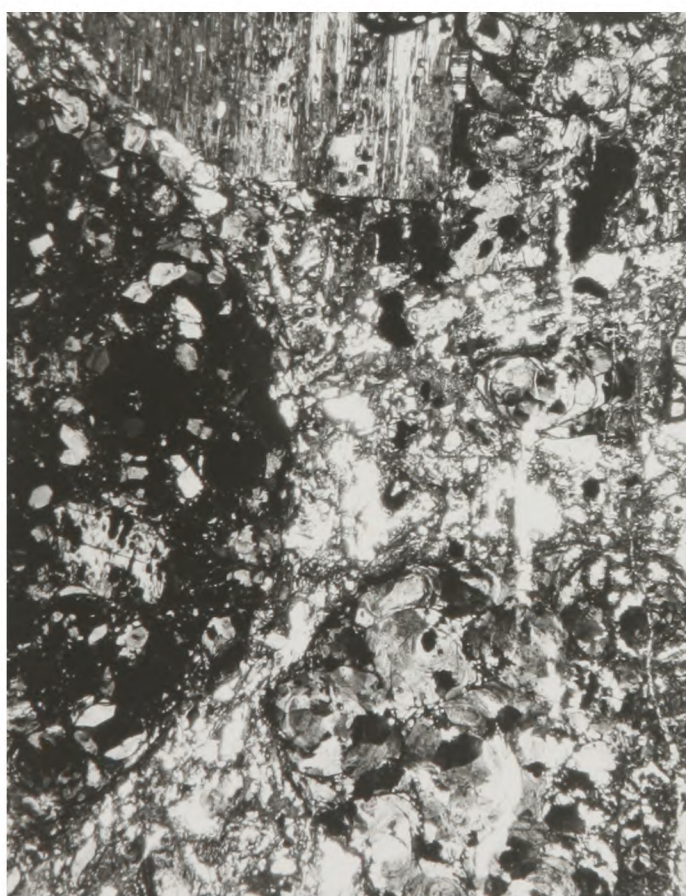
Polymict breccias

Several thick polymict breccia horizons are present at the tops of flows. One extensive horizon is present in the J.G. Strydom Dam section and crops out over 230 m horizontally, representing a vertical thickness of about 70 m. It contains a variety of clast types (80 cm to 0.5 cm in diameter) set in a weathered, fine-grained, rhyolitic matrix. The clasts include massive, laminated and spherulitic rhyolites, vesicular 'honey-comb' rhyolite, ignimbritic and eutaxitic fragments, pitchstones and breccias. All the clasts are very angular; none show signs of rounding or fragmentation. Clasts of different types are frequently found in zones so that, progressing along the outcrop, horizons with concentrations of certain clasts are found (Plate IV.22).

Vesicular, ignimbritic clasts account for half the observed clasts towards the base of this breccia unit and are abundant in polymict breccia horizons elsewhere. They were not observed on outcrop elsewhere in the Lebombo but



PLATE IV.22a Part of a thick polymict breccia exposed in the Usutu valley. Rhyolite clasts up to 1 m in diameter in a fine rhyolitic matrix.



1 mm

PLATE IV.22b Pitchstone fragments in devitrified rhyolitic matrix. Large plagioclase crystal at left. Contrast with the tuffisitic matrix in Plate IV.24.

Wachendorf (1973), though these extrusive breccias do not seem to be present in Mozambique, stated that 'the uppermost part of the rhyolitic lava flow has a scoriaceous surface and is highly vesicular ...'. These clasts, therefore, may represent the partly welded and vesicular uppermost portion of a rhyolite flow, usually removed or destroyed by erosion, or similar material forming in the vent and later erupted as pyroclastic material. The concentration of these vesicular clasts at the base of the breccia unit implies that a later, more cosmopolitan breccia flow covered, and mixed with, the brecciated flow top.

The contact with the underlying rhyolite is steep but indistinct; the upper contact is obscured by vegetation but the breccia appears to be overlain normally by the flow above. Bristow (1976) suggested that this breccia is probably a dyke or vent form on the basis of the steep lower contact. The shallow, concordant upper contact, however, suggests that the breccia is at least part extrusive. Similar polymict breccias are found throughout the rhyolite succession (the larger ones are marked on the maps). In all cases there is a coincidence of these breccias with flow contacts recognised by other means (e.g. change of phenocryst type, intervening tuff horizon, characteristic basal features of a flow, etc.). If these polymict breccias represent fissure vents, a more random distribution of them throughout the flows would be expected. The polymict breccias are gradational into the underlying flow autobreccias and seem to be a final extrusive event of the eruption of a rhyolite flow. The heterogeneous collection of clasts may arise from sampling processes at the vent, but these vents are not observed. In one flow the breccia was observed to grade into the eutaxitic zone (Section (b)), but in most cases a tuff unit intervenes.

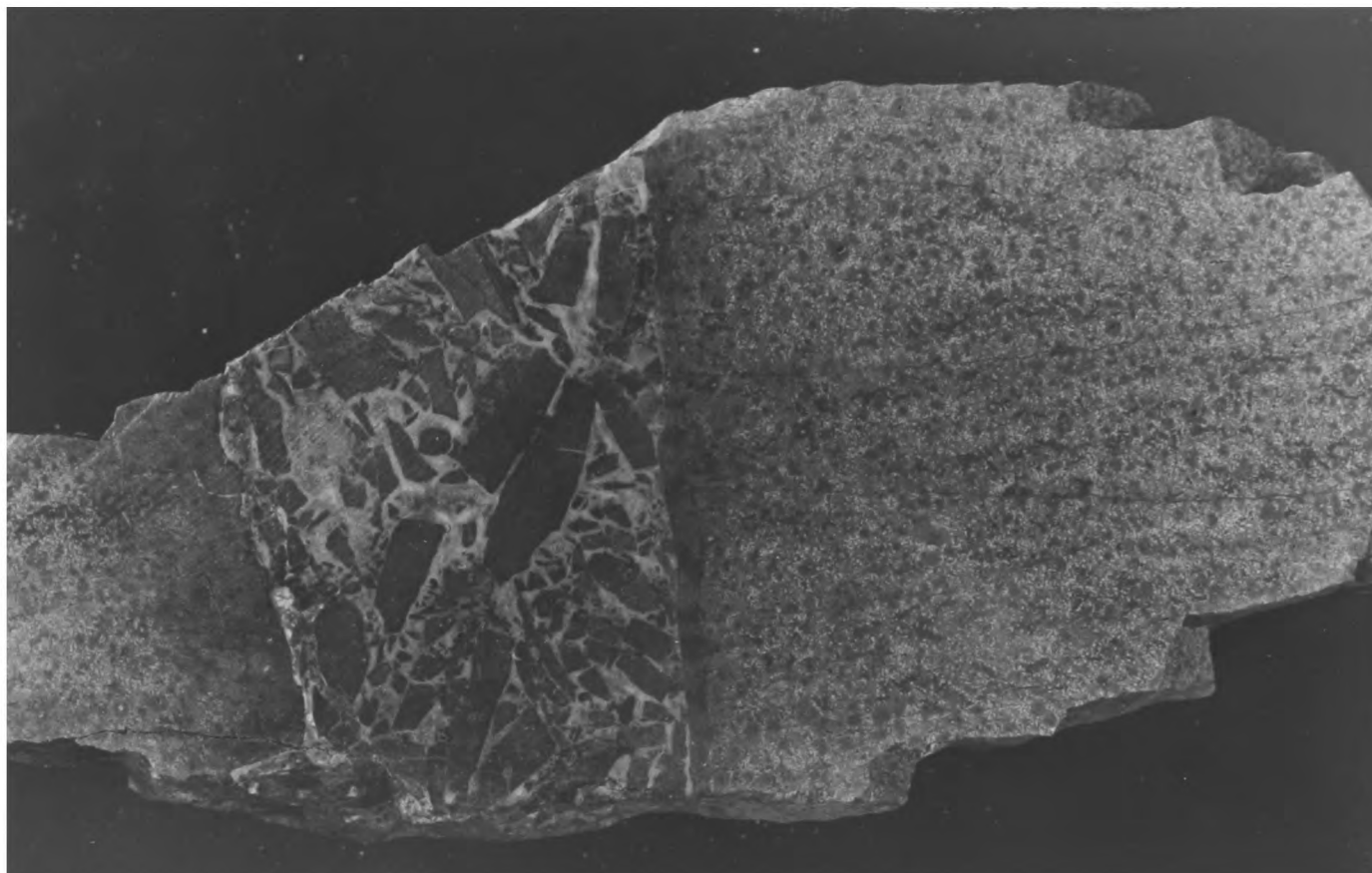
Intrusion breccias

Intrusion breccias form cross-cutting veins which occur throughout the

rhyolite flow-units. They vary in width from 0.5 cm to several metres. They consist of extremely angular fragments of rhyolite, identical to the host rock, set in a fine-grained, buff, sandy-textured matrix (Plate IV.23). The size of the clasts ranges from 15 cm to microscopic and, in general, the maximum size is proportional to the width of the vein. Veins of the matrix material are sometimes seen, with no visible clasts, cutting through the rock adjacent to the main breccia vein. In thin-section the matrix is seen to consist of fine, angular, quartz, feldspar and opaque grains set in a fine, irresolvable, haematite-stained groundmass (Plate IV.24). The matrix, containing finely ground rock fragments, is typical of cataclastic rocks and very similar to the pseudotachylytes from tuffisite pipes described by Reynolds (1954). The process envisaged is one of gas-brecciation, whereby large volumes of gas, passing through structural breaks in the rock, have sufficient energy to fracture the rock. As the energy increases, the fragments are supported or carried upwards by the rising fluid phase. The resulting rock contains angular or slightly rounded fragments with a matrix of finely comminuted rock fragments. Some examples quoted by Reynolds (op. cit.) have interstitial glass (and hence the name pseudotachylyte), but although Monkman (1961) described breccias with the margins modified by remelting, no glass was observed in the Swaziland localities. Vugs and cavities, usually where gas-bubbles have been trapped between two clasts, are completely or partially infilled with chalcedony or goethite.

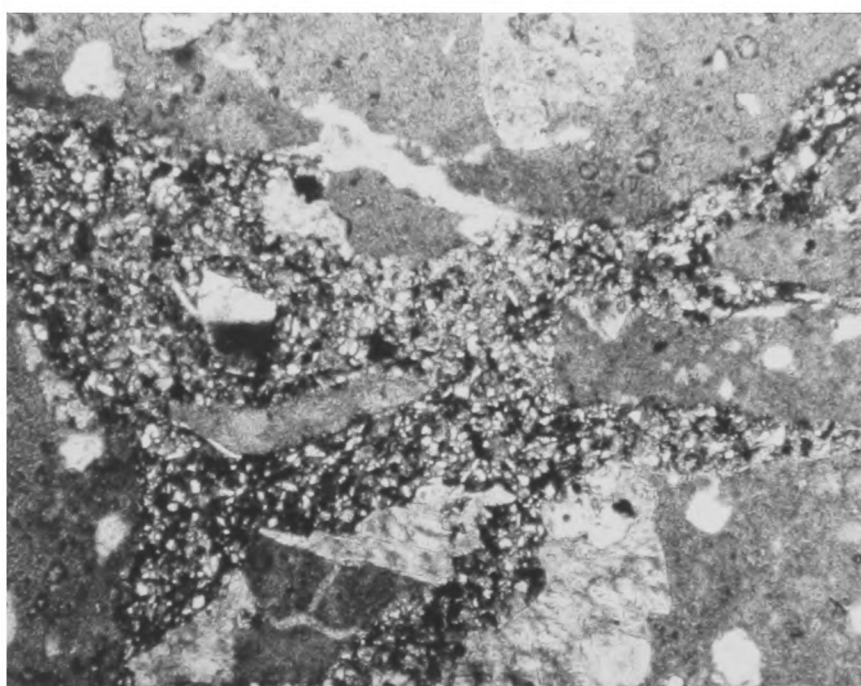
Similar dyke-like intrusion breccias were reported from the Little Lebombo by Wachendorf (1973) but he only found them penetrating the flow-folded horizon.

Large breccia dykes crop out in Zululand and intersect the basalt outcrop. They consist of large, angular fragments (sedimentary and volcanic) set in a very fine-grained siliceous or rhyolitic matrix and are frequently charac-



1cm

PLATE IV.23 Cross-cutting tuffisitic breccia vein. Angular fragments of the wall rock in a flinty matrix.



1 mm

PLATE IV.24 Thin section of tuffisitic breccia. Fragments of massive rhyolite in a matrix of finely ground rhyolite and quartz.

terised by a network of calcite veins and lenses. They only occur in a restricted area to the west of the Bumbeni Complex. Bristow (1976) interpreted these breccia dykes as fissures through which volcanic debris was ejected; the rounded nature of the fragments and the absence of lava indicating that the mechanism creating the fissure was a highly fluidized, explosive system. He added that they may also have acted as fissures for the main rhyolite flows or, due to their restricted distribution they may be remnants of the volcanic event that produced the Bumbeni and Cooyana Complexes.

Pervasive breccias

The pervasive breccias form a number of small areas in Swaziland (and five larger areas denoted on the maps by roundish areas of breccia (coloured green)). They have the outward appearance of a breccia but on the fresh surface no distinction between clasts and matrix can be made, except perhaps one of weathering. Structures within the clasts are generally undisturbed relative to the surrounding rock and large 'islands' of unbrecciated rock are frequently found in areas of intense brecciation (Plate IV.25). The contacts are always extremely vague and ill-defined and a gradation from extreme veining to brecciation can be observed in places. It is apparent that this pervasive breccia is really a 'pseudobreccia' caused by a network of veins; differences in weathering between 'clasts' and 'matrix' produce a brecciated appearance (Plates IV.26, 27). In extreme cases, areas up to 1 km across have their structures completely obliterated by pervasive breccia (Plate IV.28).

The evidence, therefore, is for a process producing a network of veins with little or no structural alteration of the rock. There are none of the autoclastic features noted in the intrusion breccias and the whole process is redolent of hydrothermal activity. There is no observed mineralisation associated with the alteration but surface sublimates and near-surface



PLATE IV.25 Breccia intruding laminated rhyolite. Large blocks of rhyolite remain in parallel orientation. 'Islands' up to 100 m across have been found in areas of intense pervasive brecciation.



PLATE IV.26 Highly veined outcrop of rhyolite. Network of anastomosing veins with near-in situ rhyolite.



PLATE IV.27 Similar to Plate 26 but here the veining has increased producing a brecciated appearance; some rhyolite fragments disorientated. All the clasts are of the same composition.



PLATE IV.28 Pervasive breccia. Part of the pervasive breccia outcrop northeast of Stegi. The original structure has been totally obliterated by the veining, leaving a massive rock containing fragments of weathered rhyolite in a weathered matrix. When broken open, the rock does not appear brecciated.

alteration would probably be rapidly reworked by water entering the flows (Smith, 1960). The five, roughly circular, larger areas of this pervasive breccia may represent areas of fumarolic activity causing gentle veining and alteration of the rhyolites in contrast to the violent gas-escape brecciation.

(f) DETAILS OF CONTACT ZONE

In the half-dozen contacts exposed in the Lebombo, the junction between the welded glassy rhyolite and underlying, usually tuffaceous, rocks is always sharp. The rhyolite is generally massive to the base, though the basal few millimetres sometimes show evidence for welding and a vitroclastic origin. Plates IV.29, 30, show the sharp contact associated with this change in physical properties between the rhyolite and tuffaceous rocks. The eutaxitic zone rocks (described in Section (b)) are rare and their relationships to the underlying flow are never seen though they are observed to grade into the homogeneous rhyolite of the massive zone.

The rhyolite at the contact frequently shows primary banding. The contact at the base of flow S in the Mbuluzi Gorge has a complex layering of rock types:

Above:	Massive rhyolite, contorted in places.
20 cm.	Black pitchstone, becoming devitrified upwards. Eutaxitic textures.
6 cm.	Weathered pumice, no structures visible.
4 cm.	Black pitchstone.
4 cm.	Weathered glass and pitchstone.
1 cm.	Finely laminated and amygdaloidal.

Breccia at base.



PLATE IV.29 Contact exposed on the road to Jozini. Massive, glassy rhyolite rests on oxidised tuff, baked by the rhyolite. The basal few millimetres of the rhyolite show a eutaxitic texture. White material is silica formed along a joint plane.



PLATE IV.30 Contact exposed in the canal cuttings below the J.G. Strydom Dam (now concreted over). Massive rhyolite with blocky jointing overlies friable, slightly bedded tuffaceous material. Approx. 5 mm of eutaxitic material was visible immediately above the contact.

This assortment of layers may reflect the complexity of an eruption cycle. The thinner glassy bands may represent small flow units in themselves, the intervening weathered zone being the equivalent of the tuffaceous zone. The whole outcrop is weathered, however, and it is not completely clear which textural types are solely due to weathering and the effects of jointing, and which are primary igneous features.

The contact surface is planar on a large scale but often small-scale irregularities are present (Plates IV.29, 30). One flow contact, observed in a road-cutting east of Komatipoort, near the Mozambique border-post, shows small warps superposed on a planar contact (Plate IV.31). The banding remains independent of irregularities in the base of the flow and is parallel to the overall dip of the contact and to the joint planes. Most eruptive processes, resulting in lavas or pyroclastic deposits, result in the laminae draping the topography, or at least responding to it. The basal few millimetres of this flow show eutaxitic textures, but the main banding, being independent of the flow contact, is probably secondary in nature. Weathering along, and parallel to, joint planes will cause devitrification and other alteration resulting in a colour banding.

The flow contact exposed in the J.G. Strydom Dam cuttings, above the thick breccia unit described above, has developed unusual botryoidal structures at the base (Plate IV.32). These structures were found at several horizons near the base of the flow and passed laterally and vertically into massive rhyolite. They were at first thought to be loadcasts formed by compaction of the fluid rhyolite into a plastic bed beneath. Lock (1971) described similar structures from the base of an ignimbrite unit in Newfoundland and proposed that they were spherulites formed at the base of the flow. A similar origin for the Lebombo occurrences seems the most feasible.



PLATE IV.31 Flow contact in road cutting 100 m west of Mozambique border, east of Komatipoort. Irregular contact of slightly banded rhyolite with pyroclastic rubble. The banding does not respond to the irregularities in the contact surface and is probably secondary, related to the joints.



PLATE IV.32 Base of flow (bottom up) from the J.G. Strydom canal section. The botryoidal structures are gradational into the main part of the flow and are probably spherulitic in origin.

(iii) Volcanological Aspects

(a) EXPLANATION OF THE PROBLEM

It can be seen from the preceding description of the Lebombo rhyolites that they have many features attributable to either lavas or ignimbrites. Contorted flow banding is characteristic of acid lavas and unusual in ignimbrites; eutaxitic textures and fiamme are indicative of ignimbrites and, except for unusual occurrences (Gibson and Tazieff, 1967), unknown in lavas.

It is suggested that, on the evidence from the eutaxitic zone, the Swaziland rhyolites were emplaced as large, hot, ash-flows. The flows are welded to the base with glassy rocks within millimetres of the contact and it seems that the heat of eruption and later devitrification were sufficient to obscure the primary features of welding.

The following sections attempt to reconcile the classic ignimbrite story with the features observed in the Swaziland rocks.

(b) DEFINITIONS

The term *ignimbrite* was first used by Marshall (1935) who defined ignimbrites as rocks which are 'thought to have been deposited from immense clouds or showers of intensely heated but generally minute fragments of volcanic magma. The temperature of these fragments is thought to have been so high that they were viscous and adhered together after they reached the ground'. This definition of ignimbrite did not make clear whether it was based on the mode of eruption or on the physical properties of the rocks.

Later workers (e.g. Fenner, 1948), because ignimbrites are characteristically welded, used the term *welded tuff* synonymously with that of ignimbrite, stressing petrographic characters rather than process. Beavon, Fitch and

Rast (1961) defined ignimbrites as those rocks deposited from nuée ardentes, making no restriction on degree of welding. Smith (1960a, b) and Ross and Smith (1961) used the term *ash-flow* to describe the eruption producing ignimbrites, and *ash-flow tuff* or *ash-flow deposit* for the resulting rocks whether welded or not.

Ignimbrite is used here (after Sparks and Wilson, 1976) to mean the rock or deposit formed from pyroclast flows composed predominantly of fragments of vesiculated magma. An ignimbrite is the product of one eruptive event and may consist of many individual flow units forming one 'cooling unit'. This definition includes all those rocks defined as ash-flow tuffs (Ross and Smith, 1961) and also other coarser breccias and pumice flow deposits. There is no inference in the usage of the term either to degree of welding or to magnitude of eruption. Welded tuffs may form all or part of an ignimbrite.

Ignimbrite, therefore, is used to describe those deposits formed by pyroclast or ash-flows as distinct from pyroclast falls and is used in a strictly genetic sense. It is useful for describing eruptions and their products but less so in palaeovolcanic terrains where weathering and poor exposure do not provide sufficient data for volcanological interpretation. In Swaziland, the presence of indurated or welded flows is usually the only evidence for an origin by pyroclast flow and the term ignimbrite is used to distinguish these from the non-welded basal tuffs which probably originated as air-fall material but may, in part, be due to pyroclast flow processes.

(c) MECHANISM OF EMPLACEMENT

General

It is generally assumed that ignimbrites are formed from pyroclast flows and that these flows move under the influence of gravity along topographic depressions (Smith, 1960a). The great extent of ignimbrites is proof of

the mobility of these flows, a mobility caused by exsolved gases or entrapped air producing a fluidized system (Reynolds, 1954; McTaggart, 1960, 1962).

The common irregular nature and poor sorting of ignimbrites have been attributed to turbulence of the ash-flow (Smith, 1960a; Ross and Smith, 1961; Fisher, 1966) but may also be due to flotation of low density pumice clasts in the denser flow (Sparks, 1976).

No eruption has ever been observed to give rise to an ignimbrite. Nuées ardentes ('glowing avalanches') have been observed and described by Anderson and Flett (1903), Lacroix (1904) and Perret (1937). They are thought to be similar to the processes producing ignimbrites but on a small scale. They are large dense clouds, composed of glowing volcanic fragments, which expand to heights of over 3000 m and rush at great speed down the slopes of a volcano. Perret (1937) described one as it passed as 'a train of ash and lava blocks not more than five metres in width or height ... during the whole time of its passage (30 minutes) there was complete silence'. He estimated the speed, conservatively, at 10 metres per second and that of a normal nuée to be something less than a kilometre *per* minute (17 m s^{-1}). Anderson and Flett (1903) reported a speed of 30 m.p.h. (15 m s^{-1}) for a nuée in the eruption of the Soufrière and velocities of 50 metres per second have been estimated by Lacroix (1904) for nuées at Pelée and by Moore (1967) for a base surge in the eruption of Taal volcano. The velocity seems to depend on the slope but, with sufficient momentum, nuées can travel great distances along the flat (Williams, 1942). Nuées seem to be able to rapidly decelerate; once slowed they are still self-supporting and one at Pelée, as observed by Anderson and Flett (1903, p. 494), 'lay almost like a dead mass on the surface of the sea'.

A prominent feature of the nuées is their rolling motion as they advance, giving the impression of boiling (Lacroix, *op. cit.*). The nuée deposits

continuously along its path, except perhaps adjacent to the source, and there is no evidence for large amounts of erosion.

Density Currents

Nuée ardentes have many similarities with high concentration flows such as mud-flows and turbidity currents. Mud flows also have the ability to carry clasts, a property attributed to the high particle concentration which results in a high apparent viscosity and density for the flow.

Both pyroclast flows and mud flows behave as density currents. A density current is a general term for relative motion under gravity of two fluids of differing densities.

One important property of density currents, both observed in nature and reproduced experimentally, is the presence of a distinct head which is at least twice the thickness of the body of the flow behind the head (Middleton, 1970). A theoretical discussion of the motion of a density current was given by Benjamin (1968) and a numerical simulation by Daly and Pracht (1968). These studies and those of the experimentalists (e.g. Middleton, 1970), have been concerned with a two dimensional head, both for numerical and practical simplicity.

Consider an empty cavity advancing beneath the upper solid boundary of an infinitely deep fluid of density ρ (Figure IV.2). The cavity develops a head and a body considered to be of infinite length. The cavity is arrested by applying an equal and opposite velocity to the ambient fluid.

The point, S (Figure IV.2), where the streamline detaches from the boundary layer, is called a stagnation or attachment point. The pressure at S is given by Bernoulli's theorem:

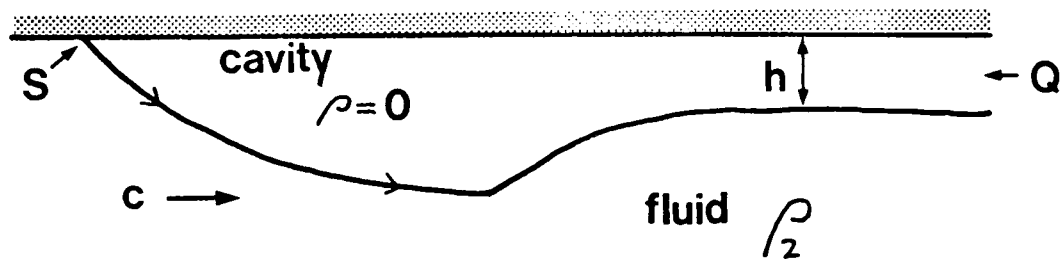


Fig. IV.2. Definition diagram for theoretical discussion of density currents.

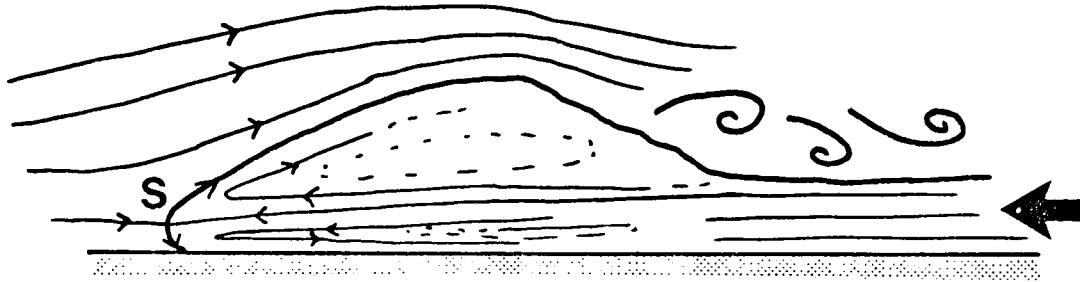


Fig. IV.3. Fluid motion in and around the head of a density current (after Allen, 1970; Middleton, 1970).

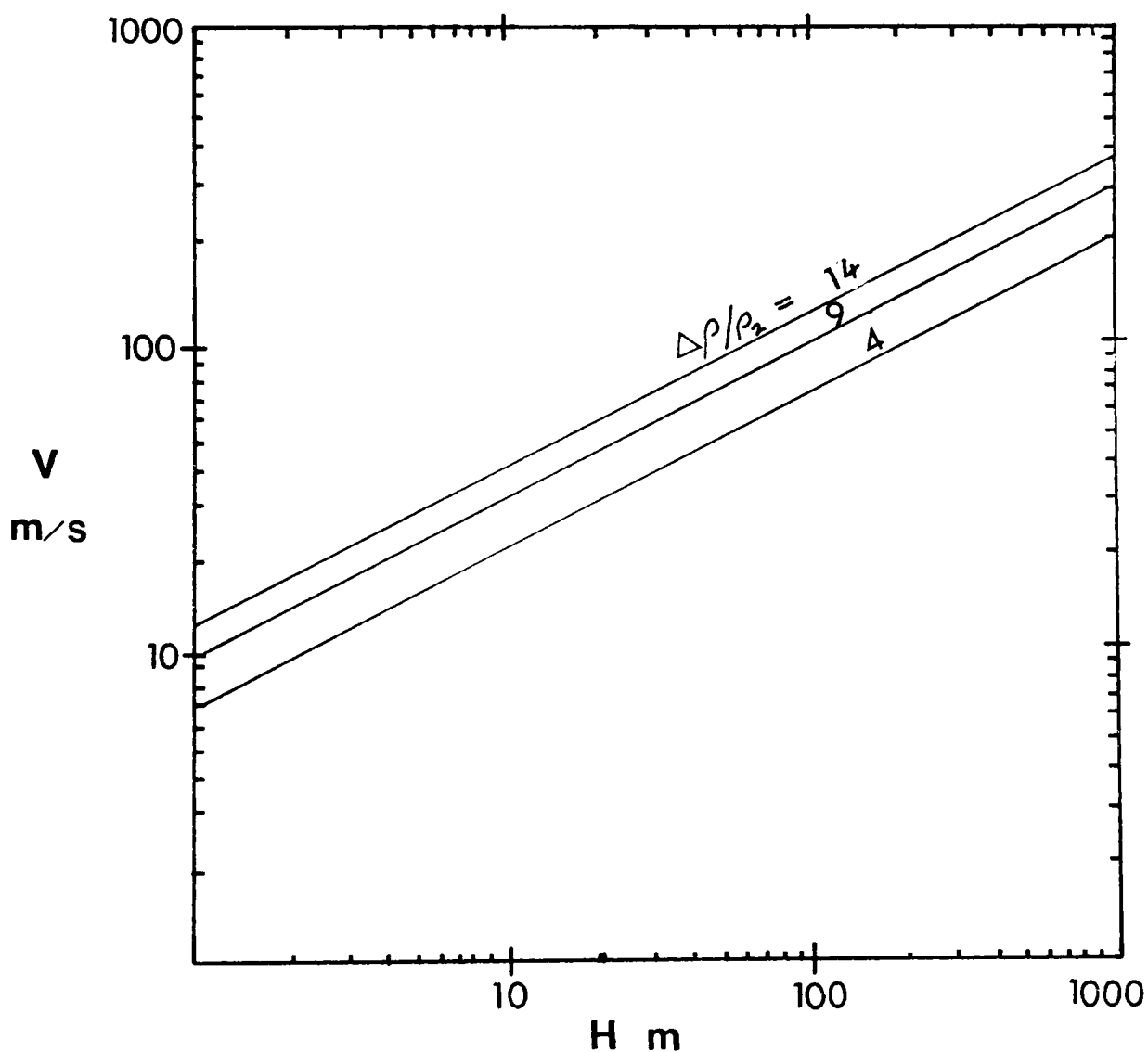


Fig. IV.4. The velocity of a density current as a function of thickness of flow and excess density (Equation IV.6). Sparks and Wilson (1976) estimated the density of a pyroclast flow where 10% by weight of the dispersion is H_2O to be five times that of the atmosphere. For more plausible gas percentages the density ratio would be much higher. Plotted for density of the flow = 5, 10 and 15 times atmospheric density. k taken as 1.07.

$$p_s = p_o + \frac{1}{2}\rho c^2 \quad (4.3)$$

where: p_s = pressure at S;

p_o = pressure on the horizontal boundary far upstream;

ρ = density of fluid;

c = flow velocity.

p_s is also equal to the pressure in the cavity due to hydrostatic pressure

$$p_s = \rho gh + p_o$$

where h is the net fall in elevation between the cavity and the fluid.

Equating these expressions:

$$c^2 = 2gh \quad (4.5)$$

or:

$$c = \sqrt{2gh}$$

When scaled for density currents:

$$c = k\sqrt{\frac{gh\Delta\rho}{\rho}}$$

where $\Delta\rho$ = density difference between the two fluids;

ρ = density of ambient fluid.

The velocity, c , is defined in terms of the supply rate:

$$c = \frac{Q}{h}$$

where Q = supply rate or volume of flow passing a fixed point per unit time, and not, as in Middleton (1970), in terms of the velocity of the head as this is modified by processes operating in and around the head. The velocity coefficient, K , in the ideal case, can be derived from equation (4.5) and is equal to 2. Experimental data give values of from 1 to 1.2 for K ; the discrepancy is explained by energy losses from the system due to viscous drag and loss of fluid due to turbulence (Benjamin, 1968). A value of 1.07 is proposed by Keulegan (1958) as a universal approximation.

Experiments show that density currents behave as predicted by equation (4.6) provided that the Reynolds number is large (a condition satisfied by virtually all natural flows). Viscous drag causes the leading point of the head to be displaced away from the boundary (Figure IV.3). The motion within the head is turbulent and diverging about a central streamline. Large rapidly circulating vortices are present at the rear of the head and fluid may be lost to the medium (Figure IV.3). Because of this the velocity of the head is generally less than the body of the flow as fluid from the body is continually replacing that lost from the head (Middleton, 1970).

Previous work, both theoretical and experimental, has been concerned with turbidity currents. The problem involved in an attempt to model *nuées ardentes* is the lack of observational data concerning the densities and sizes of the flows. Sparks and Wilson (1976) estimated the initial density of an eruption column to be at least five times that of the atmosphere and this value has been used as a guide. The flow velocity is plotted as a function of height of flow and excess density in Figure V.4. Perret's observations of a flow 5m thick travelling at about 10 metres per second, are clearly of the right order of magnitude to agree with theoretical consideration. Similarly Lacroix's estimate for *nuées* of 50 metres per second would be compatible with flow thicknesses of 20 - 50 m for the density

contrasts considered. The thickness of ash-flows producing ignimbrites is not known but estimates of 100 m would imply velocities of the order of 100 metres per second.

Fluidization

In the preceding paragraphs the assumption has been made that nuées ardentes behave as dense fluids. In this section the fluidization hypothesis, based on the industrial process, is discussed in regard to nuées and its implications considered.

In the industrial process of fluidization, gas is passed through a bed of fine-grained solid particles. As the gas velocity increases, the bed expands and the individual particles become free to move. The bed of particles adopts a fluid-like appearance and is said to be fluidized. Fluidization occurs when the velocity of the gas is sufficient to support the particles. As the velocity increases, bubbles of gas form, entraining the particles, and the bed is said to be boiling (Reynolds, 1954).

The expanded bed has a definite, and roughly horizontal, upper surface, a measurable viscosity and a density analogous to that of a liquid so that objects of lower specific gravity will float. Both the density and viscosity of the expanded bed increase with the size of the particle. For a given material, the viscosity is lower when the particles are uneven in size and the addition of fine-grained material to a bed of coarser-grained material increases the mobility markedly.

The minimum fluidization velocity for a bed of particles, assuming constant sphericity and constant voidage at minimum fluidization state, is given by the expression (after Sparks, 1976):

$$adV_{mf}^2g + bV_{mf} = \frac{d^2\rho_g(\Delta\rho)}{\mu^2} \quad (4.1)$$

where : V_{mf} = minimum fluidization velocity;

d = mean particle diameter;

ρ_g = density of gas;

$\Delta\rho$ = density difference between gas and solids;

μ = gas viscosity;

g = acceleration due to gravity;

a, b = constants.

V_{mf} is, for relatively small particles, a function of d^2 . Departure from spherical shape can reduce V_{mf} substantially for a given particle size (Sparks, op. cit.).

The terminal fall velocity, V_t , is given by the expression (Sparks, op. cit.)

$$V_t = \left[ag \cdot \frac{\Delta\rho \cdot d}{\rho_g} \right]^{\frac{1}{2}} \quad (\text{for Reynolds Number } > 500) \quad (4.2)$$

V_t is independent of viscosity for high Reynolds' numbers.

V_{mf} and V_t are plotted as a function of grain size (Figure IV.5). It is apparent that for a wide range in grain size three differing states of fluidization for a given gas velocity will exist: non-fluidized, fluidized, and entrained. The maximum extent of the fluidized phase can only cover approximately one order of magnitude of grain size, whereas ignimbrites can be expected to cover at least four. As the gas velocity exceeds V_t , the particles become entrained and may be removed from the system.

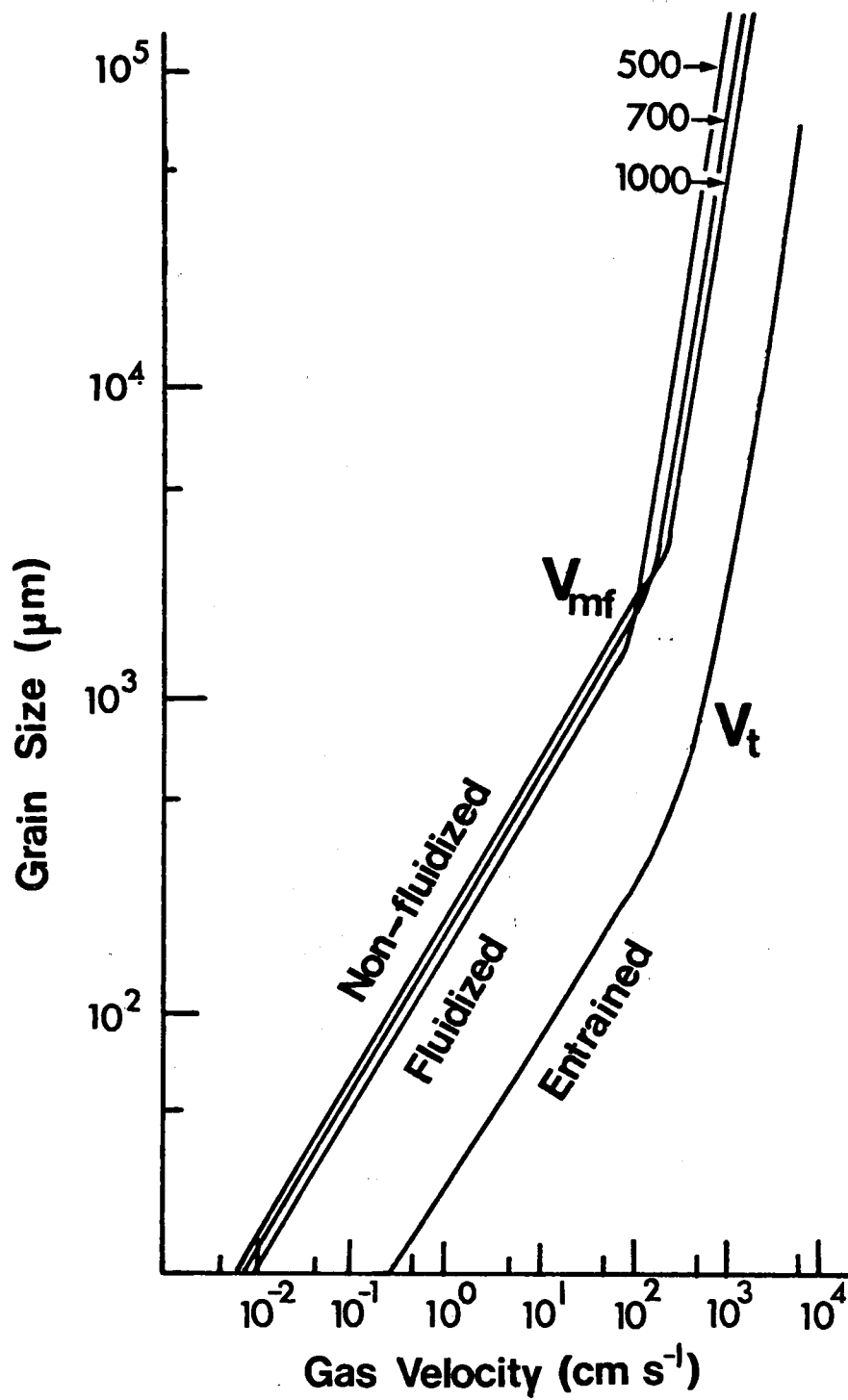


Fig. IV.5. Theoretical curves of the relationship of Minimum Fluidisation Velocity (V_{mf}) and Terminal Fall Velocity (V_t) with grainsize. Curves are calculated for spheres of density 1 gm/cm³ in CO₂ at 500, 700, 1000°C. The V_t curve is calculated at 700°C. (From Sparks, 1976.)

The estimates for V_{mf} based on theory will probably be too large for natural systems. The degree of angularity and the presence of particles of differing densities will lead to a number of particles of different sizes with the same V_{mf} . Fine-grained particles may, when fluidized, increase the apparent density of the gas and reduce V_{mf} still further (Sparks, 1976). The formation of bubbles in a fluidized system destroys the steady state and tends to produce a mixing of particles of differing densities (Rowe et al., 1962), and hence a lower apparent V_{mf} . In spite of these reservations it is hard to envisage a pyroclast flow consisting of particles with a size range of four orders of magnitude becoming completely fluidized.

McTaggart (1960) concluded that gas emission from the entrained magmatic particles was insufficient to maintain the fluidized flow and proposed a mechanism of rejuvenation by incorporation of air by the rolling motion of the flow. His experiments with hot andesitic sand showed a large dependence of mobility on temperature. He postulated that the dense lower part of the nuée was continuously engulfing cold air which on heating expanded progressively and explosively as it tore through the flow. The boiling appearance of the nuée is, presumably, due to the entrapment of air in a mechanism akin to waves breaking on a sea-shore. Further incorporation of air can arise by eddies on the rearward side of the head.

The evidence from observations and theoretical considerations implies that nuée ardentes are probably semi-fluidized. Turbulence of the ash-flow, however, will result in irregular air intake and hence a non-steady state with an irregular velocity distribution. Explosive bursts of gas will enable a flow to behave as if fluidized even when the mean gas velocity is less than the minimum fluidization velocity.

(d) FEATURES OF IGNIMBRITE DEPOSITS

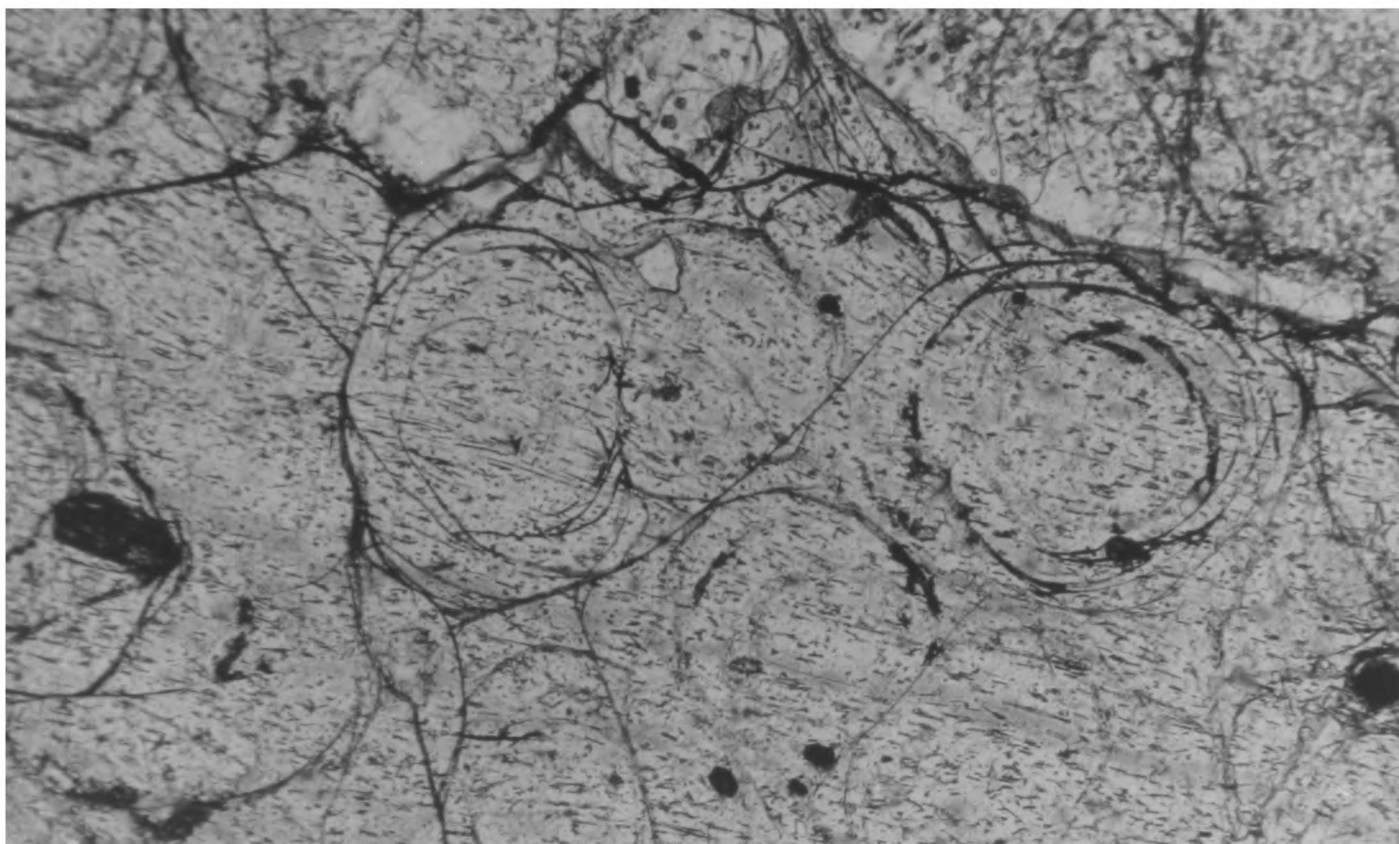
The usual and accepted model for ignimbrite emplacement, as discussed above, involves a glowing cloud of hot magmatic particles and associated pyroclasts advancing rapidly and over great distances from the eruption site. There tends to be a vague sorting of the supported clasts with pumice fragments floating to the top and lithic fragments sinking down. Some ignimbrites are completely homogeneous and unsorted (Sparks and Walker, 1973).

As the flow slows or the supporting gas pressure is reduced the suspended particles are dropped. If the temperature of the fragments is hot enough welding will occur. A temperature of 535°C has been quoted as the minimum welding temperature of a gas-rich glass and 750°C for an anhydrous glass (Smith et al., 1958). Some ignimbrites are non-welded; the great majority are all or partially welded.

The presence of welded shards is indicative of ignimbrites. With sufficiently high temperatures complete homogenisation of the shards may be expected and hence an obliteration of the primary vitroclastic features may be observed in the centres of flows (Ross and Smith, 1961). Ignimbrite sheets tend to be thick and the temperatures within would be expected to be high as ash-flows have been shown to be a thermally efficient method of emplacement (Boyd, 1961).

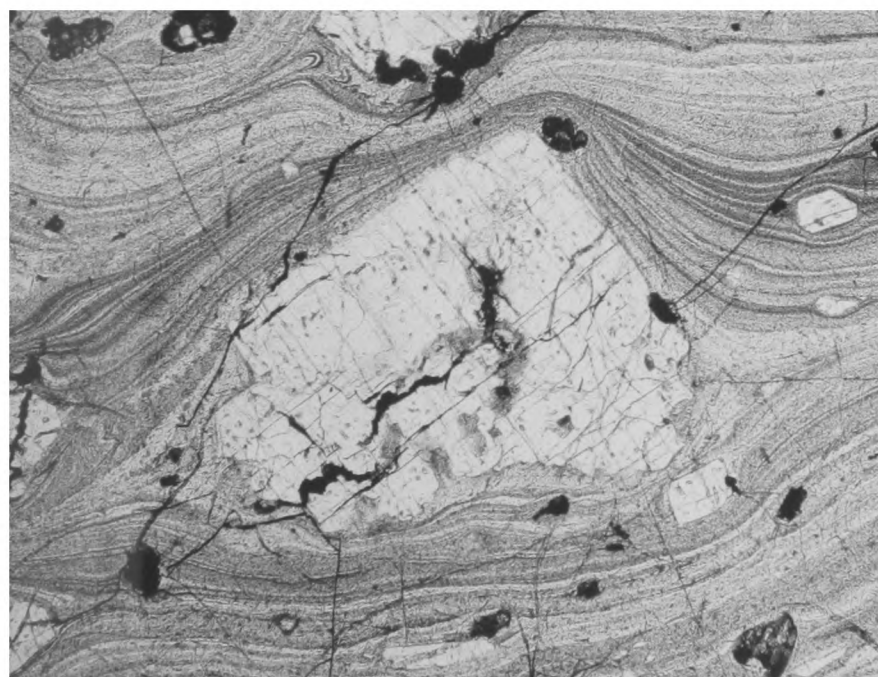
Many more recent ignimbrites can be divided into textural zones caused by differences in degrees of welding and compaction (see Briggs, 1976 for details and references), and mineralogical zones related to concentration of phenocrysts (Lipman et al., 1966). The Lebombo flows are generally extremely massive, showing none of the textural features typical of recent ignimbrites, and it seems that their large thicknesses may have served to obliterate the primary features by welding and compaction. Preliminary studies of the phenocryst abundances (this study and Bristow, 1976) have not revealed any obvious modal zonation; this may be due to these large cooling units not preserving their initial mineralogical stratification during eruption, or to the presence of a number of flow units within a cooling unit.

The majority of the Swaziland flows are completely massive and show few primary textures. In the eutaxitic zone rocks and in a banded pitchstone (Plate IV.33) the groundmass is often finely banded. This banding is slightly discontinuous and is probably due to an advanced degree of welding as are parataxitic textures (Plate IV.10); extreme compaction and elongation of the shards produces the fine banding.



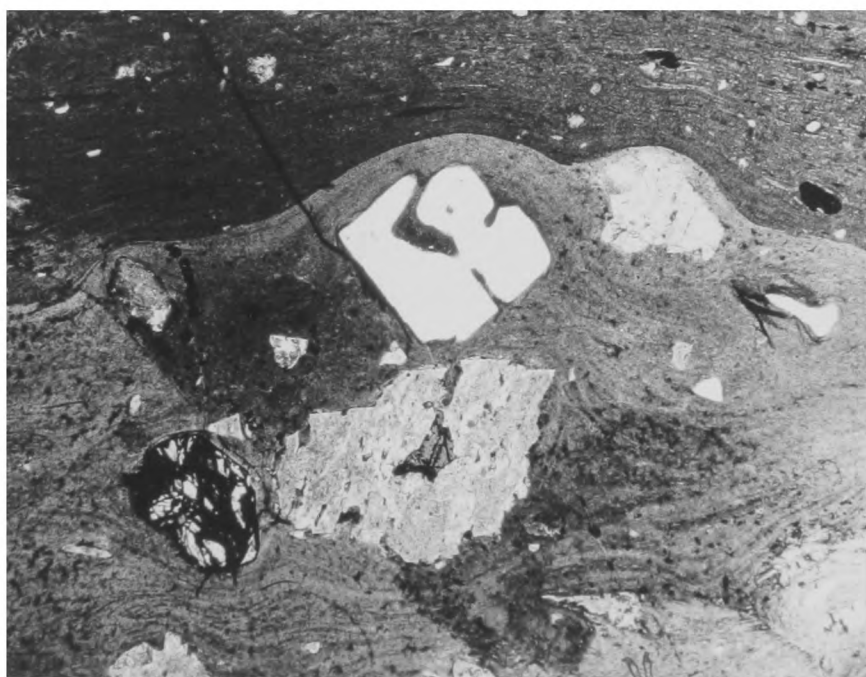
1 mm

PLATE IV.33 Rhyolitic pitchstone showing perlitic cracks. Faint darker, discontinuous bands can be seen in the groundmass; the acicular microlites are generally aligned parallel.



1mm

PLATE IV.34 Asymmetrical drape structures round feldspar phenocryst in pitchstone. (RC 144)



1mm

PLATE IV.35 Symmetrical drape structures at base of parataxitic band. Central quartz crystal with antiperthite and pseudomorph after fayalite below. (MZ 35)

Ash-flows usually carry pumice clasts which when deposited, collapsed and elongated, form the characteristic fiamme, diagnostic of ash-flow deposits (Ross and Smith, 1961). Fiamme were found in one horizon in Swaziland and represent the only direct evidence for an origin by pyroclastic flow. As described in section IV.ii.b, these fiamme grade into a well-laminated zone which becomes progressively more finely laminated and more streaky in nature until it grades into a pitchstone or its devitrified equivalent. No single flow shows the whole sequence but the complete gradation can be pieced together from the basal zone of several flows. The eutaxitic zone is seldom more than a few millimetres in thickness and eutaxitic textures are not found above this horizon. Apart from the breccia zone there are no pumice or lithic clasts observed in the massive zone.

Above the massive zone, flow folds appear. There are present throughout some flows and at the tops of the majority of flows. Flow folds are characteristic of viscous liquids and are ubiquitous in rhyolite lava flows. Many authors have described flow laminations in ignimbrites but more as exceptions rather than the rule (e.g. Rittman, 1962; Smith, 1960a; Rutten and Van Everdingen, 1961; Schmincke and Swanson, 1967; Lock, 1972) and in most cases the contortion is attributed to secondary remobilisation.

Rittman (1962), describing ignimbrites from Monte Amiata, Tuscany, proposed the term *rheoignimbrite* to describe the unusual flowage features. These rheoignimbrites form broad flat streams on the flanks of the mountain projecting lobe-like from an ignimbrite sheet. Evenly spaced arcuate ridges occur on the surface of the rock. They are formed by upthrusting of the upper more viscous portion of the flow due to flow of the more fluid basal part. The uppermost layers are folded, broken and pushed up, while the more fluid material from the deeper layers has intruded into the resulting anticlinal structures and into the planes of fracture.

This slow flow of material completely destroys the characteristic ignimbrite textures. There is no trace of welded glass splinters in the groundmass; it is homogeneous and often contains perlitic cracks. Flow texture is fairly common, recognised in thin vesiculated layers formed along shear planes. These deposits can be distinguished from lava flows by their absence of a slaggy upper surface and by their mode of occurrence; they form uniform sheets, even over undulating country, while lava streams are confined to topographic depressions (Rittman, 1962).

This concept has been applied by Rutten and Van Everdingen (1961) to a structure within the Permian Ramnes Volcano, Oslo Graben (Figure IV.6a, b). They interpreted the observed structures, previously thought to be part of a laccolith, as rheoignimbritic. A central volcanic plug is surrounded by outward dipping ignimbrite flows of syenite porphyry and quartz porphyry. Normal ignimbrite textures grade laterally into rheoignimbrites with strong developed fluidal lamination and pronounced centripetal dips, often up to 70° . Their interpretation was similar to that of Rittman's and the flow features were considered to have formed when the ash flow came to rest, remelted and flowed downhill. Lock (1972) described a rheoignimbrite from Otter Cove, Newfoundland. He accepted the theories for its formation proposed by Rittman but believed the Otter Cove ignimbrite to have suffered more gas exsolution, the final product being more brittle. This brittleness led to the development of arcuate shear zones (Figure IV.6c, d), analogous to Rittman's 'ruga' or wrinkles, along which a high proportion of the gases escaped and tuffisites were deposited and fumaroles (not preserved) developed at the surface. The banding in the Otter Cove rheoignimbrite is extremely regular and dips towards the postulated source region at angles of 50° - 70° (after correction for tilt). The shear planes are parallel with the banding.

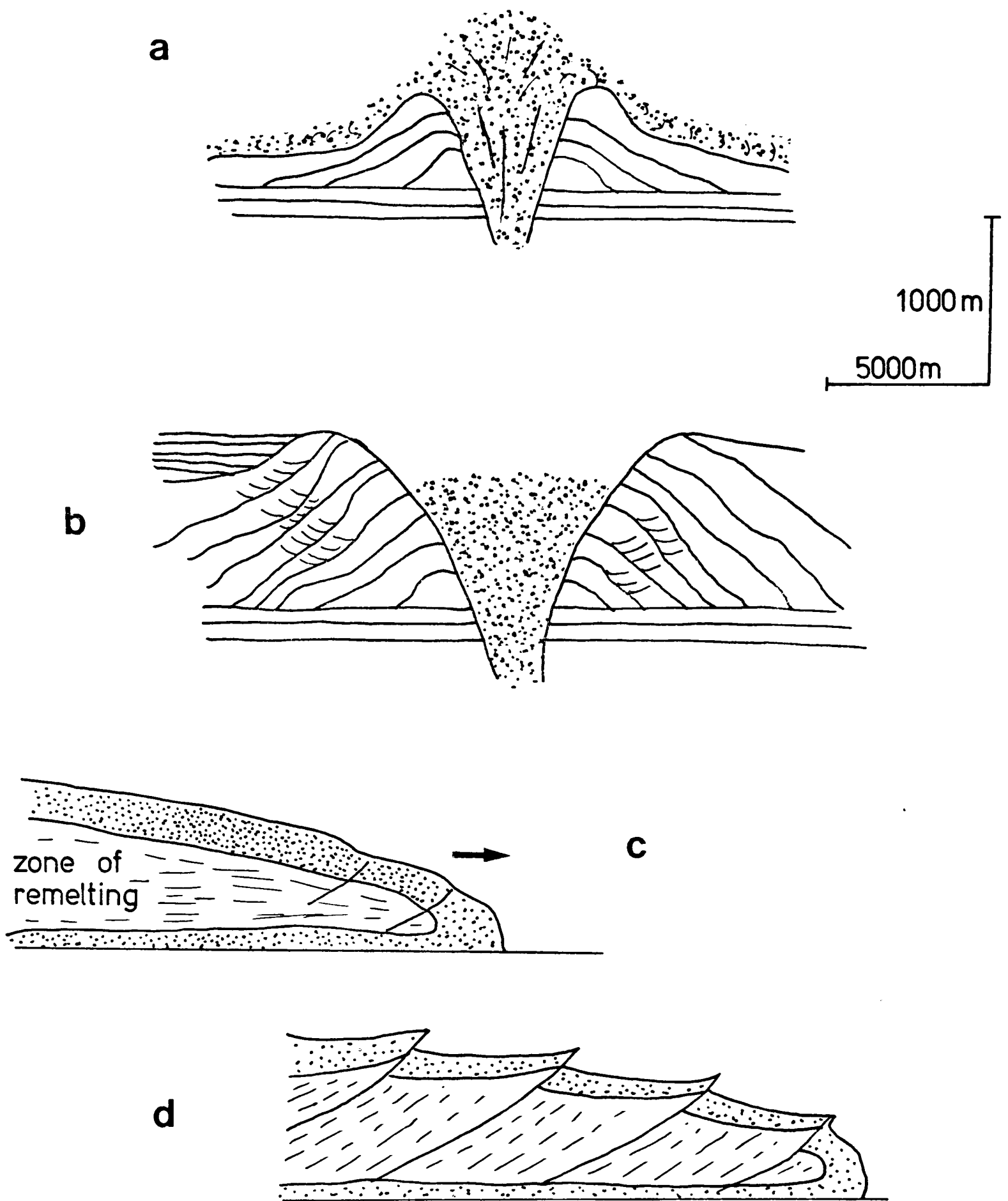


Fig. IV.6. Formation of rheoignimbrites

- a, b. Ramnes Volcano, Oslo Graben (from Rutten and Van Everdingen, 1963). (a) A vent fills with fluidized magma and erupts ignimbrites. (b) The complete volcano showing centripetally dipping rheoignimbritic banding.
- c, d. Detail from a single rheoignimbrite showing steepening of dip and the development of shear planes (from Lock, 1972).

Schmincke and Swanson (1967), describing flowage structures in ash-flow tuffs from Gran Canaria, concluded that the flow features are caused by primary laminar flow in the ignimbrite. The described stretched, broken and pulled apart pumice fragments, tension cracks in the matrix, hollows round inclusions, folds, imbricated pumice fragments and ramp structures. Most of the structures are directional and yield consistent movement directions for the ash-flows. The ash-flows are interpreted as moving as viscous, inhomogeneous liquids immediately before coming to rest. The flow is explained as primary due to the presence of fold structures throughout each flow, regardless of slope of substrate, and are even found in the flat-lying beds exposed in the caldera. Rittman (1962) hinted that the Monte Amiata rheo-ignimbrites were local 'facies' variants of a main ignimbrite sheet. Schmincke and Swanson (op. cit.) consider a coincidence of a slope and a hot magma, which caused the rheoignimbrites, to be unlikely to occur in all fifteen flows. They proposed an eruption mechanism whereby the ash flow, probably never greatly inflated, becomes slower and denser with degassing. As the density increased, compaction welding occurred to form a viscous fluid; the momentum of the flow was then sufficient to produce the flow structure.

The flow folds observed in the Lebombo rhyolites are very irregular and no consistent flow directions could be calculated in Swaziland, Zululand (Bristow, 1976) or Mozambique (Wachendorf, 1971, 1973).

While the folds at the tops of the flows are very irregular, laminar flow has taken place at the base and isoclinal folds and large amounts of extension are the result. The structures are present in virtually all the flows and, though often obliterated by devitrification, are probably present throughout each flow. The absence of flow autobreccias at the base of the flow, a characteristic of lava flows (Christiansen and Lipman, 1966), and

the presence of primary ignimbrite textures implies that the fold structures were formed in a viscous liquid deposited as an ignimbrite.

The presence of asymmetrical drape structures around phenocrysts has been cited as proof of flow (Christiansen and Lipman, *op. cit.*) and symmetrical structures as proof of compaction (Plate IV.34, 35). They are both present in the Lebombo Rhyolites and it seems that compaction has definitely occurred at the base of most flows. Higher up flow laminae are generally asymmetrical about phenocrysts.

One unusual feature about the Swaziland rocks is the presence of primary banding inclined steeper than the dip of the flow. This was observed throughout the succession and is especially obvious in the Mbuluzi and Usutu gorges. This may be explained by considering the discussion by Lock (1972) on rheo-ignimbrites where he stated that the banding of the flows is regular and dips steeply towards the centre of eruption. If the same mechanism was operative in Swaziland it would imply that the rhyolites were remobilised and flowed downhill to the west (Figure IV.7). For a likely volcanic topography a source to the east seems probable; an idea supported by the absence of volcanic necks to the west of the outcrop.

The other features described by Lock (*op. cit.*): ramp structures (one observed by Bristow (1976)), tuffisite veins and fumarolic alteration, are also observed in Swaziland but are not common and do not bear the same relationships to the flows. The applicability of a rheoignimbrite hypothesis to the rhyolites is questionable and depends on the flows being erupted on to a slope throughout the succession. Rittman's original rheoignimbrites were small in extent and part of a much larger sheet.

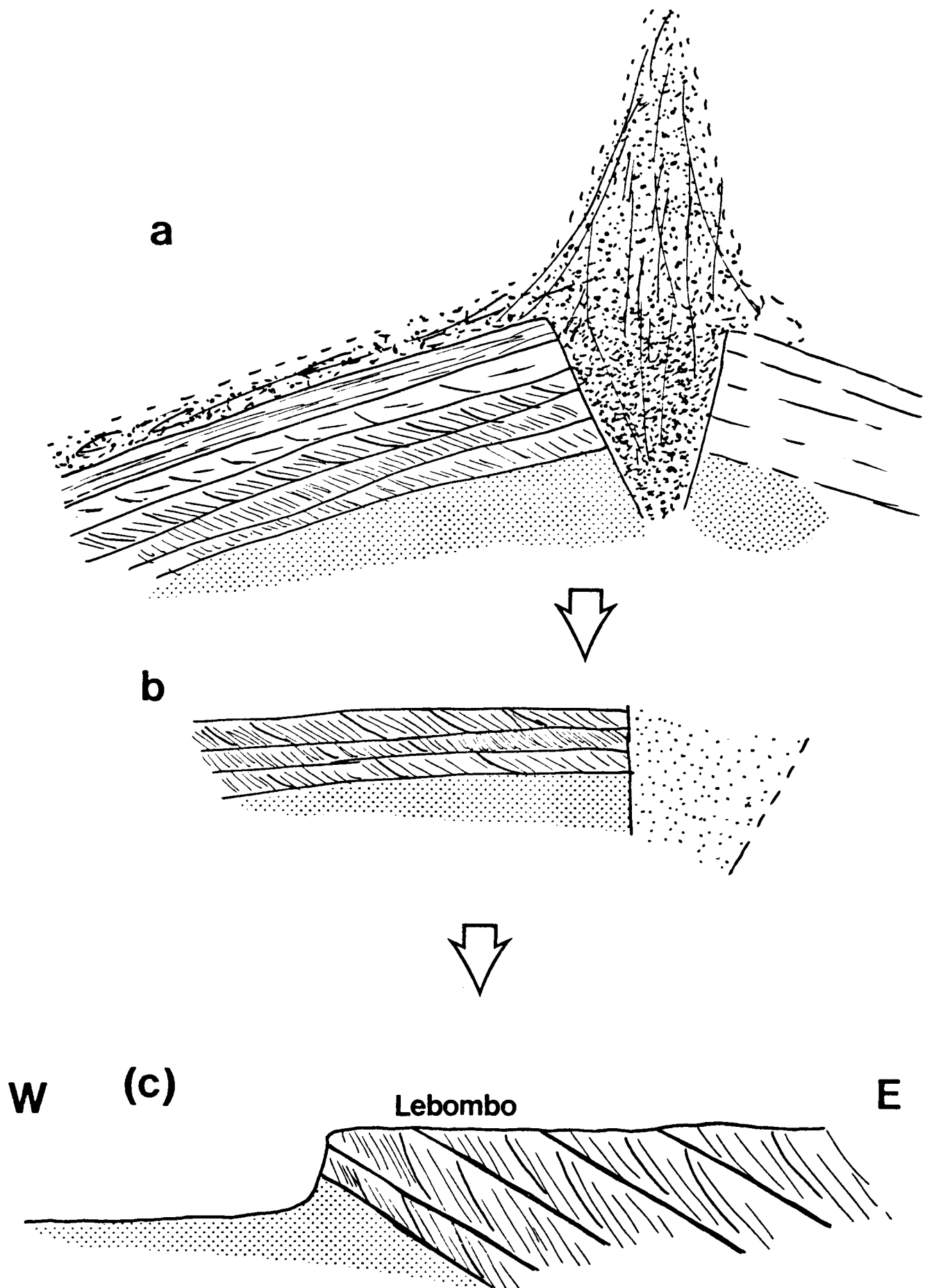


Fig. IV.7. Rheoignimbrite model for the Lebombo

(a) Eruption of ignimbrites on a westerly-dipping slope followed by remobilisation to produce centripetal dips (after Rutten and Van Everdingen, 1963).

(b) Downwarping to produce the present-day structure (c).

The source area is postulated to be to the east of the Lebombo as eruption centres are generally associated with up-doming or positive topographic features.

(e) THE ORIGIN OF FLOW BANDING : A PHYSICAL SYNTHESIS

The depositional mechanism, as outlined above, involves the transport of particles in a fluidized or semi-fluidized state. As the mean gas velocity becomes less than the fluidization velocity for a given particle, that particle will be dropped from the system. In an ideal system, this would lead to a graded bed; turbulence and varying properties of the particles and suspending fluid will produce lesser amounts of sorting. A hot layer of magmatic particles will be produced which may have primary lamination. Any contorted flow within the ash-flow will not be preserved as contortions in the resulting deposit.

Any structures observed in the rhyolites, other than primary depositional lamination must, therefore, be produced by post-emplacement processes.

These can be divided into three types:

- 1) Body forces;
- 2) Externally applied stress;
- 3) Thermal processes.

1) Body forces

Gravitational instability due to deposition on a slope or to irregular flow thickness will result in flowage down slope or lateral spreading respectively. Flow will occur throughout the rhyolite producing visible banding if the flow laminae are preserved. The flow folds at the top of the rhyolite will tend to be irregular, due to inhomogeneities within the deposit; those at the base, due to shear from the overlying moving flow, will be near-horizontal and more uniform in nature.

The velocity of a viscous lava flow can be described by the equation for flow in a wide channel (Friedman, Long and Smith, 1963):

$$V = g \sin \alpha \, d^2 \rho / 3\mu$$

where V = velocity of flow;

g = gravitational acceleration;

α = angle of slope;

d = flow thickness;

ρ = density;

μ = viscosity.

For a 200 m ignimbrite on a slope of 5° composed of rhyolite glass at 900°C with a viscosity of 10^{10} poise (Figure IV.8) and a density of 2.2 gm/cc, the velocity of movement would be 2.5 cm s^{-1} (0.1 km/hr). This will be a maximum value as the viscosity used is a minimum value and will increase with fall in temperature towards the top and bottom of the flow. At 800°C $V = 0.001 \text{ km/hr}$.

The Reynolds number for this flow:

$$\text{Re} = \frac{Vd}{\nu} = \frac{Vd\rho}{\mu}$$

where ν is the kinematic viscosity, is 10^{-5} for 900°C and 10^{-9} for 800°C (the critical value for turbulent flow is around 10^4) and the flow within the ignimbrite should be laminar, a conclusion borne out by field evidence.

Boyd (1961) estimated the cooling rate of a 30 m pyroclast flow after emplacement to be about $5^\circ/\text{hr}$. A 200 m flow would cool at about a fifth of the rate, i.e. $1^\circ/\text{hr}$. The distance travelled before cooling reduces the flow rate is of the order of 10 km for a flow initially at 900°C and of the order of 0.1 km for a flow at 800°C (for a 5° slope).

These flow rates are sufficient to produce the large-scale flowage features observed but will not greatly affect the distribution of each ignimbrite. Lateral spreading of flows due to hydrostatic pressure will take place at similar rates, dependent on the height difference.

2) Externally Applied Stress

Shear applied externally to the flow could be either due to friction from the movement of the overlying flow or to processes involving the initial ash-flow.

Shear is present in the basal zones of the ignimbrites but is not transmitted through the contact and the tuffaceous zone rocks are not affected. Applied stress from the ash-flow itself will be more complex but if deposition from a fluidized flow is assumed, the main effect will be the deposition of glass particles with a forward momentum. Though producing a slight shearing in the uppermost band and perhaps some rotation of phenocrysts and development of laminae, shear stress applied by the ash-flow cannot be envisaged to have a large effect.

3) Thermal Processes

The stability of a fluid layer heated from below or cooled from above is dependent on the ratio of the gravitational force (due to thermal expansion) to the viscous force. The criterion for stability is given by the Rayleigh Number:

$$Ra = \frac{\alpha \Delta T g z^3}{\nu k}$$

where α = coefficient of thermal expansion;

ΔT = temperature difference across fluid layer;

g = gravitational acceleration;

z = thickness of fluid layer;

ν = kinematic viscosity ($= \mu/\rho$)

k = thermal diffusivity ($=$ thermal conductivity/specific heat/density).

Taking the values for fused quartz (Handbook of Chemical and Physical constants):

$$\alpha = 1.65 \times 10^{-6} \text{ } ^\circ\text{C}^{-1}$$

$$k = 8.3 \times 10^{-2} \text{ cm}^2 \text{ s}^{-1}$$

the viscosity for rhyolite glass from Figure IV.8 (from Friedman, Long and Smith, 1963) and $\rho = 2.2 \text{ gm/cc}$ the Rayleigh Number can be calculated as a function of z (Figure IV.9).

The major variable is that of viscosity which is a function of temperature, water content and, to a lesser extent, composition. Once over a thickness of 100 m Ra is relatively independent of flow thickness. The critical value for convection is dependent on the boundaries of the fluid layer but is of the order of 2×10^3 . The temperature is assumed to be maximum at the base of the flow and zero at the top but the Rayleigh Number is not sensitive to variations in temperature gradient.

From Figure IV.9 it can be seen that for a flow of 200 m thickness at 900°C the Rayleigh Number is 30,000 and the flow should be thermally unstable.

The value of Ra , and any other viscosity dependent parameter, is critically dependent on the temperature of the emplaced ignimbrite. Pre-emplacement temperature changes of the magma are mainly due to exsolution and expansion of the gas phases; other heat changes due to conduction, vesiculation, crystallisation, oxidation, etc., have been shown to be negligible (Boyd,

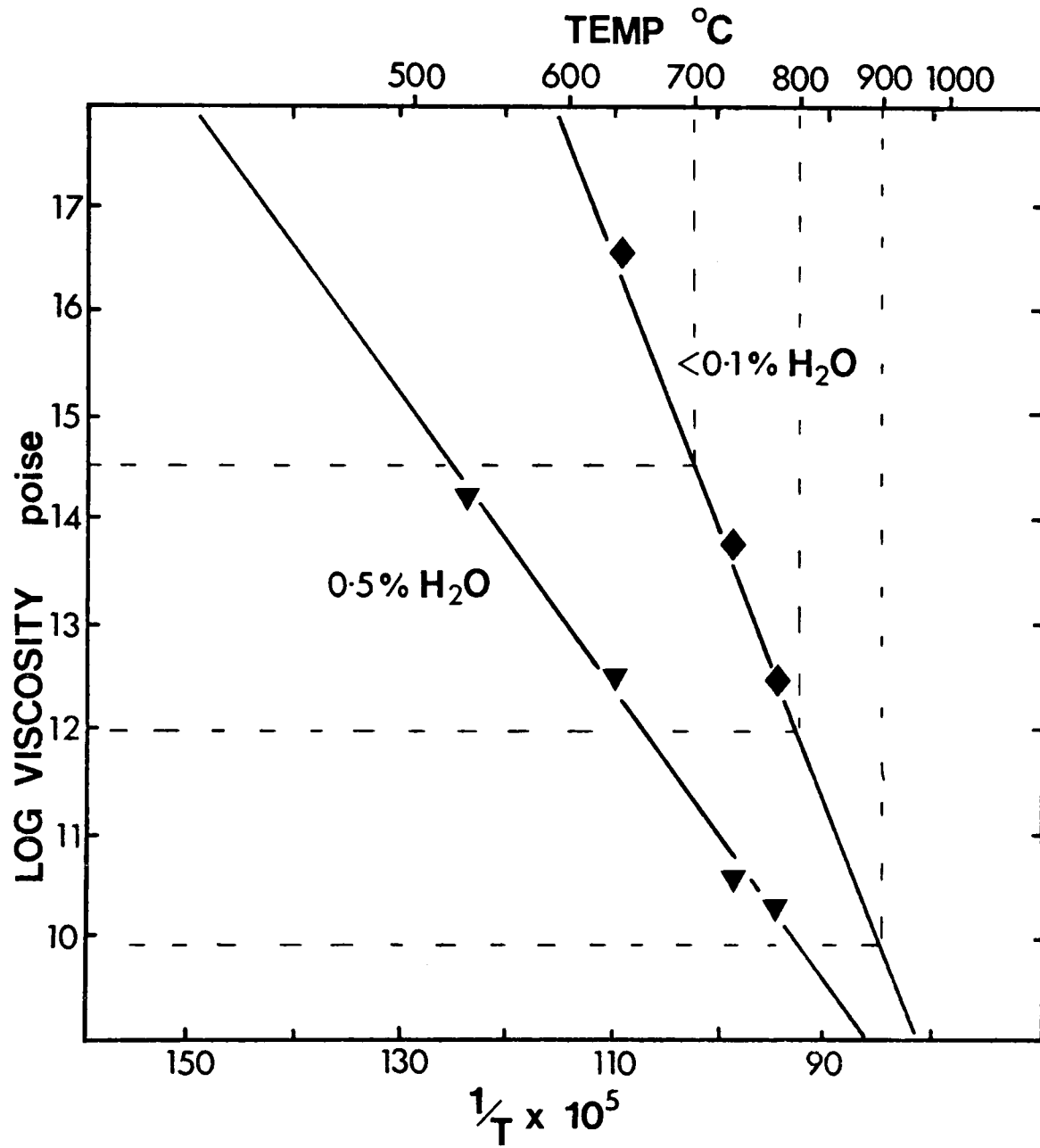


Fig. IV.8. Variation of the viscosity of rhyolite glass with temperature and water content (from Friedman, Long and Smith (1963)).

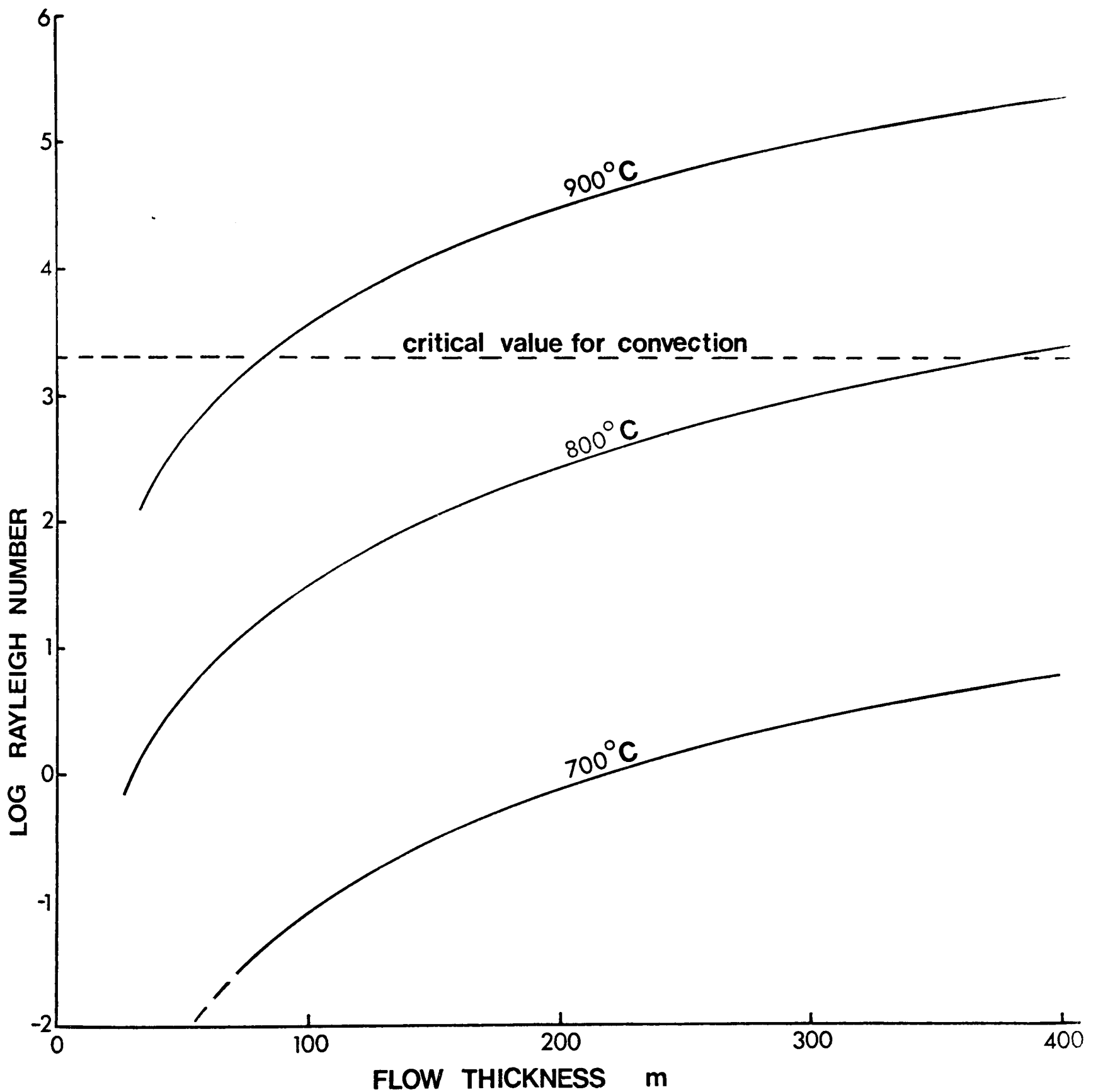


Fig. IV.9. Variation of Rayleigh Number with flow thickness and temperature for rhyolite glass. Viscosities estimated from Fig. IV.8 for 0.1% H₂O. These viscosities are maximum values and hence the Rayleigh Numbers represent minimum values. Ra = 2000 is taken as the limit for marginal convection.

1961). Boyd (op. cit.) calculated the total cooling of a rhyolitic magma assuming reversible, adiabatic expansion of the separate gas phase and allowing for the heat of evaporation of the gas and heat losses on eruption due to radiation and conduction (Figure IV.10). Boyd's calculations were for H_2O only; ΔH_{vap} for CO_2 can be calculated from the data on CO_2 solubility given by Mysen et al. (1976) and at 10 kb, ΔH_{vap} for CO_2 in an albite melt = 200 cal $s\ deg^{-1}\ g^{-1}$, and is of the same order of magnitude as that for H_2O . CO_2 solubility is insignificant in acid melts below 5 kb.

The initial temperature of the magma prior to eruption is difficult to estimate but the absence of hydrous phases and the presence of resorbed quartz and feldspar phenocrysts suggests that the magma was fairly dry and near or above the liquidus. Figure IV.11 shows the liquidus temperature for rhyolite magma as a function of pressure, temperature and water content. For low water contents and at low pressure the rhyolite liquidus is around 1000°C.

The cooling curve presented in Figure IV.10 is calculated assuming equilibrium between vapour, crystals and melt for each water content and hence represents a maximum heat loss for the system. If the magma erupted from 10 km (3 kb) to the surface the liquidus temperatures, amount of cooling and hence emplacement, temperatures can be calculated as a function of volatile content (Figure IV.12).

The emplacement temperatures are above 900°C for volatile contents of less than 1% (Figure IV.12). The viscosities used in the calculations are based on 0.1% H_2O and hence are maximum values. An increase in water content from 0.1% to 0.5% makes an order of magnitude difference in the viscosity (Figure IV.8) and hence also in the Rayleigh number.

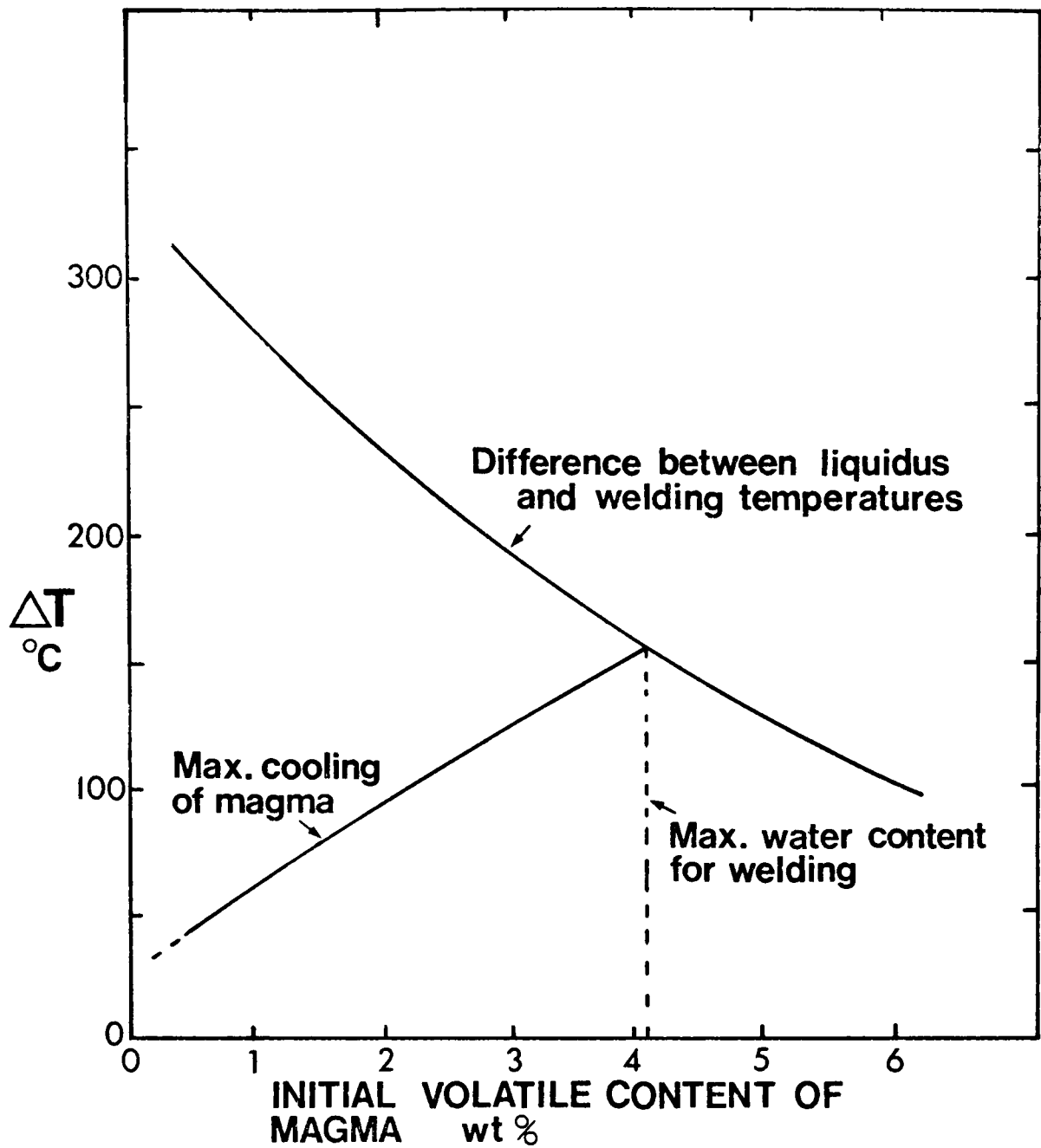


Fig. IV.10. Maximum total cooling of an ash flow during eruption due to exsolution and expansion of gas phase. Cooling due to heat losses during emplacement taken to be 25°C (estimated for 200 m flow from Boyds' data). Volatile content of magma assumed to be H₂O:CO₂ mixture ($\Delta H_{\text{evap}}^{\text{H}_2\text{O}} \approx \Delta H_{\text{evap}}^{\text{CO}_2}$) (After Boyd (1961).)

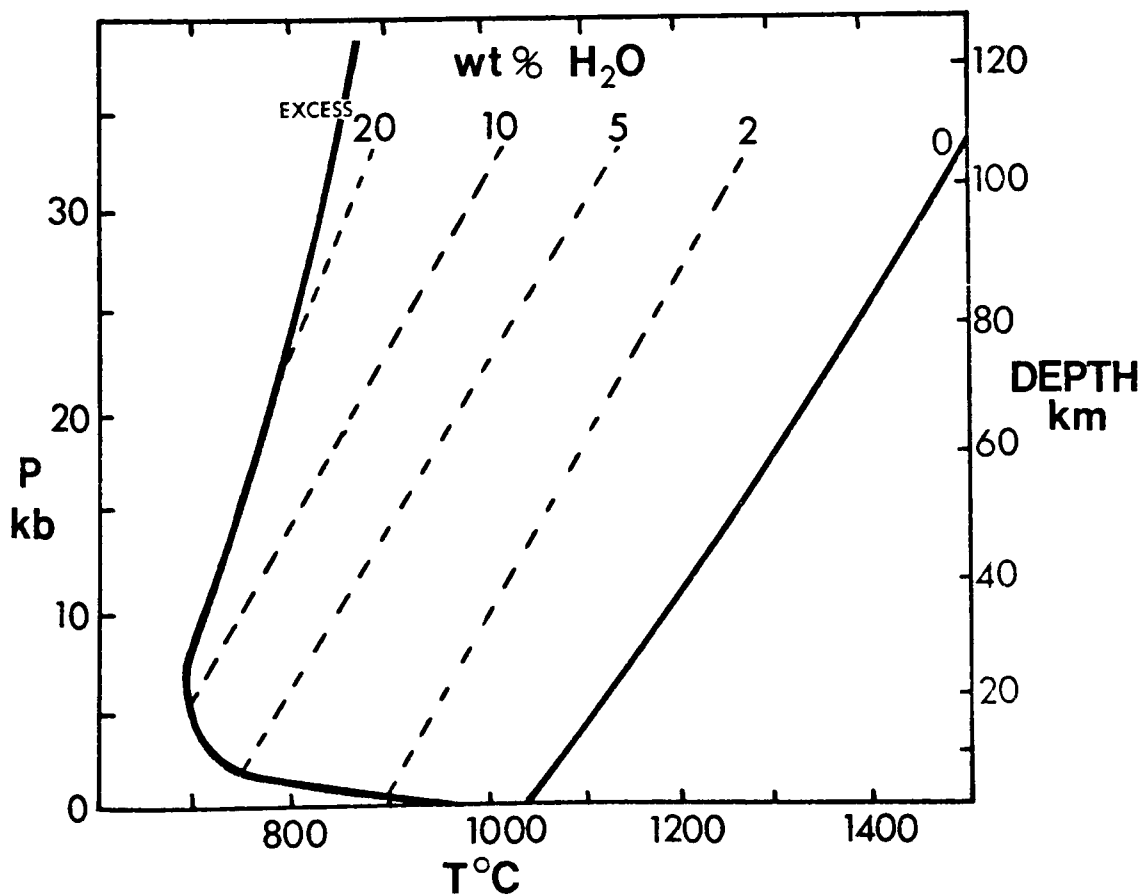


Fig. IV.11. Water undersaturated liquidus surface for rhyolite liquid, contoured for water content (from Stern et al., 1975).

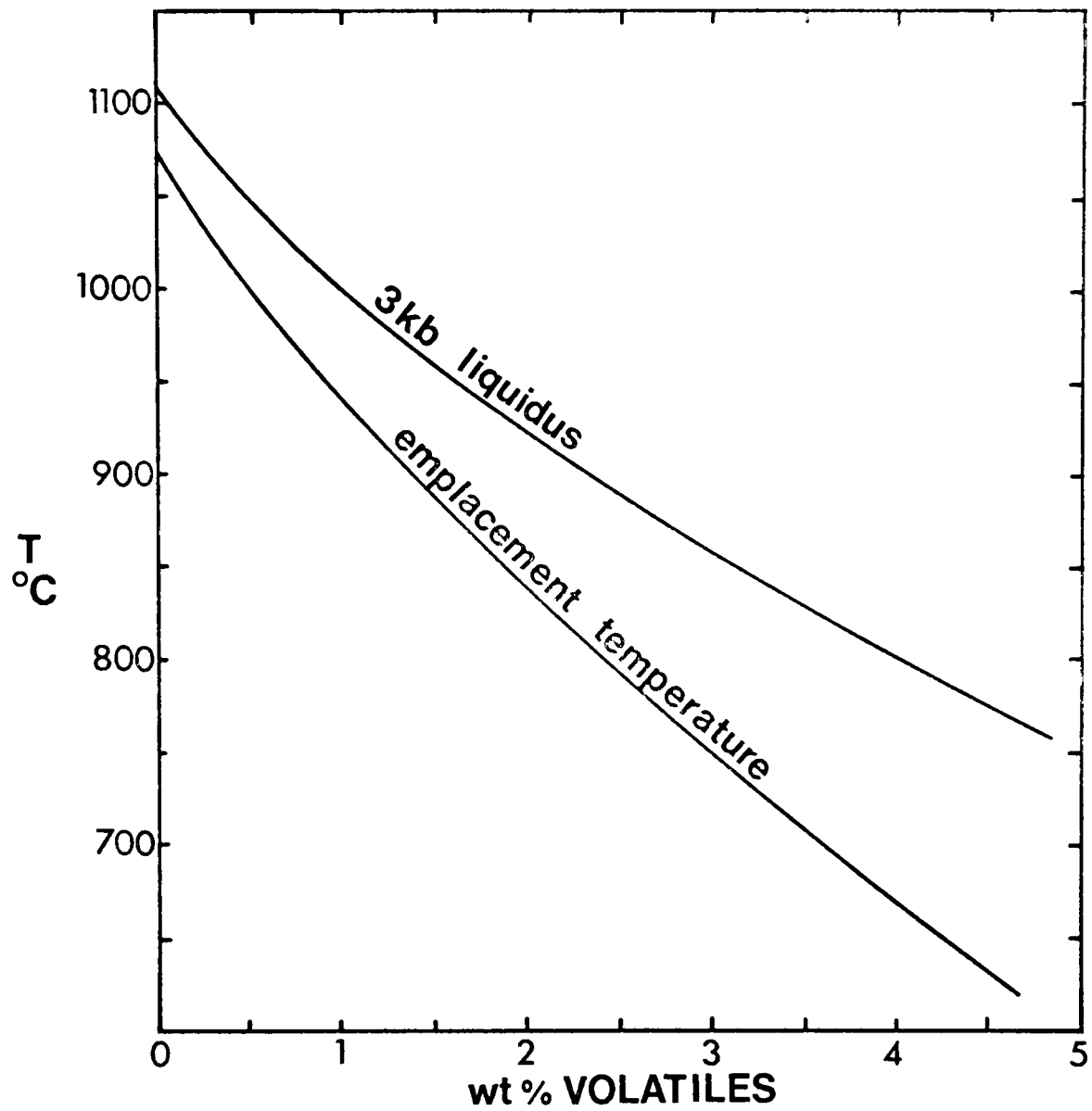


Fig. IV.12. Variation of liquidus and emplacement temperatures with volatile content. Liquidus temperatures from Fig. IV.11; emplacement temperatures calculated from liquidus temperatures and Fig. IV.10. Magma assumed to be at the liquidus at the start of eruption.

It is concluded that the flows are thick enough and hot enough for thermal instability and marginal convection to occur. The slow movement of the flow centres should be sufficient to obliterate primary textures and, depending on the preservation of flow laminae, produce some form of convection-related banding.

Summary

The Lebombo rhyolites form flows on average 200 m thick and extending for some 50 km along strike. Each flow can be divided into a series of textural zones, some or all of which are present in each flow. The basal eutaxitic zone rocks contain fiamme and other features indicative of an origin by pyroclastic flow. The overlying massive zone has had the primary features obscured by intense welding and devitrification. The contorted zone shows laminar fluidal textures which grade into the massive zone but are present throughout some flows. The contorted banding is probably caused by gravitational effects, either due to flow down a slope (as a rheognimbrite) or to lateral spreading due to initial hydrostatic pressure differences. Four types of breccia are found: autobreccias due to brittle fracture of contorted folds, polymict extrusive breccias emplaced as pyroclastic flows, tuffisitic breccias, caused by gas escape, forming veins and dykes, fumarolic breccias due to high degrees of veining and hydrothermal activity. Inter-flow tuffs have none of the features of base surge deposits and may represent pyroclast fall material associated with pyroclast flows. They contain a high percentage of derived quartz grains.

The rhyolites are considered to have been emplaced as fluidized density currents with the fine particles increasing the apparent density of the fluid phase. Calculated velocities for a density current model are consistent with observations.

A discussion of the thermal history of the magma involved gives an idea of emplacement temperatures and the possibility of post-emplacement thermal convection.

The source of the eruptions is presumed to be to the east of the present outcrop; no fissures were observed in the rhyolites and the rheoignimbrite model suggests a west-facing slope to be compatible with accepted topographies for volcanic areas.

(iv) Intrusive Rocks

(a) Granophyre

Six separate granophyre bodies are found in Swaziland: Mananga Ridge (North of Mbuluzi), Lusoti (two bodies), Stegi (two large bodies) and Maphungwane. They all crop out as north-south elongated bodies, intruding the basalt close to the base of the rhyolites. Those at Lusoti and Stegi may represent single larger intrusions. Apart from the granophyre at Maphungwane which has no topographic relief, they form hills of prominent relief, that at Mananga forming a steep escarpment. The granophyre is rarely exposed in situ and tends to form large rounded boulders.

Texturally the granophyres are all very similar. They are coarse- to medium-grained with quartz, two feldspars and a mafic phase (hornblende, augite, and/or biotite) as the main constituents. Fluorite, apatite and zircon are the main accessories. The quartz is always intergrown with the feldspar in a granophyric texture which is fine near the margins of the intrusion and coarser in the centre. There is a large colour variation in the granophyres, both between and within intrusions. When fresh the granophyres are commonly bluish-grey though reddish varieties are also found.

No contacts of the granophyres were seen exposed and their temporal relation-

ship with the rhyolites is uncertain. Small granophyre dykes cut the rhyolites and south of Stegi the southern of the two granophyre bodies appears to cut into the base of the rhyolites. In the Kruger Park, however large bodies of granophyre are found within the rhyolites (Saggerson, pers. comm.) and it seems probable that the granophyres in Swaziland are also later than the rhyolites.

Gravity and magnetic data show the Stegi granophyre to be an elongate dyke dipping to the east (Burley et al., 1970). Further detailed studies have not been done though the magnetic anomaly associated with the Maphungwane intrusion suggests that the granophyre may be considerably more extensive than the surface outcrop (aeromagnetic map of Swaziland, in Burley et al., 1970).

Small outcrops of granophyre occur to the east of the main Mananga intrusion. These form small topographic prominences and may be smaller associated bodies or parts of the top of the intrusion covered elsewhere by a basalt roof.

Granophyre dykes are common around the intrusions, the one east of the Maphungwane intrusion showing good contact relationships with the basalt and rhyolite flows. The dyke rocks are similar to, though finer grained than, the main intrusions; the largest one to the north of Stegi is indistinguishable from the north Stegi granophyre.

The orientations of the granophyre dykes are irregular in contrast to the uniform northwards elongated nature of the major intrusions. The dykes have a tendency to radiate from the main bodies but still have a dominant northerly trend.

(b) Dolerite Dykes

A large dolerite dyke swarm is present throughout most of the Swaziland lowveld. The main swarm trends slightly east of north intruding the Karroo sediments in the south of Swaziland and cuts obliquely across the basalt outcrop. It reaches its maximum development in the south of Swaziland, forming the Rooirand range of hills, but becomes less prominent further north, finally petering out around Stegi. Isolated dykes are found further north but no swarm could be identified.

The dolerites are resistant to weathering relative to the basalts and virtually all the outcrops in the lowveld are formed by dykes or basalt baked by dykes. Dolerite dykes in the rhyolites are extremely rare, even in the Mbuluzi Gorge exposures only four were found.

The thickness of individual dykes is hard to measure as the full width of the dykes is rarely exposed, dykes often forming lines of boulders. Where exposed in road cuttings the thickness of individual dykes ranges from 100 m to 1 m. Urie and Hunter (1963) reported an average thickness of between 25 and 30 metres. Bristow (1976) found that the thickness varied from 25 m to very thin dykelets and stringers.

The percentage of intrusive material is often very high in the axis of the swarm. Road cuttings in the Rooirand to the south of Swaziland contain over fifty per cent of dolerite intrusive into basalt. Bristow (op. cit.) quoted an average value of forty per cent for the Mkuze area. No estimate could be made for the extension in Swaziland.

The dip of the dykes is generally steep to the west, usually in the range of 60° - 70° though values as low as 45° are found. The structural aspect are described in Chapter V.

Bristow (op. cit.) recognised several phases of intrusion though with no significant difference in age or attitude. It seems unlikely that the two distinct periods of dyke intrusion found in the similar coastal monocline of east Greenland, one before and one after flexuring (Nielsen, 1976; Wager, 1947), exist in the Rooirand swarm.

Petrographically the dykes are all very similar. They are fine- to medium-grained and generally aphyric, though rare plagioclase phenocrysts are found. An ophitic intergrowth of plagioclase and augite forms the ground-mass. Amygdales are very rare and their presence was used to distinguish basalts from dolerites.

Large numbers of dolerite sills were emplaced into the sediments and basement beneath the basalts (Figure III.1) but no sills have been recognised in the basalt succession.

CHAPTER V

STRUCTURE AND STRATIGRAPHY

V STRUCTURE AND STRATIGRAPHY

V.1 Introduction

In the previous chapter the features observed within the rhyolites were described out of their regional context. In this chapter the rhyolite flows are treated as separate stratigraphical units and described in relation to the overall structure of the Lebombo. All the features referred to are found on the 1:50,000 map sheets, supplemented by a larger scale map of the Mbuluzi Valley (Diagram 1), the joint trend map (Diagram 4) and the geological cross-sections (Diagram 3).

V.2 Subdivision of the rhyolites

The rhyolites can be divided into three groups on the basis of hand-specimen mineralogy:

- (i) Basic Group The basic group is characterised by the rarity or absence of quartz phenocrysts. Rocks of this group are found predominantly at the base of the rhyolite succession though a few flows are present higher up. The main escarpment, except for a short stretch west and south of Stegi, is mainly formed by these basic rhyolites. Similar rocks are found in the Lebombo to the north and south of Swaziland. Two further subdivisions of the basic groups have been made:

Interbedded rhyolites These form narrow north-south strips within the upper part of the basalt succession to the south of Stegi and to the north of the Stegi granophyres to the Mbuluzi. They are petrographically very similar to the rocks at the base of the rhyolite succession proper.

Rhyolitic interbeds These form several thin flows in the lower part of the basalt succession. They are unremarkable in the field, being weathered and poorly exposed, but are chemically different from the main body of the rhyolites (see Chapter VIII).

- (ii) Acid Group Rocks of the acid group form three flows of approximately the same stratigraphical height. They are very distinctive in the field with an abundance of large quartz phenocrysts. These acid rhyolites frequently have well-developed fluidal banding throughout the thickness of the flow (e.g. Flow J in the Mbuluzi cuttings).
- (iii) Intermediate Group The intermediate group accounts for the bulk of the succession. The rocks are commonly pale-grey with quartz and white feldspar phenocrysts in a fine-grained matrix. They are often completely massive, rarely showing flow-banding or lamination. Except for some quartz-porphyrries in the Bumbeni Complex, quartz-pyric flows have not yet been recognised in the Lebombo within South Africa, but are found in Mozambique (Wachendorf, 1973; Assuncao et al., 1962).

Though these three groups are gradational into one another, the flows are usually easy to classify and are homogeneous enough for flows to be correlated over large distances purely on the basis of hand-specimen details. Adjacent flows belonging to the same group are generally separable on minor mineralogical and textural details (e.g. amount of quartz, colour of feldspar and presence of glomeroporphy); for clarity, they are given slightly different colours on the maps - these colours have no other significance.

In most cases the flow contacts are well-defined by topographic and textural features but in areas of low relief and poor exposure the flow subdivision is based purely on petrographical evidence. The area between Stegi and Groenpan contains a number of thin flows (IV, V, VI and VII). It is an area of low relief and little exposure and the mapping and subdivision of the flow is based entirely on small but significant differences in hand-specimen.

The mapping was mainly done by making east-west traverses and interpolating and extrapolating using air photos. The small extents of breccia- and tuff-horizons are a reflection of where they were encountered rather than their true distribution.

The true dip of individual flows is difficult to measure for reasons discussed in the previous chapter. Dips marked on the maps are, in general, dips on primary, non-contorted banding; dips calculated from the examination of flow contacts veering downstream are in general less than the dip measured on outcrop. Dip angles on the geological sections (Diagram 3) are thought to be the best possible estimates using both measured and calculated values.

V.3 General features of the rhyolite succession

Individual flow or cooling units have been mapped out and the sequence of flows and their stratigraphy has been determined. The structurally simple nature of the monocline has produced parallel, near north-south trending, outcrops for the rhyolite flows and the order of superposition can be easily seen.

The variability of the thicknesses of individual flows is partly caused by the identification of cooling units rather than flow units; where the exposure is better (in the Mbuluzi gorge) the average thickness of the flows

mapped is much less. Along the line of Section 4, flow XI is completely massive and homogeneous over the outcrop width of 6 km. Further south slight textural differences enable it to be divided into three thinner flows: XI, XIII and XIV. Where tuff and breccia horizons are absent, subdivision of large cooling units becomes more difficult.

The length along strike varies from 60 km (Flow XI) to 10 km (Flow XII) but in general is between 40 and 50 km. Because of the monoclinal structure no areal extent can be estimated although analogies can be drawn with the equivalent rocks in the Nuanetsi syncline where flows can be traced for up to 25 km along the length of the syncline and 13 km across it (Monkman, 1961). There is no reason to suppose that the lateral extent of the rhyolites in Swaziland is not of the same order of magnitude as their length parallel to strike.

The rhyolite outcrop has a pronounced sigmoidal warp with a broad synform and antiform centred on Stegi and the Mbuluzi respectively. The warping is most pronounced near the base and is gradually flattened out by the successive rhyolite flows so that the uppermost flows strike north-south.

There are numerous minor unconformities and overlaps within the rhyolite succession. Flow XXI effectively marks the end of the synformal warp and rests on six lower flows. None of the overlapped flows shows any large-scale truncation due to erosion, but often the contacts are not exposed. Flow XI overlaps Ia and IIa and the interbedded rhyolites and comes to rest against basalt, though with minor faults along the contact. The outcrops of the flows within the synform seem to be controlled by the synform and wedge out to the north and to the south. South of latitude $26^{\circ}45'$ the rhyolites resume a north-south strike. Some of the thinner flows may represent lobes of a larger flow- flow VI, for example, has a larger and a

smaller outcrop sandwiched between flows V and VII.

The two synforms, that centred around Stegi and that to the north of the Mbuluzi, do not seem to have any flows in common; the flows in the Stegi area, except perhaps Ia and IIa, are stratigraphically higher than those to the north. The centre portions of each of the synforms contain thick sequences of rhyolite which have a certain amount of interfingering but are basically of different ages.

The interbedded rhyolites act as marker horizons within the uppermost part of the basalt succession. The three parallel flows in the Big Bend area strike almost due north and can be traced 100 km south to the Pongola river. North of Big Bend they start curving to the west in an apparent reflection of the rhyolite outcrop, but are overlapped by the basal rhyolite flows and disappear just south of the southern Stegi granophyre. North of Stegi interbedded flows reappear, but here they are steeply dipping, forming narrow parallel ridges, and repeatedly faulted, both by north-south strike faults and smaller normal oblique faults (Plates V.1, 2). It was not possible to correlate the flows over the faulted zone, although in the Lusoti area (lat. $26^{\circ}16'5''$) a triple group of flows, two thick and one thinner, can be traced to the north and south.

These interbedded flows continue with a northerly strike and finally disappear to the north of the Mbuluzi. They are affected by the broad domal structure in the Mbuluzi poort and are cut by a number of small faults. They do not appear further north.

Trace element data (Figure VIII.3) show that the interbedded flows of the northern and southern areas (i.e. those to the north and south of Stegi) have similar Zr/Nb-Y characteristics although the uppermost, or most



PLATE V.1 View of area of interbedded flows north of Stegi. High ground on left is the western Stegi granophyre; that on extreme right background and in far distance is the main rhyolite plateau. Complex ridges and hills are the interbedded flows, generally steeply dipping and faulted. At the end of the interbedded flows the high point is the top of the Lusoti granophyre. Road in foreground is the 'red-line' foot-and-mouth track.



PLATE V.2 View looking north across the mouth of the Mbuluzi. Rhyolite plateau forming high ground on right; Lusoti granophyre highest point in middleground. Hills in distance belong to the rhyolite plateau on the north side of the Mbuluzi where the rhyolites trend north west. Picture taken from due west of the dip at Groenpan, from sample locality Sk 23.

eastern, of the interbedded flows from the two areas are more similar to the basal flows of the rhyolite succession. These data imply that the interbedded flows of the two areas are geochemically similar and may also be stratigraphically equivalent.

The sigmoidal warping can be attributed to a number of possible causes. It mainly affects the rhyolite/basalt contact and the lower part of the rhyolite succession; neither the top of the rhyolites nor the base of the basalts is affected. Post-basalt folding would have affected the base of the basalts and the interbedded rhyolite flows towards the top of the succession; post-rhyolite folding would also have affected the top of the rhyolites. The downwarping, therefore, appears to be a primary feature related to the emplacement of the rhyolites. A sub-rhyolite topography can be produced by erosion or the development of a primary dip on the surface of the basalt flows, as might be caused by the contact between two basalt 'shields'. The lack of parallelism between the interbedded flows and the base of the rhyolites, however, makes the latter argument implausible.

Overlap within the rhyolite succession is common, Flow XI transgressing over the lower flows to rest on basalt, and this has served to emphasise the synformal structure. Gravity data (Section V.9) show that a combination of a synformal nature for the upper surface of the basalts and overstep of the rhyolites eastwards can explain the anomalies. The differences between the basalt outcrop in the north and in the south of the country may be a function of erosional level: the rhyolites may have been transgressive over the basalts over the whole area; erosion to deeper structural levels in the south has caused the development of the contact east of that in the north.

V.4 Faults

(i) Strike faults

Many large north-south strike faults are observed in the sediments beneath the basalts, and displacing the base of the basalts in the south of the Swaziland (Figure III.1). None is observed within the basalts but the lack of marker horizons makes recognition of faulting difficult. No strike faults are observed within the rhyolites; the flow contacts, when observed, were of a stratigraphic nature and could be traced out regularly.

The strike faults occur in a narrow zone in the south of Swaziland coincident with, or to the west of, the main dyke swarm. The base of the basalts has a vertical displacement of 600 to 1000 m in a 5 km zone of faulting along the southern Swaziland boundary, the Cave Sandstone being repeatedly faulted up. Without the Cave Sandstone/basalt contact as a marker horizon, it is impossible to estimate the effect of this faulting in the main part of the basalt outcrop.

This zone of faulting disappears approximately level with the Usutu but more strike faults cut the interbedded rhyolites and the rhyolite/basalt contact in the area north of Stegi. The contact of Flow XI with basalt (lat. $26^{\circ}32'S$) is also affected by small north-south faults. One strike fault is exposed in a railway-cutting 1.5 km northeast of Lusoti ($26^{\circ}16'S$) where it faults rhyolite against basalt; the displacement is unmeasurable but is greater than the height of the outcrop (6 m) and presumably down-thrown to the west. West of this point a solitary fault is recorded from the base of the basalts (published geological map No. 14, Geol. Survey and Mines Dept., Swaziland).

It seems that the strike faulting remains more or less coincident with the dyke swarm, trending obliquely across the basalt outcrop. The effect of

this faulting will be to increase greatly the apparent thickness of the basalt succession. The narrowing of the outcrop to the north may be caused by the reduction in the frequency of the faulting and also the reduced crustal extension caused by the reduction in dyke density.

(ii) Oblique faults

Small northwest- and northeast-trending faults are abundant in the interbedded rhyolites and in the escarpment throughout the length of the Lebombo in Swaziland. They are only rarely observed within the rhyolites. The throws are small and never more than a few metres. The northeasterly faults are parallel to the main north-northeast joint set and many of the faults, especially in the largely unmapped area to the north of the Mbuluzi, may be prominent joint planes. One transcurrent fault, south of Stegi, displaces Flow XI on to the escarpment.

None of the complex faulting reported by Beater and Maud (1960), and Maud (1961) from southeast Zululand seems to be present in Swaziland. The structures reported seem to be confined to the southern termination of the monocline.

V.5 Joints

All the rocks in this area are jointed. However, the basalts rarely show jointing because of the poor outcrop in the lowveld, and the outcrop of the dolerites is generally insufficient to show joints other than the columns due to thermal contraction. The rhyolites are well-jointed, showing jointing on two main scales: small irregular columns due to cooling (see Plate IV.12) and larger, steep, tectonically-controlled joints, often indistinct on outcrop but clear on aerial photographs. The minor columnar joints are of local significance only; the major joint sets control the

major drainage patterns. In areas of dense jointing a pronounced topographic lineation is observed, accentuated by vegetation growth along the aquiferous joint-planes. Some major joints form large rubble-filled cracks.

Diagram 4 shows the main joint trends, inferred from the aerial photographs, and their angular distribution. Two prominent trends emerge: one at about 25°E and a second approximately east-west. There is probably a further set, in some areas, due north and a minor peak northwest. An inspection of the ERTS (LANDSAT) imagery reveals a prominent basement trend approximately west-northwest and the major joints in the centre of sheet 3 and in the north of sheet 2 may be basement-controlled. The southerly of these two joints also coincides with a prominent magnetic anomaly in the basement. An examination of the joint map reveals a proportion of joints parallel to the strike of the rhyolites and, especially in the Mbuluzi area (area 1.3, Diagram 4) the joint trend swings round as the rhyolite contacts are affected by the antiformal structure. This apparent stratigraphic control may be an artefact of the use of drainage patterns on aerial photographs to define joints; secondary streams flow along the softer interbeds and bedding planes and are hence controlled by the strike of the rhyolites rather than the superimposed jointing.

There do not seem to be any significant areal differences in the joint sets though in the extreme north, the northerly joint set is absent and the joint density appears to be less. The major north-northeast joint set is oblique to the dyke swarm and the monoclinial axis, and seems independent of basement trends. Monoclinial flexuring would be expected to produce two main tensional joint sets, one parallel to the axis along the convex portion of the fold and one normal to it in the concave part, and two associated sets of conjugate shear joints at an acute angle to the tensional joints. These four main joint sets may be able to account for the joints observed, with the notable

exceptions of the basement trend, discussed above, and the prominent north-northeast trend which, with no conjugate pair, appears unrelated to the formation of the monocline. The minor joint sets are complex and difficult to interpret without detailed measurements of hade and displacement.

The southeastern South African continental margin is characterised by a steep continental slope with a persistent, and near straight, northeast-southwest trend (Scrutton et al., 1973). It continues into the Agulhas Fracture Zone south of the Cape (Talwani and Eldholm, 1973), and it seems that the original continental fracture may have been caused by a shearing couple initiated by movement of the Falkland plateau southwestwards (Scrutton et al., 1973). Detailed interpretation of the coastal structure of Natal (Beater and Maud, 1960; Maud, 1961; Hardie, 1962) revealed that the complex fault pattern observed is explicable in terms of tension normal to the coastline (i.e. in a southeasterly direction), developed subsequent to the shearing stress (Scrutton et al., 1973). These large scale features are parallel or sub-parallel to the dominant east-of-north joint direction in Swaziland, and it is proposed that the jointing represents an overprinting of the late- or post-Karoo tensional regime on the preexisting monocline.

In summary, the Lebombo joints can probably be divided into four main components: minor columnar joints related to thermal contraction; infrequent major joints paralleling basement structures; tensional and shear joints due to monocline formation; and a regional overprint of a later tectonic event.

V.6 Dykes

The dykes are discussed in detail in Section IV.4.iv and only structural aspects are considered here.

There is an overall association of dykes and strike faults, both reaching their maximum development along the southern Swaziland border and fading out north of Stegi. There is no suggestion of strike faulting in the rhyolites and dykes are extremely rare, even where the dyke swarm trends into the rhyolite outcrop in the Lusoti area (Figure V.I).

The outcrop of the dykes in Swaziland is insufficient to interpret the temporal relationships of the dykes, though Bristow (1976) recognised several generations in Zululand. The two sets, one before and one after the formation of the monocline, recognised in east Greenland (Nielsen, 1976) do not seem to exist.

When exposed, the dykes are observed to have steeply to the west. The angle of hade seems, from the available outcrops, to be lower near the monoclinial axis; only away from the axis are the dykes approximately at right angles to the strata. In general, the strata have been tilted more than the dykes; an observation that led du Toit (1930) to conclude that the warping occurred during the eruptive period (see Section IV.1). The rare dykes cutting the rhyolites all seem to be vertical.

The dyke swarm seems to be related to the extension and faulting produced to the west of the monoclinial axis in the south of Swaziland. It is concentrated in a narrow zone in the basalt or underlying sediments which does not persist into the rhyolites. It is considered unlikely that the dykes entirely represent feeders for the basalt flows; they trend obliquely across the basalt outcrop and seem to post-date most of the basalt succession. The lack of dolerite in the rhyolites suggests that either the rhyolites are later than the dykes or are impenetrable to dolerite, or, more likely, the structural environment for dyke intrusion is restricted to the basalt outcrop and has largely faded out by the latitude of Stegi.

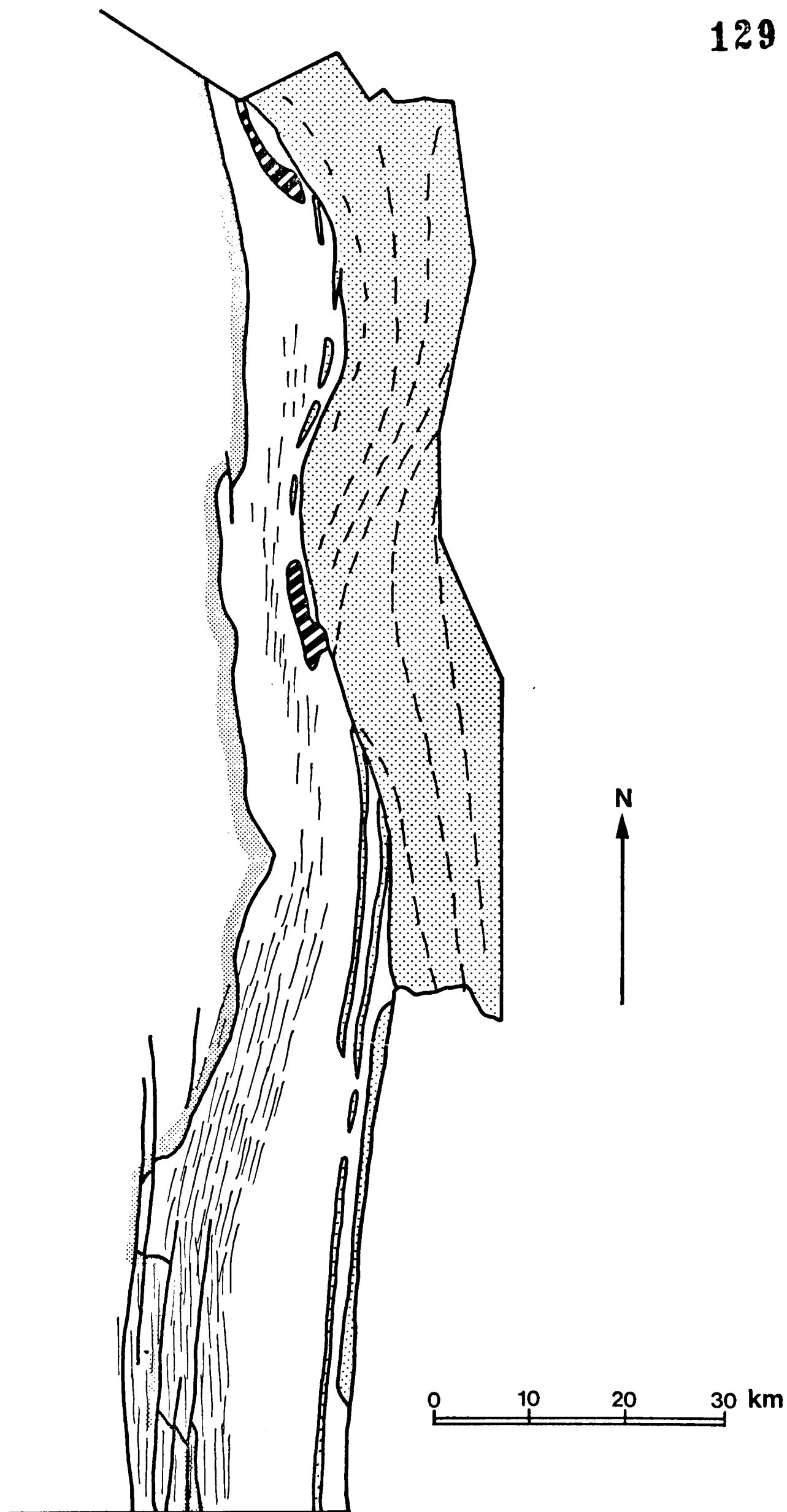


Fig. V.1. Skeletal geological map of the Lebombo in Swaziland showing the distribution and trend of the dyke swarm. Coarse stipple = rhyolite; fine stipple = limit of pre-volcanic rocks; diagonal hatch = granophyre. Major faults and contacts marked by bold lines, dolerites depicted by fine lines.

The slight change of direction south of Big Bend (Figure V.1) seems to have no other surface feature associated with it, and is not readily explicable by the Karroo tectonic framework. It may be a reflection of underlying basement structures.

V.7 Nature of the Monocline

The monoclinial structure of the volcanics has produced shallow dips in the sediments and at the base of the basalts, increasing to values of 45° and higher towards the monoclinial axis (roughly corresponding to the basalt/rhyolite contact), and decreasing towards the top of the rhyolites.

In the south of Swaziland the structure is comparatively simple; the escarpment is straight, striking almost due north-south. It is parallel to the strike of the interbedded flows and, presumably, the basalts and is coincident with the base of the rhyolites. To the south of Stegi, the escarpment bulges to the west, and is developed in the basalts on the western side of the southern Stegi granophyre. North of Stegi, it follows the rhyolite/basalt contact, but in the Lusoti area (lat. $26^{\circ}16'S$) it is developed away from the contact at the base of Flow II (the escarpment can be approximately defined on the contour-less maps by the position of the watershed); the area between the escarpment and the contact is occupied by steeply dipping rhyolites, forming a hummocky terrain (Plate V.3). Further north the escarpment continues along the base of Flow II, (or Flow A in the Mbuluzi Valley), passing just west of Nfunfu (lat. $26^{\circ}07'S$), through the dipping-tank at Duma and up to the border at Namaacha where it is 6 km east of the basalt outcrop.

In the Mananga area, a triple scarp is generated: on the west side of the Mananga granophyre; just west of Nkalishane, and the third through Namaacha.



PLATE V.3 View looking south towards Groenpan from sample locality MW 47, 4 km south-southeast of Mlaula station. The escarpment is developed at the base of the rhyolites in the background, and swings to the east so that the rhyolite-basalt contact runs along the hummocky ground on the right. A steeply dipping rhyolite unit in left middleground has a parallel foliation picked out by the foliage. The interbedded rhyolites and the lowveld beyond can be seen in the extreme right background.

The escarpments are erosional features and, as at Stegi, the resistant granophyre bodies have had an important effect on the erosional history. There is a small basic intrusion in the Nkalishane Valley, between the dip and farm of the same name (Urie, unpub. data; Snelling and Davies, unpub. data; Betton, pers. comm.); it is deeply eroded and may have caused the plateau between the second and third escarpments. The zone of steep dips, observed in the Mbuluzi poort, does not seem to persist north of the Mbuluzi.

Bristow (1976) located the monoclinal axis in the southern Lebombo by finding the locus of maximum dip along river sections (the locus of maximum dip, rather than maximum curvature, is taken, for ease of measurement, to represent the monoclinal axis). In the area between the Mkuze and Pongola rivers, he estimated this axis to trend 5° west of north and to intersect the southern Swaziland border 3 km west of the base of the rhyolites and about 17 km east of the Rooirand dyke swarm. The axis is more difficult to define in Swaziland because of the poorer exposure. The rhyolite/basalt contact in the south of Swaziland has a dip of about 30° (Plate V.4); the interbedded flows south of Big Bend have dips of 35° and a dip of 45° has been recorded in a basalt quarry in Big Bend (Cox, pers. comm.). High dip values are also recorded from Flow IIa, and it seems that the locus of maximum dip is coincident with, or slightly to the west of, the base of the rhyolites. North of Stegi the interbedded flows have near vertical dips, though this may partly be caused by the presence of north-south strike faulting. The axis runs approximately along these interbedded flows and then seems to pass into the zone of low relief and steeply dipping rhyolites between Lusoti and the Mbuluzi, passing east of Mlaula station. These steep dips do not appear to exist north of the Mbuluzi.

The structure of the Lebombo in Zululand seems to be that of a faulted monocline; north-south faults repeat the succession in a series of fault-



PLATE V.4 Outcrop of the base of the rhyolites at Cecil Mack Pass, 3 km south of the Ngwavuma Poort. The basal rhyolite, dipping east at 30° rests on basalt. The white spot by the road is a car parked at the border-post. Sugar-cane plantations on the basaltic soil of the lowveld.

blocks, forming a 'factory-roof' structure (Bristow, 1976). At the extreme southern end of the monocline a horst-and-graben structure develops, faulting up basement on the east of the volcanics, and the monocline terminates in a series of arcuate faults (Maud, 1961; Matthews, pers. comm.). Further north, in the north of Zululand and south of Swaziland, the faulting is associated with the dyke swarm and concentrated in a narrow zone across the which large displacements have occurred. Both the dyking and large-scale faulting die out to the north, though single dykes and small faults are present throughout the area. Similarly, the north-south joint set (Diagram 4) is almost completely absent from the northern part of Swaziland and the north-south structural component seems to be generally diminishing towards the north.

V.8 Thickness of the volcanics

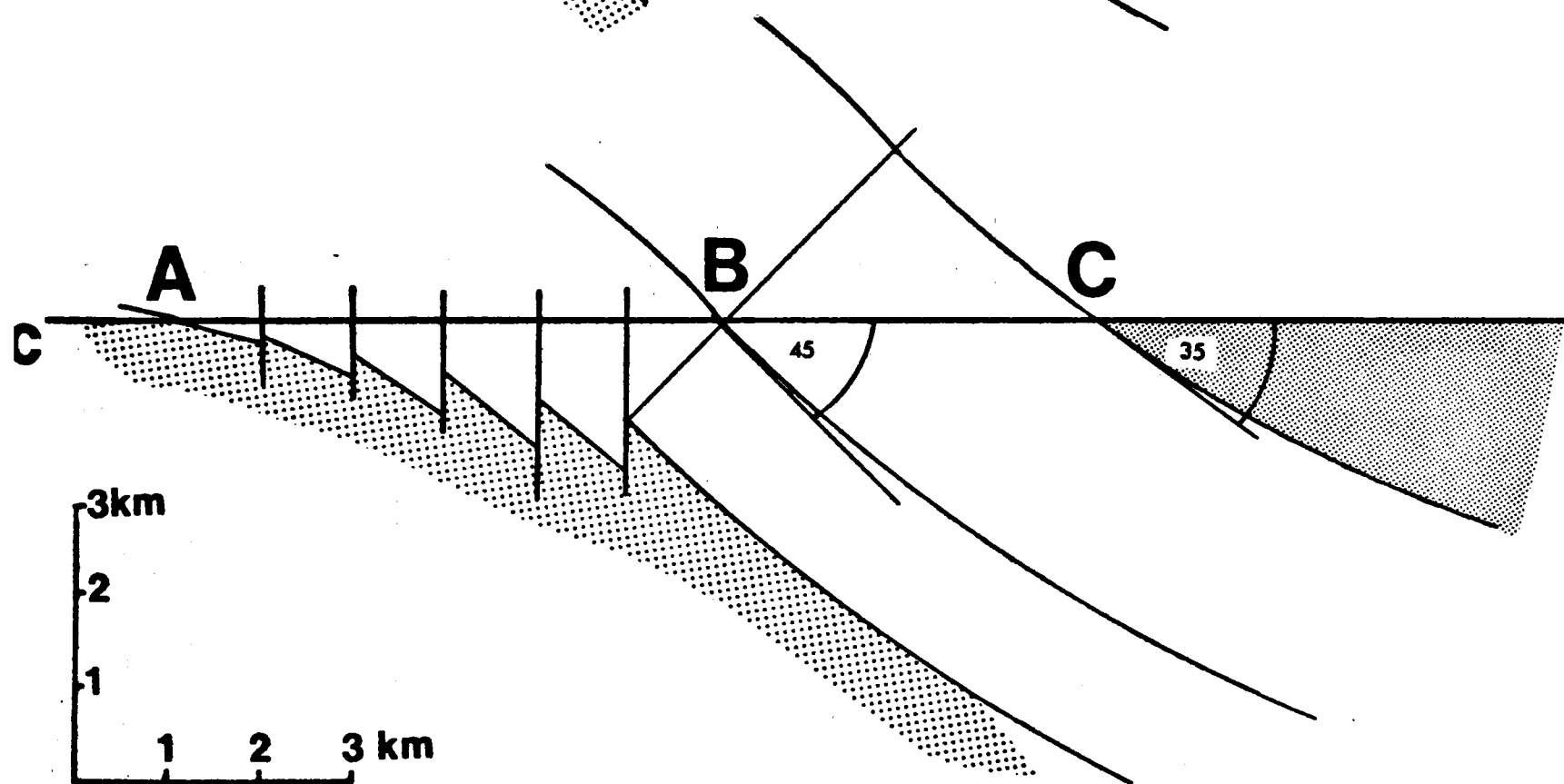
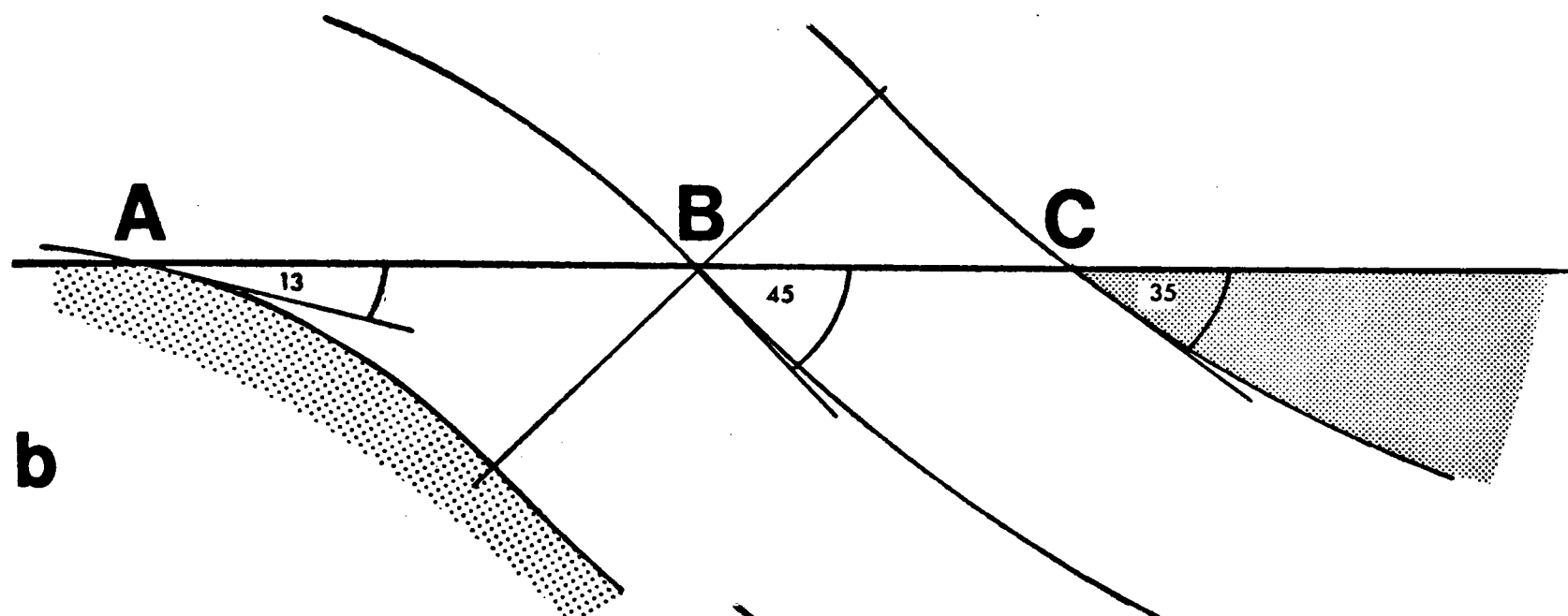
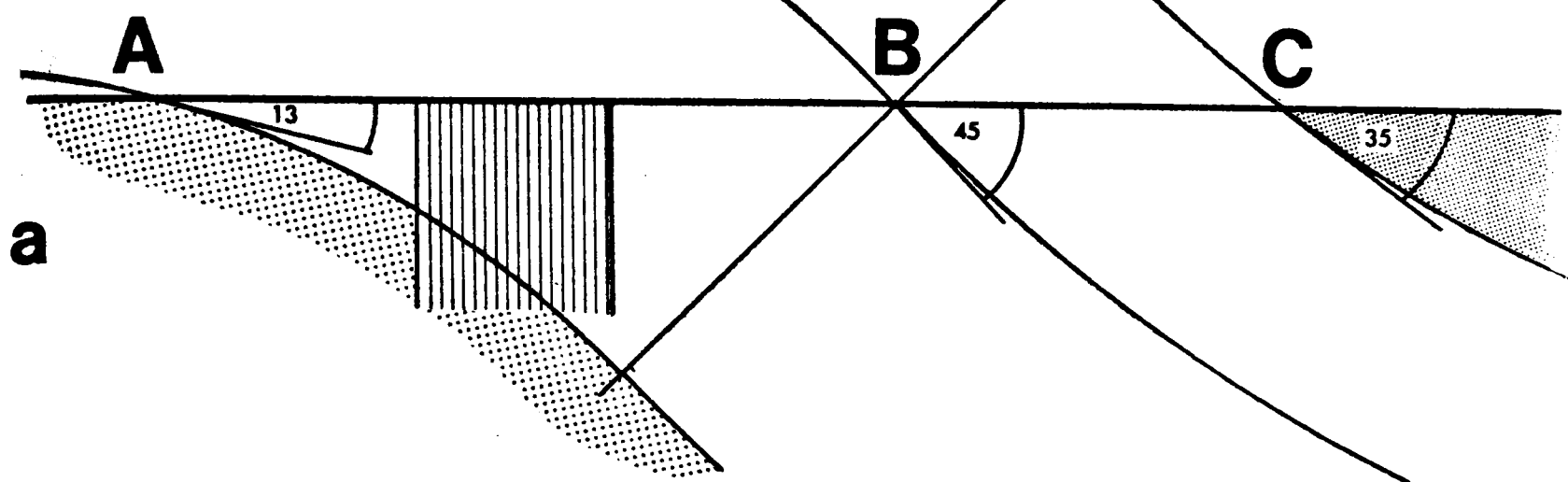
Figure V.2 shows a geometrical construction of a cross-section approximately through Big Bend. The section is based on measured dips of $12\frac{1}{2}^{\circ}$ at the base of the basalts, 45° at Big Bend and 35° in the interbedded flows east of Big Bend. If the dyking and faulting are ignored, the thickness of the basalts can be estimated at about 6.5 km. A minimum estimate of 2 km can be obtained by supposing that the shallow $12\frac{1}{2}^{\circ}$ dips extend eastwards to Big Bend and then suddenly increase to 45° . Similar models can be constructed for the basalt further north, making appropriate estimates of the angle of dip. This method produces thickness estimates proportional to the width of outcrop.

These thickness estimates, however, will have to be substantially modified if extension due to dykes, and repetition due to faulting are to be considered. Estimates can be made of the density of dykes in the swarm from exposures in road cuttings in the Rooirand. Although, in places, the dykes

Fig. V.2. Geometrical cross-sections at approximately the latitude of Big Bend. Coarse stipple = sub-basalt rocks; fine stipple = rhyolite; white = basalt.

(a) shows the apparent thickness of the basalt to be about 7 km. If the dilation due to the dyke swarm is removed (b) the thickness reduces to 6 km. (c) shows how a zone of strike faults can reduce the thickness to only 4 km.

(c) is assumed to be the closest to reality.



account for over fifty per cent of the outcrop, the total dilation west of Big Bend is probably only 25 per cent. Extension due to dykes in the northern half of the basalt outcrop is thought to be insignificant. The effect of strike faulting on the apparent thickness of the basalt is unknown and difficult to measure. Some estimates can be obtained from the extreme south of Swaziland where the Cave Sandstone/basalt contact is repeatedly faulted to the surface over a zone of 5 km, making a total vertical displacement of about 2 km. The integrated effect of the strike faulting is to reduce the effective dip and hence increase the apparent thickness of the basalts (Figure V.2). The area between Big Bend and the escarpment does not seem to be faulted on a large scale; the length and parallelism of the two or three interbedded flows implies that they are genuine intercalations rather than a single flow repeated by faulting. The estimate of 3 km for the thickness of the basalt east of Big Bend, which is calculated from measurements of dip and width of outcrop, may therefore be close to the true thickness.

The outcrop width narrows progressively towards the north. This reduction may be partly due to the relative scarcity of faults and dykes in the northern area, producing less extension and repetition of the outcrop, and partly to the overlap of the rhyolites over the upper part of the basalt succession. There may also be a reduction in the basalt thickness, or a steepening of dip; gravity data have been interpreted by Burley et al. (1970) as representing a twofold increase in the thickness of the basalt in the north, and by Darracott (1974) as not representing a significant change in thickness.

In conclusion, it seems that the thickness of the basalt is approximately 5 km at the latitude of Big Bend; there is no reason to suppose that this thickness cannot persist over the rest of the lowveld.

The rhyolites have a much simpler structure; the better exposure has enabled the tracing out of contacts and demonstrated the lack of strike faulting and the rarity of dykes. There remains the problem of the overlap of the rhyolites over the basalts, making the apparent thickness at Stegi much greater than that at, for example, Tikuba (lat. $26^{\circ}38'S$) or along the Usutu. Reasonably accurate estimates of the rhyolite thickness can be obtained away from the major bulges where the width of outcrop in Swaziland is about 11 km, with a further 4 km to the top of the rhyolites in Mozambique (Wachendorf, 1970). An assumption of a minimum average dip of 20° gives a minimum thickness of about 4 km for the rhyolites in Swaziland, or 5 km if those in Mozambique are included. Only in the extreme east of the outcrop are dips of 20° recorded however; dips in the western part reach values as high as 60° and 30° may be a more reasonable average dip, making a total thickness for the rhyolites of 7.5 km.

The thickness estimates for both the rhyolites and basalts are dependent on the assumption that the flows have infinite eastwards extent. Any wedging out or lenticularity of the flows will result in increased apparent thicknesses relative to the true thickness.

V.9 Geophysical Data

Because of the uncertainty and non-uniformity of the extension due to dyke intrusion, and repetition of strata by strike-faulting, any attempt to calculate the thickness of the lavas from dip and outcrop measurements is inevitably subject to uncertainty. Of the geophysical methods available for solution of this problem, only gravity and, to a lesser extent, aeromagnetic data have been used.

The gravity data of Burley et al. (1970), Hales and Gough (1962), and Smit

(1962) show that there is a large positive isostatic gravity anomaly corresponding with the surface exposure of the volcanics along the entire length of the Lebombo. In Swaziland, the gravity anomalies are linear along the length of the basalt outcrop, the maximum occurring within the outcrop, with the exception of an embayment to the north of Stegi (Figure V.3) (Burley et al., 1970). This embayment persists eastwards to south of Lourenço Marques (Maputo).

Hales and Gough (1962) made the first quantitative interpretation of the gravity anomaly. Assuming the rhyolites to have a density contrast with the basement of 0, the basalts + 0.3 g/cm³, and the Cretaceous cover -0.4 g/cm³, they concluded that the most realistic geological cross-section consisted of some 8000 m of basalt and 5000 m of rhyolite overlain by 2500 m of Cretaceous cover, producing a total depression of the sub-Karoo surface of at least 13,000 m. The lavas were considered to flatten out and extend infinitely to the east. No anomaly was associated with the possible development of the upper basalt.

Burley et al. (1970), working with more detailed data from the Swaziland gravity survey, proposed a model similar to that of Hales and Gough (1962). They concluded that the basalts thicken from 4000 m in the south to 8000 m in the north of Swaziland, despite the reduction in outcrop width, and estimated the depression of the sub-Karoo strata to be between 12,000 and 16,000 m.

Darracott and Kleywegt (1974) proposed that only part of the observed isostatic anomaly was due to the lower basalts, the remainder being a 'regional' or 'crustal' effect. After making allowances for this regional effect, the anomalies are much smaller; they calculated, for northern Natal, a vertical section of 2000 m of post-Karoo sediment, 2000 m of rhyolite and 1500 m of

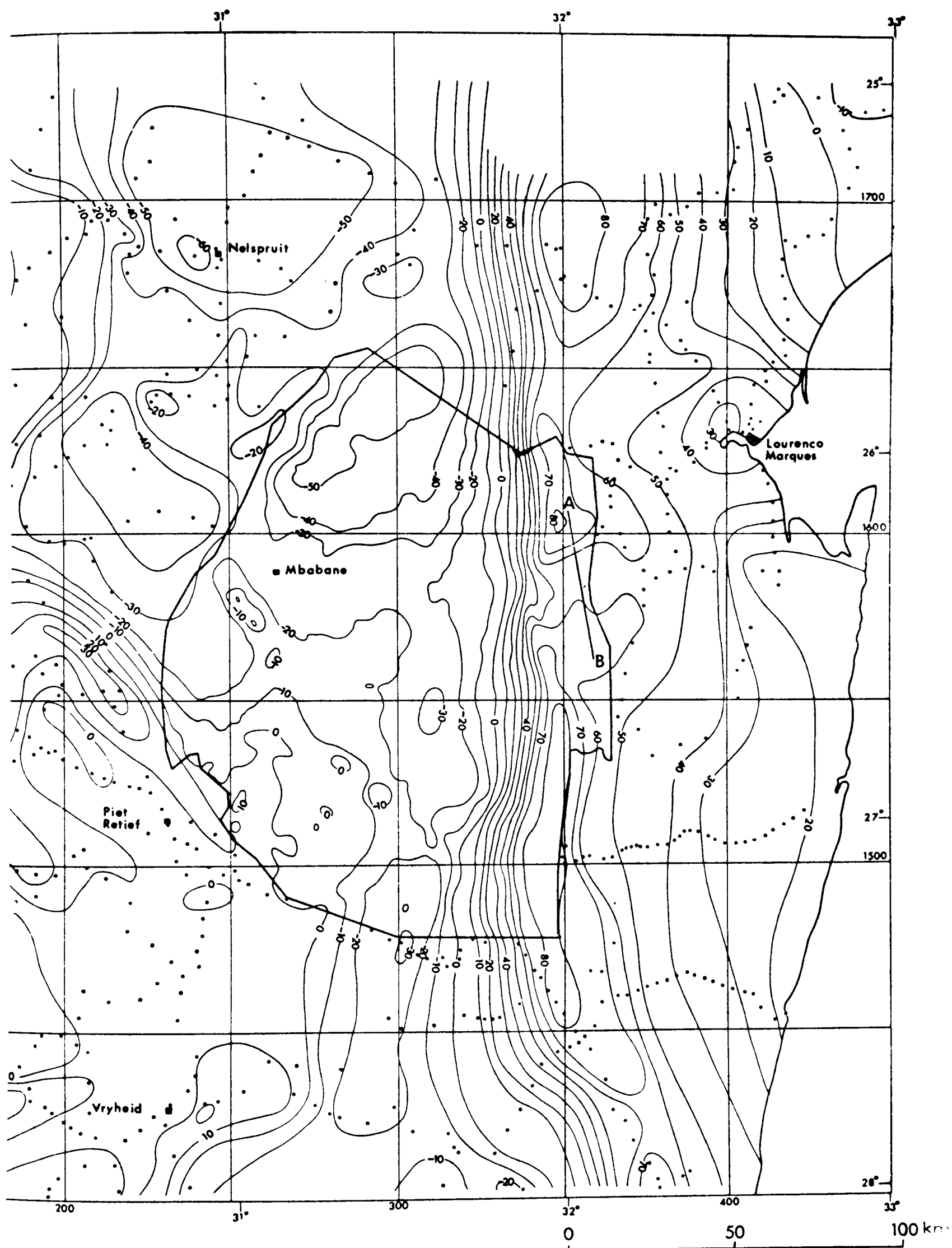


Fig. V.3. Isostatic anomaly map of Swaziland and surrounding areas (from Burley et al., 1970). A-B gives line of section for Fig. V.4.

basalt; and for northern Swaziland, 800 m of Tertiary and Cretaceous sediment, 2500 m of rhyolite and up to about 900 m of lower basalt. The thickness of the lower basalt at the flexure was shown to be significantly greater than that under the coastal plains.

The northernmost of the sections calculated by Darracott and Kleywegt (1974) that passing east-west through Namaacha, shows considerable overlap of the rhyolites over the basalts. The gravity anomaly is greater than that expected if the surface outcrop of the basalt is representative of its thickness.

One of the major problems concerning the outcrop of the rhyolites is the nature of the bulges in the base of the rhyolites, whether they are caused by an erosional or topographic unconformity, downwarping of the volcanics due to superincumbent loading, or overlap of the rhyolites over the upper part of the basalts. The gravity anomalies over these bulges show a departure from two-dimensionality with a minimum and maxima approximately corresponding, respectively, to the antiform and synforms of the rhyolite outcrop and a trough of gravity anomalies extending east-west approximately along latitude $26^{\circ}30'S$. A sketched gravitational cross-section across this trough shows a Bouguer anomaly relief of some 25 milligals (Fig. V.4). These data can be interpreted as being representative of a depression of the basalt along the gravity anomaly trough though no attempt is made to model this quantitatively in the present study. If the displacement of the rhyolite outcrop, relative to the

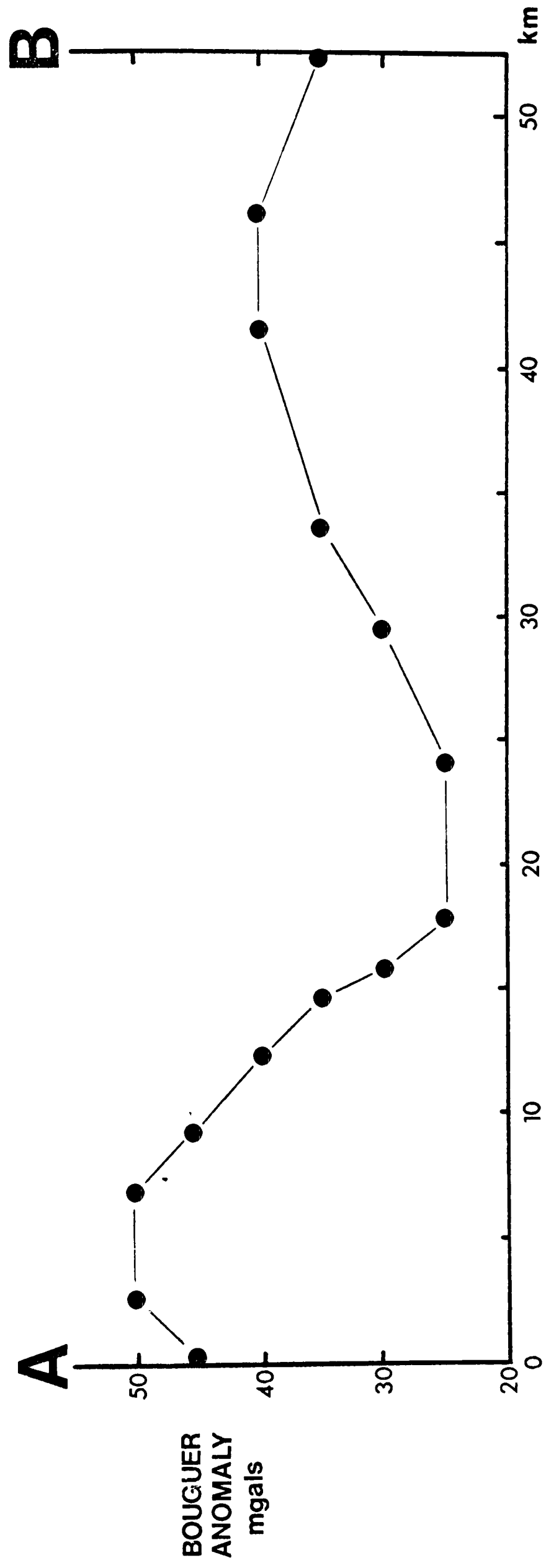


Fig. V.4. Bouguer anomaly section across the gravity embayment in the Lebombo along the line marked on Fig. V.3.

horizon defined by the interbedded flows, is approximately 8 km; using an average-to-low value of 25° for the dip, this represents a vertical displacement of some 3 km. This seems to be larger than expected for an erosion surface of this nature and it is suggested that some of the observed warping is attributable to depression of the top surface of the basalt. The initial sub-rhyolite topography was blanketed by the ignimbrite flows and eventually smoothed out; the effect of the initial topography being emphasised by overlap of the lower rhyolite flows over the upper flows of the basalt.

The aeromagnetic data (Bacon, 1970) show a complex pattern of anomalies over the basalts, attributed to a series of flows with varying magnetic properties. Larger anomalies are associated with the granophyres, interpreted as being due to an abundance of magnetite in the granophyres (Bacon, *op. cit.*). It is found, in the field, that the granophyre contacts are often highly magnetic, sufficient to affect a compass held ten metres away. The Nyetane granophyre has a relatively large magnetic anomaly and may be a much larger body than suggested by the surface outcrop. Magnetic profiles for the granophyres are generally compatible with dyke-like forms for the intrusions, dipping steeply eastward (Bacon, 1970).

Seismic refraction data (Ludwig et al., 1968), from off the southeast African continental margin, show a layer of velocity 6.49 km/s at a depth of 4 to 10 km which may be correlated with the Lebombo basalts on the grounds of velocity (Burley et al., 1970). The layer appears to terminate 130 km east of Durban and thus can be considered to support the assumption of infinite extension in an easterly direction.

V.10 Conclusions

Du Toit (1930) first recognised that the Lebombo Monocline is a region of tension. The faulting, dyking, flexuring, and most of the jointing are compatible with an east-west tensional regime. Several of the joint sets can be attributed to pre- or post- Karroo tectonic features. The structure of the Lebombo changes progressively from a graben structure in the south, through a zone of fault-blocks in Zululand and repeating strike faults in the south of Swaziland, to a more flexural monocline in the north of Swaziland and further north in South Africa.

The Lebombo runs along the eastern edge of the Kaapvaal craton (Figure I.1) but does not seem to correspond to any Kaapvaal lineament. It is, however, parallel to the postulated southerly extension of the Mozambique belt (Cox, 1970) and may be constrained in its northerly trend by the tectonic grain of the metamorphic belt.

CHAPTER VI

PETROGRAPHY OF THE RHYOLITES

VI PETROGRAPHY OF THE RHYOLITES

VI.1 Introduction

The rhyolites of the Lebombo are all porphyritic, containing phenocrysts of feldspar, clinopyroxene, quartz, ore, zircon, apatite and occasionally fayalite. The groundmass is fine-grained and micro-granophyric, spherulitic or glassy. Minor differences in the relative amounts of different phenocrysts, the presence or absence of glomeroporphyritic aggregates and, especially, the amount of quartz phenocrysts serve to distinguish between the various flows. As described in Chapter V, the flows have been divided into three major groups on the basis of hand-specimen characteristics; thin-section study supports this classification but there is generally too much intra-flow variation for adjacent flows to be separated purely on the basis of thin-section petrography.

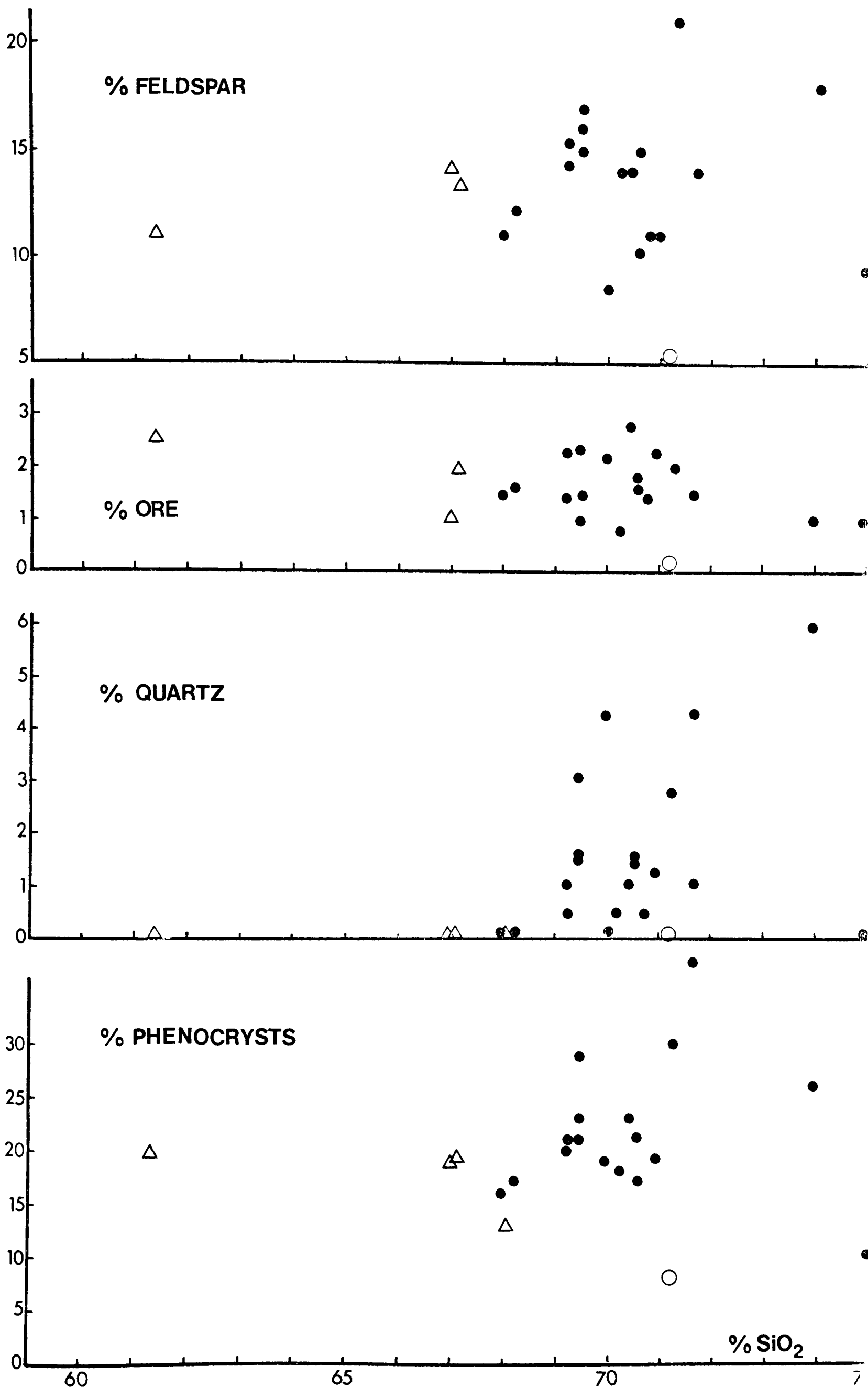
Over half of the 450 samples collected, including all of the samples analysed, have been studied in thin section. The following discussion is not exhaustive but illustrates some of the more important points.

VI.2 Phenocryst phases

The modal abundances of the phenocrysts are shown in Figure VI.1. They are plotted against silica as a convenient index of fractionation. Apart from the abundance of quartz, there is little correlation between phenocryst content and chemistry.

The individual phenocrysts are discussed below. The different types of rhyolite are discussed together as there are no significant textural differences between the various types.

Fig. VI.1. Modal abundances for Swaziland rhyolites as a function of silica content.



(a) FELDSPAR

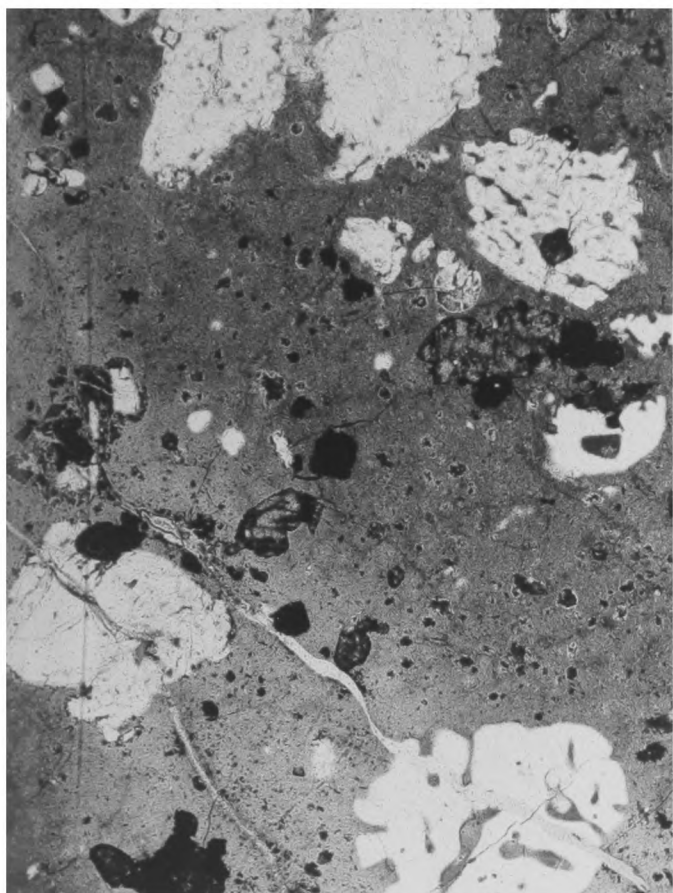
Plagioclase is ubiquitous in these rocks forming crystals of 1-5 mm in size. It is frequently altered, becoming cloudy with clay minerals; some samples developing epidote. The crystals are generally euhedral but are commonly broken and resorbed, showing marginal corrosion and sieve textures. The resorption textures are often more pronounced where the crystals are fractured, and crystals in many differing stages of corrosion can be seen in a single thin section (Plate VI.1).

The phenocrysts are oligoclase-andesine in composition; zoning is usually observed but not extreme. Sanidine overgrowths are common in the more evolved samples, forming discrete rims on the plagioclase (Plate VI.2).

The phenocrysts are often present as glomeroporphyritic clusters; some forming clots 2 cm across (Plate VI.3). Some of these clusters have a central crystal which is usually more sodic and may be xenocrystic in origin (Plate VI.4, 5). The origin of these is unknown. Trapped liquid within these aggregates contains curved quench crystals and has suffered less devitrification than the rest of the groundmass (Plate VI.6).

Some of the rocks contain plagioclases with abundant glass inclusions (Plates VI.7, 8, 9). These are isotropic and contain quench crystals or daughter crystals. They do not seem to be arranged in zones, are often rectangular and do not appear to be resorptional features; rather they seem to be primary inclusions related to the crystallisation of the feldspar. The nature of these inclusions is not clear.

Sanidine forms rims and overgrowths on the plagioclase phenocrysts. Occasionally discrete sanidine crystals are noticed in thin-section; some of these may be sections through the edge of rimmed plagioclases.



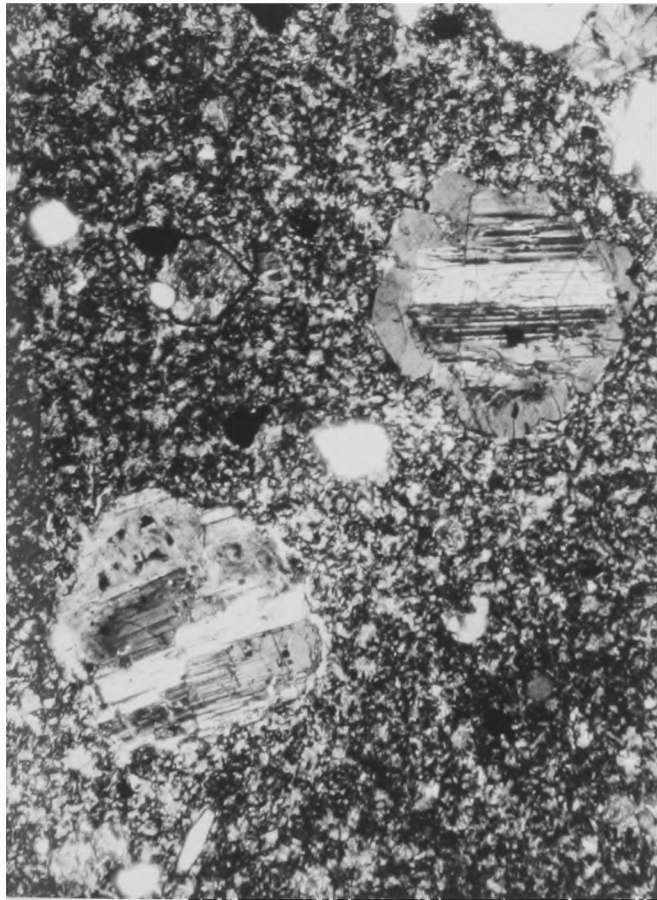
1 mm

PLATE VI.1 Quartz (bottom right) and feldspar exhibiting various stages of marginal corrosion and sieve textures. Black ore and semi-opaque pyroxene in glassy matrix (SK 153).



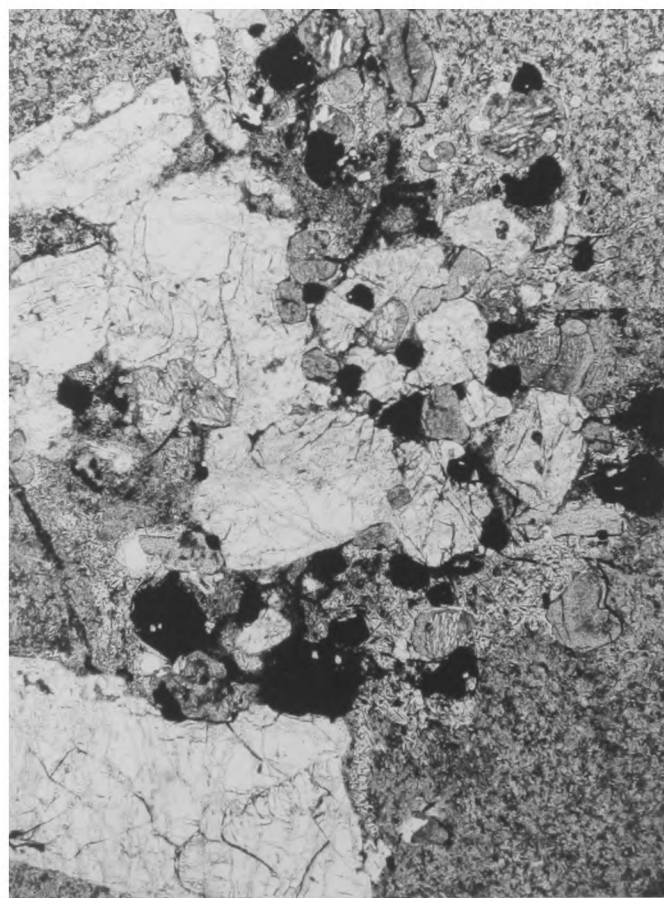
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PLATE VI.1a Broken feldspar and quartz phenocrysts in eutaxitically banded matrix. Upper of two feldspar phenocrysts shows faint perthitic exsolution (MZ 35).



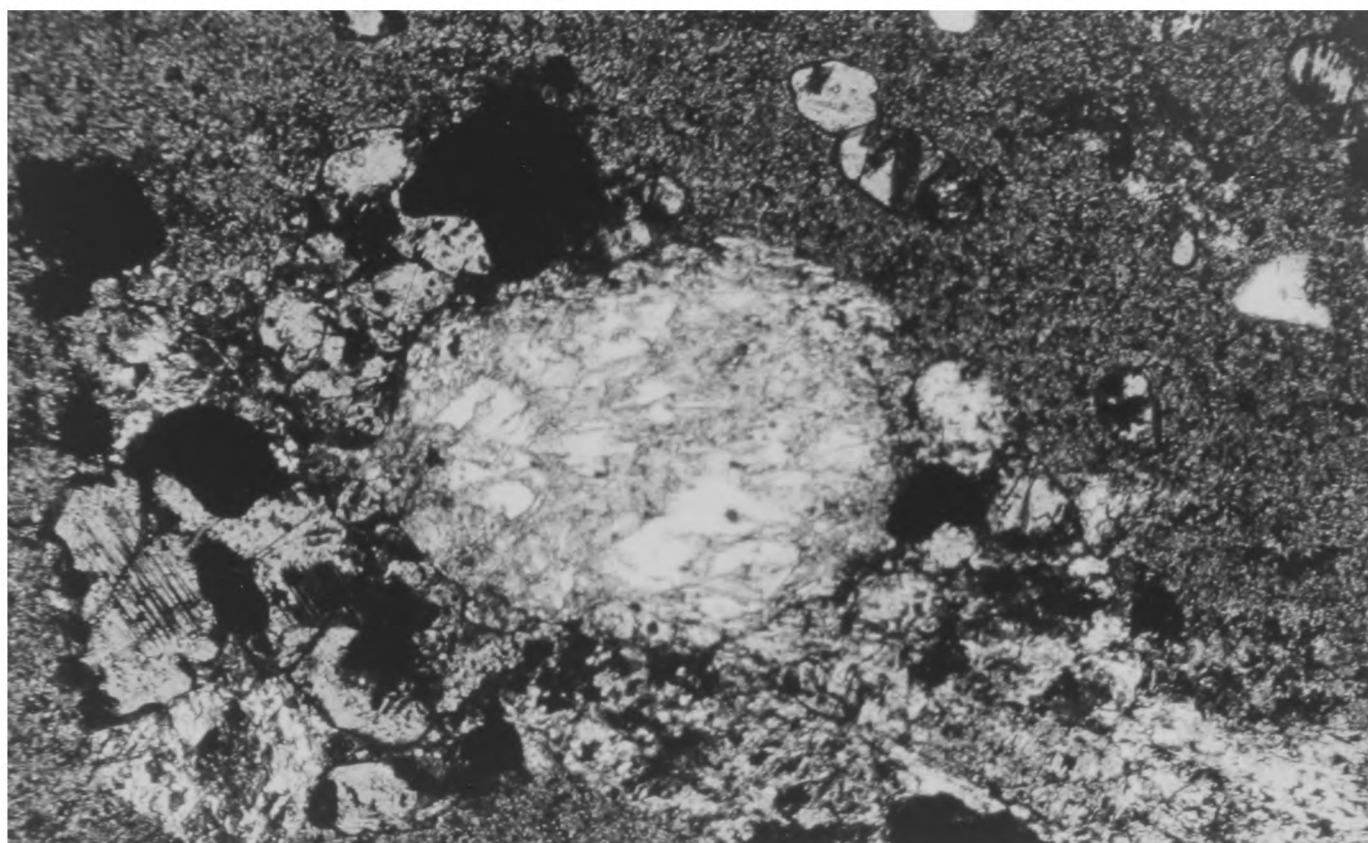
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PLATE VI.2 Twinned plagioclase with sanidine rims. With quartz, ore and pyroxene in fine felsitic groundmass (RC 4).



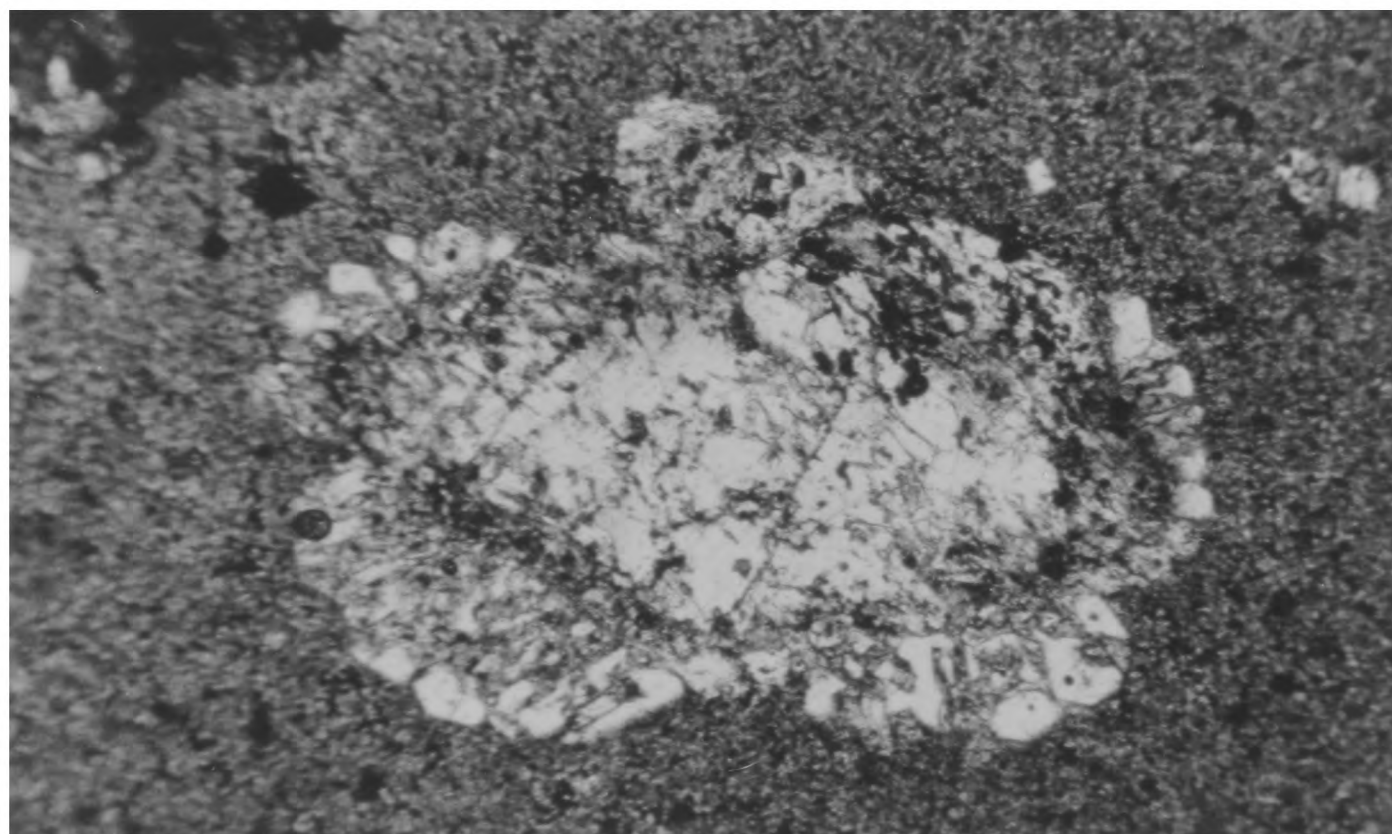
1 mm

PLATE VI.3 Large glomeroporphyritic aggregate formed of plagioclase (pale), pyroxene (grey) and ore. Small euhedral crystals are apatite. The groundmass within the aggregate is less devitrified than that outside and curved quench crystals can be seen (RC 36).



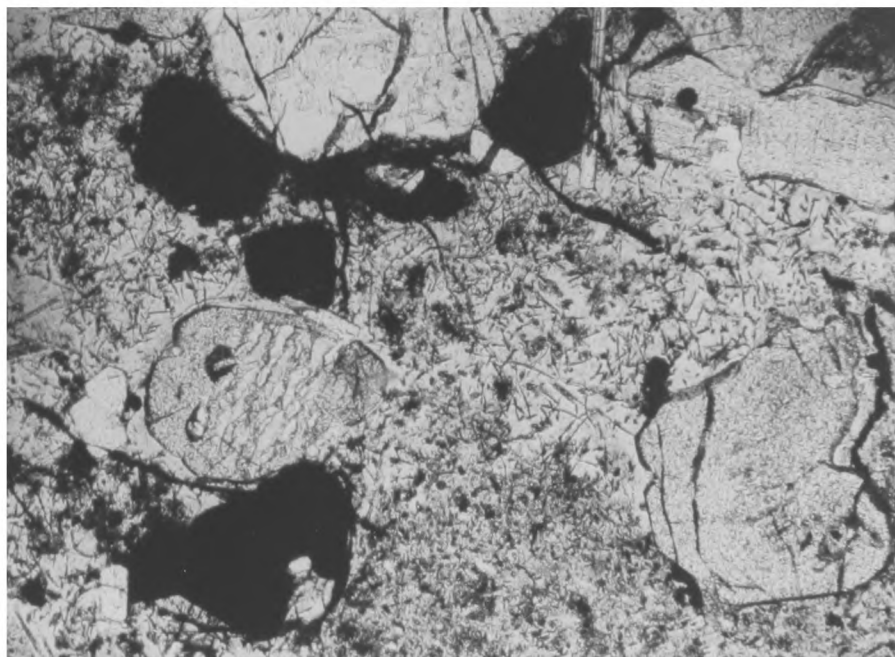
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PLATE VI.4 Labradorite xenocryst in aggregate with plagioclase, exsolved pyroxene and ore. (Photograph: P. Betton)



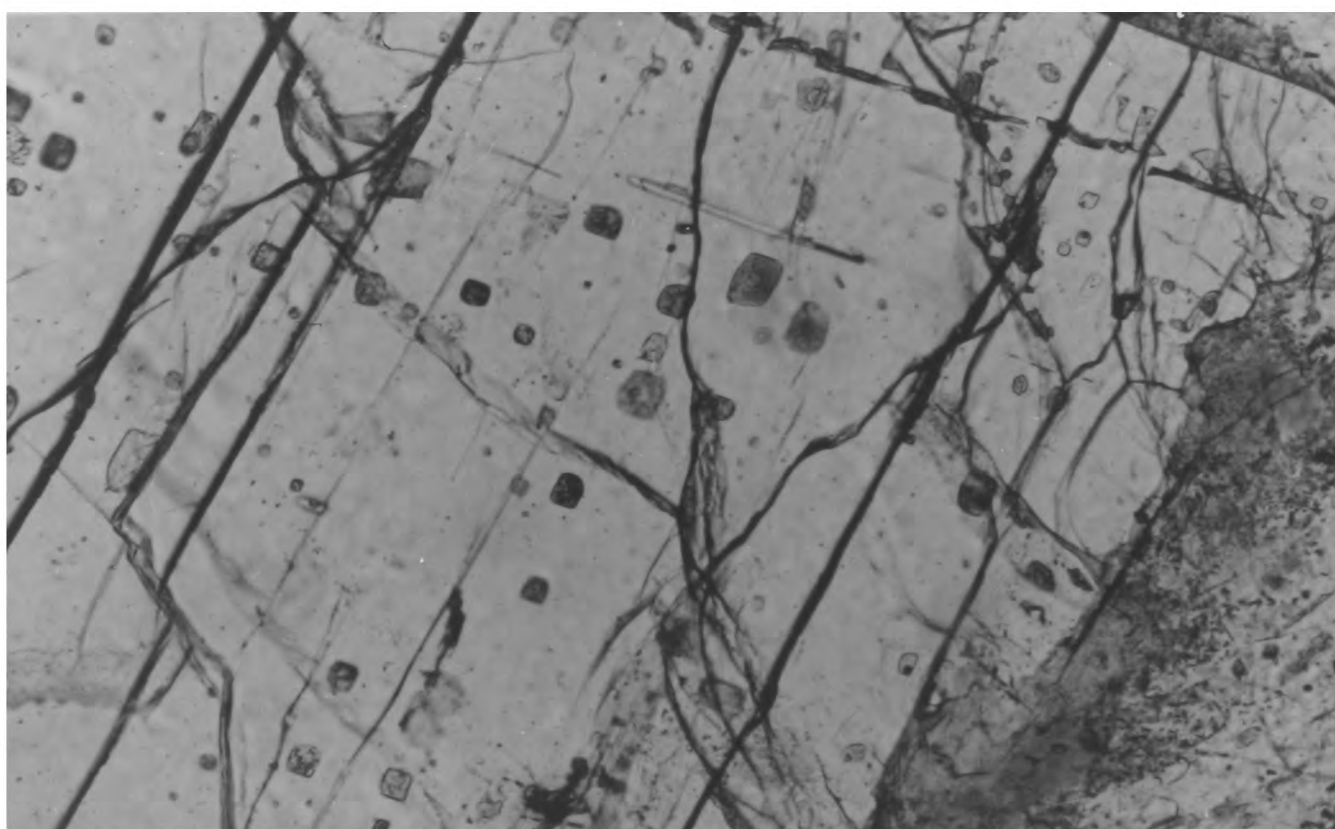
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PLATE VI.5 Albitic core with plagioclase overgrowth. (Photograph: P. Betton)



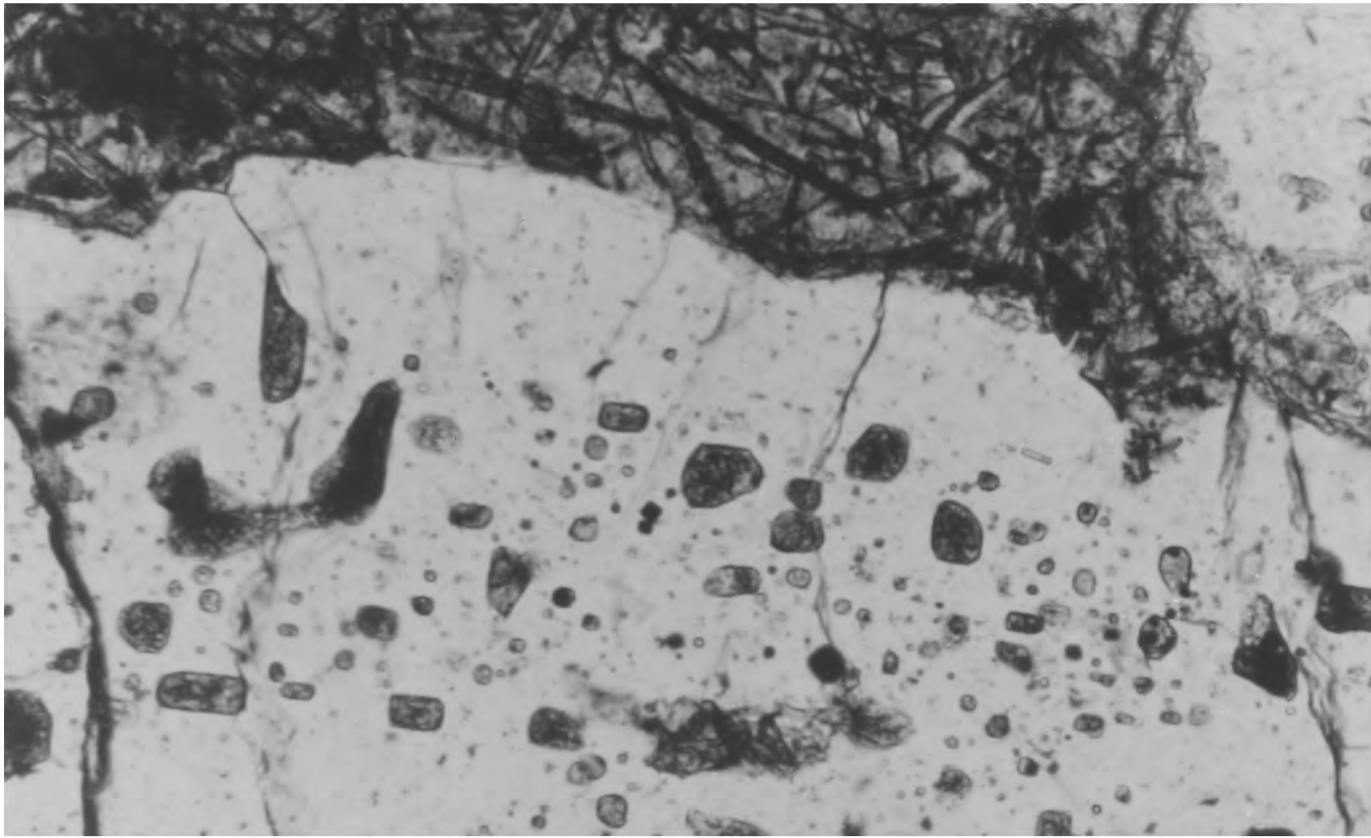
1 mm

PLATE VI.6 Close-up of Plate VI.3, showing detail of curved crystals in groundmass. Pyroxene, ore and apatite crystals (RC 36).



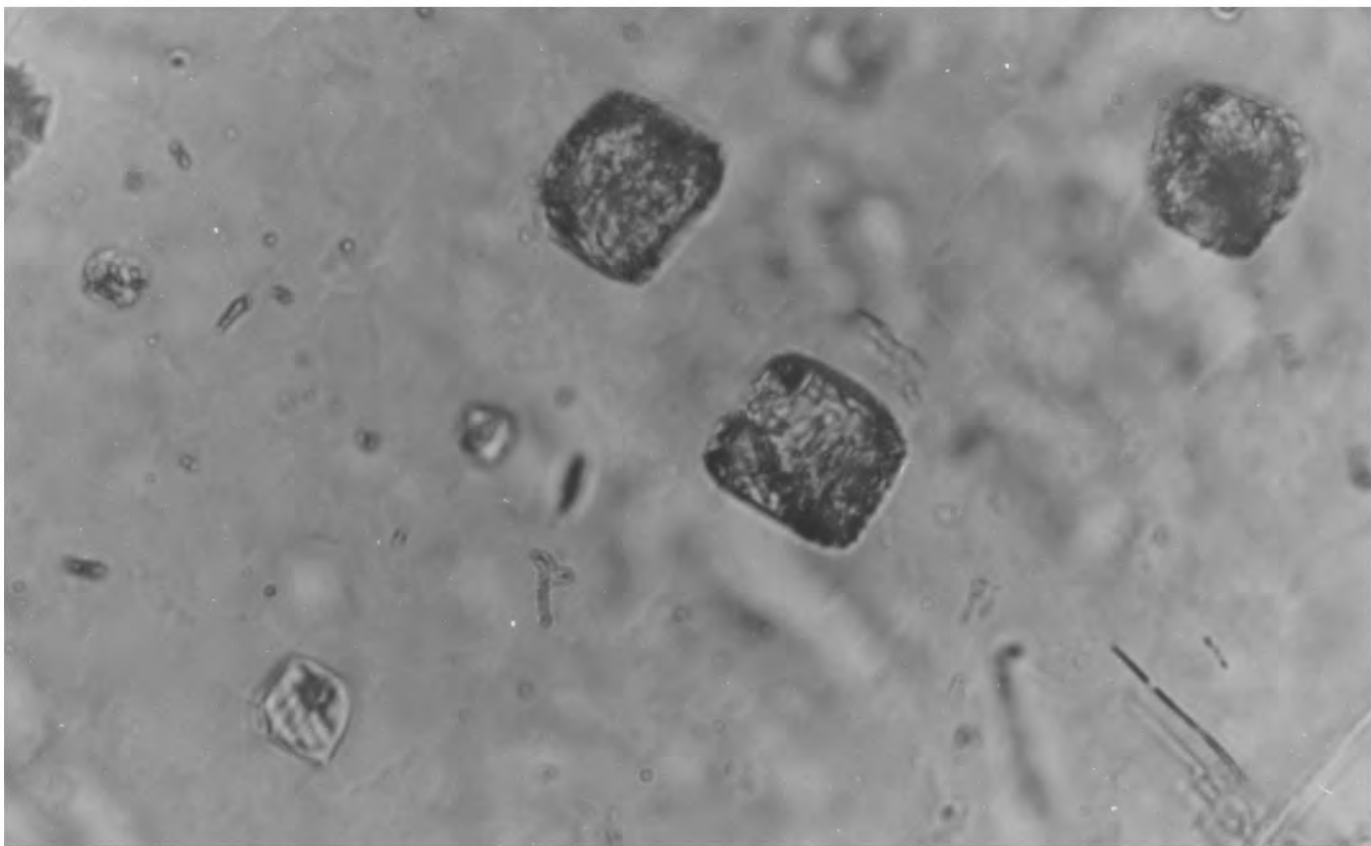
0.5 mm

PLATE VI.7 Plagioclase crystal with glass and fluid inclusions. (Photograph: P. Betton; RC 144)



0.5 mm

PLATE VI.8 Glass inclusions in plagioclase; two at left beginning to coalesce. (Photograph: P. Betton; SK 138)



100 μ m

PLATE VI.9 Close-up of glass and fluid inclusions in plagioclase. (Photograph: P. Betton)

Potassium feldspar is presumably present as a groundmass phase in intergrowth with quartz but positive identification has not been made.

Several of the acid rhyolites contain antiperthite phenocrysts (Plate IV.35). The perthitic intergrowth is coarse and invariably has a wide sanidine rim. Perthites as phenocryst phases are rare in volcanic rocks; they are generally associated with slowly cooled plutonic rocks. Assuming that the antiperthite is cognate and not accidental, it may be that the large size of these ignimbrite flows has caused high temperatures to be reached and maintained for periods long enough to grow antiperthites.

(b) CLINOPYROXENE

Clinopyroxene is nearly universally present; a few rocks do not contain pyroxene, these are generally rather aphyric. The pyroxene present is a pale-green hedenbergitic or ferroaugitic one, non- or only slightly, pleochroic, and 0.1 to 1 mm in diameter. Twinning is rare but it is commonly exsolved, developing thin parallel opaque lamellae (Plate VI.5). It is sometimes corroded but generally forms anhedral crystals with sharp edges. Weathering and hydrothermal alteration has resulted in the production of amphibolised rims. The pyroxene is often found in glomeroporphyritic clusters with plagioclase, sometimes included round the edge of larger crystals (Plate VI.10). It also forms aggregates with magnetite, zircon and apatite (Plate VI.11).

(c) OLIVINE

Pseudomorphs after olivine (probably fayalite) are found in some of the more basic rhyolites and also in some of the most evolved. They form rounded 1 mm patches, composed of silica and skeletal magnetite, often surrounded by radiating cracks probably caused by the volume change on pseudomorphing (Plate VI.12).

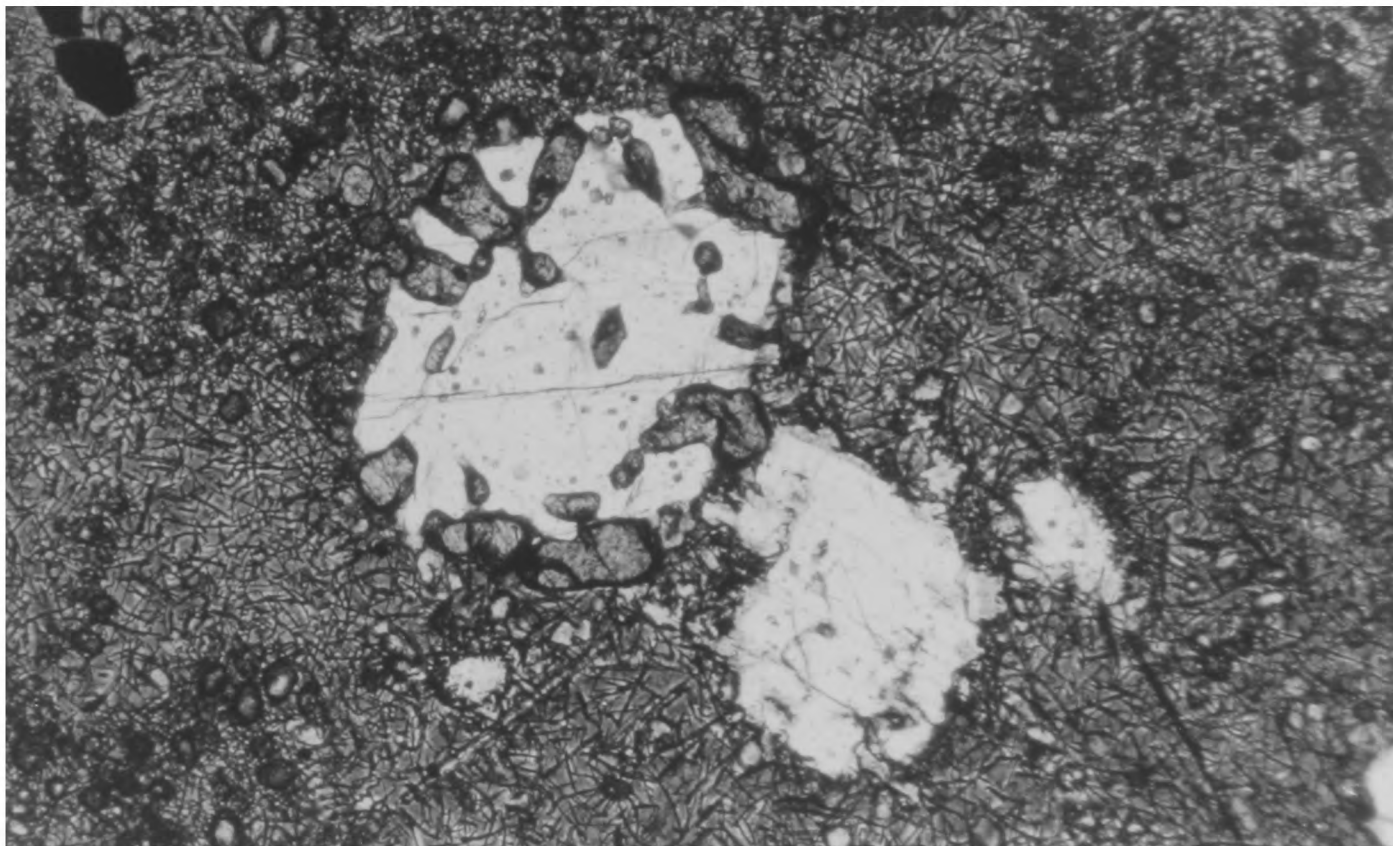


PLATE VI.10 Plagioclase crystal with inclusions of pyroxene round the edge. Curving crystallites and incipient spherulites in the groundmass. (SK 138)

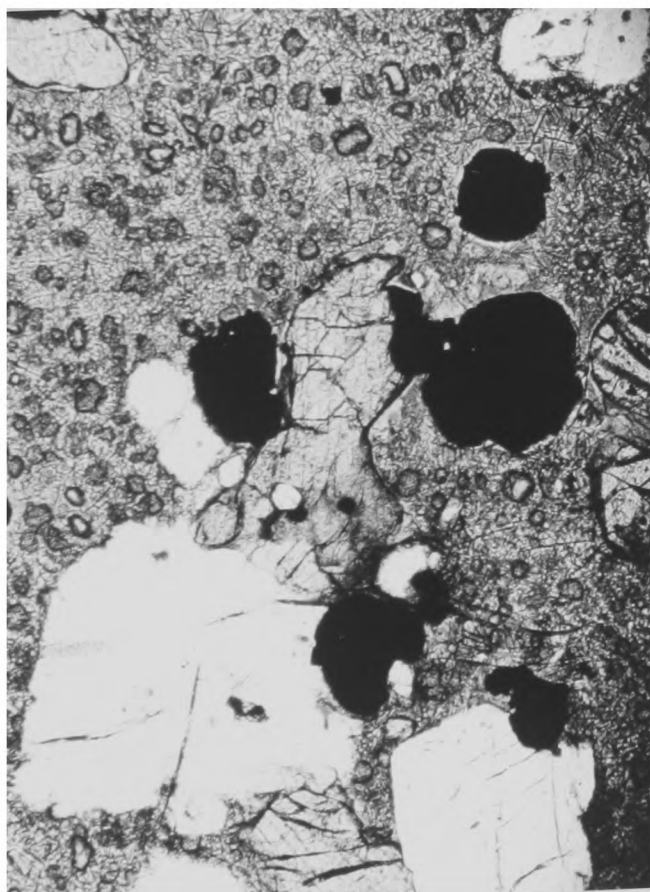


PLATE VI.11 Pyroxene, ore, apatite and plagioclase in a glomeroporphyritic aggregate. Round structures in the groundmass are small spherulites consisting of very fine radiating intergrowths of quartz and feldspar representing incipient devitrification. (SK 138)



PLATE VI.12 Two large rounded pseudomorphs after olivine, consisting of magnetite and quartz. Associated with plagioclase and pyroxene. (RC 66)

(d) QUARTZ

Quartz is a common phenocryst in all but the basic rhyolites and its abundance has been used as the chief classificatory criterion. It is present as euhedral, rounded and resorbed crystals and is a principal constituent of the groundmass (Plate VI.13). Some of the groundmasses have long blade-like quartz crystals which may be paramorphs after tridymite. Quartz crystals are often broken; some rounded and resorbed phenocrysts have been shattered into many angular pieces (Plate VI.14).

(e) ORE

Titanomagnetite is a ubiquitous phenocryst phase and forms smaller crystals in the groundmass. It forms rounded anhedral crystals, 0.1 to 1 mm in size often in clusters with pyroxene, apatite and zircon. It contains inclusion of apatite and zircon (Plate VI.12).

(f) ZIRCON

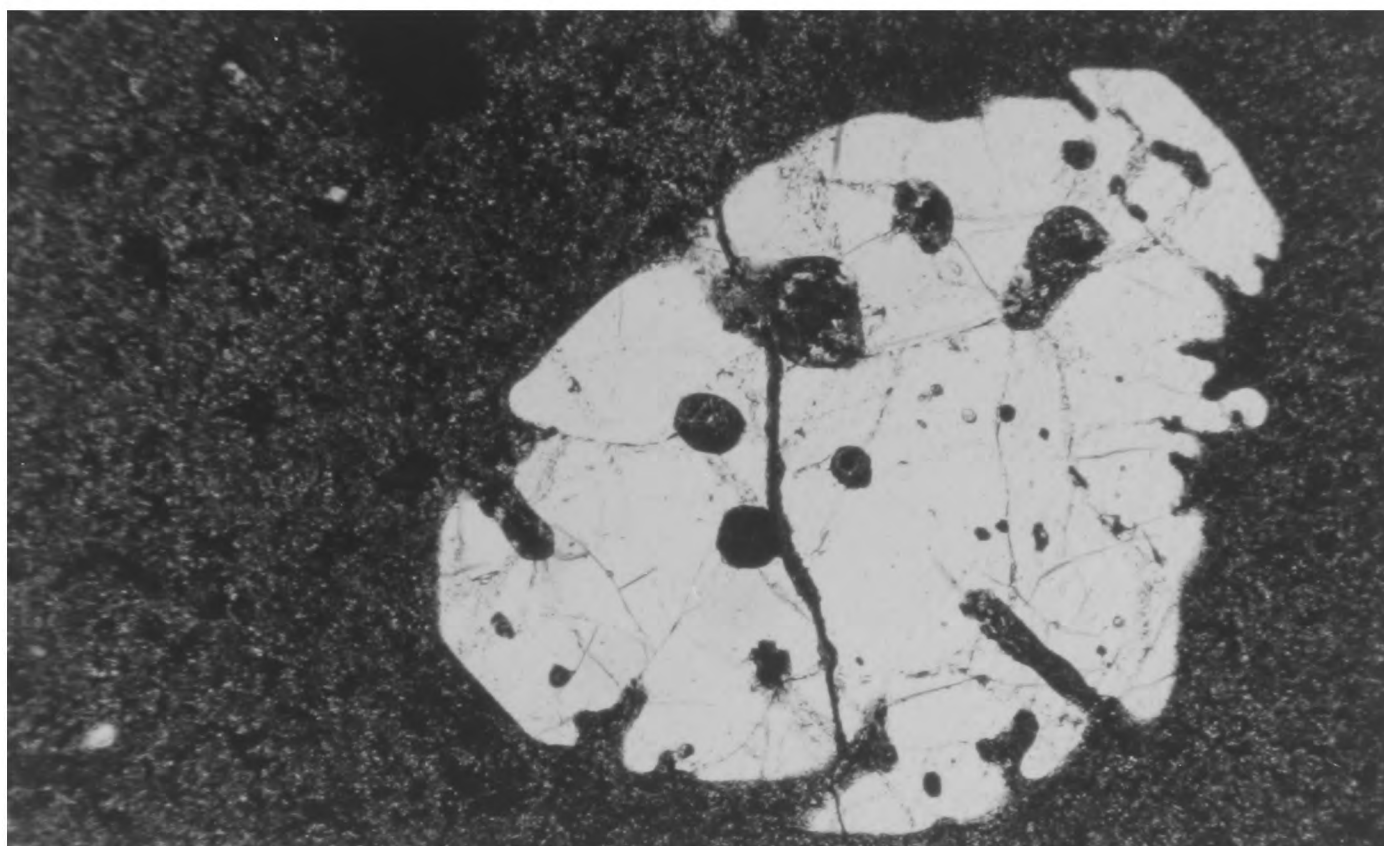
Zircon is a common phenocryst phase in the rocks with greater than 68 per cent SiO_2 . It forms small euhedral crystals often included in or stuck to ore crystals.

(g) APATITE

Apatite is an ubiquitous phase, present in virtually all the slides examined. It is generally associated with ore and pyroxene crystals and is present as needles in the groundmass.

VI.3 Glomeroporphyritic aggregates

Glomeroporphyritic aggregates are common in these rhyolites. They are generally composed of plagioclase grains; some also contain pyroxene and ore, others contain just pyroxene and ore. Apatite and zircon phenocrysts



1 mm

PLATE VI.13 Large quartz crystal showing marginal and central corrosion. (Photograph: P. Betton)



1 mm

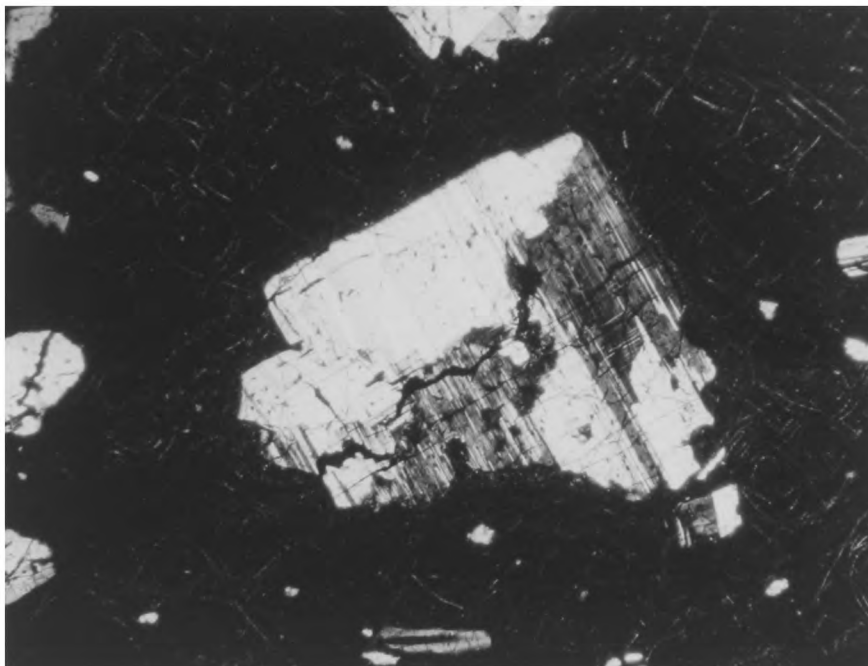
PLATE VI.14 Shattered quartz crystals in an eutaxitic matrix. (MZ 35)

are predominantly associated with ore, and to a lesser extent pyroxene, in these aggregates (plates VI.3, 11).

Crystal aggregates are not present in all the flows. Often they are characteristic of certain flows; adjacent flows can be distinguished by the relative abundances and natures of the composite phenocrysts. The more basic rhyolites tend to be more glomeroporphyritic; those containing quartz tend to be less so.

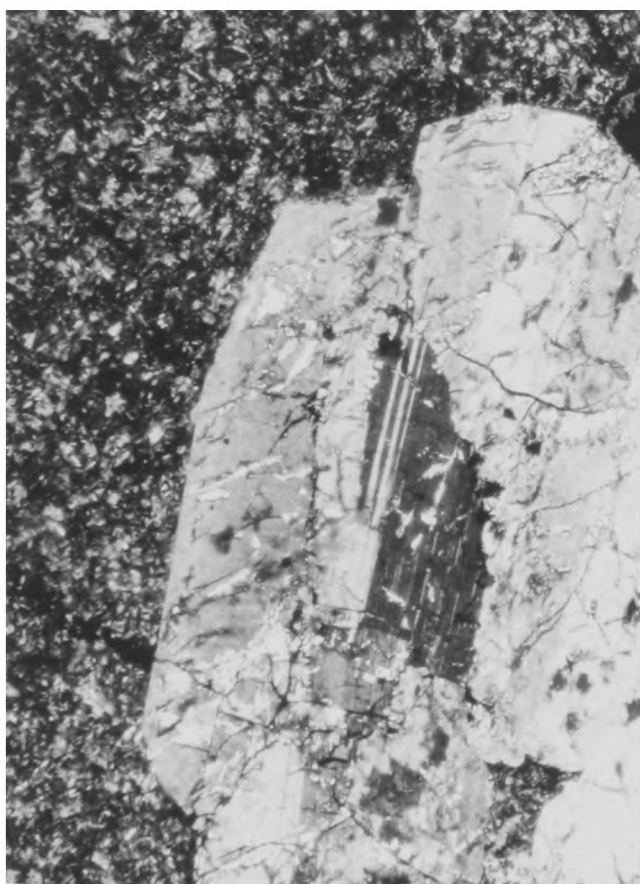
Some of the larger aggregates may be pieces of cognate cumulitic material; the similarity of phenocrysts within them to those separate rules out an accidental origin for them. Many of the aggregates show characteristics of synneusis, which is a word coined by Vogt (1921) to mean the mutual drifting together and attachment of crystals in a magma. Similar crystals often join in parallel orientation by the attachment of the broader crystal faces, giving rise to a 'synneusis twin' (Vance, 1969); the whole might develop crystal faces and appear like a crystallographic twin were it not for the irregular nature of the 'twin plane' (Plates VI.15, 16). Synneusis also affects dissimilar crystals; they will form aggregates but with a random orientation of the component crystals. The characteristic features of synneusis, the mutual parallelism of the crystals and freely developed overgrowths, is indicative of a free-floating history; aggregates representing part of a cumulate would not necessarily show parallelism or crystal faces after adcumulus growth.

Synneusis crystals are common in the Swaziland rocks. The diagnostic characteristics of synneusis are demonstrated clearly in plagioclase aggregates because of the ease of recognising crystallographic parallelism and growth lines due to the multiple twinning and zoning respectively. Larger aggregates are often composites of synneusis crystals; plagioclase



1 mm

PLATE VI.15 Well formed plagioclase synneusis twin. The two halves of the crystal are in parallel orientation but are not in optical continuity; the irregular nature of the 'twin plane' is indicative of an origin by synneusis. Note irregular broken fragments of plagioclase. (RC 144)



1 mm

PLATE VI.16 Synneusis twin of plagioclase. Three or maybe four crystals are joined in near parallel orientation; the whole developing moderately well formed crystal faces. (RC 36)

aggregates generally contain pyroxene but quartz is very rarely seen in synneusis with crystals other than quartz.

It is possible that in flows where the phenocrysts have become resorbed that these synneusis aggregates will begin to become disaggregated; the original plane of attachment being a line of weakness. The result will be a number of crystals with irregular faces, very similar to broken crystals. An abundance of fractured crystals, therefore, may not be indicative of a violent mode of eruption, but may be the result of partially resorbed synneusis clusters.

Synneusis must be largely restricted to the earlier stages of crystallisation as most clusters show extensive overgrowth (Vance, 1969). The formation of aggregates will be presumably favoured by low viscosity, a high phenocryst-liquid ratio and movement or turbulence of the magma. Once formed, the large size of the aggregate will cause it to sink more rapidly than the smaller discrete crystals and hence produce more effective crystal fractionation. The fractionation of minor phases such as apatite and zircon will only be appreciable when they are attached to a larger crystal or crystal aggregate.

VI.4 Devitrification and groundmass textures

Devitrification is the change from a metastable glassy state to a crystalline one. Glass is rare in rocks older than Cretaceous (Lofgren, 1971) and is rare in the Lebombo rhyolites.

The products of devitrification of a rhyolite glass are predominantly fine-grained silica (tridymite or cristobalite, usually inverted to quartz) and alkali feldspar which form fibres either in bundles or radiating spheres

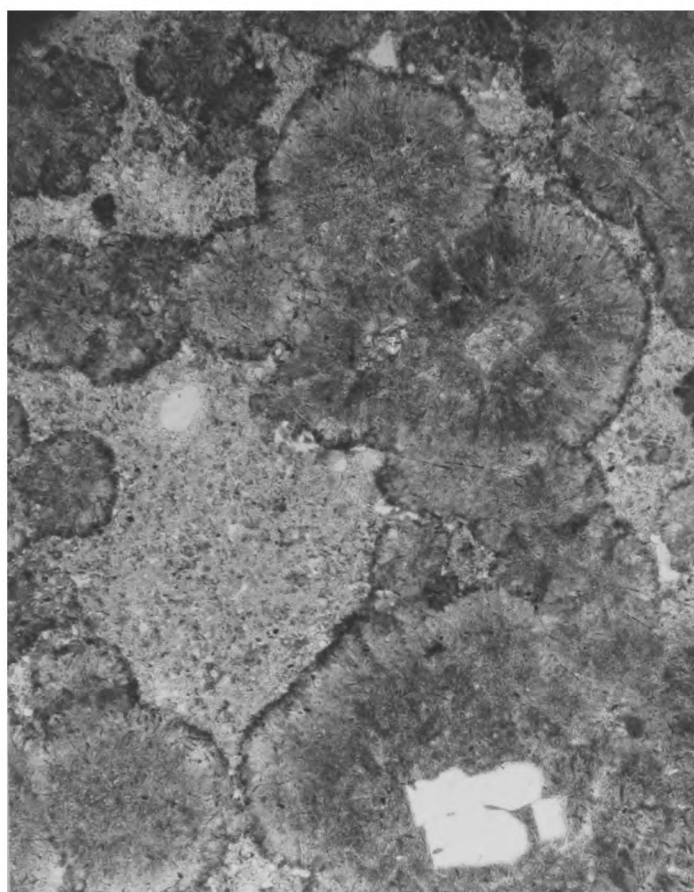
(spherulites). Spherulites are formed when the growth rate (or nucleation rate) is in excess of the diffusion rate (Keith and Padden, 1964) and are common in acid lavas and ignimbrites. Rejected impurities are trapped between the fibres within the spherulites or present as impurity-rich rims to the spherulites (Plate VI.17).

A few localities of pitchstone are found in Swaziland, often showing perlitic cracks (Plate IV.33) or incipient devitrification (Plate VI.10, 11). The majority of the rhyolites, however, are completely devitrified; some showing well-developed spherulites, others microgranophyric, snowflake or fine-grained groundmass textures.

The commonest type of simple spherulite is that consisting of radiating fibres of quartz and feldspar. The spherulites often nucleate on the phenocrysts and grow outwards, mutually interfering (Plate VI.17). The patches forming between spherulites are often coarsely granophyric. Each spherulite has a dark, often chloritic, rim of material rejected from the crystallizing fibres. These radial spherulites vary in size from 0.5 mm to 1 cm and completely destroy the primary groundmass textures.

The banding in the rhyolites is often formed by layers of spherulites. The fiamme or flow lines act as nuclei for the spherulites and devitrification is facilitated by a higher concentration of volatiles in the collapsed pumice or where released along flow shear planes. The flow lines may also possess more short range order which induces more rapid nucleation and crystal growth (Lofgren, 1971). Spherulites also grow along joint planes due to the increased water content; the resulting layers are very similar to those produced by spherulite development along primary laminar features.

Smooth, round, red spherulites are found in some rocks; these can range in



1 mm

Plane Polarized Light



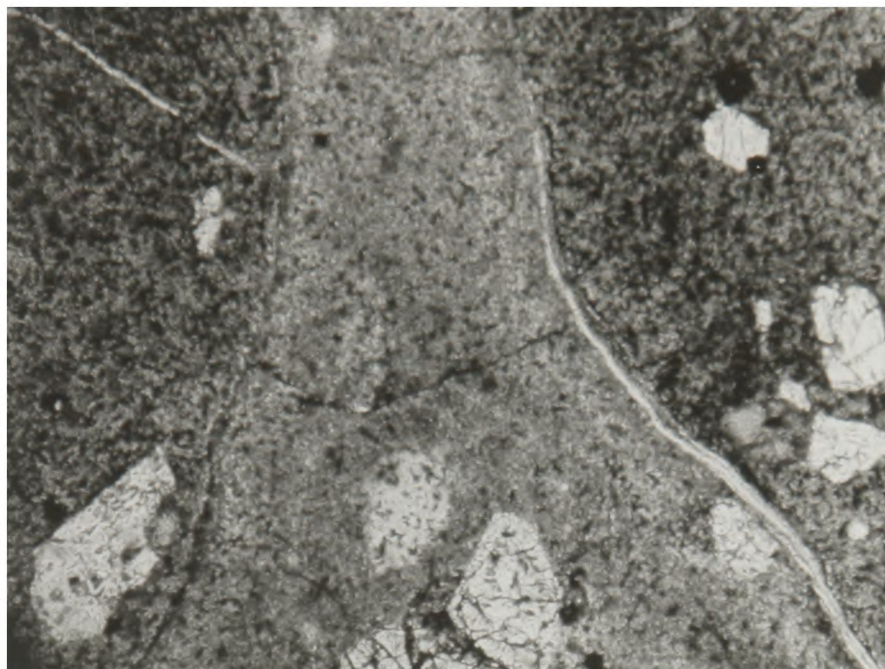
Crossed Nicols

PLATE VI.17 Radiating spherulites forming round phenocrysts and mutually interfering. Dark impurity-rich rims develop round the spherulites. Mesostasis granular. (SK 21)

size from 3 mm to 6 cm, the larger ones developing hollow centres. They are characterised by a fine-grained matrix, generally haematite-stained. The groundmass within these spherulites is not significantly different from that outside and primary banding can be traced through them; they have not obliterated the groundmass textures to the same extent as the radiating spherulites (Plate IV.8). As their size increases they develop hollow centres and become lithophysae (Plate VI.18); when infilled with quartz or agate they are known as 'thunder eggs'. They are often cracked and fractured.

Lithophysae are common in gas-rich lavas and less so in ignimbrites (Smith, 1960b). Their hollow centres are partly due to the reduction in volume or devitrification and partly to the release of volatiles during crystallisation. Spherulites are not an equilibrium crystal habit because of their large surface area: volume ratio due to the fibrous nature of the component crystals, experimentally formed inorganic ones have been known to shatter spontaneously (Keith and Padden, 1963). This instability may be a factor in producing brecciated lithophysae.

The process of formation of these fine-grained spherulites is in some doubt; they postdate the primary banding but only partially obscure it. Where they occur in glassy rocks, the matrix within the spherulites is identical to that found in the normal fine-grained rhyolites. The process producing these spherulites may be similar to that producing the normal fine-grained groundmass in the rhyolites, except, in the latter case, the devitrification has gone to completion. Radiating spherulites form where the growth rate is faster than the diffusion rate (Keith and Padden, 1963); if the diffusion rate were faster, or the growth rate slower, different spherulite morphologies might be produced. The presence of dissolved volatiles may have a significant effect on devitrification; when released during devitrification



1 mm

PLATE VI.18 The margins of two massive spherulites. There is no internal structure and the matrix is texturally very similar within and without the spherulites. (MW 42)



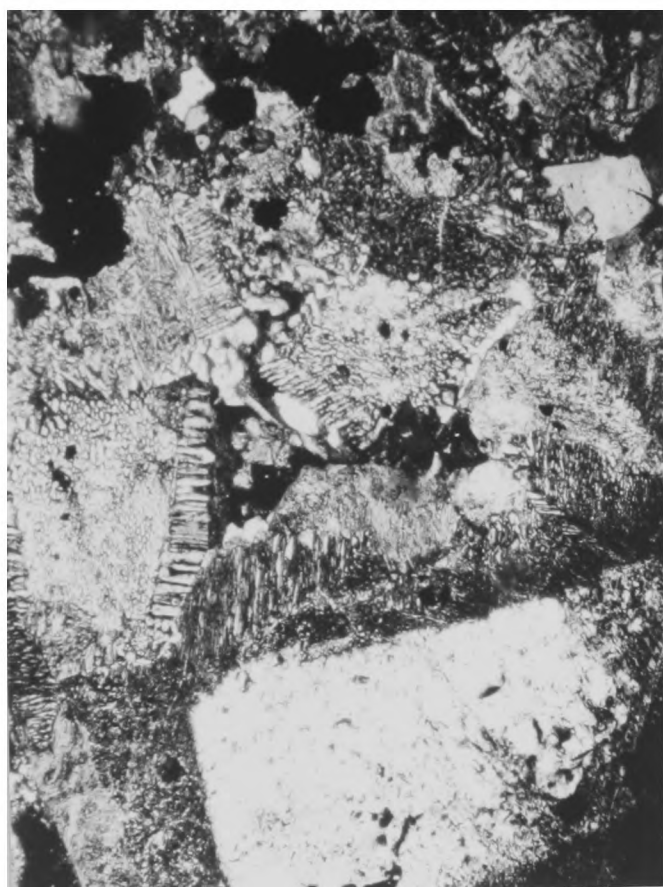
PLATE VI.19 Lithophysae in glassy rhyolite. Spherulites have developed hollow centres. Note their cracked and broken nature. (Kruger National Park)

they will migrate and aid devitrification elsewhere (Smith, 1960b). Volatiles released from the first nucleus in a glass to crystallise may migrate outwards, causing crystallisation, and hence produce the spheroidal structures observed.

The groundmass of the majority of rhyolites is completely devitrified, often showing no trace of either the original structures or any spherulite development. Many of the flows show a microgranophyric texture (referred to as 'felsitic' by Monkman (1961)) showing a fine-scale quartz-feldspar intergrowth which often seems to form rounded spherulite-like shapes (Plates VI.19, 20). This texture is present in the samples from the thicker flows, on a slightly coarser scale in the granophyre dykes, and on a coarse scale in the granophyres. It is probably due to recrystallisation of an original spherulite or the primary development of coarse intergrowth textures as the spherulite formed.

Snowflake texture (Anderson, 1969) is developed in many of rhyolites. Patches of poikilitic quartz in optical continuity enclose feldspar crystals, giving the effect of many rounded irregular patches in the groundmass (Plate VI.24). Some of the snowflakes contain platy crystals of quartz probably paramorphing trydymite (Plate VI.25). Anderson (1970) has proposed that snowflake textures are diagnostic of texturally altered welded tuff, although there seems no reason why they cannot also form in rhyolite lavas (cf. Green (1970)). Snowflakes may be analogous to spherulites with quartz recrystallised in optical continuity.

Many of the rhyolites, if not showing obvious devitrification characteristics, have a fine-grained homogeneous groundmass formed by very small mutually interfering spherulites. In contrast, the trachyandesite sample, MZ 6, has a fine ophitic-type groundmass with small randomly arranged feld-



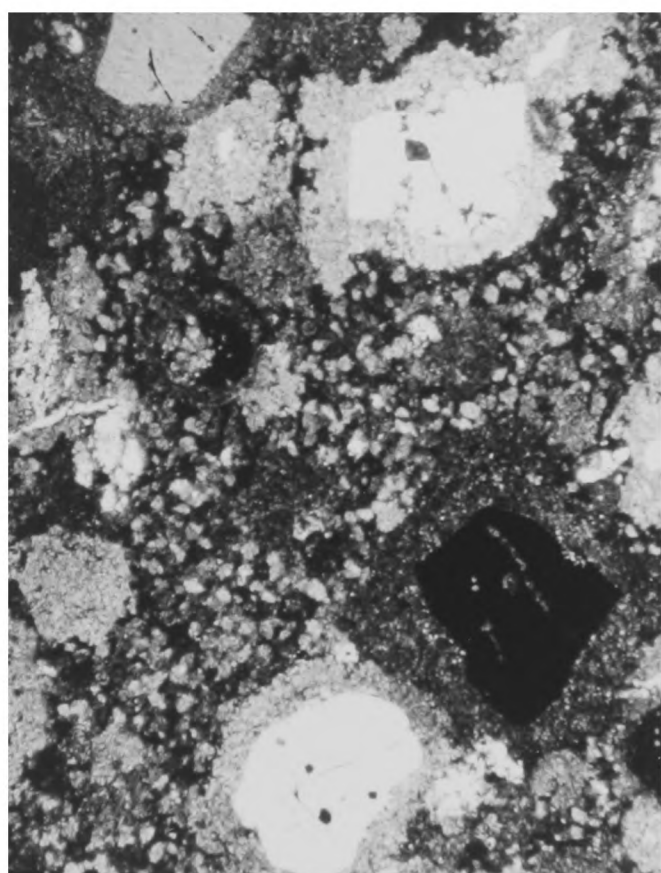
1 mm

PLATE VI.20 Microgranophyric texture in the groundmass of a rhyolite. Fine-grained quartz feldspar intergrowth, often nucleating round the phenocrysts. (RC 61)



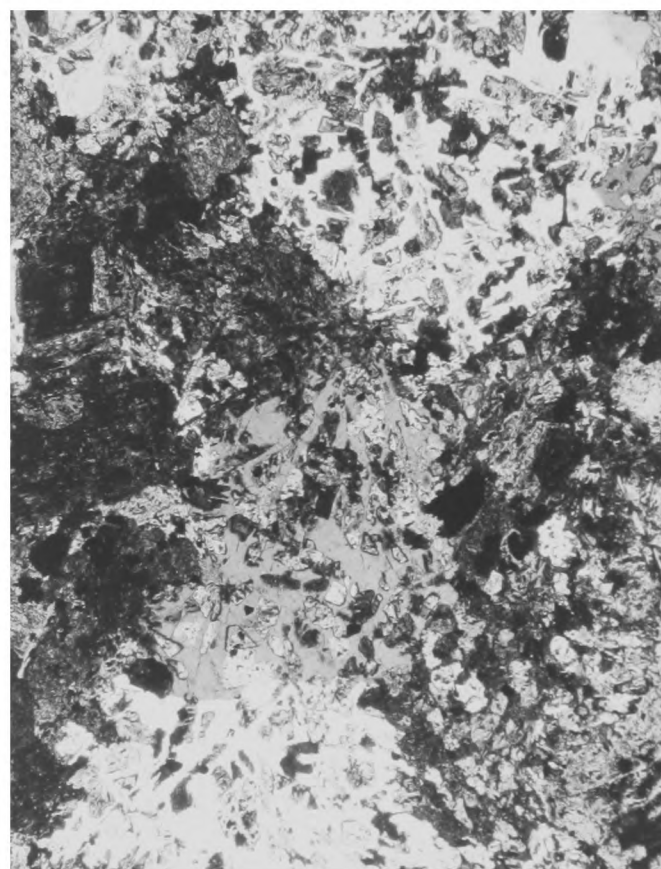
1 mm

PLATE VI.21 Relict spherulitic texture. Radiating quartz-feldspar intergrowth largely recrystallized to a microgranophyric texture. (RC 48)



1 mm

PLATE VI.22 Snowflake texture forming in optical continuity round quartz phenocrysts. Poikilitic patches of quartz enclose feldspar laths. Developed on a smaller scale in the mesostasis. (SK 131)



1 mm

PLATE VI.23 Coarse-grained snowflake texture. Poikilitic quartz enclosing feldspar grains. Lath shaped quartz may be paramorphing tridymite. (MW 66)

spars and may have originated as a lava rather than as an ash-flow tuff.

The rhyolites have had a complex devitrification history and show a large variety of groundmass textures. As discussed above, spherulites are themselves unstable and in these Karroo rocks, the original radial texture often seems to have been completely recrystallised to a fine quartz-feldspar intergrowth. Cyclic sub-solidus recrystallisation may take place as the groundmass progressively re-equilibrates. Volatiles released during devitrification will rise through the flow and promote devitrification elsewhere. One result of devitrification is that the elements excluded from quartz and feldspar migrate outwards forming an impurity-rich rim round the spherulites. These elements will thus be inhomogeneously distributed in the groundmass and may be more prone to leaching. Devitrification is an exothermic process and will help to maintain the temperature within the large rhyolite cooling-units.

CHAPTER VII

GEOCHRONOLOGY OF KARROO VOLCANICS FROM SWAZILAND

VII GEOCHRONOLOGY OF KARROO VOLCANICS FROM SWAZILAND

VII.1 Introduction

Rubidium-strontium, potassium-argon and fission-track dating techniques have been used in this study to date the rhyolites and dolerites and to confirm the correlation with other areas. From the data the isotopic characteristics of the source region can be defined and the effect of devitrification on the isotope systems can be assessed.

In the following sections the new data are discussed and compared with those previously published. Analytical techniques are described in Appendix B.

VII.2 Constants

The following decay constants are used in this work, published ages are recalculated where necessary and the original ages given in parentheses:

Rb-Sr:

$$\lambda_{87\text{Rb}} = 1.42 \times 10^{-11}/\text{y} \quad (t = 4.88 \times 10^9 \text{ y})$$

(Recommended by the I.U.G.S. Subcommittee on Geochronology, 1977.)

K-Ar:

$$\lambda_{\text{E}} = 0.585 \times 10^{-10}/\text{y}$$

$$\lambda_{\text{B}} = 4.72 \times 10^{-10}/\text{y}$$

$$^{40}\text{K}/\text{K} = 0.0119 \text{ at. \%}$$

$$(^{40}\text{Ar}/^{36}\text{Ar})_{\text{atmospheric}} = 288 \text{ (measured at I.G.S., London)}$$

1 σ errors are quoted unless stated otherwise.

MSWD (mean squared weighted deviation) values are quoted for the calculated isochrons. Isochrons with MSWD values less than 2 are considered sufficiently good fits to the data.

VII.3 Rb-Sr

Rubidium-strontium whole-rock age determinations on Karroo rocks from Nuanetzi and the southern Lebombo were published by Manton (1968) and his data are shown in Figure VII.1. The data from rhyolites and granophyres from the two areas define two near parallel isochrons giving ages of 202 (206) $\pm$ 13 my* (initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio = 0.708) and 198 (202) $\pm$ 14 my (0.704) for Nuanetsi and the southern Lebombo respectively.

Strontium isotope analyses were carried out as part of this study to confirm Manton's data, to determine the isotopic composition of the source region for the rhyolites and to see if devitrification had any effect on the strontium isotopes.

Six samples were selected for Sr-isotope analysis: two devitrified 'normal' rhyolites (RC 131 {intermediate} and RC 74 {acid}), two pitchstones with little or no devitrification (Sk 138 and Sk 153) and the two rhyolitic interbeds (RC 42 and RC 119). The data, together with three analyses of Swaziland dolerites (Table VII.2) from Norry (1977), are presented in Table VII.1 and Figure VII.2.

The rhyolites and pitchstones fall on an isochron with an age of 189 $\pm$ 7 my (MSWD = 1.1). The two pitchstones are not significantly different from the devitrified samples. The initial ratio (0.7038 $\pm$ 3) is within error of that of the basic rocks.

Manton (1968) calculated his southern Lebombo isochron using only seven of the ten data points. Only the more radiogenic of the two basalts, which is interlayered with rhyolites near the Jozini dam, falls significantly off

* The ages are recalculated with $\lambda_{^{87}\text{Rb}} = 1.42 \times 10^{-11}/\text{y}$, the originally published ages are given in brackets.

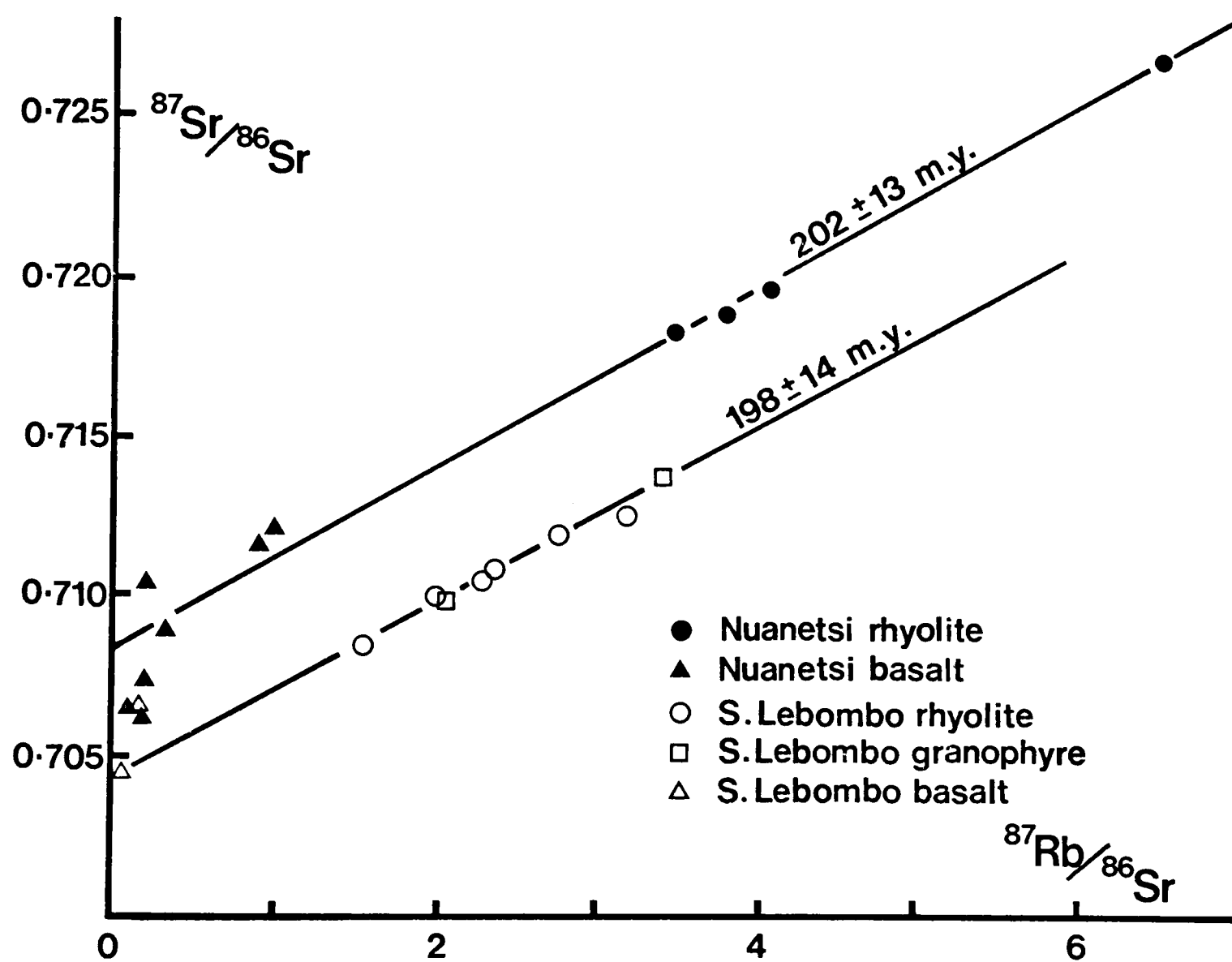


Fig. VII.1. The data from Manton (1968) for Karroo volcanics from the Lebombo and Nuanetsi volcanic provinces. The ages are recalculated with $\lambda_{^{87}\text{Rb}} = 1.42 \times 10^{-11}/\text{y.}$

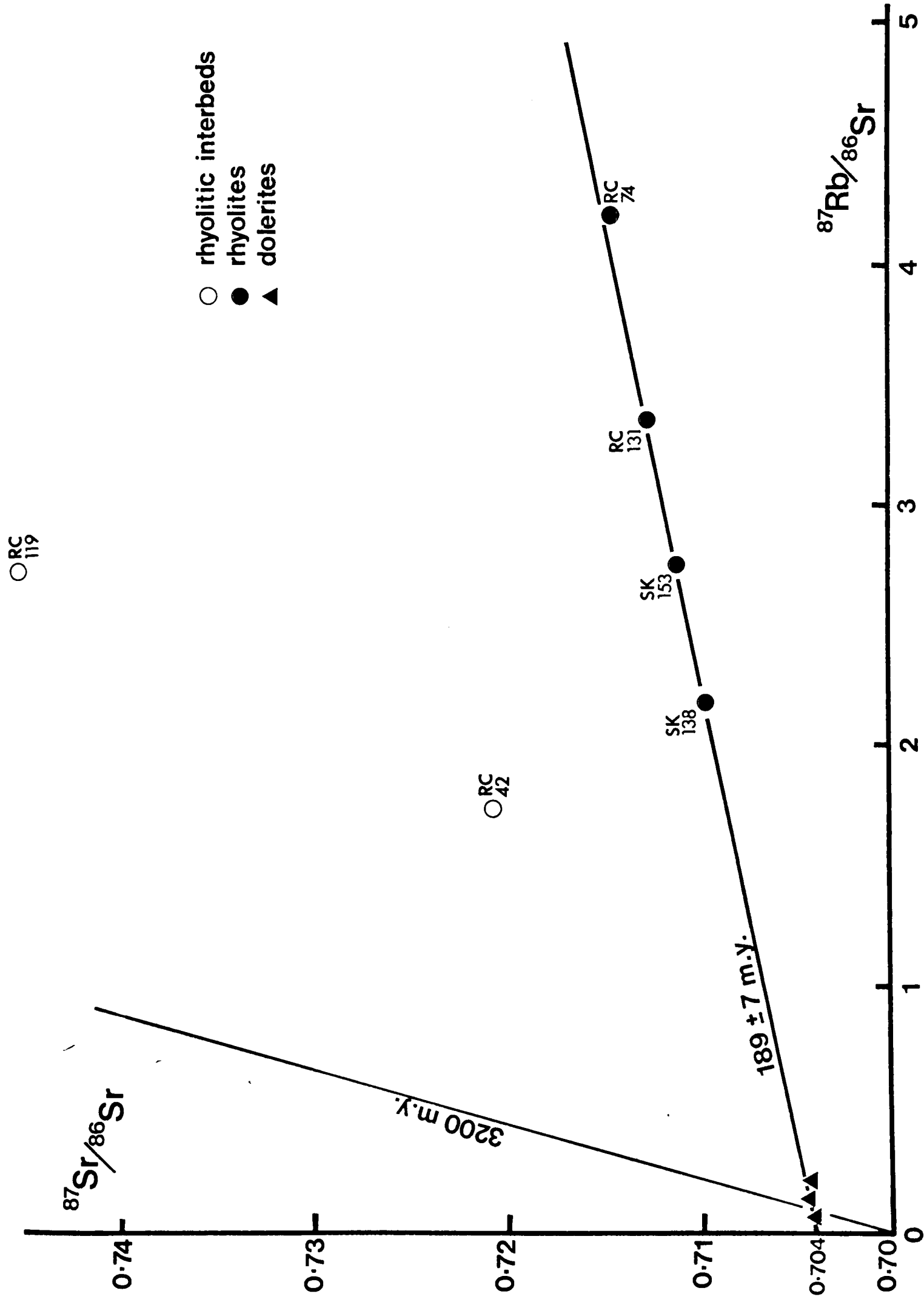


Fig. VII.2. Rb-Sr isochron plot of data from Swaziland rhyolites.

TABLE VII.1

New Rb-Sr data for Karroo rhyolites from Swaziland (analyst : H.J. Chapman)

Sample Details	Rb/Sr (error = 2σ)	⁸⁷ Rb/ ⁸⁶ Sr (error = 2σ)	⁸⁷ Sr/ ⁸⁶ Sr (error = 2σ)	(⁸⁷ Sr/ ⁸⁶ Sr) ₀ at 190 my.
RC 42 Rhyolitic interbed. 1.75 km SW Magwangana dam. 26°37'S 31°56'E.	0.60 ± 0.02	1.74 ± 0.05	0.71994 ± .6	0.7158
RC 74 Acid rhyolite. Flow IX. 1 km SE Oribi Farm. 26°23'S 32°00'E.	1.395 ± 0.04	4.03 ± 0.12	0.71457 ± 10	0.7035
RC119 Rhyolitic interbed. Lubuli-Nsokg Rd. 8 km W Lubuli. 27°00'S 31°50'E.	0.94 ± 0.03	2.73 ± 0.08	0.74571 ± 6	0.7385
RC131 Intermediate rhyolite. 1 km east Groenings Farm. 26°33'S 32°05'E.	1.152 ± 0.02	3.33 ± 0.06	0.71276 ± 8	0.7039
SK138 Intermediate pitchstone. Flow XVII. 6 km E Oribi Farm. 26°22'S 32°03'E.	0.746 ± 0.03	2.16 ± 0.08	0.70953 ± 6	0.7039
SK153 Intermediate pitchstone. Flow XXI. 7 km E Oribi Farm Farm. 26°22'S 32°04'E.	0.955 ± 0.04	2.76 ± 0.11	0.71142 ± 8	0.7041

TABLE VII.2

Sr-isotope data from Norry (1977) for Swaziland Karroo dolerites.

<u>Sample details</u>		<u>Rb/Sr</u>	<u>$^{87}\text{Sr}/^{86}\text{Sr}$</u> (error = 2σ)
RC 110	Dolerite. Lubuli-Nsoko Rd. 1 km W. Lubuli. $27^{\circ}01'S$ $31^{\circ}54'E$.	0.045	0.70446 ± 4
RC 120	Dolerite. 11.5 km W. Lubuli. $27^{\circ}01'S$ $31^{\circ}47'E$.	0.073	0.70425 ± 6
RC 121	Dolerite. 14.5 km W. Lubuli. $26^{\circ}59'S$ $31^{\circ}46'E$.	0.021	0.70409 ± 4

the line. An isochron through the remaining nine data points gives an age of 188 ± 12 my (MSWD = 0.2 for the quoted errors, which appear to be over-estimates). If the new data are added to Manton's, the resulting sixteen point isochron gives an age of 187 ± 3 my (MSWD = 1.5) with an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.70397 ± 4 . If the rhyolites and granophyres only are considered, the age obtained is 185 ± 6 my (MSWD = 1.2) with the initial ratio = 0.7041 ± 3 .

The two rhyolitic interbeds (RC 42 and RC 119) are enriched in $^{87}\text{Sr}/^{86}\text{Sr}$ relative to the main body of the rhyolites (Figure VII.2). This enrichment is presumably due to the incorporation of radiogenic crustal strontium; hydrothermal or groundwater alteration would also have affected the basalts. Rb-Sr data from the Swaziland basement yield ages around 3,200 my with initial ratios (for the Ancient Gneiss Complex) as low as 0.7006 and Rb/Sr ratios in the range 0.05 - 2.0 (Hunter, 1974a). The present day $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the crust is defined by the 3,200 my isochron in Figure VII.2. If the interbeds were derived entirely from the crust, crustal $^{87}\text{Rb}/^{86}\text{Sr}$ ratios of at least 0.35 and 0.9 would have been required to produce their high initial ratios (0.7158 and 0.7385 respectively) in 3,000 my. These $^{87}\text{Rb}/^{86}\text{Sr}$ ratios correspond to Rb/Sr ratios of 1 and 2.5 and are average-to-high values for the Precambrian crust. If contamination of a 'normal 0.704-type' magma took place, there would have needed to be virtually complete strontium isotope exchange between the magma and the crust. The rhyolitic interbeds are also anomalous in their trace element contents (see Chapter VIII) and it is considered unlikely that they were derived from the same source as the rest of the rhyolite magma.

The equivalence of the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the dolerites and rhyolites is generally accepted as indicating their comagmatic nature; the low value (0.704) is indicative of an origin from the mantle. The lower

crust, however, with sufficiently low Rb/Sr ratios can also have low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios: to produce a value for the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.704 200 my ago the $^{87}\text{Rb}/^{86}\text{Sr}$ ratio of the crust would have needed to be of the order of 0.1 (i.e. Rb/Sr = 0.03). The Rb/Sr ratios of the rhyolites are of the order of 1 and it is considered unlikely that magmas with Rb/Sr = 1 can be produced from a granitic (or granulitic) source region with Rb/Sr = 0.03 with any significant degree of melting (see discussion on melting models in Chapter VIII).

Lead isotope data (Betton, pers. comm.) confirm the conclusions reached from the strontium data. The rhyolitic interbeds contain typical Archaean crustal lead whereas the rhyolites and dolerites contain less radiogenic lead more typical of the mantle.

VII.4 K-Ar

Fitch and Miller (1971) published potassium-argon age data from Lesotho basalts and dolerites. They obtained an age of 187 ± 7 my from the base of the Drakensberg succession and this has recently been confirmed by K-Ar isochron dates of 185 ± 4 (2σ) my and 187.9 ± 5.6 (2σ) my for samples respectively 6 m and 1 m above the Cave Sandstone (Fitch, unpublished data, 1977). From the spread of their data, Fitch and Miller (op. cit.) were able to estimate the time span of Karroo volcanism as being from 190 - 150 my.

Potassium-argon determinations were carried out as part of this study with the original object of defining the age range of the rhyolites and the extent of any north-south diachronism. Ten of the freshest samples were selected covering the whole range of rock-types (including three of those defining the Rb-Sr isochron) to see what precision could be obtained. The data are presented in Table VII.3 and summarised in Figure VII.3.

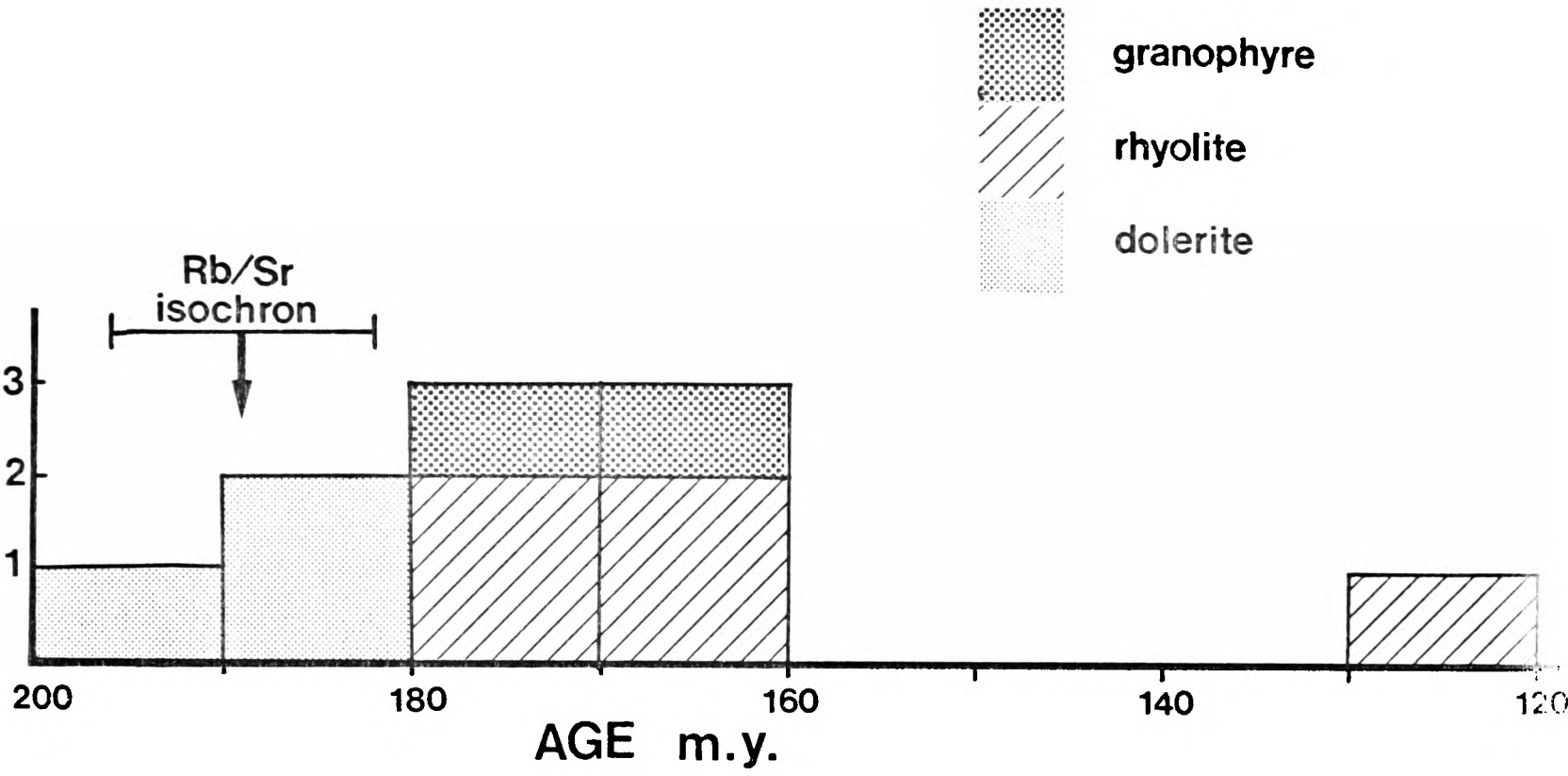


Fig. VII.3. Histogram of individual K-Ar ages for samples from Swaziland
See text for discussion.

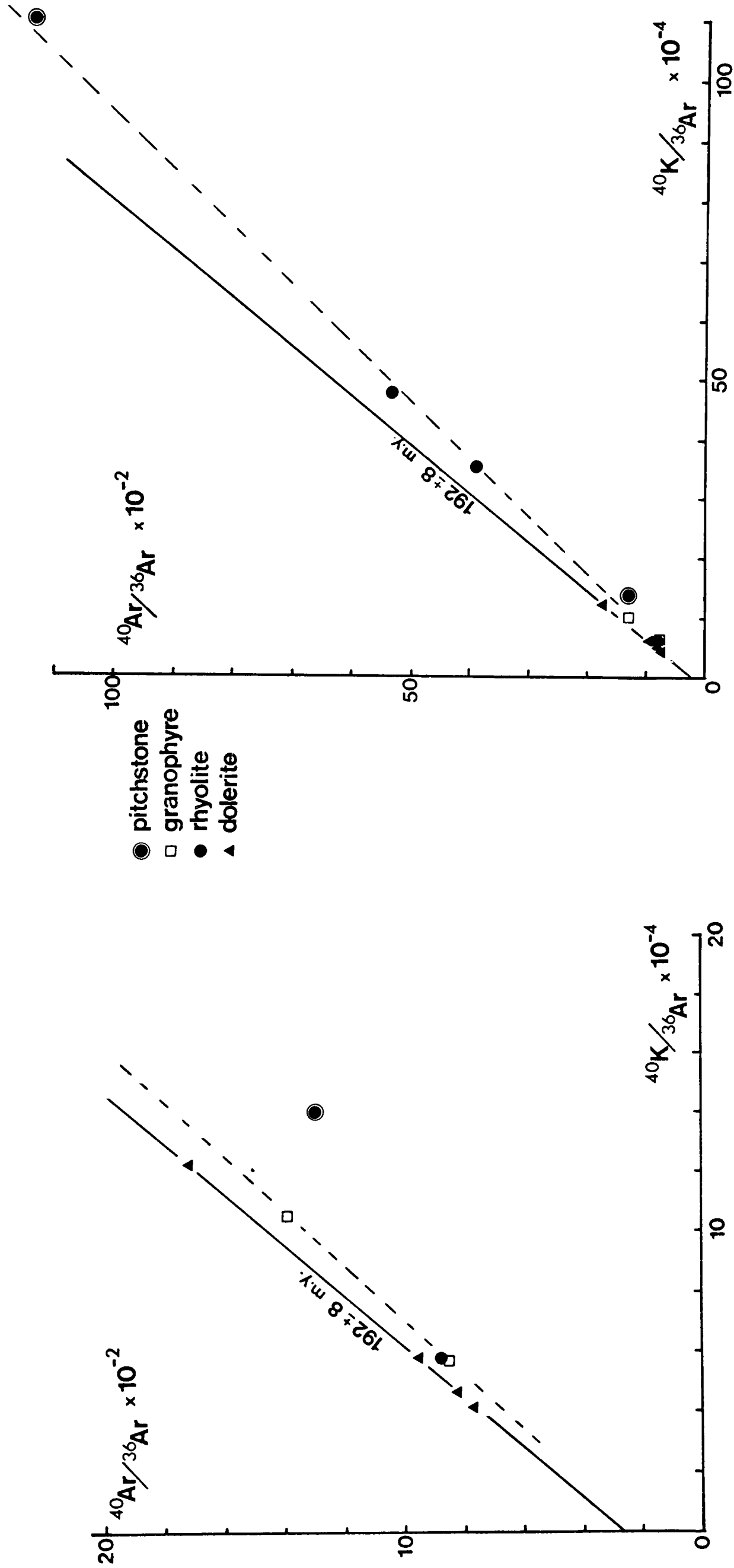


TABLE VII.3

New whole rock K-Ar data from Karroo igneous rocks from the Lebombo in Swaziland

Sample References		K %	Atmospheric contamination %	Vol ₄₀ Ar na/g	Age+ error (2σ) m.y.
<u>Dolerites</u>					
RC 4 115	Dolerite. Lubuli-Nsoko road. 3.5 km west of Lubuli. 27°00'S 31°52'E.	0.451	30.28	3.49	185 ± 9
RC 4 116	Dolerite. Lubuli-Nsoko road. 4.5 km west of Lubuli. 27°00'S 31°51'E.	0.279 0.279	34.93 36.99	2.17 2.23	185 ± 9 190 ± 13
RC 4 120	Dolerite. Lubuli-Nsoko road. 11.5 km west of Lubuli. 27°01'S 31°46'E.	0.539	16.35	4.35	192 ± 9
<u>Rhyolites</u>					
RC 4 52	Intermediate rhyolite. Flow XI. Maphyngwane store. 26°35'S 32°01'E.	3.24	7.21	12.05	173 ± 8
RC 4 55	Intermediate rhyolite. Flow XI. 2 km NW Maphungwane store. 26°34'S 32°00'E.	3.36	5.41	24.50	174 ± 8
RC 4 74	Acid rhyolite. Flow IX. 1 km SE Oribi Farm. 26°23'S 32°00'E.	4.17	32.76	28.62	164 ± 8
SK 138	Intermediate pitchstone. Flow XVII. 6 km E Oribi Farm. 26°22'S 32°03'E.	2.48	21.90	12.36	121 ± 6
SK 153	Intermediate pitchstone. Flow XXI. 7 km E Oribi Farm. 26°22S 32°04'E.	3.59	2.59	24.32	162 ± 7
<u>Granophyres</u>					
RC 70	East Stegi granophyre. 4 km NW Stegi. 26°25'S 31°55'E.	3.85	20.64	27.72	172 ± 8
SK 2		3.92	33.58	26.63	162 ± 9

188 ± 5

The three dolerites analysed form a tight group with an average age of 188 ± 8 (2 σ) my. The rhyolites and granophyres form a broad group on the young side of the dolerites, the maximum age of any individual sample being only 174 my. One of the pitchstones (Sk 138) gives an age of 121 ± 6 (2 σ) my.

The data are plotted on a K-Ar isotope ratio plot in Figure VII.4. The use and misuse of these plots have been discussed by Hayatsu and Carmichael (1970, 1977). The dolerites fall on a line corresponding to an age of 192 ± 4 my (initial $^{40}\text{Ar}/^{36}\text{Ar} = 272 \pm 14$) which is within error of the ages calculated using an assumed initial ratio of 288 (measured at I.G.S.). The rhyolites and granophyres form a vague line with a lower gradient than that of the isochron. This 'errorchron' is felt, in the light of the Rb-Sr data on the same samples, not to be significant and to represent ^{40}Ar loss in proportion to the amount present, resulting in rotation of the isochron. Radiogenic ^{40}Ar is present in potassium sites and can diffuse more easily than ^{36}Ar out of the crystal lattice.

The processes of spherulite formation and recrystallisation are envisaged to result in argon loss and ages calculated from devitrified rocks may give an age of recrystallisation rather than formation. The pitchstones, however, give the youngest ages and argon diffusion out of glass may be more significant than the effects of recrystallisation and subsequent diffusion.

No conclusions concerning the relative ages of crystallisation and devitri-fication of the acid rocks can be drawn from these data. The problems of argon loss render this technique unsuitable for the dating of whole-rock samples though the analysis of separated phases may produce more concordant dates.

VII.5 Fission-track dating of zircons

Fission-track dating of separated zircons has been carried out by Fitch and Hurford (unpublished data). The results are given in Table VII.4. The Zululand rhyolite gave an age of 188 ± 2 (2σ) and subsequent determinations on Swaziland samples confirm this age though with higher errors due to fewer determinations. No older zircons have been found in these samples.

VII.6 Summary

Three separate techniques give an age for the Karroo in Swaziland of about 188 my. All the determinations are within error (except for some of the k-Ar dates for reasons discussed above) and this is the first available comparison of Rb-Sr and fission-track dating techniques.

No significant age differences can be measured between the rhyolites and dolerites or between the rocks of Swaziland and those at the base of the Lesotho succession. No reliable dates have been obtained from the granophyres.

TABLE VII.4

Fission-track dating of zircons

<u>Sample details</u>	<u>Source</u>	<u>Date (error = 1σ)</u>
Rhyolite, Ngweni Quarry	Fitch (1976, unpubl.)	189 ± 21
Zululand		187 ± 16
		189 ± 20
RC 74**	Hurford (1977, unpubl.)	193 ± 8
SK 153**	" "	189 ± 11
SK 138**	" "	*211 ± 11

* anomalously high age because of a technical fault;

** for sample localities, see Table VII.3.

CHAPTER VIII

GEOCHEMISTRY

VIII GEOCHEMISTRYVIII.1 Introduction

Fifty-three whole-rock samples have been analysed for major elements (analytical techniques are summarised in Appendix A). The first batch analysed consisted of some twenty rhyolites and seven dolerites and represented the freshest rocks available. Consequently the compositional spread was small, most of the samples coming from the intermediate rhyolites and falling between 68 and 72 per cent silica. For the second batch, the rocks identified in the field as falling away from the centre group were chosen and the observed compositional spread increased. The data, therefore, are not representative of the rhyolites as a whole but are biased away from the 'average rhyolite'. The data are presented in Appendix B.

All the samples analysed for major elements have also been run for the trace elements: Nb, Zr, Y, Sr and Rb and a selected few for Ba, La, Ce and Nd. These trace element data show more variation than the major element data and also reveal two samples with anomalous trace element contents (later also shown to have high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios). Fifteen more samples from the inter-bedded flows were run for trace elements only, in an attempt to identify any further exceptions.

P. Betton of this Department is near completion of a research project concerned mainly with electron microprobe studies of rocks collected for this study. No probe work was undertaken as part of this study; unpublished data on the phenocrysts has been made available by Betton and used in the least squares mixing calculations.

The rhyolite data are plotted in Diagrams 5 and 6, those for the dolerites in Figure VIII.4.

VIII.2 Classification of rock-types

The taxonomy of igneous rocks is in a state of flux. The use of the modal Q-A-P-F (Quartz-Alkali-feldspar-Plagioclase-Femics) double triangle for classification (Streckeisen, 1967) is gaining acceptance by many petrologists. For fine-grained volcanic rocks, however, this classification is dependent on the calculation of mineral modes; a procedure fraught with difficulties. Published tables of average rock analyses (Le Maitre, 1976) show that a given rock name can cover a range of compositions. In the Swaziland rhyolites there is a natural division at about 68 per cent silica between the interbedded rhyolites and the main group of rhyolites (cf. Diagram 5) and for convenience those with silica contents below 68 per cent will here be called *dacites* and those above *rhyolites*. The terms *acid*, *intermediate* and *basic*, when used to qualify the rhyolites, refer to the subdivision discussed in Chapter V and have no implications outside this context. The term *rhyolite* is also used in a general sense to describe the acid volcanics of this and other areas. The granophyres do not fall significantly off the major element trends and will be considered along with the rhyolites. Two of the interbedded rhyolites, cropping out towards the base of the basalt succession, have similar major element chemistry but depleted trace element abundances; they are called *rhyolitic interbeds* and are considered to have a separate origin to the rest of the rhyolites.

Two of the samples have less than 65 per cent silica: one of these, MZ.6, has only 56 per cent silica, but relatively high amounts of alkalis, and is similar to a trachyandesite in composition; RC 128, at 61 per cent silica comes from the same flow as specimens RC 45 and RC 36 and is more obviously related to the dacites. MZ 6 comes from a flow exposed in a quarry near Mlaula in the Mbuluzi poort and its field relations could not be ascertained. Le Maitre's averages are compared with the Swaziland data in Table VIII.1.

TABLE VIII.1

Comparison of Lebombo analyses with
average analyses from Le Maitre (1976).

	1	2	3	4	5	6	7	8
SiO ₂	72.82	65.01	58.15	75.6	70.62	66.78	56.25	70.10
TiO ₂	0.28	0.58	1.08	0.17	0.48	0.75	1.45	0.57
Al ₂ O ₃	13.27	15.91	16.70	12.05	12.17	13.31	14.88	12.77
Fe ₂ O ₃	1.48	2.43	3.26	1.61	2.60	4.83	8.28	3.68
FeO	1.11	2.30	3.21	0.31	2.68	1.44	2.93	1.06
MnO	0.06	0.09	0.16	0.03	0.12	0.14	0.21	0.08
MgO	0.39	1.78	2.57	0.09	0.21	0.64	1.21	0.32
CaO	1.14	4.32	4.96	0.36	1.78	2.48	3.36	1.70
Na ₂ O	3.55	3.79	4.35	3.16	3.41	3.78	3.84	3.15
K ₂ O	4.30	2.17	3.21	5.30	4.33	4.33	4.22	4.61
P ₂ O ₅	0.07	0.15	0.41	0.03	0.11	0.22	0.50	0.11
Vol	1.49	1.25	1.91	1.10	1.03	1.88	3.41	1.85
Total	99.96	99.78	99.97	99.81	99.52	100.56	100.52	100.00

- 1. Average rhyolite (Le Maitre, 1976).
- 2. Average dacite (" " ")
- 3. Average trachyandesite (" " ")
- 4. Typical Swaziland acid rhyolite (RC 73)
- 5. Typical Swaziland intermediate rhyolite (RC 55)
- 6 Typical Swaziland basic rhyolite (RC 40)
- 7 Swaziland trachyandesite (MZ 6)
- 8 Average Lebombo rhyolite from Wachendorf (1973)

VIII.3 Major element variation

The data are plotted in Diagram 5. Silica is used for the abscissa as a convenient index of fractionation. The analyses are divided into the five groups outlined above:

- (a) the rhyolitic interbeds from the base of the basalt succession;
- (b) the one intermediate "trachyandesite" flow;
- (c) the rhyolites from the main part of the succession;
- (d) the intrusive granophyres.

The dolerites are considered in Section VIII.6.

Of the major element trends, TiO_2 , P_2O_5 , total Fe, MnO and K_2O show tight correlations with silica whereas Al_2O_3 , MgO, CaO and Na_2O have looser trends. The errors in the silica analyses are proportionately larger than those for the lesser oxides and this may be reflected in, for example, TiO_2 and P_2O_5 by the presence of a slight 'plateau' between 68 and 72 per cent silica where there is a wide variation in silica content for a constant amount of the minor element.

The five granophyre analyses show no significant differences from the rest of the analyses and are considered as part of the same fractionation trend. The interbedded rhyolites (dacites) are also similar to the main group of rhyolites but are poorer in silica and, perhaps, in iron and significantly richer in Al_2O_3 and K_2O . The rhyolitic interbeds are similar in the major elements but poorer in K_2O and very depleted in the trace elements.

Figures VIII.1 show the whole-rock variation relative to average phenocryst analyses (data summarised in Table VIII.2)

TABLE VIII.2

Phenocryst Data

1. Basic rhyolite (RC 36)

	Plagioclase	Clinopyroxene	Ore
	<u>PL 36</u>	<u>CP 36</u>	<u>OR.36</u>
SiO ₂	60.72	53.60	0.16
TiO ₂	0.08	0.35	16.14
Al ₂ O ₃	23.61	1.09	2.04
FeO	0.54	19.70	75.41
MnO	0.01	1.07	1.26
MgO	0.04	11.11	0.09
CaO	7.25	16.54	0.01
Na ₂ O	6.66	0.35	0.03
K ₂ O	0.91	0.01	0.01
	<u>99.82</u>	<u>103.82</u>	<u>95.15</u>

2. Intermediate rhyolite (RC 55)

	Plagioclase	Clinopyroxene	Ore
	<u>PL 55</u>	<u>CP 55</u>	<u>OR 55</u>
SiO ₂	60.60	50.32	0.14
TiO ₂	0.04	0.04	16.81
Al ₂ O ₃	24.86	0.94	1.48
FeO	0.42	23.57	76.45
MnO	0.01	1.19	1.58
MgO	0.02	6.31	0.15
CaO	6.48	15.09	0.01
Na ₂ O	6.40	0.13	0.01
K ₂ O	1.29	0.01	0.01
	<u>100.12</u>	<u>97.6</u>	<u>96.64</u>

In addition analyses of pure quartz and magnetite were used.

3. Dolerites

	Plagioclase	Clinopyroxene	Ore
	<u>PL 11</u>	<u>CP 11</u>	<u>OR 11</u>
SiO ₂	55.85	51.89	0.15
TiO ₂	0.08	0.68	48.89
Al ₂ O ₃	26.72	1.19	0.24
FeO	0.65	15.77	44.26
MnO	0.01	0.35	1.11
MgO	0.13	14.44	0.08
CaO	10.71	14.93	0.06
Na ₂ O	4.96	0.26	0.01
K ₂ O	0.30	0.01	0.01

Analyses for labradorite and olivine (Fo 80) were taken from Deer, Howie and Zussman (1966).

Figures VIII.1 show the whole-rock chemical variation relative to average phenocryst analyses (data summarised in Table VIII.2). The envelopes of the rhyolite compositions and their approximate trends are also shown.

An idealised straight-line variation diagram can be considered to represent a mixing of two end-members: the phenocryst 'extract' whose bulk composition is enclosed by the triangle or polygon linking the phenocryst compositions; and the liquid composition falling at the end of the trend (generally the most evolved liquid). In a cumulus enrichment or partial melting trend the 'extract' is added to an evolved liquid to produce intermediates; in a crystal fractionation trend, removal of the phases constituting the extract from the bulk composition will produce a range of liquid compositions trending away from the extract. In both cases the variation will be linear if the extract is of constant composition. The use of the term 'extract' to describe one end of the mixing line, while having implications of fractional crystallisation, should not be taken to exclude the other possibilities for producing these linear trends.

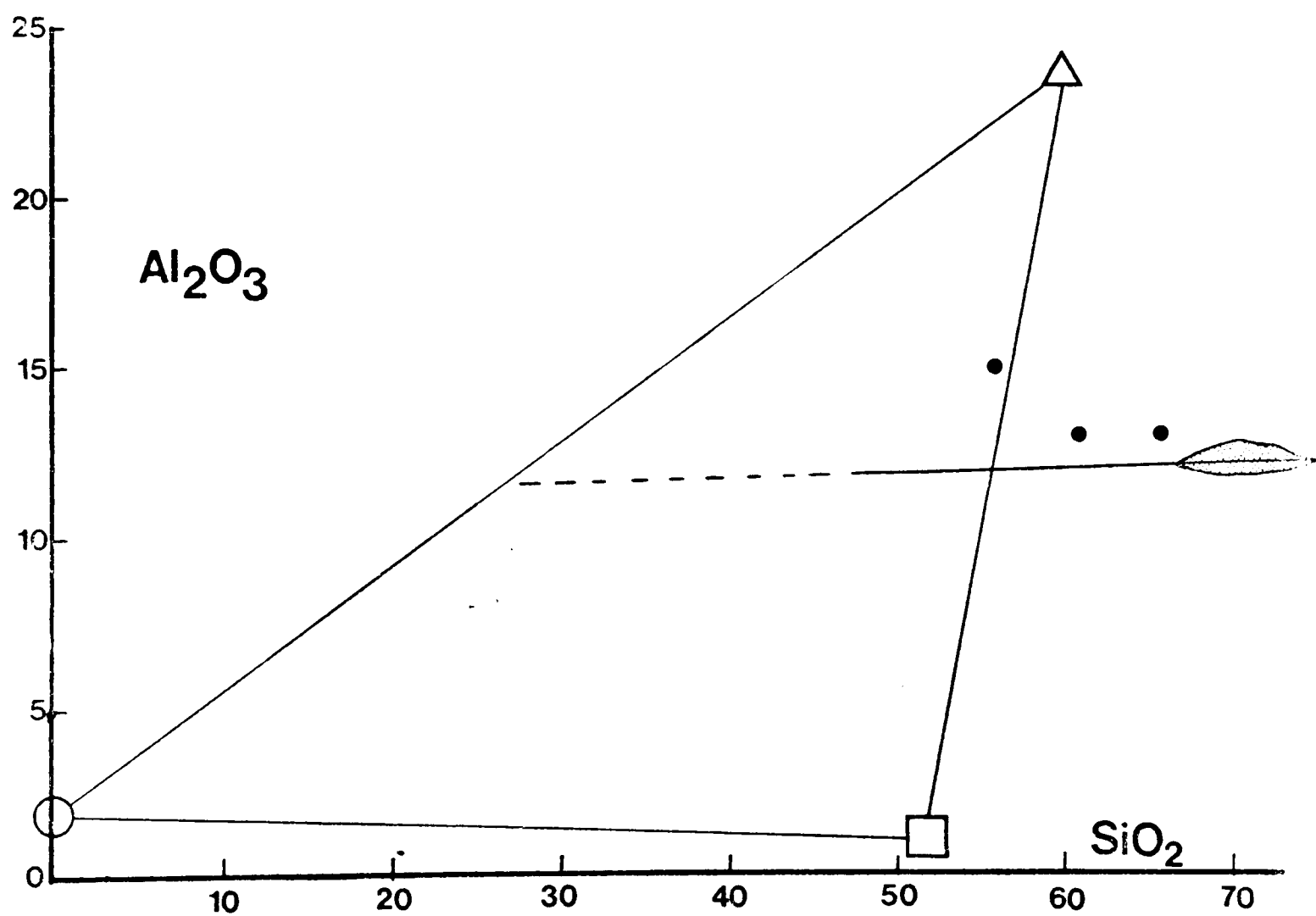
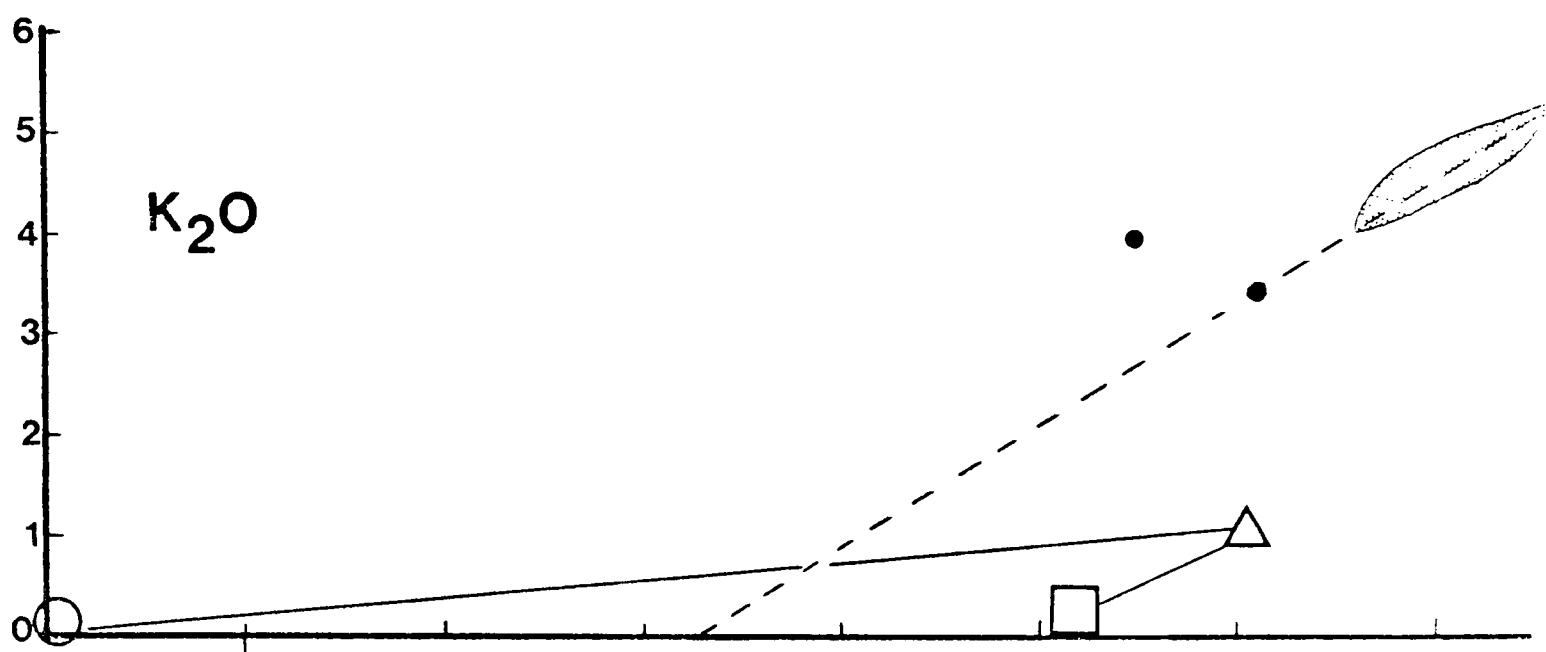
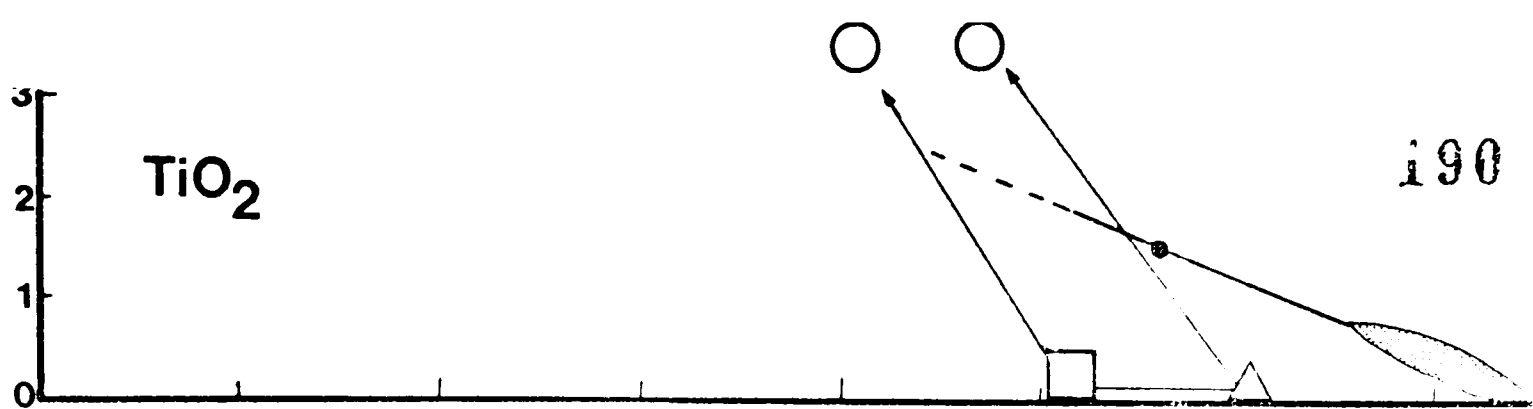
The back projections of the envelopes of the rhyolite analyses would, in a perfect system, intersect with the bulk composition of the extract.

With a two phase extract on a two-dimensional variation diagram, the extract is specified uniquely; with a three phase extract, the bulk composition will fall on the back-projection of the trend line within the three phase triangle. A unique solution for the extract composition can be obtained where the number of components is equal to, or greater than, the number of phases. Multi-dimensional analysis of variation diagrams is considered further in Section VIII.8.

Inspection of these 'extract variation diagrams' can yield qualitative esti-

Fig. VIII.1. Phenocryst extract variation diagrams for the rhyolites. Stippled area denotes main envelope of rhyolite compositions; single points are the chief exceptions to this. Best fit trend lines are solid to approximate extract composition and extrapolated as a dashed line.

Phenocrysts: circles = ore; triangles = plagioclase; squares = clinopyroxene.



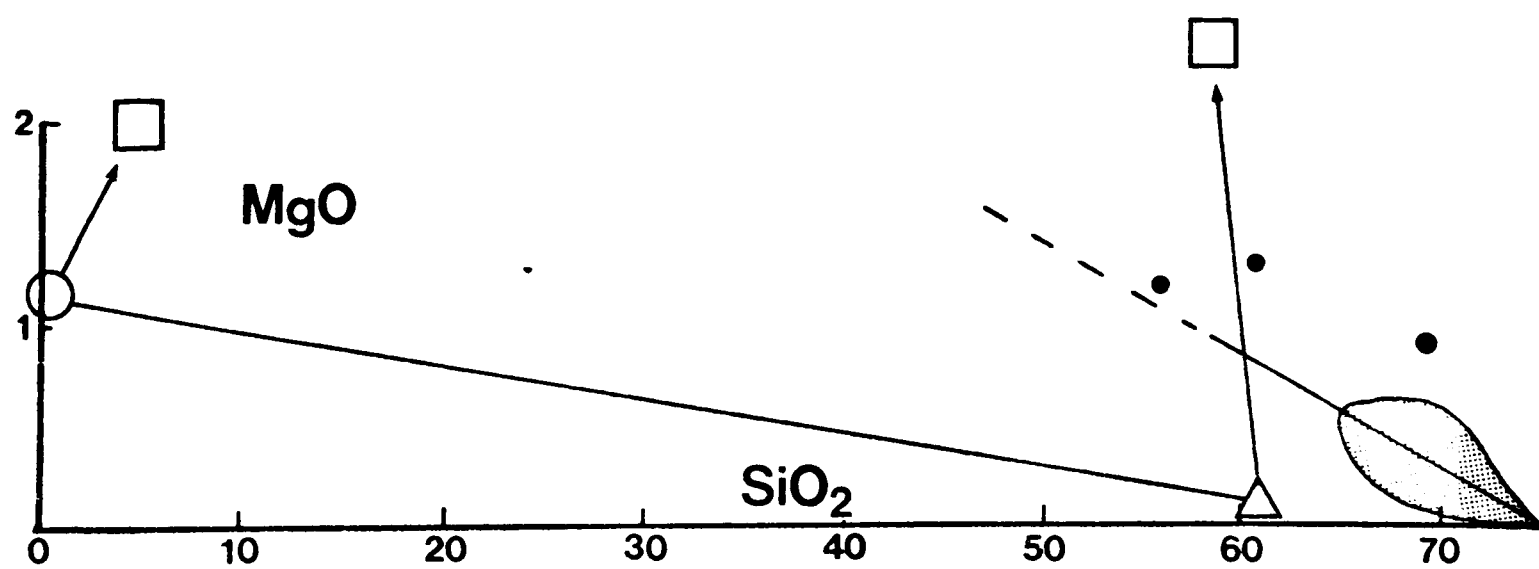
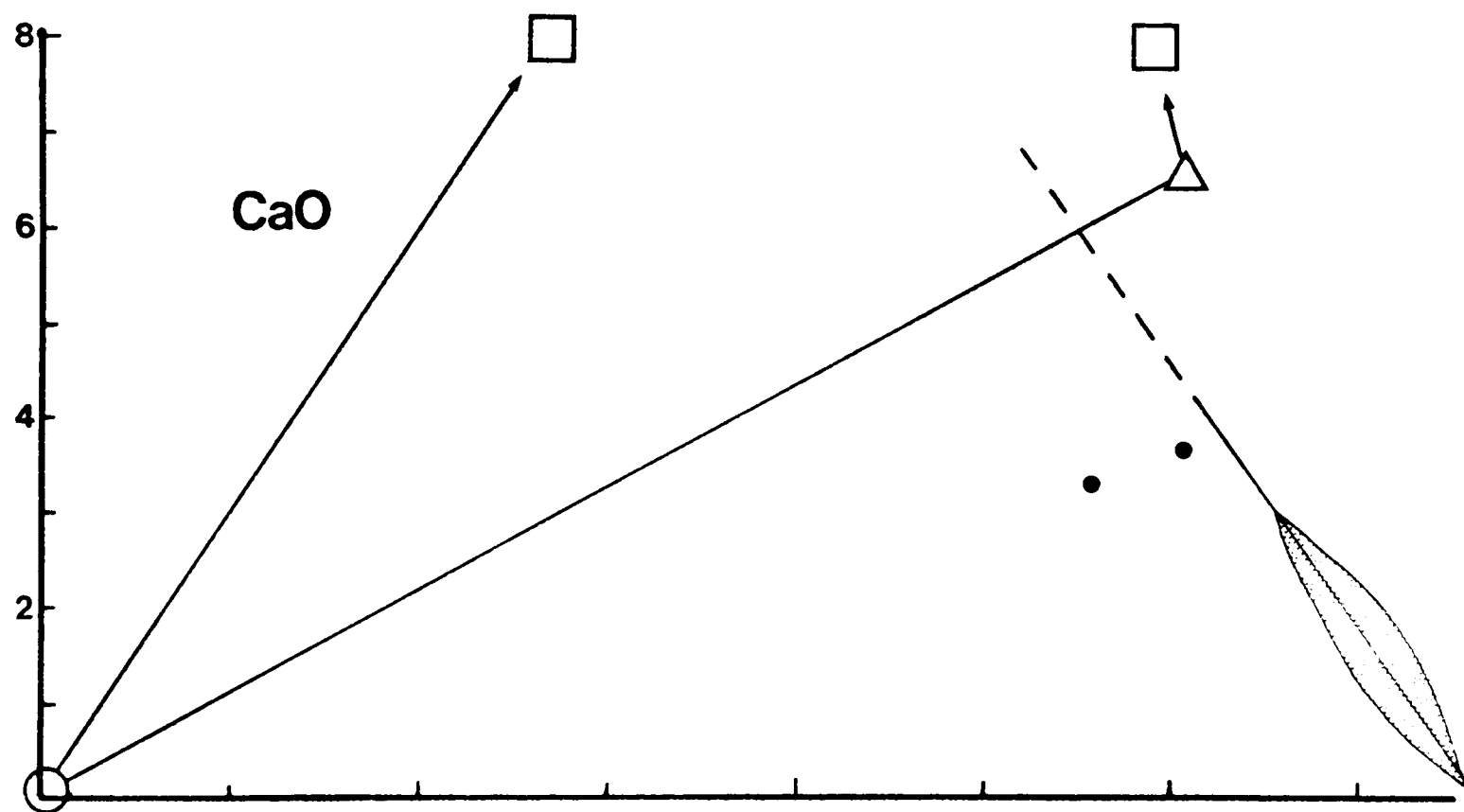
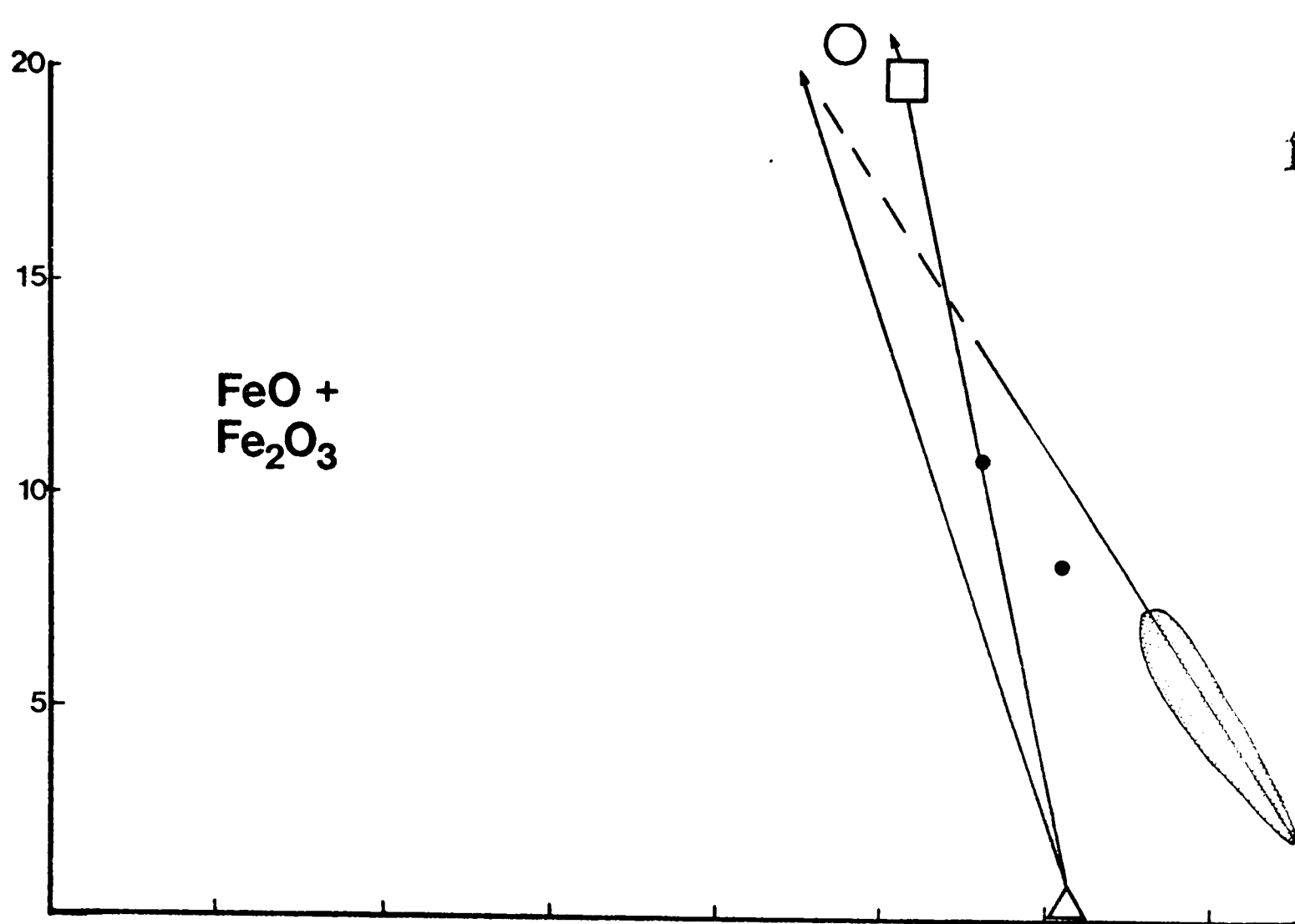


Fig. VIII.1. (Continued)

mates of the extract composition. Because each phase is generally enriched in a characteristic element, the amount of that phase in the extract is related to the abundance of its characteristic element in the extract: thus the modal proportions of apatite and zircon in the calculated extract are largely controlled by the predicted amounts of P_2O_5 and Zr in the extract; MgO is dominantly associated with, and hence controlled by the amount of, clinopyroxene; FeO and TiO_2 are controlled by the ore phases; Al_2O_3 and K_2O by plagioclase; and CaO and SiO_2 by clinopyroxene and plagioclase.

An extract with a small percentage of ore and a dominance of plagioclase over clinopyroxene can explain the variation for most elements. Discrepancies occur for iron, where the extract is too rich in clinopyroxene; aluminium where the plagioclase may be slightly too aluminous and for potassium where the extract is too rich in ore, or generally too low in potassium. These inconsistencies may be reduced if further phases, for example magnetite, fayalite and potassium feldspar, are considered.

The variation is considered in more detail in Section VIII.8 where least-squares petrographic mixing models will be discussed.

VIII.4 Trace element variation

The major element variation diagrams show that the Swaziland rhyolites are a fairly homogeneous group of rocks with little overall variation in bulk composition. Most trace elements are more sensitive to petrogenetic processes than major elements and most of the petrogenetic modelling has been based on them.

Trace element data for rhyolites are scarce but Weaver et al. (1972), Jenks and Goldich (1956), El Hinnawi et al. (1969), Ewart and Taylor (1969), and

Gibson (1972), all quoted trace element data for acidic extrusives. The Swaziland rocks are enriched in Nb, Zr and Y relative to these published data - often by a factor of three. The Swaziland data are presented in Diagram 6.

a) NIOBIUM

Niobium has a very constant abundance in the rhyolites with a range of 65 - 80 ppm. Nb in the granophyres is on average 5 - 10 ppm lower for a given silica content. The data fall into two groups with a dividing line at 72% SiO_2 . Within each group there seems to be a depletion of the Nb with silica, the trends for the two groups displaced by 30 ppm Nb. The upper group contains the four acid rhyolites and one of the granophyres. The two rhyolitic interbeds are depleted by a factor of four relative to the main group.

Very little is known about the geochemistry of niobium; it may substitute for Ti, Sn, Zr, Mo and W on the basis of similar ionic radii but the bulk of it, however, may exist as $(\text{NbO}_4)^{3-}$ complexes and be enriched in residual melts (Taylor, 1965). The substitution of Nb^{5+} for Ti^{4+} is probably the major solid solution effect and Nb would be expected to have the highest distribution coefficients in pyroxenes, rutile and perhaps the ore phases (Taylor, 1965). Cann (1970) used the Nb abundances and Erlank and Kable (1976) Zr/Nb ratios as indicators of petrogenetic environments but these studies considered hydrothermal mobility and melt chemistry respectively rather than solid solution chemistry.

Niobium does not show the residual enrichment predicted by Taylor (1965) and the lack of similarity between the TiO_2 and Nb plots and the closer relation with Zr perhaps suggests that the substitution of Nb in zircon is the major factor affecting the niobium distribution in these rocks (see discussion on zircon fractionation below).

b) ZIRCONIUM

The rhyolites form a tight grouping for zirconium, apparently, insofar as data are available, remaining constant till about 68 per cent silica and then falling as silica increases. Zircon is an abundant accessory in the rhyolites, often attached to larger ore phenocrysts, and it seems likely that the observed fall in Zr has been caused by zircon fractionation. The rocks with less than 68 per cent silica very rarely contain zircons. Four of the samples are richer in Zr; they do not appear to be enriched in zircons and the Zr levels of 1350 ppm probably represent the limit of zircon solubility for that magma composition. When the liquid is saturated with ZrSiO_4 , crystallisation of zircon begins, sufficient to halve the Zr content of the magma by 75 per cent silica. Similarly to Nb, the acid rhyolitic interbeds are fourfold depleted. The two separate groups identified on the basis of Nb abundances are less pronounced or non-existent.

The substitution of Zr for Ti, Fe^{3+} , Fe^{2+} as suggested by Taylor (1965) and its subsequent fractionation in pyroxenes and Fe-Ti oxides are considered to be insignificant relative to its fractionation in zircon.

c) YTTRIUM

Yttrium is very enriched in these rhyolites relative to most other rhyolites for which data are available and ranges from 100 - 180 ppm. The general trend within the rhyolites is one of enrichment with silica; Y in the granophyres is 60 ppm lower than in the acid rhyolites and apparently fall on a trend decreasing with silica. The rhyolitic interbeds are depleted by a factor of three.

The geochemistry of Yttrium is governed by its association with calcium and medium-heavy rare earth elements (e.g. Dysprosium) in rock-forming minerals. Y-minerals are rare; Ca-minerals poor in Y are: calcite, augitic

clinopyroxene and plagioclase; those rich in Y are: hornblende, garnet and apatite. Of the Ca - poor minerals, orthopyroxene and K-feldspar are relatively rich in Y (Lambert and Holland, 1974). Zircon contains large quantities of heavy rare earths and can contain up to 1.5% Y (Krasnobayev et al., 1976).

The variation of Y is considered in greater detail in Section VIII.10 where its variation is partly explained by fractionation of Y-rich apatite and zircon and partly by variations in the source region.

d) STRONTIUM

Strontium falls off rapidly and regularly with silica from over 200 ppm to 30 ppm. Some of the acid rhyolites are enriched relative to the granophyres but the rhyolitic interbeds do not show the same depletion as in Y, Zr and Nb. Both clinopyroxene and plagioclase have high partition coefficients for Sr and it is shown later that the observed variation can be accounted for by the fractionation of a clinopyroxene-plagioclase mixture.

e) RUBIDIUM

Rubidium has a tight trend with little variation about a mean value of approximately 120 ppm. The granophyres are enriched relative to the acid rhyolites, especially the most silica-rich one. Most common rock-forming minerals have very low distribution coefficients for Rb and this is compatible with the gradual enrichment with silica observed.

f) BARIUM, LANTHANUM, CERIUM AND NEODYMIUM

Ba, Ce and Nd show similar patterns and are depleted with increasing silica. La shows less variation and a gradual enrichment with silica. The two acid rhyolites have high contents of all four elements relative to the granophyres. Some of the similarities observed may to some extent reflect

the small number of analyses. Ba, similarly to Sr, is enriched in plagioclase and the variation in Ba can be explained by predominant plagioclase fractionation. The light lanthanides have moderately high distribution coefficients for apatite and hornblende, less so for clinopyroxene, plagioclase and zircon (Arth, 1976). The variation of these elements is considered in more detail in Section VIII.10.

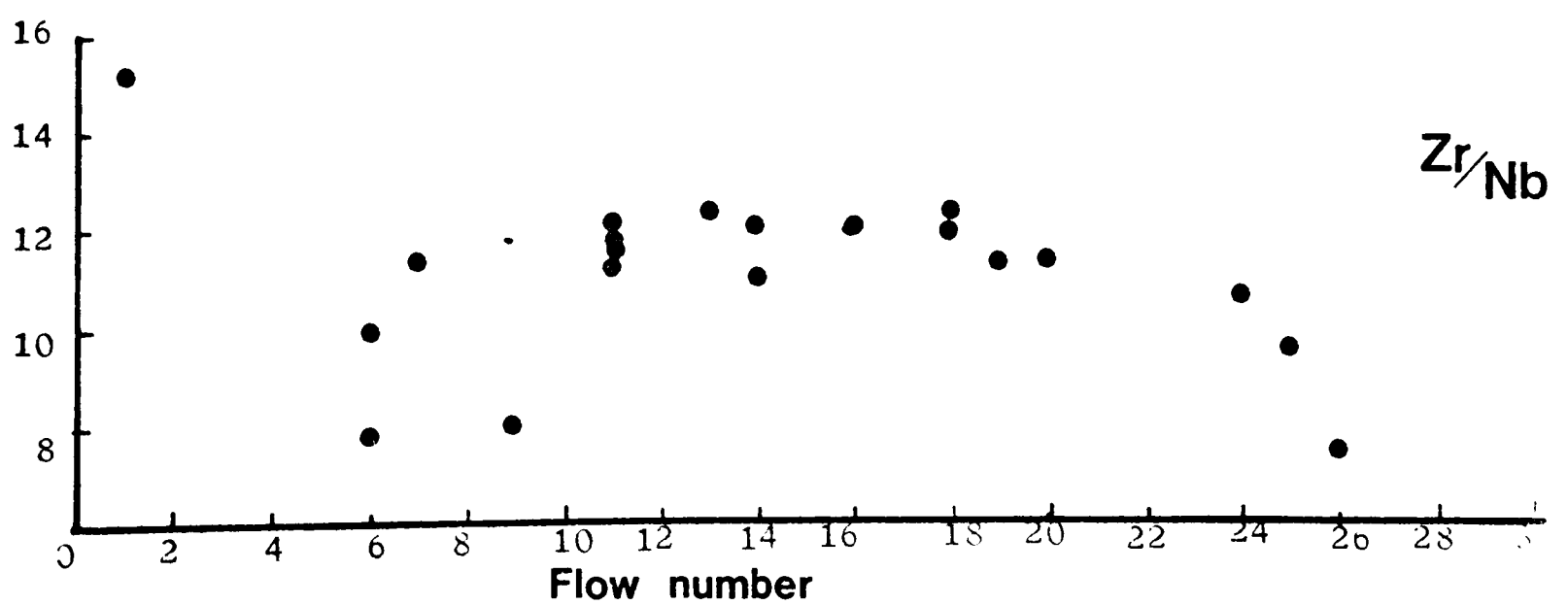
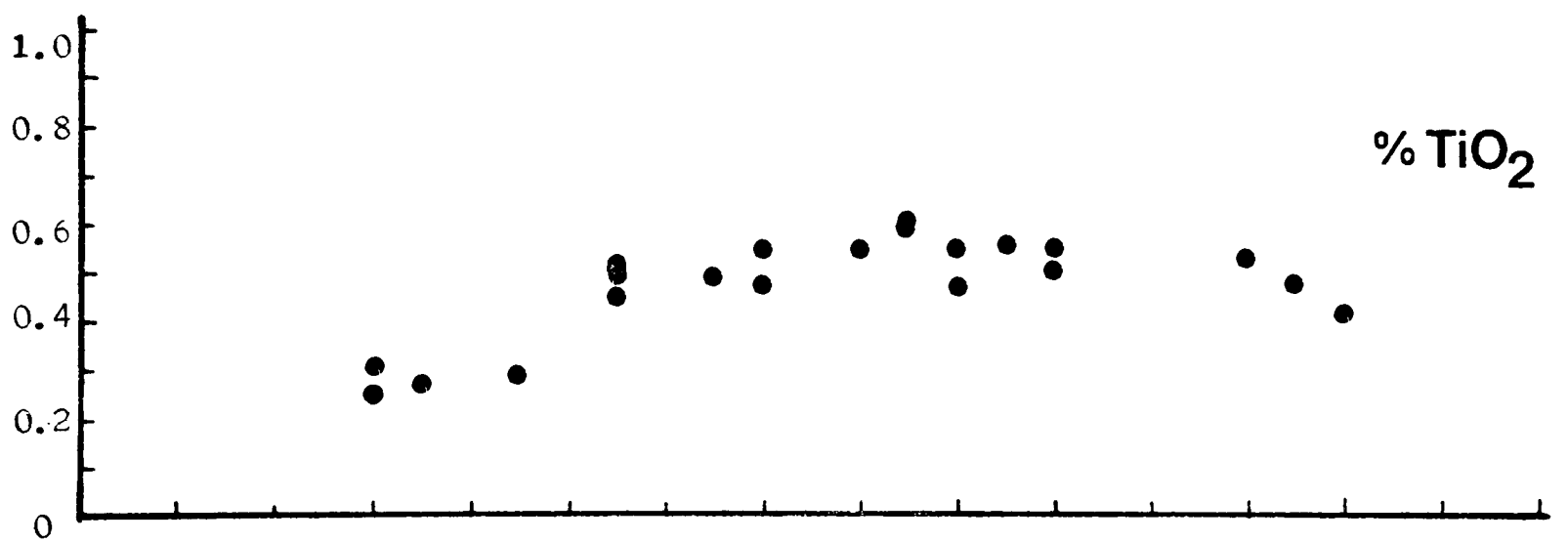
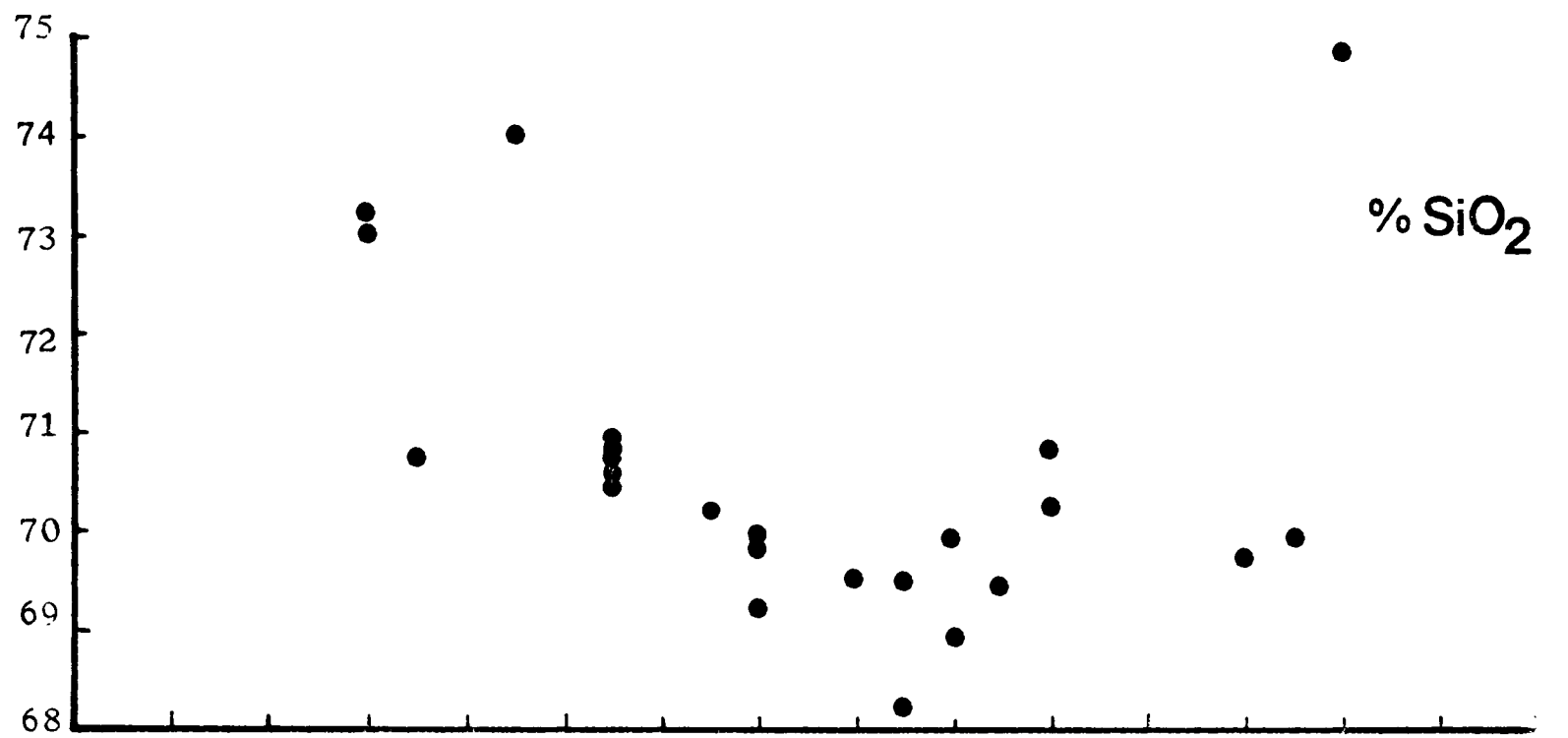
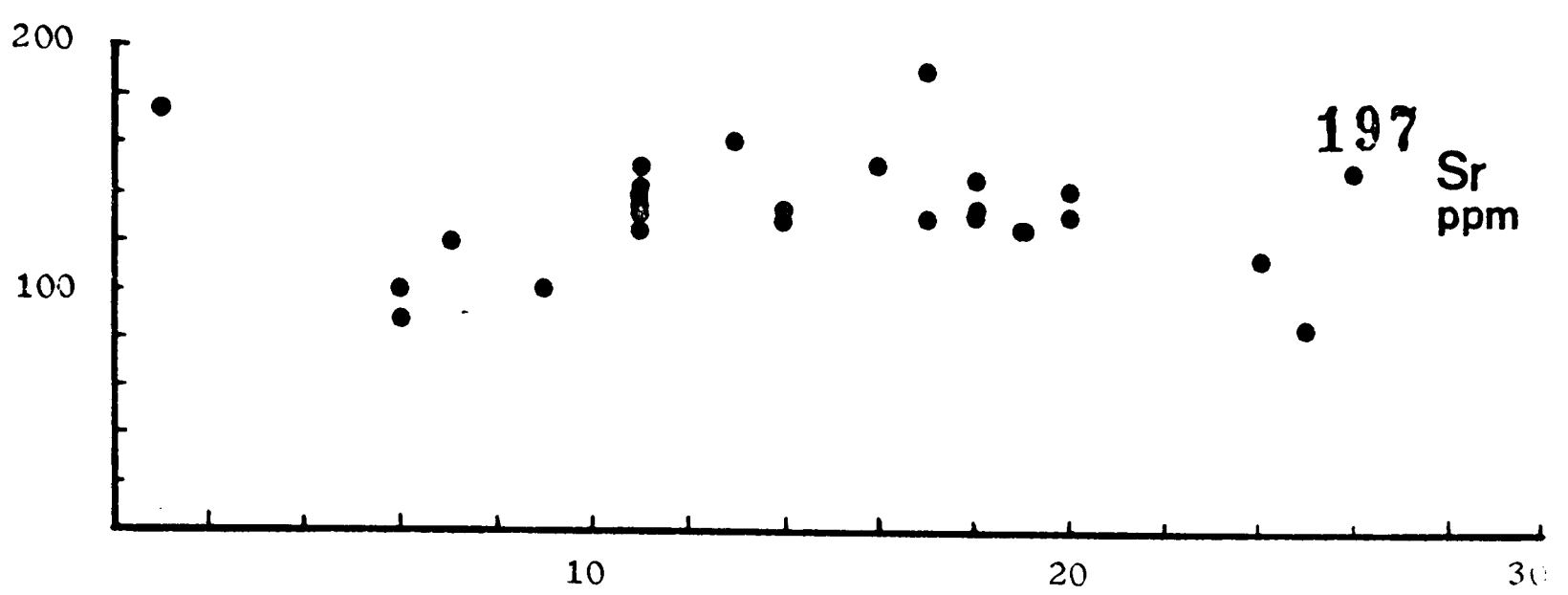
With most of the trace element patterns there is some doubt as to the true trend as the most evolved rhyolites are enriched relative to the granophyres at the same silica level. It is difficult to envisage how fractional crystallisation of a single magma-type can produce residual liquids both enriched and depleted in certain trace elements; processes such as crystal accumulation or contamination, or variations in the melting mechanism or source chemistry may thus be involved in the creation of bimodal trace element distributions.

The two samples of the rhyolitic interbeds have anomalous trace element abundances (except for Rb and Sr) and must be considered to have a separate origin to the main group of rhyolites. Most of the trace elements analysed are resistant to weathering, and the observed depletions may well reflect the chemistry of the source.

VIII.5 Variation with stratigraphic height

Detailed mapping has enabled the flows to be correlated over large distances and samples 30 km apart can be assigned to the same flow from field evidence. Little or no vertical or lateral intraflow variation was observed in the field in Swaziland; the southern Lebombo also shows chemical and mineralogical homogeneity (Bristow, 1976).

Fig. VIII.2. Variation with stratigraphic height of selected elements.
Flow numbers are taken from the maps.



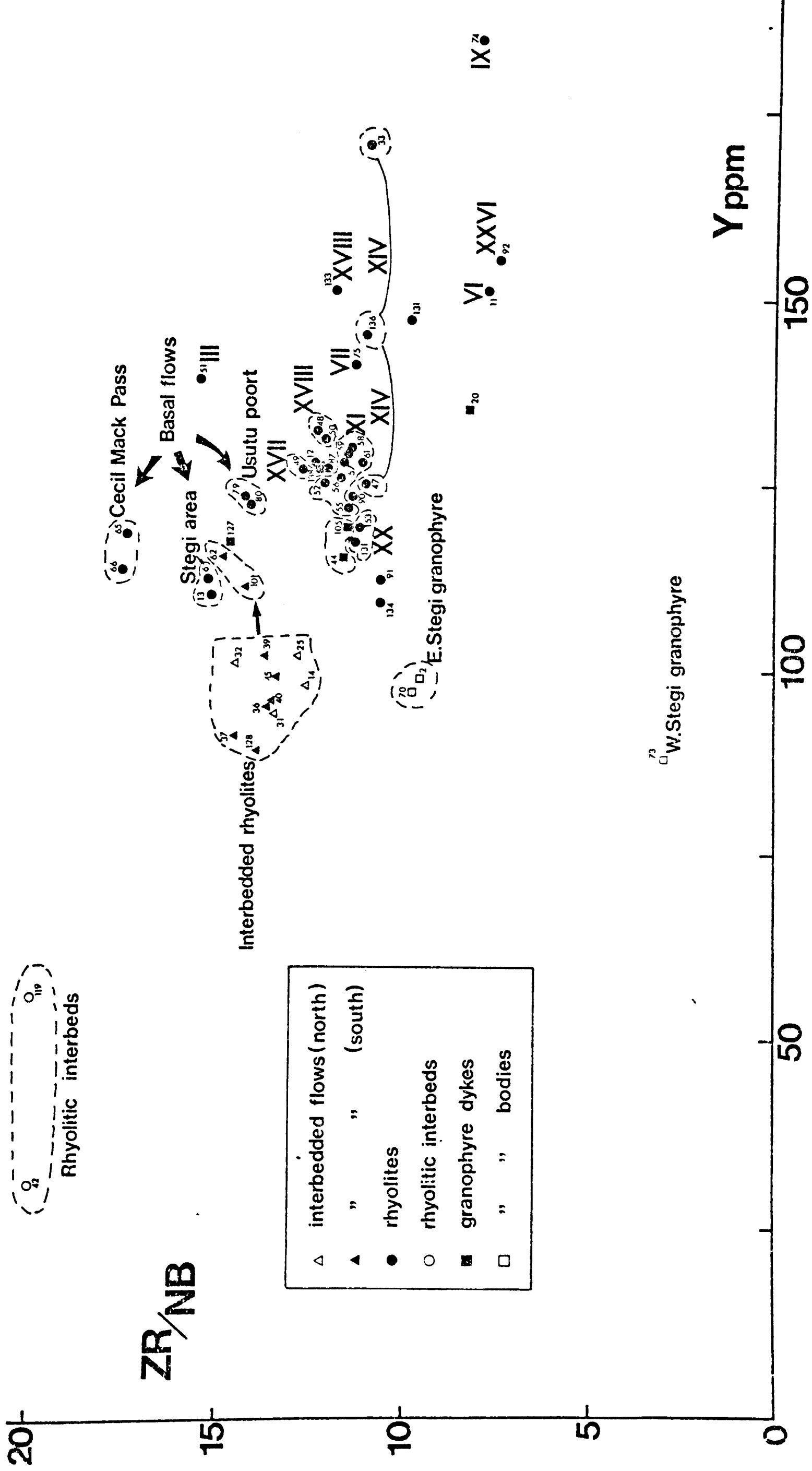
The samples are divided into flows, and their chemical variation with height plotted in Figure VIII.2. Flow XI is represented by seven samples and the data give an indication of the intra-flow variation. Only the elements with tight trends were used so that the small variation with height was not obscured by analytical and petrogenetic noise. Apart from the basal flow which is considerably more basic, the data form a gentle curve with the most basic of the rhyolites in the middle of the succession.

VIII.6 Chemical 'fingerprinting' of acid flows

The chemical 'fingerprinting' of lava flows has been studied using various discriminant analyses by Cann (1970), Pearce and Cann (1973) and Jack and Carmichael (1969). These studies have been concerned with separating and distinguishing rocks from different tectonic environments, or different areas or eruptive centres within one tectonic setting. The problem in the Lebombo is in distinguishing between flows which are part of the same fractionation sequence to find a chemical method of verifying the field subdivisions or, in areas of inadequate field-mapping or exposure, to divide the rocks into separate flow units.

The Zr/Nb ratio has been used by Erlank and Kable (1976) as an index to interpret mantle differentiation processes. Both Zr and Nb have low crystal-liquid distribution coefficients in basic systems and hence, after only small degrees of melting, the Zr/Nb ratio of the melt should be equal, or nearly so, to that of the source. In the Swaziland rhyolites, however, the presence of zircon as a discrete phase and the consequent fractionation of Zr means that only the initial liquids may have a Zr/Nb ratio equivalent to that of the source; the Zr/Nb ratios of liquids affected by zircon fractionation will be indicative of the amount of that fractionation as well as their source chemistry. Zr/Nb is relatively unaffected by weather-

Fig. VIII.3. Zr/Nb-Y plot for the rhyolite data. Sample numbers are given. Broken lines define approximate limits of composition for each flow.



ing and rocks too weathered to merit a chemical analysis may still be usable in 'fingerprinting'

The use of Zr/Nb as a discriminant is shown in Figure VIII.3 where it is plotted against Y. Yttrium was used as the other index principally because there are a number of rocks which have only been analysed for the elements Nb, Zr, Y, Sr and Rb, and is also relatively unaffected by weathering. This plot can, ideally, discriminate between liquids formed from sources with differing Zr/Nb or the rare earth characteristics, and also between rocks on a crystal fractionation trend which have suffered differing amounts of zircon fractionation.

On the Zr/Nb-Y plot the rhyolite analyses fall in a broad zone with negative slope. The rhyolitic interbeds with high Zr/Nb ratios and low Y contents fall away from the bulk of the analyses. In nearly all cases samples from the same flow occur in well-defined areas; the seven analyses from flow X1 give an indication of the precision of the discriminant. The interbedded rhyolites fall into two groups, separated by their Y content. The flows from the northern and southern areas fall in the same group, thus supporting the tentative correlation based on field evidence, while the uppermost of the interbedded flows (RC 62, 101) are more similar to samples from the base of the rhyolite sequence proper (SK 13, MZ 6). The main body of rhyolites form a small grouping of data points with a broad left-to-right division into flow units. The basal flows at Cecil Mack Pass, Usutu Poort and the Stegi area have significantly higher Zr/Nb ratios than the main group. The granophyre dykes are indistinguishable from rhyolites whereas the granophyre bodies are lower in Zr/Nb and Y and may be more fractionated. Three, or maybe four, of the rhyolites and one of the granophyre dykes are enriched in Y and have low Zr/Nb ratios; this enrichment in Y relative to Zr is later shown to be a reflection of the source composition.

The value of this 'fingerprinting', while supporting the field subdivision of flows, is in identifying the principal groups of rocks and being able to distinguish between flows from different areas. The precision is, however, insufficient to enable the flows to be subdivided purely on the basis of Zr/Nb-Y.

VIII.7 Dolerite chemistry

The data for dolerites are unfortunately limited but, hopefully, representative of the dolerite swarm. Virtually all the basalt in Swaziland is deeply weathered, except where baked by dolerite or granophyre intrusions, and only the fresh dolerites were selected for analysis. The dolerites and basalts are assumed to be cogenetic and the petrogenetic conclusions reached for the dolerites will be assumed, in the absence of further data, to be applicable to the basalts.

The major and minor element data for the Swaziland dolerites are summarized on Figure VIII.4. All the rocks are quartz-tholeiites, containing between 1 and 6 per cent of normative quartz. Most of the variation diagrams show linear trends: CaO and Al_2O_3 have positive correlations and Na_2O , K_2O , P_2O_5 and the trace elements negative correlations with MgO. Fe (total) and TiO_2 have kinked trends, both with maxima around 5.5% MgO, possibly representing the onset of ore fractionation.

Regression lines for the incompatible elements (i.e. K_2O , P_2O_5 , Zr, Nb, Y, Rb and TiO_2) have intersections with MgO between 7.3 and 7.8 (Table VIII.3). The order of the elements reflects the relative sizes of their distribution coefficients and the amount in the extract. The maximum MgO value for the extract to have positive values for all the trace elements is 7.2% (for Rb). The values for the major elements at that MgO content

Fig. VIII.4. Variation diagrams for Swaziland dolerites. Best fit trend lines are given where possible.

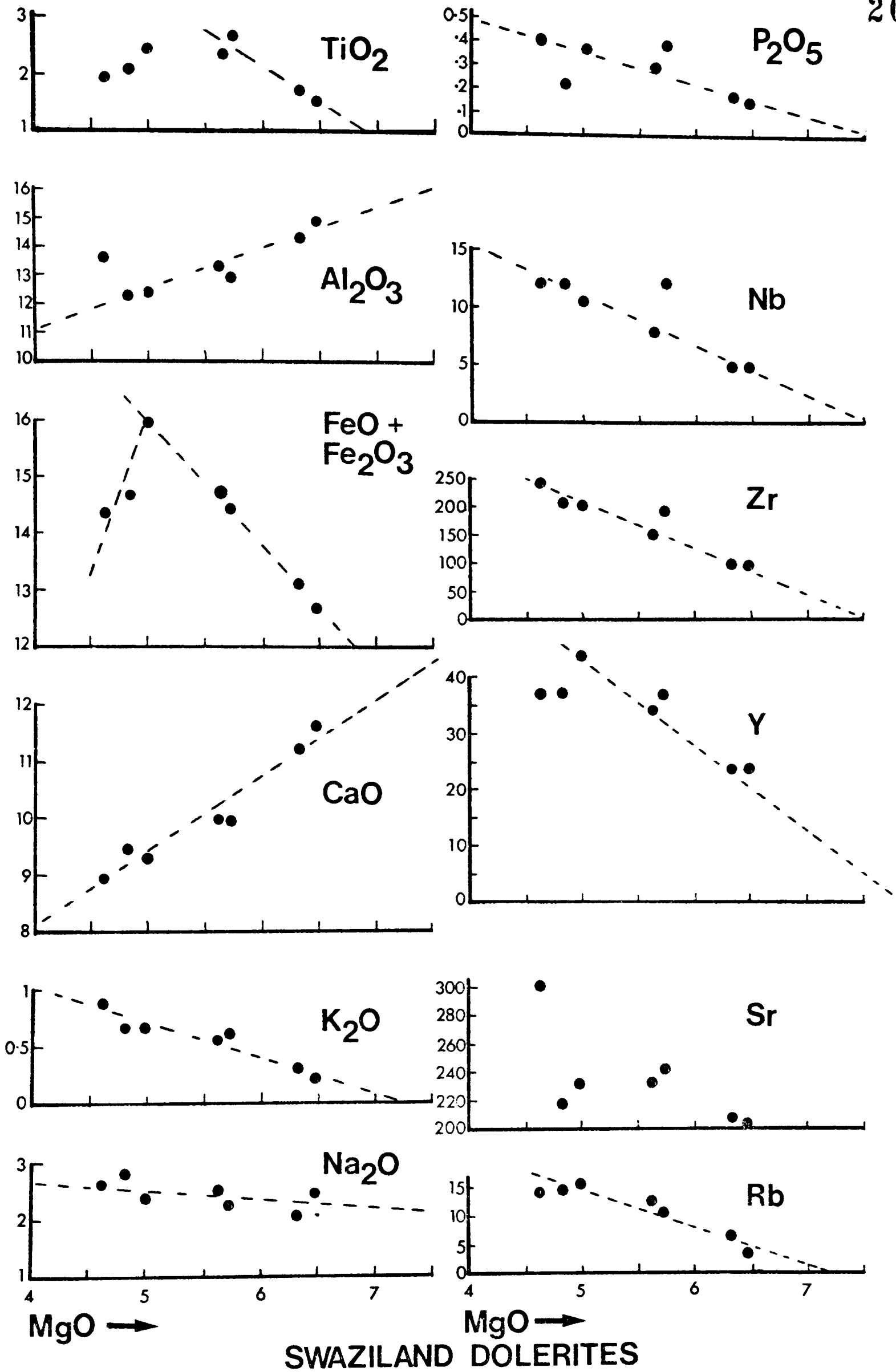


TABLE VIII.3

Extract composition for dolerites

<u>Intersections of regression lines with MgO</u>		<u>Composition of extract for 7-2% MgO</u>	
Element	% MgO	SiO ₂	50.0 %
Rb	7.2	TiO ₂	0.8
K ₂ O	7.25	Al ₂ O ₃	15.6
Zr	7.5	MgO	7.2
Nb	7.5	MnO	0.2
P ₂ O ₅	7.6	CaO	12.5
Y	7.8	P ₂ O ₅	0.02
TiO ₂	7.9	Na ₂ O	2.2
		K ₂ O	0.0
		FeO (tot)	11.2
		Total	99.1

Mineral composition of extract

<u>a) Normative minerals</u>		<u>b) Modal minerals</u> (calculated by least squares fitting)	
Q	0.0 %	Labradorite	49%
Ab	18.6	Clinopyroxene (CP 11)	44%
An	33.2	Ore (OR 11)	3%
Di	23.7	Olivine (Fo 80)	3%
Hy	16.9		
Ol	4.0		
Mt	2.17		
Il	0.4		

are given in Table VIII.3 together with the calculated mineral composition of the extract. The modal proportions of this calculated extract are consistent with the observed mineralogy and it is concluded, on the basis of the available data, that the chemical variation can be explained by fractionation of an approximately 50:50 mixture of plagioclase and clinopyroxene.

VIII.8 Petrogenesis : general

The Lebombo rhyolites have been shown to be a relatively coherent group; the thickness of the succession, uniformity of rock-type and geochemical features imply a similar origin for all the samples analysed (bar the two rhyolitic interbeds). Any theories for their genesis must take into account the intimate association between basalt and rhyolite, and the presence of underlying continental crust.

Petrogenetic hypotheses to account for the rhyolites and their variation can be classed into the following categories:

- (1) Fractional crystallisation of a dacitic magma, the latter produced by
 - (a) fractional crystallisation of a basaltic magma, or
 - (b) partial melting of basaltic source material, or
 - (c) partial melting of mantle source material;
- (2) Varying degrees of partial melting of a basic parent;
- (3) Partial melting of acid crustal materials, with or without subsequent fractionation.

Direct derivation from the mantle will not be considered here as (a) it is very unlikely that the quartz-bearing rhyolites were ever in equilibrium with a mantle assemblage (Stern et al., 1975), and (b) if they formed by differentiation of a magma produced from the mantle the situation becomes analogous to case 1(a), (b) or 2 above and need not be considered separately as an origin direct from the mantle.

Of the three mechanisms outlined above, case 1 is considered the most likely and will be discussed in detail in the following sections. Cases 2 and 3 will be discussed in terms of the trace element data in Sections VIII.13 and 14 respectively.

VIII.9 Major element modelling

(a) Method

A least-squares approximation method (after Bryan et al., 1969) has been used to model the observed major element trends. A Fortran IV program 'Mixer' (written by A.R. Duncan and adopted to run at Oxford by P. Betton) based on this method was used for the computation.

The objective of extract calculations is to determine whether an assumed residual liquid might be derived from an assumed parent magma by removal or addition of components which may be mineral phases or 'whole-rock' components such as other magmas, wall-rock or xenolithic material. The variation of two, three or four components can be depicted graphically but for multi-dimensional analysis where all components are considered simultaneously the least-squares method can be used.

The program works in the reverse way to Nature in that it mixes the residual liquid and mineral phases (the extract) to approximate the starting (or

bulk) composition. The method is also suitable for the calculation of moes, given the mineral and whole-rock compositions.

Consider a system of n oxides and k components where x_{ij} is the concentration of the ith oxide in the jth component; y is the observed amount of the ith oxide in the composition to be approximated and $\hat{y}_i$ the calculated amount; b_j is the estimated amount of component j required.

For each of the n oxides a linear equation can be written:

$$y_i = \sum_{j = 1,k} x_{ij} \times b_j \tag{1}$$

or in matrix notation

$$Y = XB \tag{2}$$

or

$$\begin{vmatrix} y_1 \\ y_2 \\ . \\ . \\ . \\ . \\ y_n \end{vmatrix} = \begin{vmatrix} b_1 \\ b_2 \\ . \\ . \\ . \\ b_k \end{vmatrix} \times \begin{vmatrix} x_{11} & x_{12} & x_{13} & . & . & . & x_{1k} \\ x_{21} & x_{22} & x_{23} & . & . & . & x_{2k} \\ . & . & . & & & & . \\ . & . & . & & & & . \\ . & . & . & & & & . \\ . & . & . & & & & . \\ x_{n1} & x_{n2} & x_{n3} & . & . & . & x_{nk} \end{vmatrix}$$

provided $n > k$ a unique solution can be obtained by matrix inversion for B and by substitution in equation (2) a vector $\hat{y}$ can be obtained which is the least squares approximation of y such that the quantity:

$$\sum R^2 = \sum_{i=1}^n (y_i - \hat{y}_i)^2$$

the sum of the squares of the residuals, is minimised.

For the modelling involved in this work the vector equation (2) can be rephrased as:

where:

$$\begin{aligned} \text{Liq Y}_{\text{calc}} &= x \text{ Liq X} + aA + bB + cC \dots \\ \text{Liq Y} &= \text{parental magma (assumed and recalculated)} \\ \text{Liq Y} &= \text{residual magma} \\ x, a, b, c &= \text{proportions of components X, A, B, C} \dots \text{ such that} \\ &x + a + b + c \dots = 1. \end{aligned}$$

The number of oxides, n , in general is a maximum of 9 and the number of components involved varies from four to eight depending on the conditions specified.

This method is only a simple mixing model and the results must be treated subjectively. Magma variation will only be linear over small amounts of fractionation; more extensive fractionation should be considered as a number of smaller steps.

The squared residuals are an indication of the goodness of fit of the data but the absolute values of the residuals are not necessarily an indication of the relative quality of fit as large absolute discrepancies in the value of SiO_2 may only be small relative errors, and small discrepancies in MnO , P_2O_5 , etc., may be large percentage errors. In general, for 9 oxides

$\sum R^2$ values greater than 5 indicate an inadequate fit, and less than 1 indicate a good fit to the data. The sum of the percentage errors, $\sum \%R^2$, is generally less than 100 for a good fit; an average relative deviation of 5% for 9 oxides would give a $\sum \%R^2$ of 225.

It must be remembered, however, that an excellent fit to the data means that the observed variation is consistent with crystal fractionation but not that that fractionation occurred.

(b) Results of the modelling

To assess the variation within the rhyolite sequence, three representative whole-rock compositions were selected representative of the acid, basic and intermediate groups of rhyolites respectively. Real rock compositions were used rather than those calculated from regression lines because of the difficulty of defining the fractionation trend towards the basic end of the rhyolites. For the computation the data were normalised to the dependent vector (i.e. the starting composition).

The extract minerals are assumed to be those present as phenocrysts in the rocks. The phenocryst data (Table VIII.2) were chosen as being typical of the phases present and were analysed in the freshest rocks available. The analyses are thought to resemble as closely as possible the actual fractionating phenocrysts.

The modelling is considered in two halves:

- 1) Basic rhyolite (RC 40) to intermediate rhyolite (RC 52).

The best fitting solution is summarised in Table VIII.4.

Using pure magnetite and quartz as part of the extract a good fit can be obtained. Inspection of the data, however, reveals

that 6% quartz needs to be added if, in the fractionation sequence, RC 52 is derived from RC 40. The variation diagrams (Diagram 5) show that, relative to RC 52, RC 40 is enriched in Al_2O_3 and K_2O and it seems doubtful that the two rocks are related by simple crystal fractionation.

2) Intermediate (RC 52) to acid rhyolite (RC 73).

A good fit is obtained between RC 52 and RC 73 using an average feldspar composition from RC 52. Plagioclase, clinopyroxene and ore (titanomagnetite) are the main phases fractionating, being extracted in the ratio 7:2:1. Quartz and magnetite are present in minor amounts, to produce an iron and silica balance. The computation was repeated using the most basic of the main rhyolite group: RC 65 (Table VIII.4). The fit of the data is good though with a discrepancy in Al_2O_3 which may be due to the plagioclase composition used; the fit may be also improved by altering the weighting of the data.

The various mixing models employed show that the observed variation within the main group of rhyolites can be approximated by crystal fractionation of a mixture of 65% plagioclase, 15% clinopyroxene and 20% ore, magnetite and quartz (values averaged over the whole fractionation range). The interbedded dacites can not be interpreted in a similar fashion and contain K_2O and Al_2O_3 in excess of that required by the crystal fractionation model.

TABLE VIII.4

1. Fractionation of basic to intermediate rhyolite (RC 40 - RC 52).

	RC 40 observed	RC 40 calculated			
SiO ₂	66.78	66.37	% components		% of extract
TiO ₂	0.75	0.70			
Al ₂ O ₃	13.31	13.05	RC 52	96.8	0
FeO*	5.78	5.78			
MnO	0.14	0.16	PL 36	3.5	50
MgO	0.64	0.58	CP 36	2.1	30
CaO	2.48	2.61	OR 36	1.2	18
Na ₂ O	3.78	3.54	QZ	-5.6	
K ₂ O	4.33	4.26	MT	-0.8	

$$\sum R^2 = 0.32$$
$$\sum \%R^2 = 460$$

2. Fractionation of intermediate to acid rhyolite (RC 52 - RC 73).

	RC 52 observed	RC 52 calculated			
SiO ₂	70.98	71.20	% components		% of extract
TiO ₂	0.52	0.54			
Al ₂ O ₃	12.58	13.13	RC 73	75.5	0
FeO*	5.41	5.41			
MnO	0.13	0.12	PL 55	15.9	60
MgO	0.36	0.38	CP 55	4.8	18
CaO	2.08	2.03	OR 55	2.4	9
Na ₂ O	3.41	3.41	MT	1.1	4
K ₂ O	4.37	4.21	QZ	2.1	8

$$\sum R^2 = 0.38$$
$$\sum \%R^2 = 143$$

VIII.10 Trace element modelling

(a) Introduction

Detailed treatments of the theory of trace element partitioning have been presented by McIntire (1963), Banno and Matusi (1973), Wood and Fraser (1976) and O'Nions and Powell (1977) and only a brief summary of relevant aspects is attempted here.

Theoretical models for trace element behaviour assume that trace elements form dilute solid solutions and behave in accordance with Henry's Law. This assumption is invalid, however, where a trace element is a stoichiometric constituent of a phase (e.g. zirconium in zircon). The assumption of equilibrium, or near equilibrium, conditions is also made.

At equilibrium the chemical potential, μ , of component i in phase α equals that of i in β :

$$\mu_i^\alpha = \mu_i^\beta \quad (1)$$

Expressed in terms of the activity, a , and the standard state chemical potential, μ^0 , this becomes:

$$\mu_i^{0\alpha} + RT \ln a_i^\alpha = \mu_i^{0\beta} + RT \ln a_i^\beta \quad (2)$$

or

$$\frac{a_i^\beta}{a_i^\alpha} = \exp \frac{\mu_i^{0\alpha} - \mu_i^{0\beta}}{RT} \quad (3)$$

This expression is a constant at given P, T and composition. Expressing the

activity in terms of the activity coefficient, γ , and the concentration, c

$$\frac{\gamma_i^\beta \cdot C_i^\beta}{\gamma_i^\alpha \cdot C_i^\alpha} = k \quad (4)$$

Henry's Law states that, for dilute solutions, the activity of a component is directly proportional to its concentration and hence, the activity coefficient is independent of concentration and is a constant at given pressure and temperature. Equation (4) can thus be restated:

$$\frac{C_i^\alpha}{C_i^\beta} = \frac{\gamma_i^\beta}{k\gamma_i^\alpha} = D_i^{\alpha/\beta} \quad (5)$$

where $D_i^{\alpha/\beta}$ is the distribution coefficient for i between phases α and β .

In practice, many trace elements are present in abundances in excess of the Henry's Law region and distribution coefficients are composition, as well as pressure and temperature, dependent. Within a given fractionation sequence however, the pressure and temperature cannot be expected to vary greatly and for moderate amounts of fractionation the variation in D should be small.

Much of the composition and P, T dependence of distribution coefficients can be eliminated by considering pairs of similar elements (e.g. $D_{Ba}^{\alpha/\beta}/D_{Sr}^{\alpha/\beta}$) as the activity coefficients for the two elements will be similar and cancel out of equation 4. McIntire (1963) discussed these compound distribution coefficients in detail; their potential usefulness is outweighed by the difficulty of their use in partial melting and crystallisation equations.

(b) Theoretical models for trace element behaviour(i) Partial Melting

Quantitative models for trace element distribution during partial melting have been devised by Gast (1968), Shaw (1970) and Hertogen and Gijbels (1976) and summarised by Arth (1976). Three basic mechanisms are proposed:

- 1) Batch melting: continuous equilibration between liquid and solid until the removal of melt from residual solid;
- 2) Fractional melting: continuous removal of melt from solid;
- 3) Integrated fractional melting: continuous removal of melt followed by collection and homogenisation in a single magma reservoir.

The three models produce similar results for small degrees of melting but not for larger amounts. Models (2) and (3) require the extraction of infinitesimal increments of melt, a process thought by Arth (1976) to be unlikely in a geological context. Only batch melting will be considered here.

BATCH EQUILIBRIUM MELTING

For partial melting where the liquid remains in equilibrium with the solid until it is removed, for a given element (Shaw, 1970):

$$\frac{C^L}{C^0} = \frac{1}{D^0 + F(1-P)} \quad (6)$$

where C^L = concentration of element in the liquid;

C^0 = initial concentration of element in solid;

F = fraction of melting;

D^0 = bulk partition coefficient in a multiphase system. For an n phase assemblage:

$$D^0 = \sum_{j=1,n} D^{j/l} \cdot X_0^j$$

where $D^{j/l}$ = crystal/liquid distribution coefficient for phase j ;

and X_0^j = initial weight fraction of j in the solid;

$$\text{and } P = \sum_{j=1,n} p^j \cdot D^{j/l}$$

where p^j = fraction of liquid contributed by phase j .

In the unlikely, but easily modelled, situation where modal melting takes place and the percentage of each phase in the solid does not change during melting, $P = D^0$ and equation 6 becomes:

$$\frac{C^L}{C^0} = \frac{1}{D^0 + F(1-D^0)} \quad (7)$$

If D^0 or $F \rightarrow 0$, then $C^L/C^0 \rightarrow \frac{1}{F}$ or $\frac{1}{D^0}$ respectively.

(ii) Fractional Crystallisation

EQUILIBRIUM FRACTIONAL CRYSTALLISATION

For equilibrium fractional crystallisation the whole of the solid portion is kept in equilibrium with the melt and

$$\frac{C^L}{C^0} = \frac{1}{F + D^0(1-F)} \quad (\text{Arth, 1976}) \quad (8)$$

where F = fraction of liquid remaining;

D^0 = bulk distribution coefficient for solid phases;

C^0 = initial concentration of element in liquid;

C^L = concentration in differential liquid.

This equation is analogous to the batch melting model considered above.

RAYLEIGH FRACTIONAL CRYSTALLISATION

In Rayleigh fractional crystallisation, only the outer part of the crystals is kept in equilibrium with the melt and

$$\frac{c^L}{c^0} = F(D^0 - 1) \quad (9)$$

symbols as for equation (8).

For both the above models, as $D \rightarrow 0$, $\frac{c^L}{c^0} \rightarrow \frac{1}{F}$.

More complex models involving cyclical crystallisation and replenishment have been proposed by O'Hara (1977).

(iii) Discussion

All the theoretical models outlined above assume that the bulk distribution coefficient remains constant - i.e. both the distribution coefficients and the amounts of the phases involved are constant. In all cases the maximum possible enrichment occurs when $D = 0$; enrichment in excess of this implies non-equilibrium or open system processes.

The values employed for distribution coefficients are critical for the successful application of these theoretical equations. Several techniques have been used to measure phenocryst/liquid partition coefficients including microprobe studies of doped experimental systems, the use of radioactive tracers and the measurement of phenocryst-groundmass compositions in volcanic rocks. Major problems include the zonation of natural crystals and the doping of experimental systems in excess of the Henry's Law region.

TABLE VIII.5

Average crystal/liquid distribution coefficients for minerals in basic, intermediate and acid rock types (from compilation by Arth (1976))

	<u>Basic Rocks</u>		<u>Intermediate Rocks</u>	
	Augite	Hornblende	Plagioclase	Plagioclase
Rb	0.031	0.29	0.071	0.048
Sr	0.12	0.46	1.83	2.84
Ba	0.026	0.42	0.23	0.36
Ce	0.15	0.20	0.12	0.24
Nd	0.31	0.33	0.081	0.17
Dy	0.68	0.64	0.055	0.086
La*	0.1	0.15	0.15	0.3

Rhyolitic Rocks

	Plagioclase	Clinopyroxene	Hornblende	K-feldspar	Apatite	Zircon
Rb	0.041	0.032	0.014	0.34	-	-
Sr	4.4	0.516	0.022	3.87	-	-
Ba	0.308	0.131	0.044	6.12	-	-
Ce	0.27	0.50	1.52	0.044	34.7	2.64
Nd	0.21	1.11	4.26	0.025	57.1	2.20
Dy	0.064	1.93	13.0	0.006	50.7	45.7
La*	0.3	0.3	0.8	0.06	30	3

*Data for La extrapolated from Ce and Nd

Rb, Sr, Ba and the rare earth elements are the best documented trace elements for distribution coefficients and the values used in this study are taken from the compilation by Arth (1976) and given in Table VIII.5.

(c) Application of trace element models

Consider a pair of trace elements, A and B, involved in Rayleigh fractionation. From equation (9) it follows that for A:

$$\frac{C_A^L}{C_A^O} = F^{(D_A - 1)}$$

and similarly for B

$$\frac{C_B^L}{C_B^O} = F^{(D_B - 1)}$$

Hence:

$$\log C_A^L - \log C_A^O = (D_A - 1) \log F$$

and:

$$\log C_B^L - \log C_B^O = (D_B - 1) \log F.$$

By dividing and eliminating F

$$\frac{\log C_A^L - \log C_A^O}{\log C_B^L - \log C_B^O} = \frac{D_A - 1}{D_B - 1}$$

or

$$\log C_A^L = \frac{D_A - 1}{D_B - 1} \log C_B^L + \text{constant} \quad (10)$$

where constant =

$$\log C_A^O - \frac{D_A - 1}{D_B - 1} \log C_B^O$$

Thus, on a plot of $\log C_A^L$ against $\log C_B^L$ a mechanism of Rayleigh fractionation will produce a straight line of gradient $\frac{D_A - 1}{D_B - 1}$. The ratio $\frac{D_A - 1}{D_B - 1}$

can now be determined and if either D is known the other can be calculated relative to it for that fractionation sequence. A straight line plot of data on a log-scale is compatible with Rayleigh fractionation and incompatible with equilibrium crystallisation, partial melting or mixing models.

Allegre et al., (1977) adopted a similar technique but used an element with $D = 0$ to calculate F directly. In these acid rocks, however, none of the elements is wholly incompatible and a slightly more complex model is justified.

There are no aphyric rocks present in Swaziland and, in the absence of groundmass analyses, whole rock compositions have been used for the modelling. The percentage of phenocrysts is fairly constant (15-25%) and it is hoped that the effects of any crystal accumulation will be small and constant over the whole suite. Where closed-system equilibrium crystallisation has taken place the whole-rock composition will be representative of a liquid composition.

Rb and Ba are plotted against Sr in Figure VIII.5. In both cases the data can be approximated to a straight line and are thus consistent with Rayleigh

Fig. VIII.5. Log-plot of Rb and Ba against Sr.

(a) Rb vs. Sr.

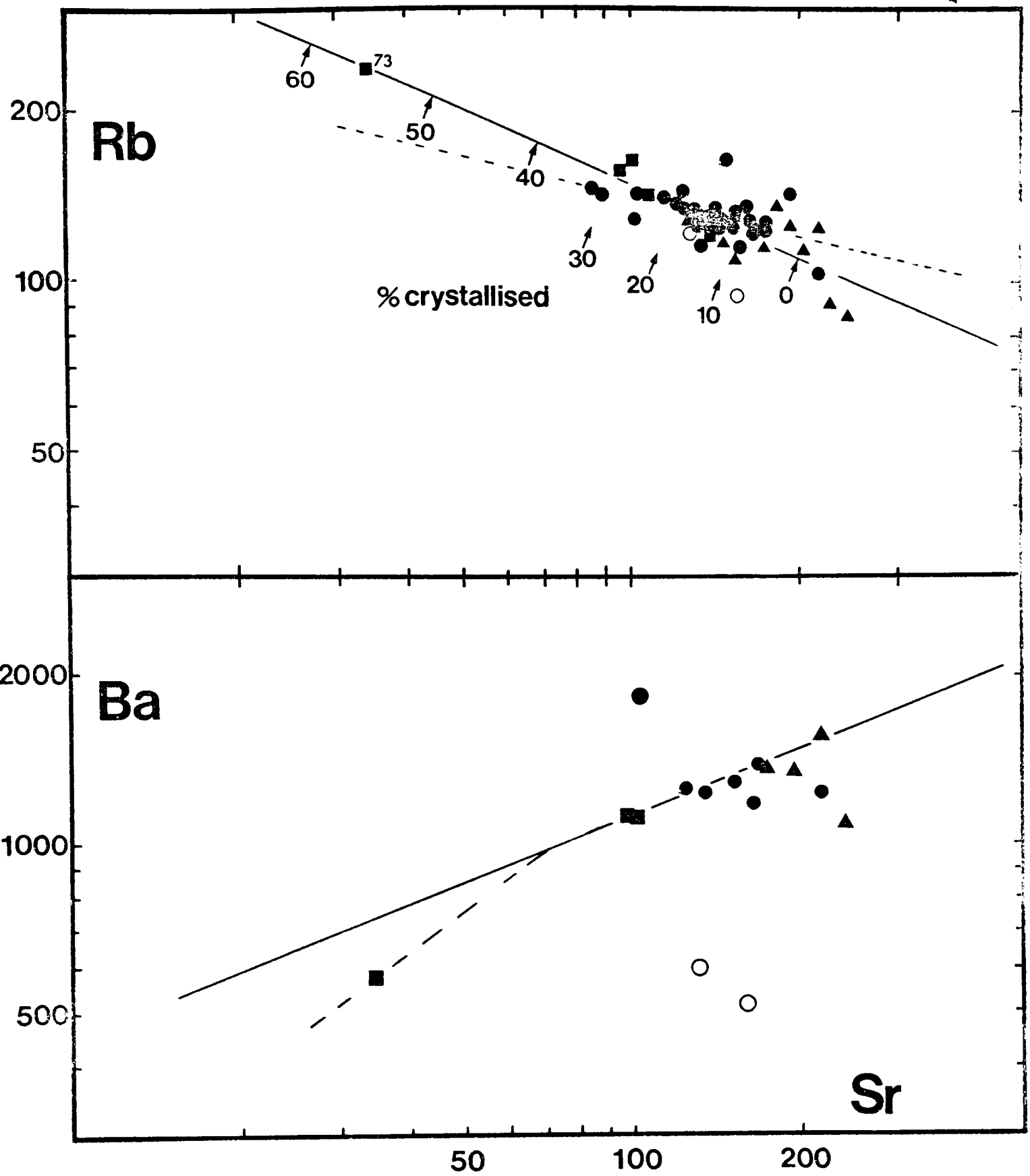
The solid line is a best fit line through all the data, with the exception of the most basic, high Sr rocks. The dotted line represents a grouping of data oblique to the main trend; the freshest samples fall on the solid line and this secondary trend is not thought to be significant. The trend is contoured for Rayleigh fractionation assuming $D_{Rb}^0 = 0.03$, from an arbitrary starting point of $Sr = 200$, $Rb = 100$.

(b) Ba vs. Sr.

An analogous plot to Rb vs. Sr. The most basic rhyolites fall below the line and belong to a more basic fractionation sequence; the most evolved sample may have suffered Ba depletion by fractionation of potassium feldspar.

(Symbols for all figures:

squares = granophyre; solid circles = rhyolites; triangles = interbedded rhyolites; open circles = rhyolitic interbeds)



fractionation. The high Rb and Ba points are ignored and also the most basic, highest Sr, points which may belong to a more basic fractionation trend. The Rb-Sr plot shows some scatter, partly due to the presence of weathered samples; the freshest rocks, defining the Ba-Sr trend, all fall or near the Rb-Sr line and the secondary trend in Rb-Sr (marked by broken line) is not thought significant.

Published distribution coefficients are summarised in Table VIII.5. D_{Rb} remains low and fairly constant irrespective of composition whereas D_{Sr} is more variable and D_{Sr}^{plag} increases from 1.83 to 4.4 between acid and basic rocks. Published values for D_{Ba} are also fairly constant.

Using the relationship formulated in equation 10, the data can be tested for self-consistency assuming a Rayleigh fractionation model. The distribution coefficient data for Rb seem the most reliable; taking the extract composition calculated from the major element data and the average distribution coefficient data from Table VIII.5 a value of 0.03 can be calculated for D_{Rb}^0 , the bulk distribution coefficient for Rb. The gradient in Figure VIII.5(a) is -0.4304 and, relative to $D_{Rb}^0 = 0.03$, $D_{Sr}^0 = 3.25$. This value is slightly higher than that calculated from published data (= 2.9) but a value for D_{Sr}^{plag} of 4.9, within experimental error, would produce a bulk coefficient of 3.25. Probe data on feldspars from these rocks (Betton, pers. comm.) give values for Sr in plagioclase of 120 - 800 ppm which, when compared with whole-rock analyses, give an average distribution coefficient of 4. Similarly, D_{Ba}^0 can be calculated to equal 1.9; published data give a value of 0.22 but the feldspars contain 1500 - 3000 ppm Ba (Betton, pers. comm.) and the value of $D_{Ba}^{plag} = 2.9$ required to give a $D_{Ba}^0 = 1.9$ (assuming $D_{Ba}^{Cpx} = 0.13$) is compatible with the probe analyses.

Using the internally consistent bulk distribution coefficients above, the

Rb-Sr plot has been contoured with the amount of crystallisation relative to an arbitrary starting composition of Sr = 200 ppm, Rb = 110 ppm. Thirty percent fractionation is sufficient to account for the variation in the main body of analyses and 55% for the most evolved composition. The major element data give values of 22% and 30% in the two cases. For the main group of rhyolites, therefore, the major element and trace element data are in reasonable agreement, but for RC 73 the trace element enrichment is in excess of that predicted by the major elements and eutectic crystallisation may have taken place to enrich the trace elements without greatly affecting the major elements. For Ba-Sr the same contours of F will apply; RC 73 falls below the main trend and may have suffered late-stage fractionation of a Ba-rich phase such as alkali-feldspar.

Y, Nb, La, Ce and Nd have been plotted against Sr (being the trace element with greatest variance) in Figure VIII.6. If the most basic points are ignored for belonging to a more basic fractionation trend, and the most enriched rocks for belonging to a different trend, the data, in general define reasonable straight lines. The gradients of these lines and the calculated distribution coefficients are shown in Table VIII.6.

The diagram for Yttrium shows some scatter but two possible trends can be drawn: (1) represents a general enrichment of Y from the basic to the most evolved rocks; (2) shows the depletion of Y in the granophyres relative to the main group of rhyolites. The first of these corresponds to a negative value for D_Y^0 and represents an enrichment of Y relative to Sr greater than that possible by Rayleigh fractionation from a single magma. The few rocks showing these enrichments do not seem to be cumulus enriched and the anomaly is interpreted as enrichment of Y in the source, reflected in the derived magmas.

Dots = rhyolites; squares = granophyres; triangles = interbedded rhyolites; open circles = rhyolitic interbeds.

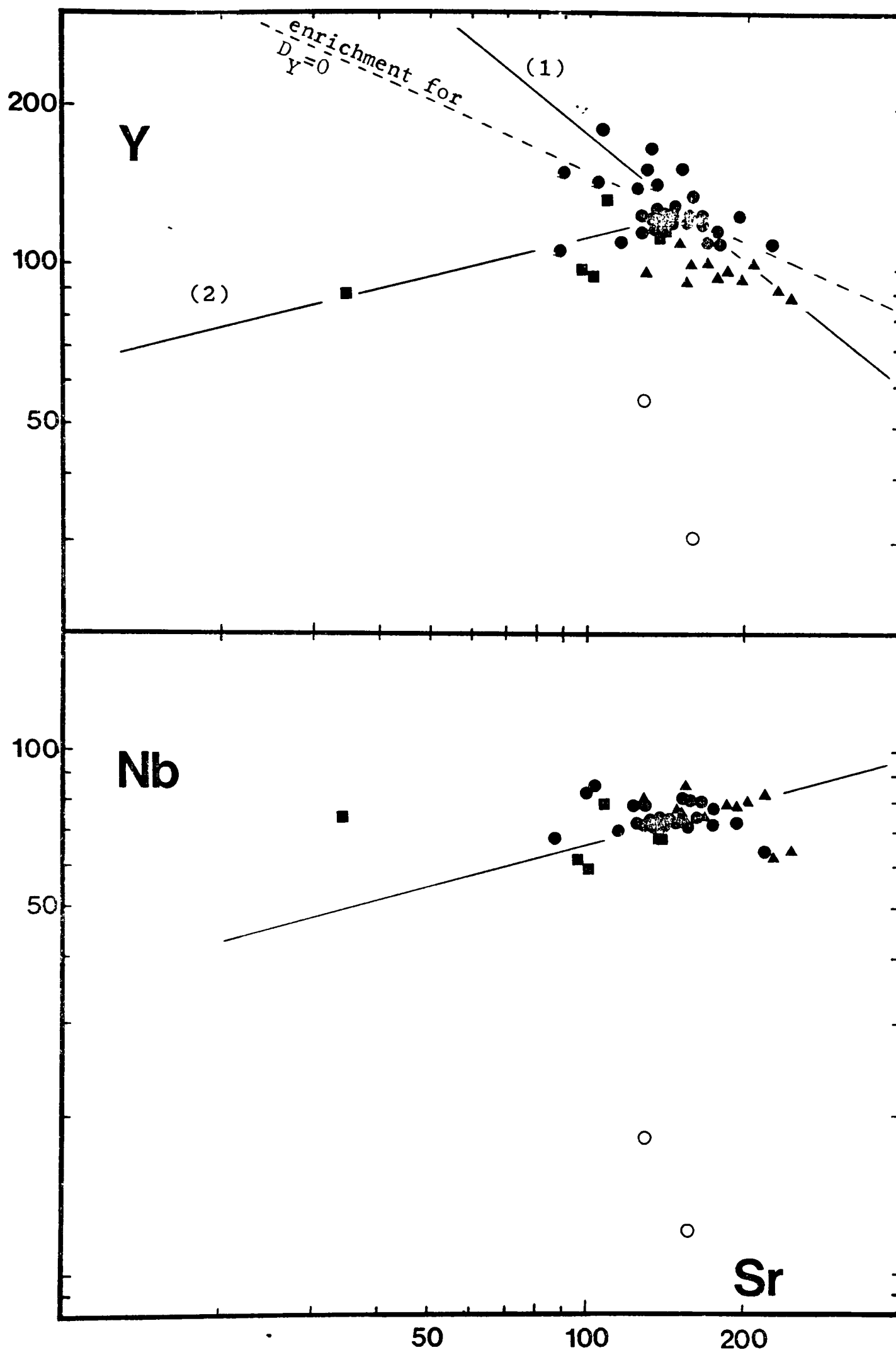


Fig. VIII.6 (a) Y vs. Sr. A plot with no well-pronounced trends. Line (1) is an approximate trend of Y enrichment; (2) is a trend corresponding to the Y depletion observed in the granophyres. The dashed line represents the maximum enrichment of Y possible by Rayleigh fractionation.

(b) A better defined trend for Nb. The group of three more basic points (high Sr) and the most Nb-rich rocks probably do not belong to the same fractionation trend.

Fig. VIII.6 (c - e) Plots for La, Ce and Nd vs. Sr. All three define reasonable straight lines if the most basic and most enriched points are excluded. See text for further discussion.

Dots = rhyolites; squares = granophyres; triangles = interbedded rhyolites; open circles = rhyolitic interbeds.

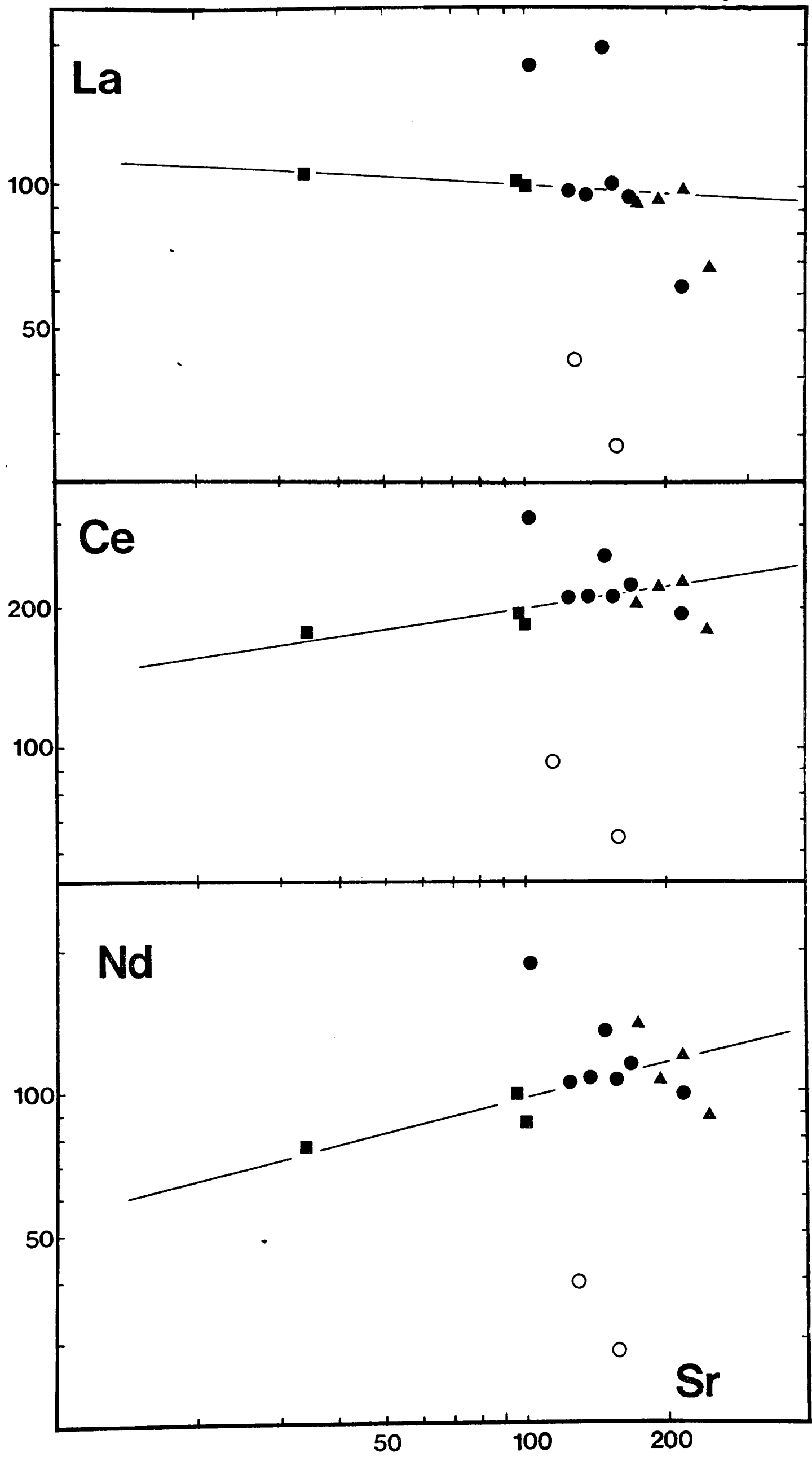


TABLE VIII.6

Calculated bulk distribution coefficients

Element	Gradient of element/Sr plot	D^0	D^0
		$(D_{Rb}^0 = 0.03)$	$(D_{Rb}^0 = 0.1)$
Rb	-0.4304	$(D_{Sr}^0 = 3.25)$	$(D_{Sr}^0 = 3.1)$
Ba	0.399	1.90	1.83
Y(1)	-0.764	-0.72	-0.60
Y(2)	0.258	1.58	1.54
Nb	0.266	1.60	1.56
La	-0.050	0.89	0.90
Ce	0.154	1.35	1.32
Nd	0.247	1.56	1.52

The second of the calculated Y coefficients and those for La, Ce, Nd and Nb are all greatly in excess of those predicted from published data. The rare earth elements and Yttrium generally have appreciable crystal/liquid partition coefficients (Table VIII.5); bulk distribution coefficients, calculated from the published data, are an order of magnitude less than those calculated for the rhyolites. The presence of accessory phases with high contents of REE may provide a mechanism for the REE fractionation.

Apatite is one of the main accessory phases in the Swaziland rhyolites: the extract at about 60% SiO₂ contains 0.4% P₂O₅ or 1% apatite. It contains 600 - 3000 ppm Yttrium (Lambert and Holland, 1974) and up to 2000 ppm of each of the light lanthanides La, Ce and Nd (Nagasawa, 1970). Yttrium shows little correlation with CaO (Figure VIII.7a) and the trends of Lambert and Holland (1974) are not applicable to these Y-rich rocks. Phosphorus shows a better correlation with Y (Figure VIII.7b) with a well-defined central group of data points showing depletion of Y and P₂O₅. Seven of the points are richer in Y but low in P₂O₅; if apatite is the main phase containing Yttrium, it is not clear why the rocks richest in Yttrium are poor in phosphorus.

Zircon is the other principal accessory in the rhyolites: the extract contains 2000 ppm or 0.4% zircon. The behaviour of Zr with respect to Sr is shown in Figure VIII.8a. The solid line is a best-fit to the data and from it an apparent D_{Zr}^0 could be calculated (it is only an apparent distribution because Zr is a stoichiometric constituent of zircon and the normal petrogenetic equations are inapplicable); the dotted line shows the expected gradient for $D_{Zr}^0 = 0$, a situation presumably present before the crystallisation of zircon. The Zr/Y plot (Figure VIII.8b) is characterised by two parallel trends. The interbedded dacites form a cluster with positive gradient and are 100 ppm richer in Zr than the main group of rhyolites. The

Fig. VIII.7 (a) CaO vs. γ CaO shows little correlation with γ and the fractionation of γ does not seem to be controlled by Ca-containing phases.

(b) P_2O_5 vs. γ . The data points can be divided into three groups: a central group showing positive correlation between P_2O_5 and γ ; a group rich in P_2O_5 ; and a group rich in γ . The behaviour of γ seems largely independent of the steady depletion in P_2O_5 .

Dots = rhyolites; squares = granophyres; triangles = interbedded rhyolites; open circles = rhyolitic interbeds.

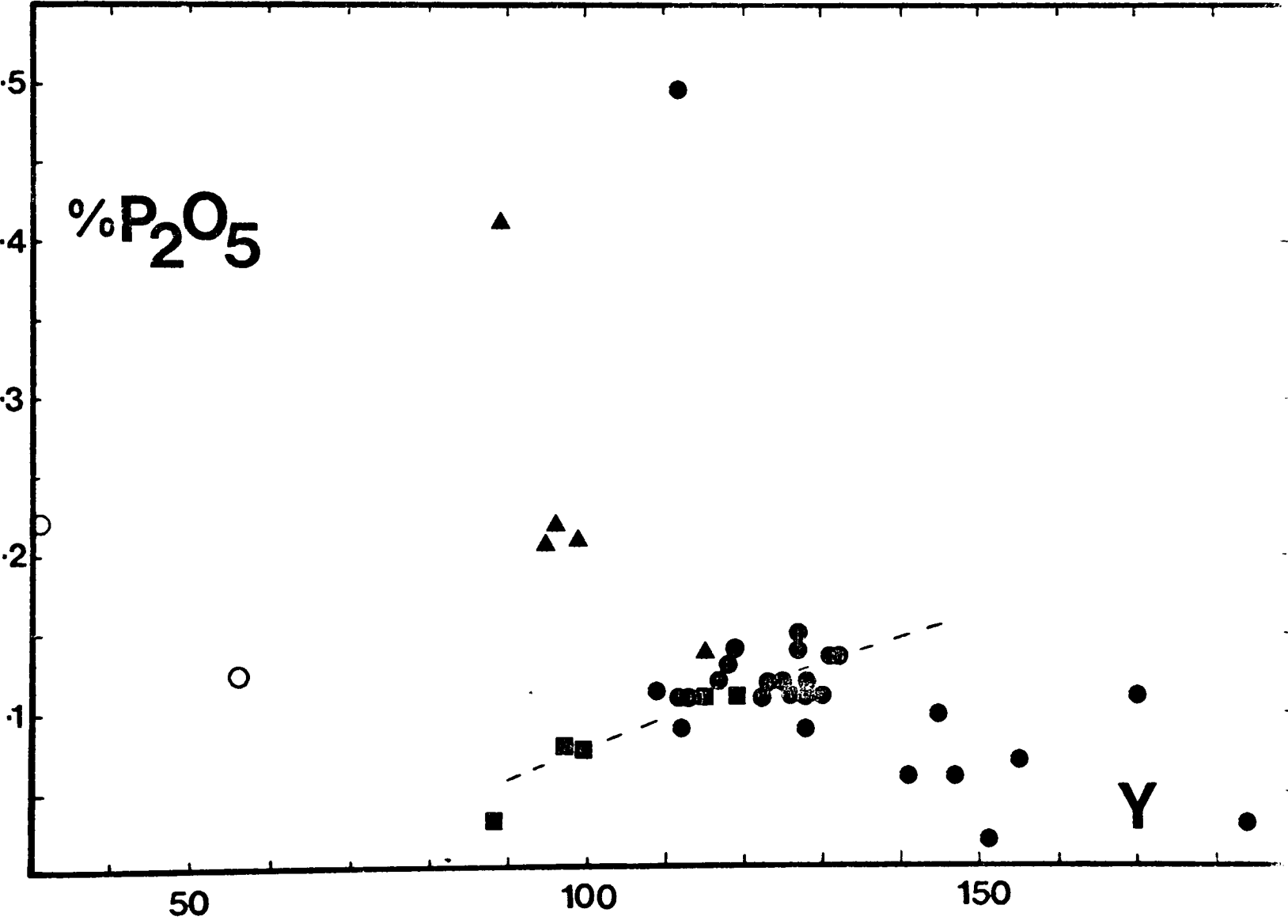
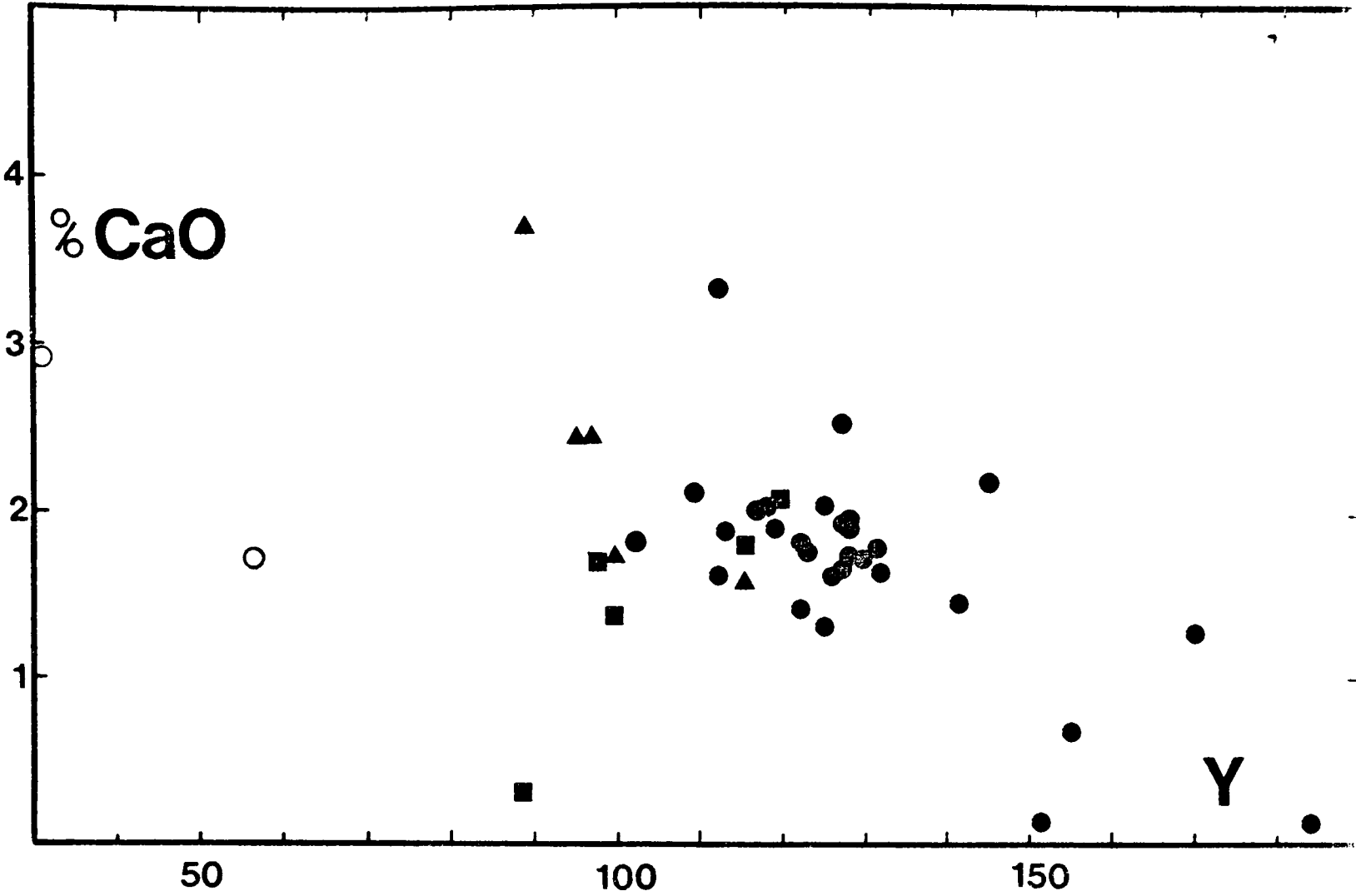
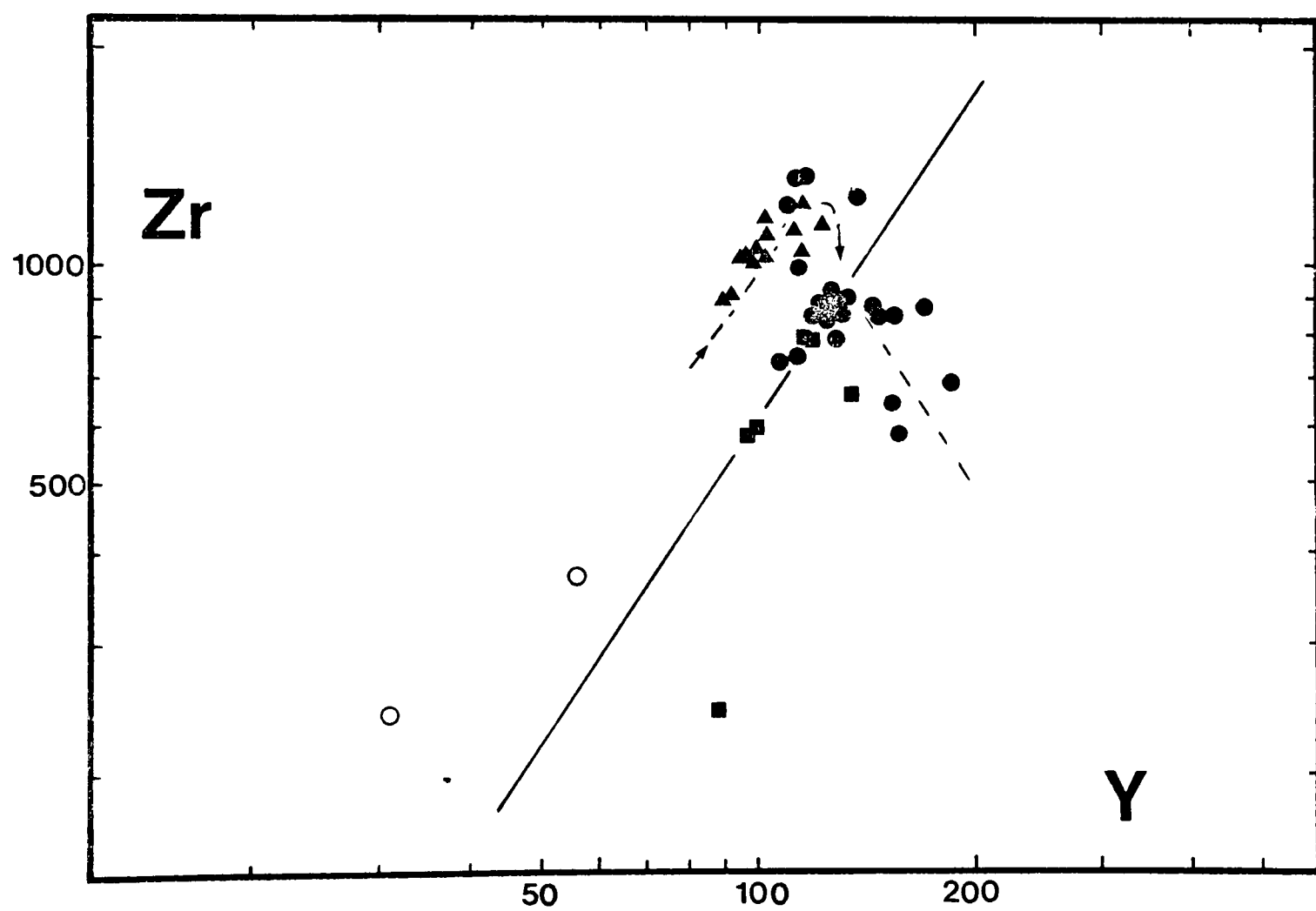
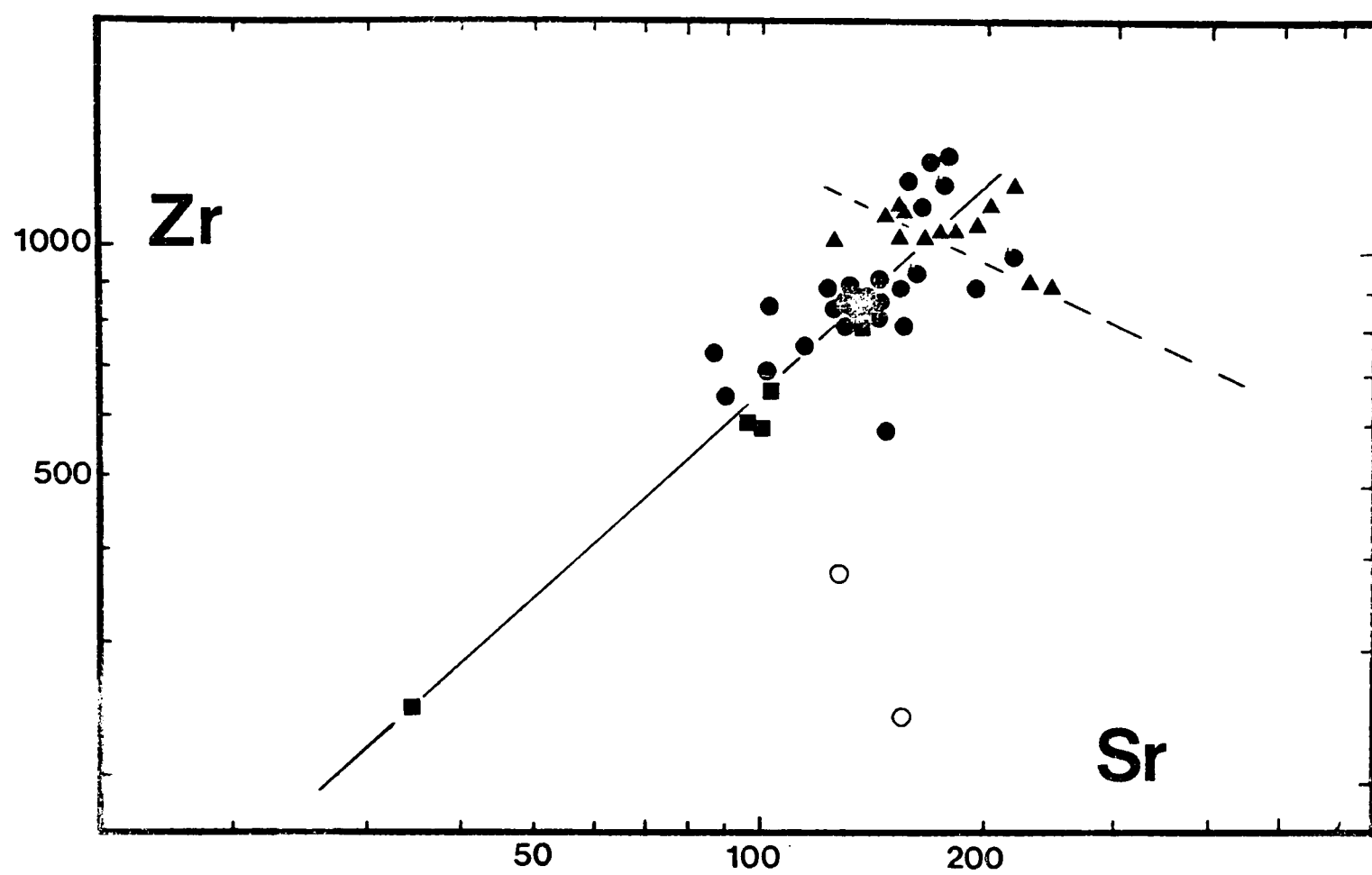


Fig. VIII.8 (a) The variation of Zr with Sr for the rhyolites. The solid line represents a best fit to the data and from it an apparent distribution coefficient can be calculated. The dotted line represents the enrichment for $D_{Zr} = 0$.

(b) A plot of Zr vs. Y shows two sub-parallel trends: that in the interbedded rhyolites; and that in the main group. The most basic of the main group fall at the top of the group of interbedded flows; the most acid flows fall on the high Y side of the main group. The coherence between the two elements is good; the fractionation of Zr and Y seem to be related.

Dots = rhyolites; squares = granophyres; triangles = interbedded
 ---rhyolites; open circles = rhyolitic interbeds.



initial trend is one of enrichment in both Zr and Y; this is followed by rapid depletion in Zr as zircon begins to precipitate. The final trend is a linear depletion of both Y and Zr. The separation of the enrichment and depletion trends may be due to initial fractionation of Y-poor zircon. The Y content of zircons is variable, ranging from 5000 to 14,000 ppm (Krasnobayev et al., 1976), and often has a broad positive correlation with SiO_2 though some massifs studied by Krasnobayev et al. (op. cit.) showed inversions in this trend. The amount of Y required in the extract to explain the variation diagrams is 200 ppm, representing a D_Y^0 of 1.57. Of the crystallising minerals, only zircon, apatite, and clinopyroxene contain any Yttrium; these will contribute to the extract as much as 60, 30 and 30 ppm respectively (these are maximum values calculated from the available concentration data for apatite and zircon and from partition coefficient data for clinopyroxene assuming $D_Y^{\text{Cpx}} = D_{\text{Dy}}^{\text{Cpx}} = 2$ and a liquid composition of 120 ppm). The covariance of Y and Zr, however, indicates that the fractionation of Y in zircon is the dominant factor and the presence of Y in apatite and clinopyroxene is of secondary importance. The extract for Y calculated above is 50% too low; in the absence of further data, it is not known whether the necessary increase is caused by higher concentrations of Y in zircon or the other phases.

Niobium shows a similar trend to Zr (Figures VIII.6b, 8a) and the substitution of Nb in zircon is expected to be of greatest importance for the chemistry of the extract. If all the Nb fractionated in zircon, it would imply a concentration in zircon of 2.5% Nb.

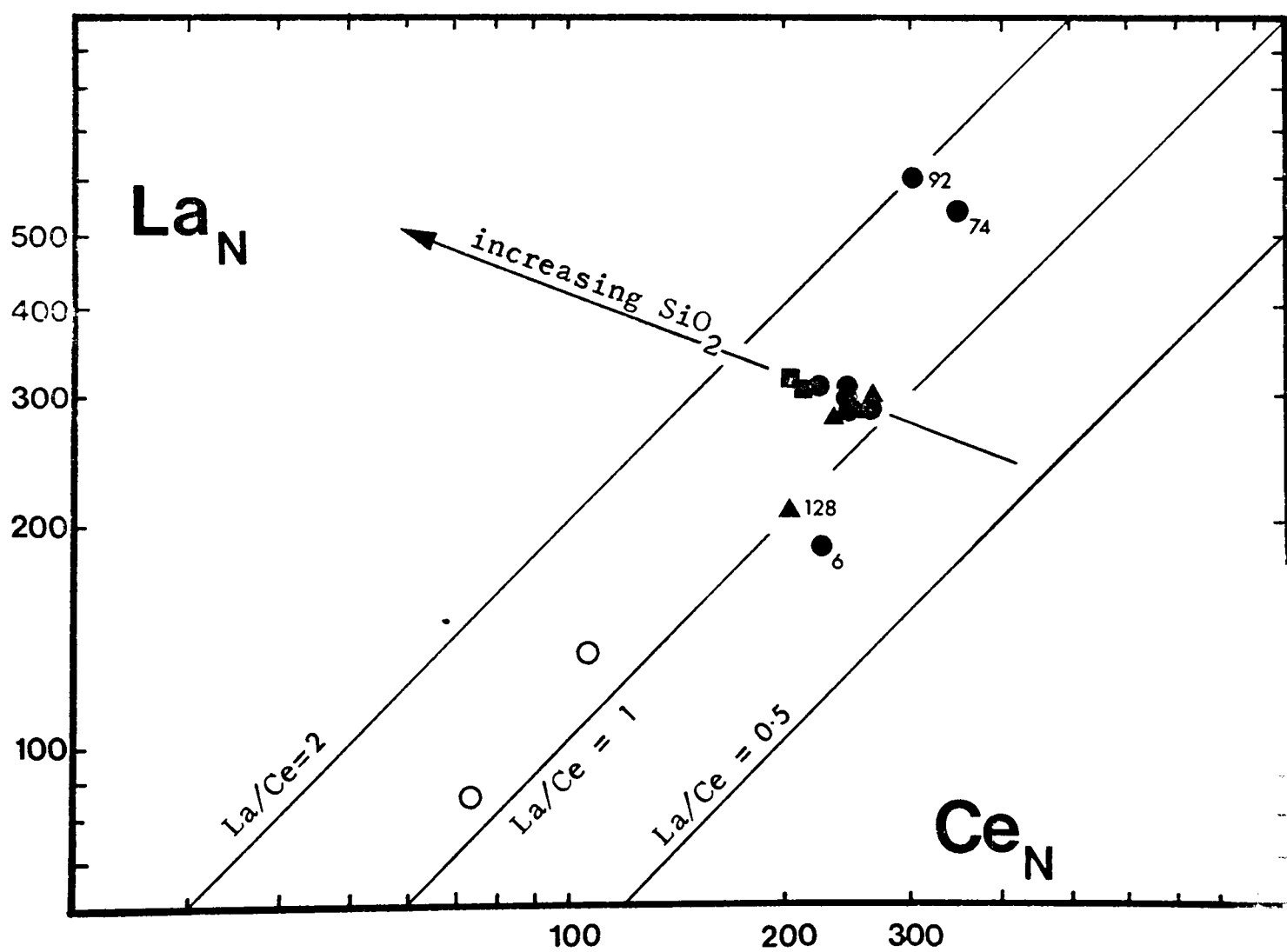
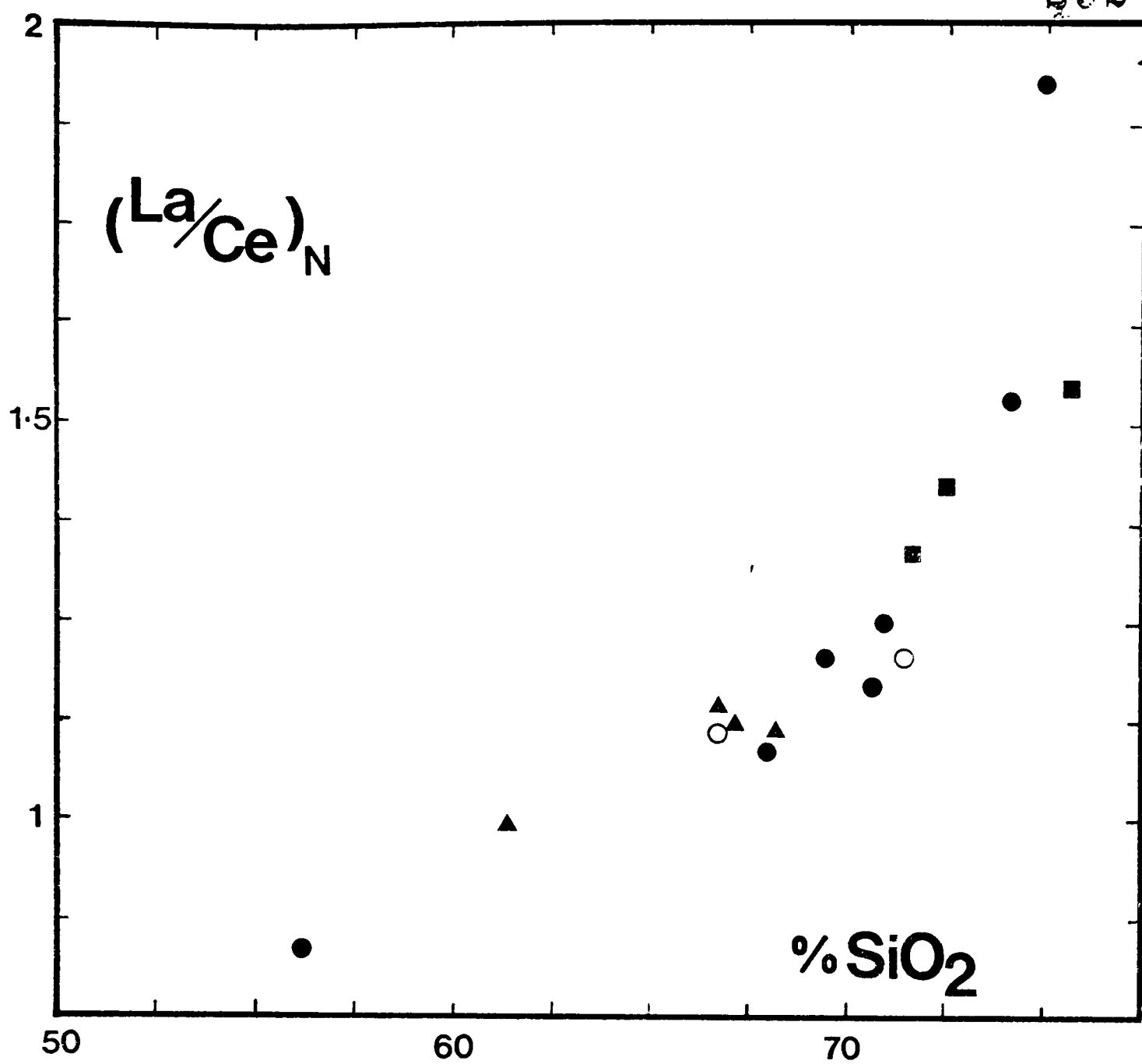
The variation in La, Ce and Nd poses some of the same problems as that in Y, in that the calculated bulk distribution coefficients are higher than predicted. A further problem is that there is fractionation between the light rare earths not predicted by the distribution coefficient data.

Fig. VIII.9 (a) The variation of La/Ce with SiO_2 . The data define a clean trend of increasing La/Ce with SiO_2 . The variation is too rational to be explained by mobility of La or anomalies in Ce due to oxidation to Ce^{4+} .

(b) La vs. Ce. The data fall into three groups (discounting the rhyolitic interbeds) separated by their La abundances. The central group defines a linear trend of increasing La and decreasing Ce with fractionation; the acid rhyolites are enriched almost twofold and the two 'trachyandesitic' flows are depleted in La relative to the middle group. Most petrogenetic processes produce little variation in La/Ce ratio; the observed variation here may be produced by the combination of an unusual fractionation trend depleting Ce, and a second process, for example partial melting of more basic material, which produces magmas with different abundances of La and Ce but a constant La/Ce ratio.

(La and Ce data normalised to chondrites)

Dots = rhyolites; squares = granophyres; triangles = interbedded rhyolites; open circles = rhyolitic interbeds.



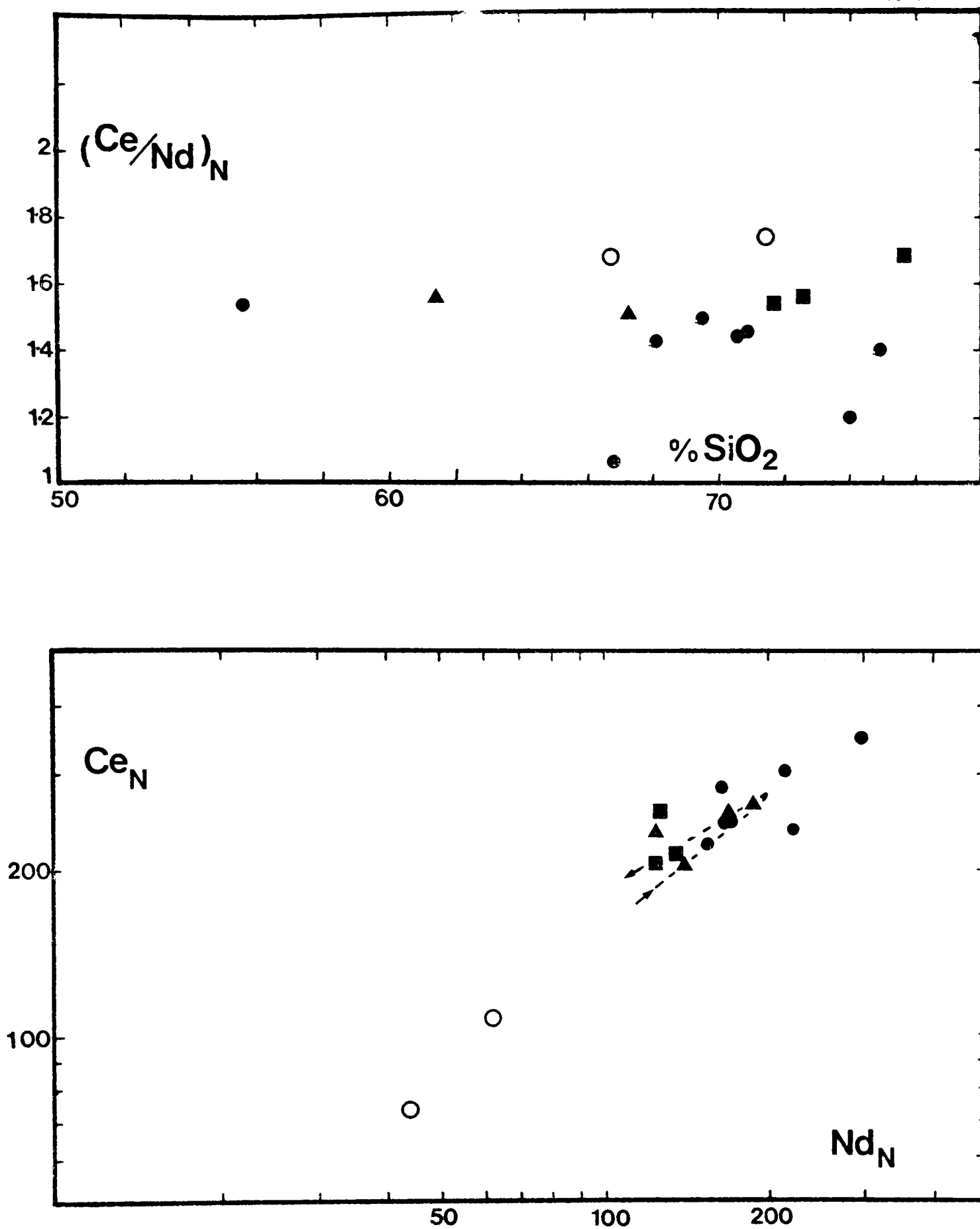


Fig. VIII.10 a,b. Analogous figures to Figure 9 for Ce and Nd. The Ce/Nd ratio remains constant. By comparison with Figure 9, it seems that La is behaving in a rational but anomalous fashion.

The La/Ce ratio has a good correlation with SiO_2 (Figure VIII.9a); this is also shown in Figure VIII.9b where the main group of rhyolites and granophyres forms a small group with negative slope. Samples 128 and 6 are poorer, and 92 and 74 are richer in La and Ce than the central group. Available distribution coefficient data for La and Ce are similar and most petrogenetic processes produce trends with equal enrichment or depletion in La and Ce.

Ce and Nd behave very similarly (Figure 10a, b) and the Ce/Nd ratio is scarcely affected by fractionation. The anomalous behaviour, therefore, is that of La, with a low D^0 relative to the other elements.

Weaver et al. (1972) reported anomalies in La and Ce abundances in Ethiopian pantellerites attributed to vapour phase activity but the variation in the Swaziland rocks seems too uniform to be explained by a fairly random process like hydrothermal leaching or differential oxidation.

One of the main problems is finding a mineral which will fractionate La relative to Ce to the extent that La is enriched during differentiation ($D_{\text{La}}^0 = 0.87$) and Ce depleted ($D_{\text{Ce}}^0 = 1.3$). The light REE content of zircon increases relative to the heavy REE with the K_2O content of the magma; La and Ce abundances range from 200 - 4000 ppm (600 - 8000 times chondrite), and the La/Ce ratios from 2 - 4 (1.6 to 0.9 times chondrite) (Krasnobayev et al., 1976). The maximum concentrations of Ce and Nd in apatite are 2000 and 1200 ppm (2000 times chondrite) respectively (Nagasawa, 1970) and a value for La can be extrapolated from these data (approximately 700 ppm, 2000 times chondrite).

The concentrations of these elements expected in the extract can be calculated from the available data (Table VIII.7). The extract is too poor in

TABLE VIII.7

Extract composition for La, Ce and Nd

Phase	% in extract	La		Ce		Nd	
		D*	Contribution to extract	D*	Contribution to extract	D*	Contribution to extract
Plagioclase	65	0.3	20	0.27	35	0.21	14
Clinopyroxene	15	0.5	8	1.11	33	1.67	25
Apatite	1	700	7	2000	20	1200	12
Zircon	0.4	4000	16	4000	16	4000	16
Total			51		104		67
Amount required in extract by variable diagrams			100		300		180

D* : distribution coefficient or concentration for each phase. Data from Arth (1970), Nagasawa (1970), Krasnobayev et al. (1976).

The contribution to extract is calculated from distribution coefficient data assuming the liquid compositions to be: La = 100; Ce = 200; Nd = 100.

Ce and Nd by a factor of three and in La by a factor of two to explain the observed variation. It may be possible that zircon, or zircon and apatite between them, are enriched in the rare earth elements, or that the published distribution coefficients are gross underestimates, but, without further data, only speculation is possible.

The La/Ce plot (Figure VIII.9b) produces good discrimination between the central group of rhyolites and the more basic and acid ones. The main fractionation produces an increase in La/Ce ratio with silica and it is apparent that samples 92 and 74 cannot belong to the same fractionation sequence. The samples provide a continuous variation in the La/Ce ratio but are discontinuous in their La and Ce abundances. There seem to be two processes operating: one enriches La and Ce with constant, or nearly constant La/Ce ratio, while the other fractionates the La/Ce ratio with little variation in absolute abundance. The variation within the main group is compatible with fractional crystallisation; the enrichment trend may reflect the source chemistry and/or the partial melting event producing the rhyolite magma.

The production of acid melts and the relative importance of partial melting and crystallisation are considered in Section VIII.13.

(d) Summary of trace element models

The variation of Rb, Sr and Ba can be approximated by a simple Rayleigh fractionation model with a plagioclase/clinopyroxene extract consistent with the major element trends. Distribution coefficients derived from the whole-rock data are slightly higher than those published but are compatible with the phenocryst chemistry. The variation of Nb, La, Ce, Nd and Y can also be explained by this simple model but the derived bulk distribution coefficients are in excess of those published or expected. Anomalously

high distribution coefficients or enrichment of the minor elements in the accessory phase zircon and apatite need to be involved to explain the observed variation.

Several of the samples are enriched in Y, La, Ce and Nd to an extent not possible by normal fractional crystallisation. At the high silica end of the variation diagrams the principal fractionation is defined by the granophyre analyses; the acid rhyolites fall above these trends and can not be formed by crystal fractionation from the same magma. Possible causes for this enrichment include crystal accumulation, formation on a partial melting rather than crystallisation trend and formation from enriched source regions. For further discussion see Section VIII.13.

VIII.11 Density considerations

For a fractional crystallisation model to be possible there needs to be a reasonable settling rate for the phenocrysts in the magma.

The settling rate of a solid sphere in a liquid is governed by Stokes's

$$V = \frac{1}{18} \frac{\Delta\rho g d^2}{\mu}$$

where V = settling rate;

$\Delta\rho$ = density difference between solid and medium;

g = gravitational acceleration;

d = diameter of a sphere;

μ = viscosity of liquid.

The density of a silicate melt can be calculated from its chemical composi-

tion and the data on partial molar volumes given by Bottinga and Weill (1970). Calculations for RC 74 and 40, representing extremes in composition, both give a density of 2.37 and this value is taken as representative of the density of a rhyolitic liquid. Densities for plagioclase (= 2.6) and clinopyroxene (3.2) are taken from Deer et al. (1966).

Viscosities can be calculated from the data of Bottinga and Weill (1972) but are critically dependent on the water content. In the following calculations the viscosity of the rhyolite liquid is assumed to be 10^5 poise; this is the value obtained by Bottinga and Weill (1972) for an anhydrous granite liquid at 1200°C . For comparison, settling rates are calculated for a basalt of viscosity 10^3 poise.

For a crystal of diameter 5 mm in a liquid of viscosity 10^5 poise the Stokes Law equation can be reduced to:

$$V = 1.4 \times 10^{-4} \times \Delta\rho \text{ cm/s}$$

where $\Delta\rho$ is the density difference.

Thus for plagioclase ($\Delta\rho = 0.23$) $V = 3 \times 10^{-5} \text{ cm/s}$

or $V = 1000 \text{ cm/y}$;

and for clinopyroxene ($\Delta\rho = 0$) $V = 1.2 \times 10^{-4} \text{ cm/s}$

or $V = 3,600 \text{ cm/y}$.

For a basalt of density 2.6 g/cm^3 and viscosity 10^3 poise the rate for clinopyroxene, $V = 0.01 \text{ cm/s}$.

These settling rates for phenocrysts in rhyolitic liquids are slow, but, with geologically reasonable times, appreciable fractionation can occur. Fractionation

tion of all phases, but especially zircon and apatite, will be aided by the formation of glomeroporphyritic aggregates, thus increasing their effective diameter. For zircon and apatite, their large densities are more than outweighed by their small size and only as inclusions in the major phases will they fractionate efficiently.

VIII.12 Aeolian differentiation

Certain crystal-rich ignimbrites have been interpreted by Walker (1972) as having suffered aeolian differentiation. The process envisaged is one of winnowing, whereby the lighter pumice fraction is selectively removed during the eruption. If this process is important, it implies that whole-rock analyses of ignimbrites may not be representative of the eruptive magma.

If the eruption represents the discharge of a single magma, the pumice fraction can be regarded as the liquid component and the crystals, the phenocrysts in equilibrium with it.

Two models for aeolian differentiation are considered, both involving a magma which undergoes Rayleigh fractionation before being modified by removal of the liquid fraction.

Model 1

The magma undergoes 50% Rayleigh fractionation with complete removal of the fractionating phases. Prior to the eruption 30% of equilibrium crystallisation takes place- the resulting liquid, with the equilibrium phenocrysts, is erupted.

After 50% Rayleigh fractionation the concentration of a trace element in the liquid is given by:

$$C = 0.5^{(D-1)} \quad (\text{for } C^0 = 1)$$

If at this point equilibrium crystallisation starts, the melt composition will evolve:

$$\frac{C^L}{C} = \frac{1}{F + D(1-F)} \quad \text{where } C \text{ is defined above.}$$

The composition of the solid, C , in equilibrium with this melt is:

$$\frac{C}{C} = \frac{1}{F + D(1-F)}$$

Figure VIII.11 shows the path of the evolving magma and solid compositions for the trace elements Rb and Sr. If, after 30% of equilibrium crystallisation the magma is erupted and the liquid partly removed, the bulk composition of the resulting rock will lie on curve E.

Model 2

The magma undergoes Rayleigh fractionation but keeps the crystals in suspension until the time of eruption. The crystals have only their margins equilibrated and are perfectly zoned.

The melt composition is given by:

$$C^L = F^{D-1} \quad (C^0 = 1)$$

and the compositions of the margins of the crystals by:

$$C^m = DF^{D-1}$$

The composition of the bulk crystalline extract can be calculated by a mass balance equation:

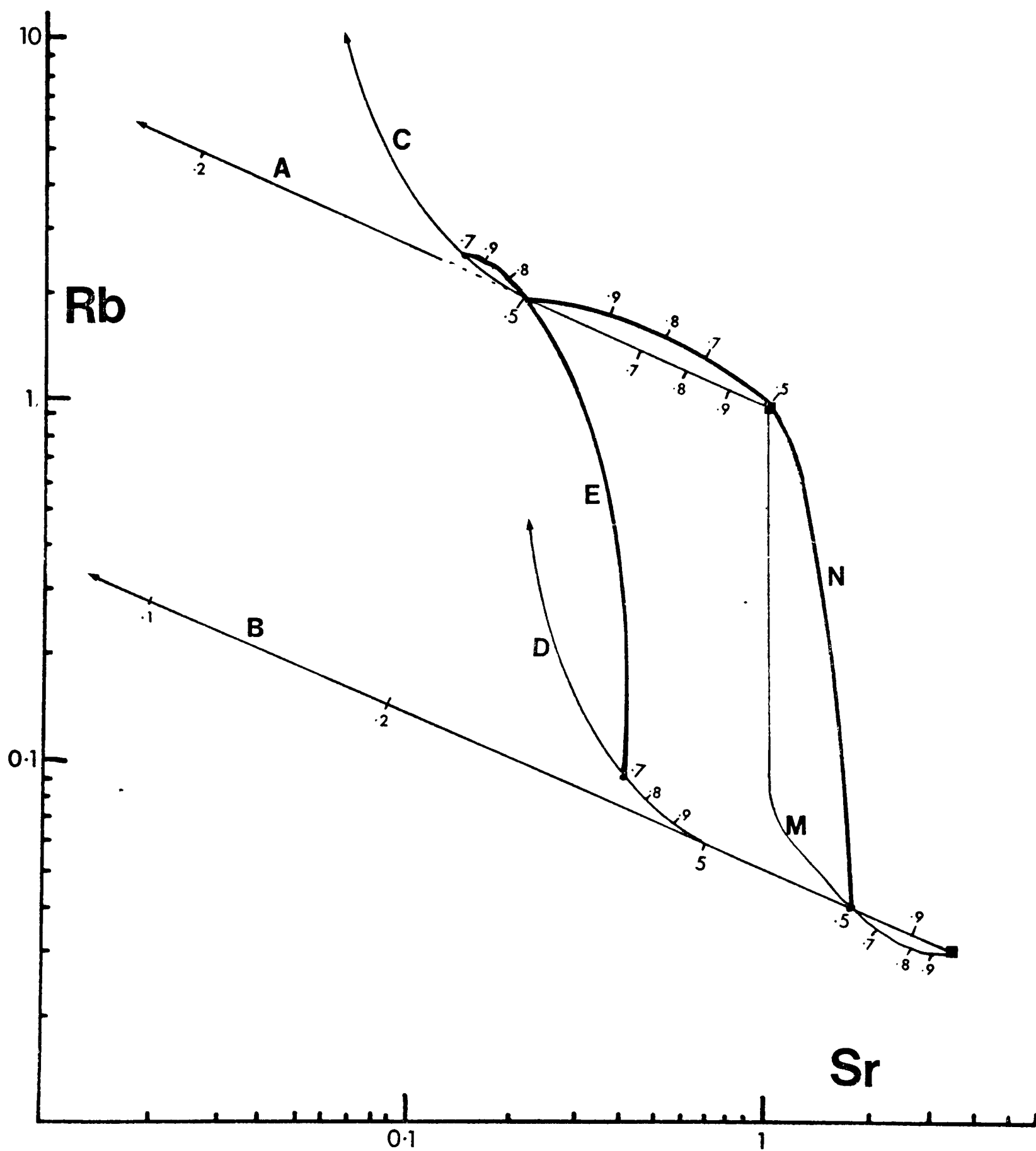
Fig. VIII.11. Theoretical models for aeolian differentiation.

Model 1 A magma undergoes 50% Rayleigh fractionation with complete removal of the phenocrysts; the melt evolves along A and the solids in equilibrium along B. Prior to eruption 30% equilibrium crystallisation occurs, the compositions of the magma and solids given by C and D respectively. If the magma is erupted and loses the melt fraction through winnowing, the final bulk composition will fall on E which is a mixing curve of the solid and liquid compositions after 30% equilibrium crystallisation.

Model 2 The magma undergoes 50% Rayleigh fractionation, but the crystals remain in the magma, becoming perfectly zoned. The liquid evolves along A; M represents the bulk solid composition. After eruption of the melt and a proportion of the crystals, and after winnowing, the bulk composition will fall on N.

In both these models the effects of incorporating crystals into the final melt composition is to produce trends effectively coincident with the fractionation trends, at least for small degrees of crystal concentration.

The curves are drawn to be analogous to those in Fig. VIII.5.



$$C^S \times (1-F) + C^L \times F = 1$$

and hence

$$C^S = \frac{1 - F^{D-1} \cdot F}{1 - F} = \frac{1 - F^D}{1 - F}$$

Curves M and N (Figure VIII.11) show respectively the variation of the bulk solid composition and the bulk composition of the system after melt removal.

The effect of aeolian differentiation on the whole-rock composition is appreciable, even for small amounts of melt removal, but up to 50% of the melt can be removed before the effects are distinguishable from the Rayleigh fractionation trend. The result of this process is to produce scatter about the fractionation curves which is probably no greater than the natural 'noise'. The Swaziland rhyolites rarely contain more than 25% phenocrysts; the small amount of aeolian differentiation permissible will not be noticeable on whole-rock variation diagrams.

A further consequence of aeolian differentiation is that, for a given magma, there will be a correlation between bulk composition and phenocryst content. This correlation is not apparent in the Swaziland ignimbrites (Figure VI.1) and it must be concluded that aeolian differentiation has not appreciably modified the chemistry of these rocks.

The calculations performed here have analogies with models for magma chamber evolution, where the liquid removed, rather than the residue, becomes important.

VIII.13 Production of acid magmas

In the previous sections the modification of an assumed primary acid magma was discussed without reference to a source region. The observed chemical variation was shown to be approximately consistent with crystal fractionation of a dacitic melt with 60 - 65% SiO_2 assuming an extract of the phenocrysts present in the rocks in their approximate modal proportions. In this Section the derivation of dacitic magmas is discussed.

As mentioned previously there are three principal ways of producing dacitic magmas:

- (1) Fractional crystallisation of basaltic magma;
- (2) Partial melting of basalts or gabbros;
- (3) Partial melting of acid crustal material:
 - a) Precambrian basement;
 - b) Solidified acid differentiates of basic magma.

The intrinsic difference between partial melting and fractional crystallisation is in the nature of the equilibrating solid phases: in ideal partial melting the liquid remains in equilibrium with the source, and the trace element chemistry of the melt is controlled by the residual mineralogical assemblage irrespective of the composition of the melt; in fractional crystallisation the solid phases form in equilibrium with the melt and change in composition as the melt composition changes. Changes of melt composition during fractional crystallisation can always be expressed in terms of a stoichiometric mineral phase or phases; during partial melting the melt increments may be also controlled by changes in solid solution and, as a result, inexpressible in terms of an actual mineral assemblage (cf. Helz, 1976).

The least-squares mixing program was used in an attempt to model the differentiation of a basic magma using major element data. The most basic rhyolite (RC 128) was used as an approximation of the "primary" dacitic magma, and an average dolerite (RC 111) of the source region.

The results of the mixing are given in Table VIII.8a. The extract approximates the source composition modally, though enriched in the ore phases. The major part of the squared residuals is due to a 1% discrepancy in Al_2O_3 ; this may be caused by the use of a plagioclase composition that is too albitic or sub-aluminous. The compositions of the plagioclase and clinopyroxene present in RC 111 have been used in the calculation. They are possibly slightly more evolved than the initial fractionating phenocrysts but may approximate the average phenocryst composition taken over the whole fractionation range.

The next stage of fractionation, from RC 128 to RC 40 (a more typical dacite), is best modelled using an andesinic plagioclase and basic pyroxene analyses (Table VIII.8b). The better fit of these basic phenocrysts suggests either a closer similarity of the pyroxenes in the basic rocks to those fractionating in the basic rhyolite or, because the relationship between the two rocks is a partial melting trend, the melt increments may correspond to these mineral compositions. The mixing model for major elements is compatible with both fractional crystallisation and partial melting if the residual solid, or cumulate, corresponds to the extract minerals.

Trace element data, because of the relatively simple laws governing trace element behaviour, can provide more precise estimates of partial melting and fractionation than major element data.

The amount of differentiation necessary to produce RC 128 (basic dacite)

TABLE VIII.8

A. Fractionation of basaltic magma (RC 111) to produce a basic rhyolite (RC 128). (Phenocryst compositions from Table 2)

	RC 111 observed	RC 111 calculated			
SiO ₂	49.35	49.38			
TiO ₂	2.68	2.72	% components		
Al ₂ O ₃	13.07	12.01	RC 128	15%	% of extract 0
FeO	14.15	14.15			
MnO	0.23	0.22	PL 11	36	43
MgO	5.72	5.85	CP 11	39	46
CaO	10.05	10.21	OR 11	4.5	5.4
Na ₂ O	2.23	2.38	MT	4.6	5.5
K ₂ O	0.63	0.63			

$$\sum R^2 = 1.18$$
$$\sum \%R^2 = 152$$

B. Fractionation of basic rhyolite (RC 128) to produce RC 40.

	RC 128 observed	RC 128 calculated			
SiO ₂	61.36	61.48			
TiO ₂	1.43	1.47	% components		
Al ₂ O ₃	12.94	12.86			
FeO	8.09	8.09	RC 40	81	% of extract 0
MnO	0.18	0.15			
MgO	1.30	1.36	Andesine*	7.7	45.3
CaO	3.74	3.48	CP 11	5.8	34.2
Na ₂ O	3.31	3.57	OR 11	1.68	9.9
K ₂ O	3.48	3.59	MT	1.74	10.2

$$\sum R^2 = 0.17$$
$$\sum \%R^2 = 373$$

* analysis from Deer, Howie and Zussman (1966)

3. Fractionation of basic rhyolite (RC 65) to acid rhyolite (RC 73).

	RC 65 observed	RC 65 calculated	% components		% of extract
SiO ₂	68.29	68.29	RC 73	75.4	
TiO ₂	0.57	0.57			
Al ₂ O ₃	12.46	13.74			
FeO*	6.96	6.96	PL 55	18.4	67
MgO	0.30	0.30	CP 55	3.61	13.2
CaO	2.07	2.01	OR 55	2.6	9.4
Na ₂ O	3.79	3.57	MT	2.7	10
K ₂ O	4.41	4.24	QZ	-1.2	

$$\sum R^2 = 1.71$$
$$\sum \%R^2 = 174$$

* Total iron as FeO

TABLE VIII.9

A. Trace element enrichment between RC 111 and RC 128

Element	Concentration RC 111 RC 128		Enrichment factor	D ⁰ estimated	Calculated % differentiation	
					Rayleigh fractionation	Partial melting
Rb	10	90	9.0	0.03	10	8
Sr	240	244	1.0	0.9	-	-
Nb	12	65	5.4	0	18	18
Zr	200	900	4.5	0	22	22
Y	35	90	2.6	0.3	25	12
Ba	125	1100	8.8	0.1	11	11
La	16	70	44	0.2	15	3
Ce	26	180	7.0	0.1	12	5
Nd	12	90	7.5	0.2	8	-

% predicted from major elements = 15

B. Source composition for 15% fractionation

Element	RC 128	<u>Rayleigh fractionation</u>		<u>Partial melting</u>		RC 111
		enrichment	source comp.	enrichment	source comp.	
Rb	90	6.3	14	5.7	15	10
Sr	244	1.2	203	1.09	223	240
Nb	65	6.7	9.7	6.7	9.7	12
Zr	900	6.7	134	6.7	134	200
Y	90	3.8	24	2.5	36	35
Ba	1100	5.5	200	4.25	259	125
La	70	4.6	15	3.1	22.6	16
Ce	180	5.5	33	4.25	42	26
Nd	90	4.6	20	3.1	29	12

from RC 111 (dolerite) can be calculated from the trace element data; if the enrichment factors and distribution coefficients are known, the amount of fractionation or partial melting can be determined and compared with that calculated from the major element data (Table VIII.9a). Conversely, the amount of differentiation calculated from the major elements can be used to estimate enrichment factors and hence the source composition relative to an assumed derived liquid (Table VIII.9b).

The agreement of major and trace element models is reasonable; the data in Table VIII.9b show that, for the assumed fractionation model, the distribution coefficients for Sr and Zr may be too low and those for Ba, Ce and Nd too high. The Rayleigh fractionation model is slightly more consistent with the data than a partial melting model but the uncertainties in the distribution coefficients do not enable the two models to be distinguished.

Figure VIII.12 shows the variation of Rb/Sr with Rb for fractional crystallisation and partial melting. The source composition is taken as unity for both Rb/Sr and Rb; the values of Rb/Sr and Rb are thus equivalent to the enrichment factor relative to the source. The fractional crystallisation models are calculated in three steps: 85 or 90% fractionation of basic phenocrysts from a basic magma to form a dacitic melt; 25% crystallisation of intermediate phenocrysts (or, at least, those with dacitic distribution coefficients) and a final stage of fractionation of rhyolitic phenocrysts. D_{Rb}^0 is assumed to remain constant throughout the fractionation; $D_{\text{Sr}}^{\text{Plag}}$ varies from 1.8 in the basic rocks to nearly 5 in the acid rocks and causes the rapid enrichment in Rb/Sr. The phenocryst modes in the extract are taken from the various least-squares mixing data.

The partial melting and fractional crystallisation models produce a similar range of Rb/Sr ratios but, because the phenocryst compositions change with

Fig. VIII.12.

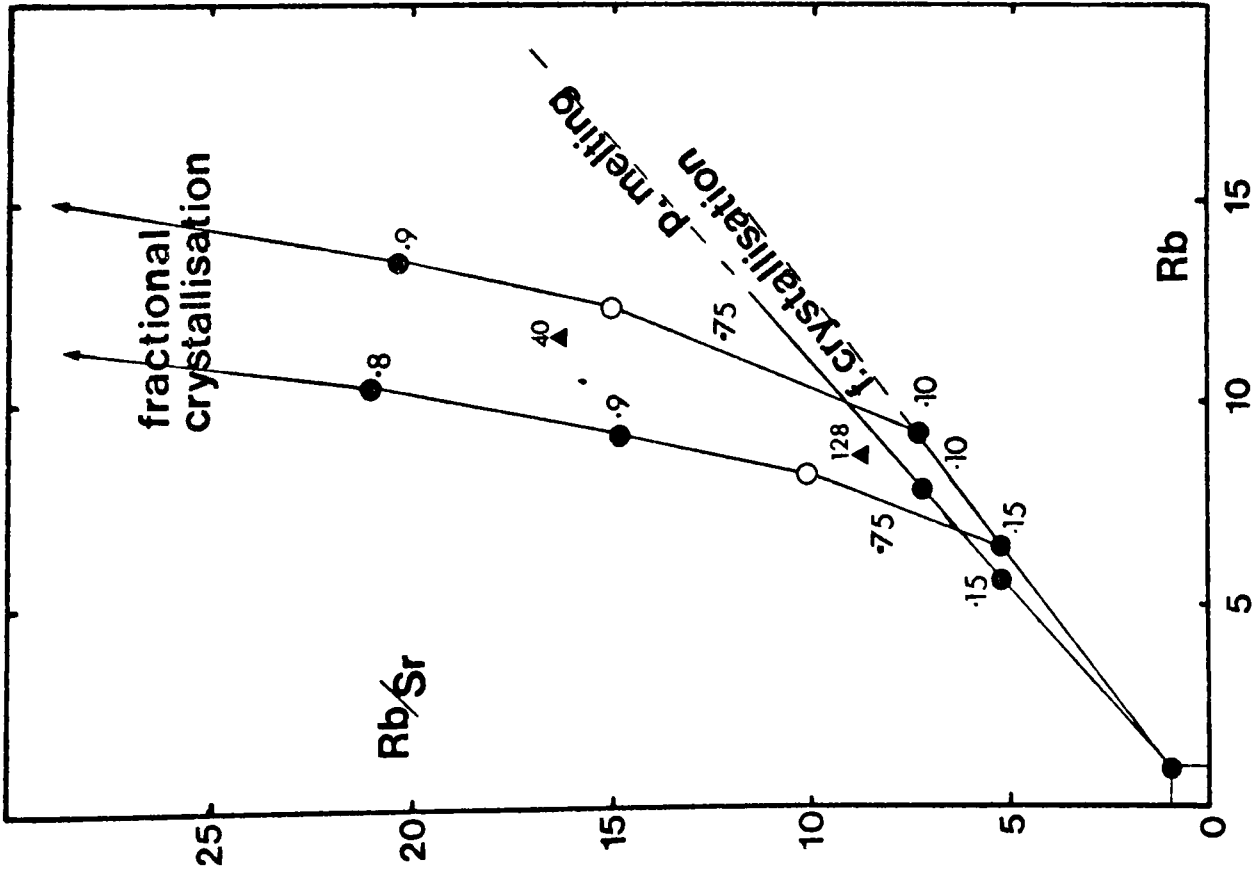
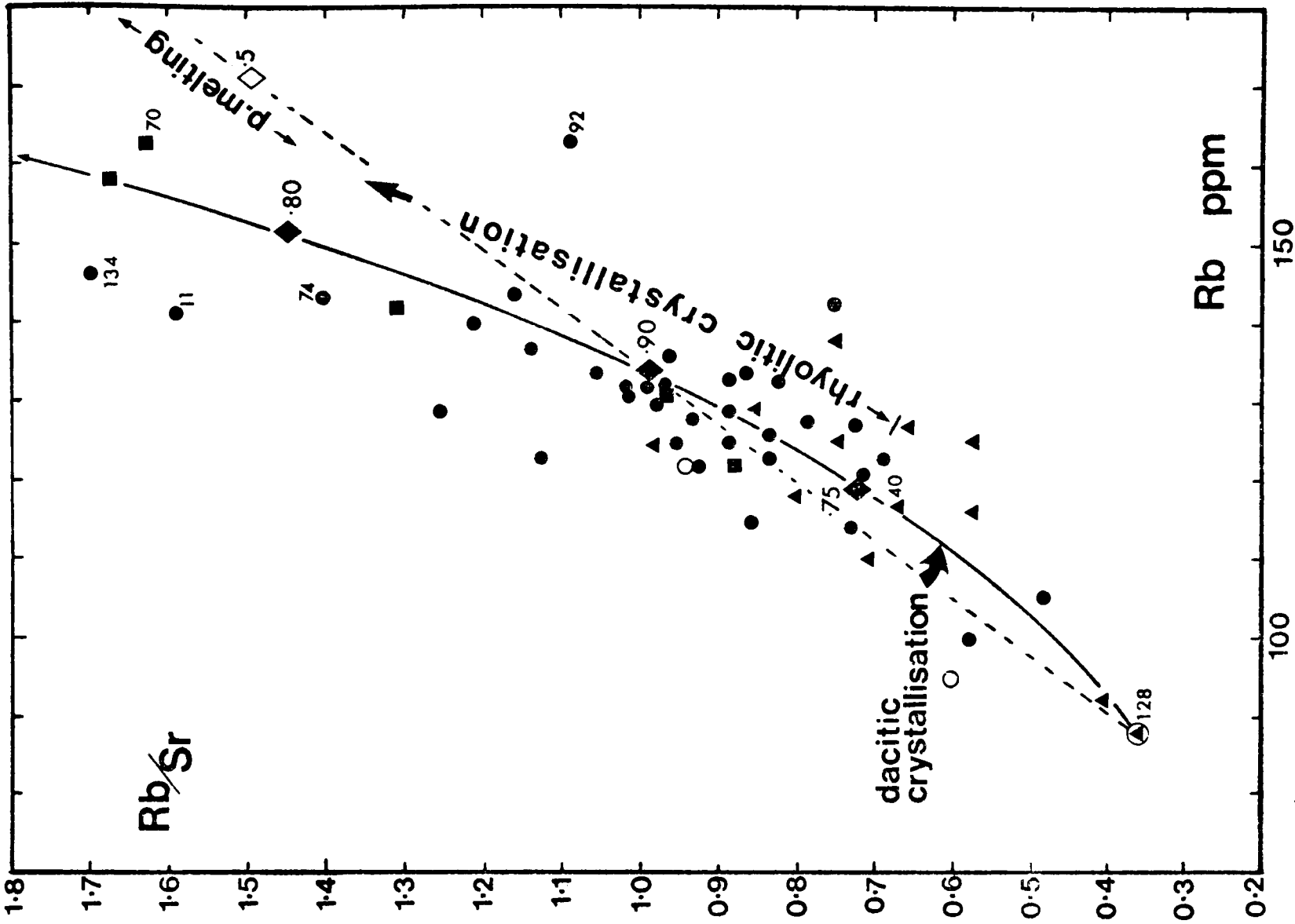
Model curves for fractional crystallisation and partial melting for a source $Rb = 1$, $Sr = 1$. Initial partial melting or Rayleigh fractionation from a basic parent is followed by 25% crystallisation of dacitic phenocrysts and further fractionation of rhyolitic phenocrysts. The two models for partial melting and fractional crystallisation of a basic source cannot be distinguished.

$$(D_{Rb} = 0.03; D_{Sr} = 0.97,$$

3, 3.25 in basic, dacitic and rhyolitic compositions respectively.)

Fig. VIII.13.

Analogous diagram to Figure 12, showing the rhyolite data points with superimposed curves for fractional crystallisation and partial melting of an assumed parent magma. For moderate amounts of partial melting the two processes are indistinguishable with respect to Rb and Sr.



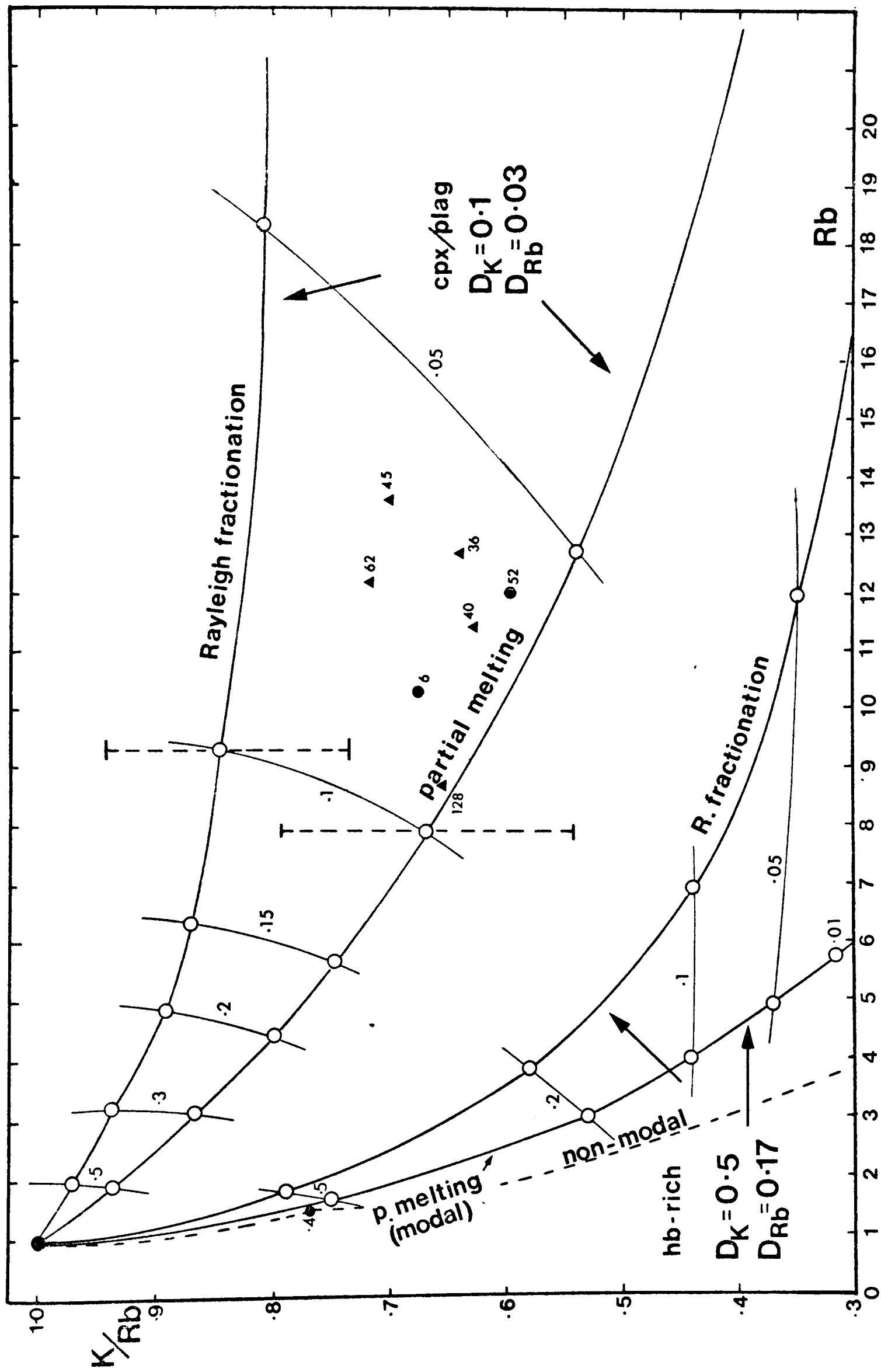
the melt composition, the crystallisation model produces liquids relatively enriched in Rb/Sr in the late stages of fractionation. While the equilibrating solids remain basaltic the two processes are indistinguishable except for extreme degrees of differentiation. The rhyolite data, relative to an assumed source chemistry ($RB = 10$, $Rb/Sr = 0.04$), fall between the two examples of crystallisation models and seem to be produced initially by a 10% extract from the source. The rhyolite data are shown in Figure VIII.13 with fractional crystallisation trends superimposed (cf. Figure VIII.5a).

The mineralogy of the source critically affects the bulk distribution coefficients and hence the composition of derived liquids. The presence of amphibole in the source is considered in Figure VIII.14. Two simple models for differentiation, Rayleigh fractionation and batch equilibrium melting are considered for a hornblende-rich and a 'normal' clinopyroxene-plagioclase source. A few data for the most basic rhyolites are plotted relative to a source with $K/Rb = 500$ and $Rb = 10$ ppm. They fall in a cluster around the anhydrous melting curve and have too high a K/Rb ratio to be produced from a source with residual amphibole. The anhydrous crystal fractionation curve produces a higher and nearly constant K/Rb ratio but, in the absence of reliable distribution coefficient data, crystal fractionation can not be entirely ruled out.

At water pressures of greater than 2 kb, the stability limit of amphibole is greater than that of plagioclase (at 5 kb with $P_{H_2O} = 0.6 P_{total}$: hornblende-out = $1025^{\circ}C$, plagioclase-out = $925^{\circ}C$; at 8 kb : hornblende-out = $1060^{\circ}C$, plagioclase-out = $880^{\circ}C$ (data for 1921 Kilauea tholeiite from Holloway and Burnham {1972}) and the compositions of the early liquids will be controlled by the breakdown of plagioclase; only when all the plagioclase has melted does amphibole start contributing to the melt (Helz, 1976). The effect of this non-modal melting on the trace element chemistry is shown by

Fig. VIII.14. K Rb-Rb systematics. Partial melting and Rayleigh fractionation curves are calculated for two source mineralogies relative to a starting composition of unity. The lower pair of curves are calculated for a source containing 60% amphibole, 40% plagioclase. The solid partial melting curve represents modal melting, the broken line non-modal melting where the first 40% of melting is entirely plagioclase. The upper pair are for a source containing 50% plagioclase and 50% clinopyroxene. Modal and non-modal melting curves are effectively identical.

The rhyolite data are plotted relative to a source with $K/Rb = 500$, $Rb = 10$ ppm. The most basic rhyolites fall close to the partial melting curve though the errors in distribution coefficient data, shown by the error bars, do not allow a fractional crystallisation origin to be eliminated.



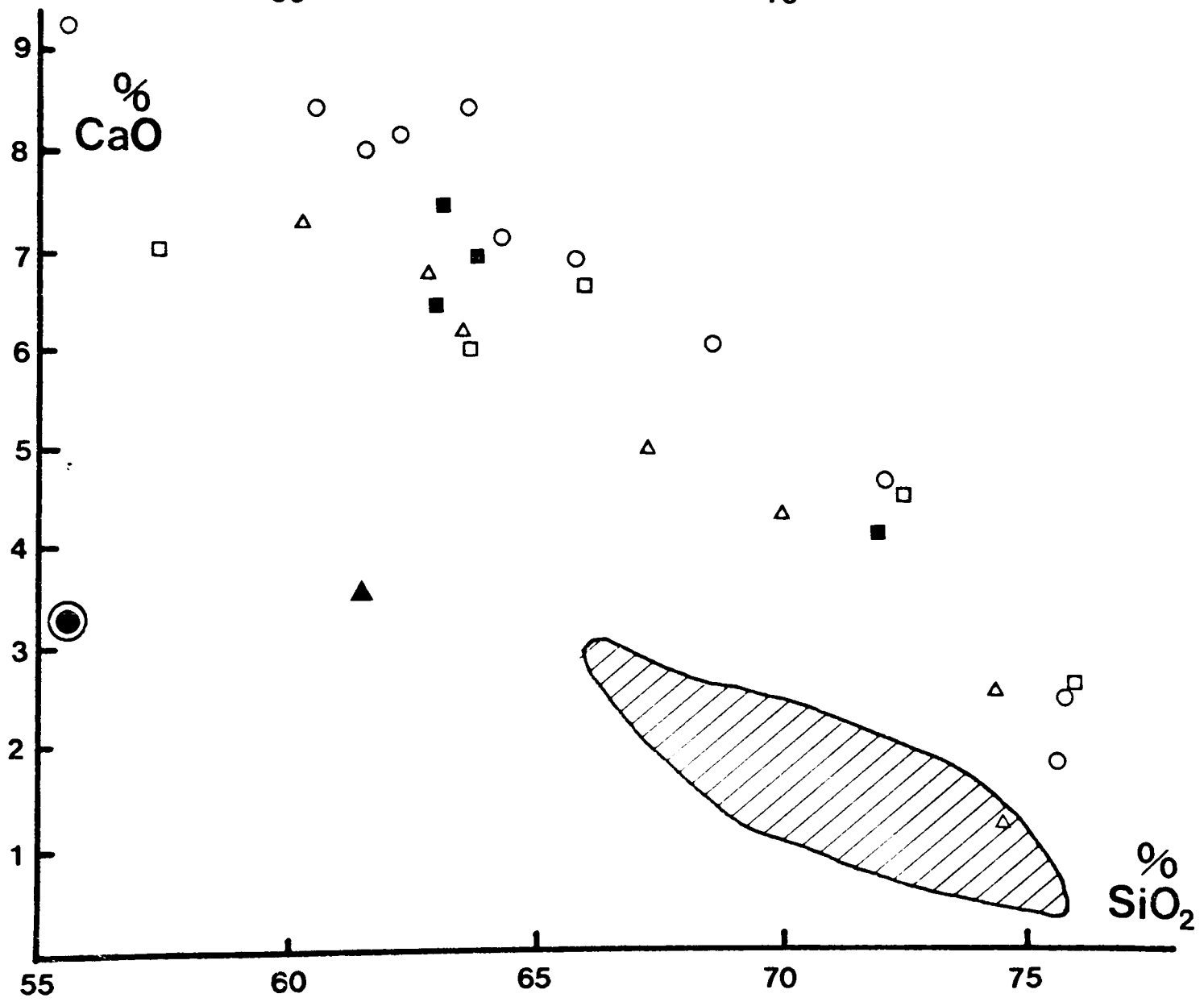
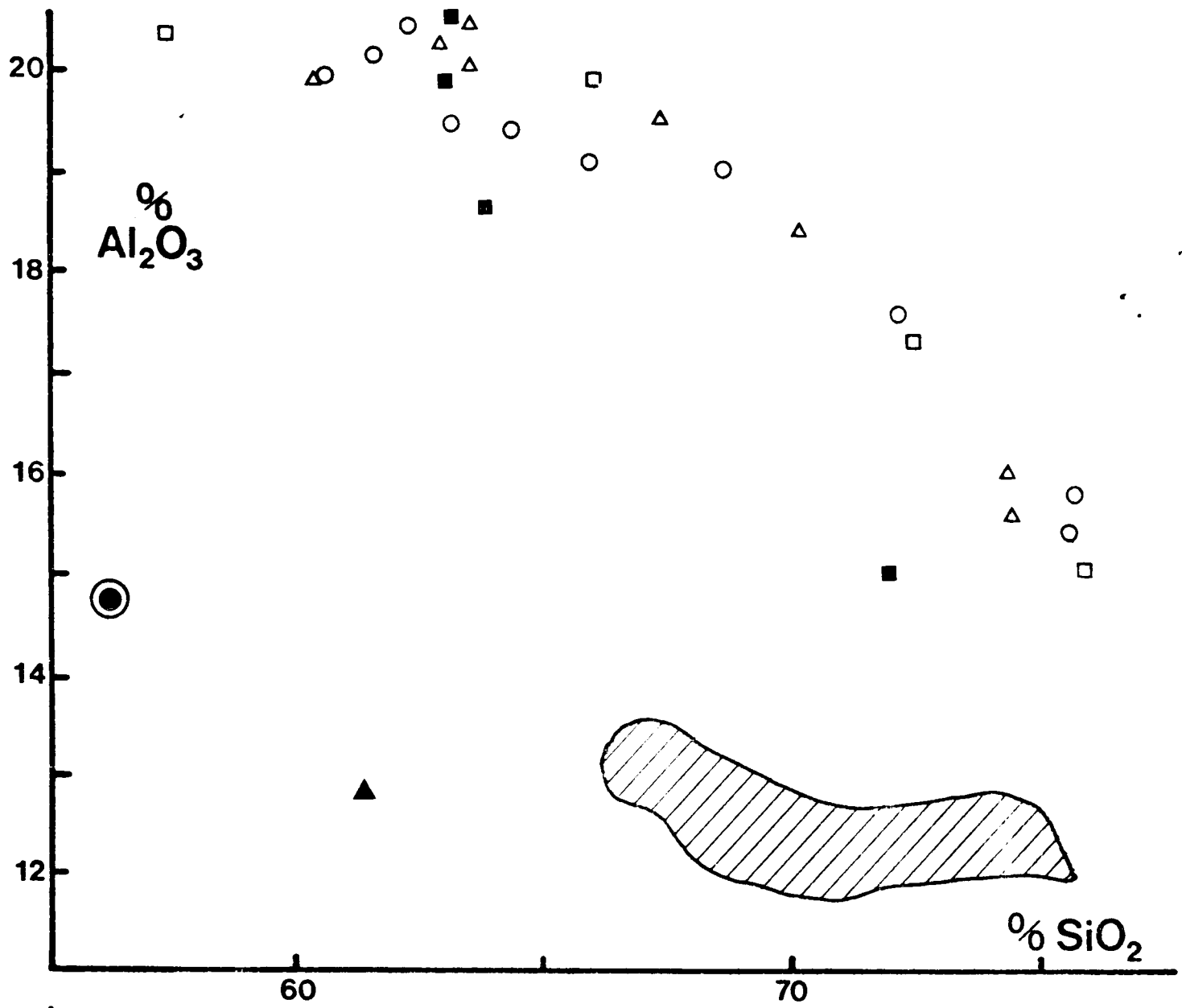
the dotted line in Figure VIII.14 which represents the melting of a source with 60% amphibole and 40% plagioclase where the first 40% of melting is assumed to be entirely plagioclase.

This early melting of plagioclase produces melts which, while being siliceous, are rich in CaO and Al_2O_3 with high normative feldspar contents (Helz, 1976; Holloway and Burnham, 1972); melts formed with residual amphibole are also corundum-normative (Holloway and Burnham, 1972). The Swazi-land rhyolites are poor in CaO and Al_2O_3 relative to these water-saturated melts (Figure VIII.15) and, in the light of the trace element data in Figure VIII.14, it seems unlikely that they were derived from a source containing amphibole.

There is a lack of experimental data on the melting relationships of anhydrous basalts with reference to melt compositions. Many basalts show cotectic, or near cotectic, crystallisation of clinopyroxene and plagioclase. Partial melting of such an assemblage, apart from an initial quartz-rich fraction, should be nearly modal; the calculated mode (by least squares) for RC 111 is: 49% plagioclase, 41% clinopyroxene and 10% ore; and with 15% melt (Table VIII.8): 39% plagioclase, 39% clinopyroxene, 9% ore and 15% melt (RC 128). The contributions of the phases to the melt are thus: plagioclase: 80%, clinopyroxene: 15%, ore: 5%. Taking this non-modal melting into account makes a negligible difference to the partial melting curves in Figures VIII.12 and 14.

Several of the more evolved rhyolites, but not the granophyres, have trace element contents in excess of those possible by simple crystal fractionation from a single parent. These enriched flows, apart from being more silica-rich, are similar to the rest of the rhyolites and appear from the major element data to lie on the same fractionation trends. They are

Fig. VIII.15 Comparison of the rhyolite compositions with those produced from partial melts of wet basalt. Cross-hatched area and large symbols represent the compositions of the Swaziland rhyolites; solid squares from Holloway and Burnham (1972); open symbols from Helz (1976): circles = Kilauea tholeiite (QFM); squares = Kilauea tholeiite (HM); triangles = Picture Gorge tholeiite. All experimental data on glasses is recalculated anhydrous.



enriched in the rare earth elements, Nb and Y, less so in Zr and depleted in Rb relative to the granophyres at the same silica level. They are not appreciably enriched in zircon or apatite and these high trace element contents seem to be primary features related either to the process of derivation or to the chemistry of the source.

It is possible to produce these enrichments by partially melting, rather than crystallising, the dacitic source. Small degrees of partial melting of a basaltic source would produce melts too enriched in Rb relative to Rb/Sr (Figure VIII.12). The composition of a partial melt is controlled by the residual mineralogical assemblage; if the dacitic source contained no residual zircon the derived melts would be enriched in those elements, normally partitioned into zircon, relative to the liquids, precipitating zircon, formed on the fractionation trend.

The acid rhyolites are enriched in Ce, Nd and Y by a factor of 1.5 relative to the basic rhyolites, and in La by a factor of 2.5. Rhyolites with less than 68% SiO₂ do not contain any zircon; 50% melting of one of these can occur without noticeably affecting the Rb/Sr systematics (Figure VIII.13) but would produce enrichments of about 1.5 (for a $D = 0.2$), sufficient to account for the variation in Ce, Nd and Y but not La.

The enrichments are, however, not all consistent with the partial melting model, even allowing for later fractionation. The Rb-Sr systematics place a close constraint on the amount of remelting possible. The major element chemistry and the one Sr-isotope analysis available for the acid rhyolites (RC 74) indicate that these acid rhyolites are genetically related to the rest; their trace element abundances suggest a partial melting trend, but some of the anomaly may be due to the chemistry of the source.

VIII.14 Role of volatiles

The rhyolites are interpreted as ignimbritic in origin and volatiles must have played an important part in their emplacement, if not in their magmatic evolution. Constraints are placed on the composition of the fluid phase by the absence of hydrous phenocrysts in the rhyolite- the enrichment of Y and the K/Rb relationships also argue against the presence of residual amphibole in the source.

It is unlikely, however, that the rhyolite source was anhydrous; water, if not in amphibole, may be concentrated along grain boundaries and in alteration products: chlorite, serpentine and clay minerals. Small amounts of water will lower the liquidus and facilitate melting. At low pressures the hornblende stability curve has a positive P/T slope (Stern et al., 1975) and it may be possible to melt a water-saturated basalt at 5 kb and 1100°C where hornblende is not stable.

Carbon dioxide has a small but finite solubility in granitic magmas at crustal pressures (Holloway, 1976). The presence of CO₂ in a granitic magma will have a small effect on the liquidus temperatures owing to its small solubility; the major effect will be in the reduction of the activity of water.

One of the major arguments against water in rhyolitic magma has been that the water-saturated solidus has a negative slope (Burnham, 1967) and hence H₂O-water saturated acid liquids will crystallise before reaching the surface. Addition of CO₂ will lower a_{H_2O} and may result in positive P-T paths for the solidus; Burnham and Davies (1974) have shown that the granite solidus can have a positive slope for H₂O:CO₂ ≤ 0.4.

Little more than this can be said about the role of volatiles. Phenocryst

studies may be able to provide estimates of magma temperature and details of small scale chemical variation attributable to vapour phase changes. Many of the quartz phenocrysts contain inclusions; many of these are glass or solid and few fluid inclusions have been seen.

Fumarolic pervasive breccias and tuffisite veins are found within the rhyolites though there is no evidence for fumarolic mineral deposits. Tazieff (1970) has argued that CO_2 is important in ignimbrite eruptions, drawing analogies with air- CO_2 density currents.

The evidence from the rhyolites is for the presence of a volatile phase but, in general, against the presence of water. A high proportion of CO_2 in the gas phase can explain many of the features observed.

VIII.15 Production of melts from acid crustal materials

It has been shown in the preceding sections that the variation in the rhyolites can largely be explained by fractional crystallisation of a dacitic magma. This magma has a composition compatible with differentiation of a basaltic source but the underlying Archaean granitic crust also provides a potential source region.

Sr-isotope considerations indicate that the source region for the main body of rhyolites can not have had an Rb/Sr ratio of greater than 0.03 (cf. Chapter VII). The most basic rhyolite has an Rb/Sr ratio of 0.36, ten times this; 25% melting of a granitic source will produce an enrichment in Rb of 3.7 and in Sr of 0.37 representing a tenfold increase in Rb/Sr. As RC 128 (the most basic dacite) has 88 ppm Rb and 244 ppm Sr, this implies a source composition of 24 ppm Rb and 650 ppm Sr.

TABLE VIII.10

Comparison of Karroo rhyolites with
Archaean basement rocks

Element	RC 128	RC 42	Tonalite
Rb	88	95	39
Sr	244	158	670
Ba	1091	500	500
La	67	28	16
Ce	177	64	20
K	3.5	3.4	1.7
Rb/Sr	0.36	0.6	0.05

Tonalite analysis from Condie and Hunter (1976)

Trace element analyses of the Swaziland Archaean rocks (Condie and Hunter, 1976) show that many of the samples with the required low Rb/Sr ratio are very depleted in Rb, and the other incompatible elements, and are hence unlikely source materials. Analyses from the Granodiorite Suite and of the tonalite diapirs have approximately the right Rb-Sr chemistry (Table VIII.10) but are depleted in the rare earths ($Tb = \sim 6$ chondrite). Data on Y, Zr and Nb are unavailable.

In contrast, the rhyolitic interbeds have high $^{87}Sr/^{86}Sr$ ratios and low abundances of La, Ce and Nd, and are therefore expected to be similar to a melt derived from a crust with the above characteristics. It is considered that, although the main group of rhyolite may possibly be differentiates of Precambrian crustal materials, the composition of the source required to produce them is atypical of the surface crustal rocks.

VIII.16 Geochemical Provinces

Cox, Macdonald and Hornung (1967) were able to demonstrate that the Karroo basalts are divisible into two provinces: a northern province covering the Rhodesian localities: Nuanetsi, Victoria Falls, Nyamandhlovu, Wankie and Featherstone; and a Southern province consisting of the basalts of Lesotho and Swaziland and the South African dolerites. They concluded that relative to the southern province, the northern province shows:

1. Absolute enrichment in K, Ti, P, Ba, Sr and Zr by factors of approximately 2-4.
2. Probable enrichment by a similar amount in Nb and Y.
3. A ratio of Ni/Mg twice that of the Karroo dolerites and Lesotho basalts but the same as the Swaziland basalts.

TABLE VIII.11

Comparison of Northern and Southern Karroo provinces

Basalts

	Zululand ¹	Swaziland ²	Nuanetsi	
			Lower Basalts ³	Upper basalts ⁴
Al ₂ O ₃	14-17	11-15	7-12	13-17
K ₂ O	0.2-1.0	0.2-1.0	0.5-3.0	0.5-3.5
TiO ₂	1.0-1.6	1.5-2.5	2-3.5	0.5-3.5
P ₂ O ₅	0.1-0.5	0.1-0.5	0.3-0.6	0.1-0.6
Ba	100-500	100-200	800-1200	-
Sr	100-400	200-300	600-1200	-
Zr	100-200	100-250	250-450	-
Y	20-35	20-45	30-50	-
Nb	2-8	5-12	12-30	-

Rhyolites

	Zululand + Swaziland ^{1,2}	Nuanetsi ⁵
Al ₂ O ₃	12-13	13-16
K ₂ O	4.5	5.0
TiO ₂	0.4-0.6	0.2-0.4
P ₂ O ₅	0.1-0.2	0-0.1
Ba	1100-1400	1000-3000
Sr	100-200	30-100
Zr	700-1200	300-700
Y	100-150	40-100
Nb	60-80	70-110

Major elements in %; trace elements in ppm.

1. Bristow (1976); 2. This study; 3. Jamieson (1969); 4. Erlank et al. (unpub. data); 5. Weaver (unpub. data).

4. Slightly reduced ratios of CaO/MgO and $\text{Al}_2\text{O}_3/\text{MgO}$."

Further data are now available on the Nuanetsi rhyolites (Weaver, unpub. data, 1976), Swaziland rhyolites and dolerites (this study), Nuanetsi Upper basalts, Zululand rhyolites and basalts (Erlank et al., unpub. data, 1977; Bristow, 1976), and data on the Kruger Park localities will soon be forthcoming from the Cape Town research group.

It has been shown in this Chapter that there is an intimate relationship between the basalts and rhyolites in Swaziland; possible crustal remelts also exist and reinforce the argument that the majority of the rhyolites are consanguineous with the basalts. For the province model to be realistic, therefore, the rhyolites should differ in the two provinces but also reflect the differences observed in the basalts.

The province argument is clouded by the presence of chemical variation within each group. In particular the Nuanetsi basalts fall into two groups the lower Mg-rich group and the tholeiitic Upper basalts. The Mg-rich group have positive MgO-incompatible element correlations and at 15% MgO can contain up to 2.5% K_2O . The Upper basalts show more normal negative MgO-incompatible element correlations. Viewed as a whole, therefore, the Nuanetsi basalts show an increase of, e.g. K_2O with MgO for MgO values greater than about 7% and an increase in K_2O with fractionation (i.e. with falling MgO) for MgO values less than 7%. The scatter in both groups is large - about 1% K_2O absolute.

The original northern province was based on analyses of the lower basalts; the new upper basalt analyses do not have the same extremes in chemistry and are more comparable with the basalts in the central and southern Lebombo. Relative to the lower basalts, the upper basalts are equally rich in K_2O ,

enriched in Al_2O_3 and slightly depleted in TiO_2 and P_2O_5 . The main difference between the Nuanetsi upper basalts and the southern province basalts is in K_2O ; the differences in P_2O_5 and TiO_2 between the two provinces are much less marked when the upper basalts are considered (Table VIII.11).

The various data available on the rhyolites show that, in general terms, the two provinces are still recognisable- the Nuanetsi rhyolites are richer in Al_2O_3 , K_2O , Rb, Ba and Nb than the southern rhyolites but poorer in TiO_2 , P_2O_5 , Sr, Zr and Y. The behaviour of these last five elements and Al_2O_3 is thus opposite to the generalised features originally proposed to distinguish the provinces. Erlank et al. (pers. comm., 1977) reported that a gradual increase of Sr with decreasing latitude is observed in the rhyolites over the length of the Lebombo exposed in South Africa. It is not known whether these data have been normalised or whether the geographical variation simply represents an increase in basicity going north. It is also not known when the Sr-enrichment observed in the Kruger Park inverts to the Sr-depleted rhyolites in Nuanetsi. The succession in Swaziland is the thickest observed and shows the most variation; the more acidic and quartz-phyric flows are only found in Swaziland.

The cross-over relationships observed for some elements, i.e. where the northern lower basalts are richer and the rhyolites poorer in an element than the southern province rocks, imply that the petrogenetic processes proposed for the southern Lebombo may not be simply applicable to the northern province. The rhyolites are, except for a short stretch near the Limpopo where the Cretaceous overlaps on to the basalts, continuous from Swaziland to Nuanetsi and it seems illogical to propose a completely different origin for the rhyolites of the two areas. Part of the problem lies in the difficulty of deciding, in suites with a large variation, the true nature of the parental material. The basalts in Nuanetsi show

considerable variation in major and minor elements and in initial Sr-isotope ratios (Manton, 1968; Norry, 1977) which is attributable to an inhomogeneous source region; this scatter is also present in the rhyolites and the data do not define the clear trends found in Swaziland. If the cross-over relationships are real, it implies that for the Nuanetsi rhyolites the residue is relatively rich in TiO_2 , P_2O_5 , Sr, Zr and Y and poor in Al_2O_3 and hence is probably controlled by the residual mineralogy of the source (e.g. amphibole). The upper basalts bear the same relationship to the lower basalts that the rhyolites do; it may be that the 'cross-over' occurs within the basalts and the ultimate source for the rhyolites may show the same depletions as the rhyolites. Trace element data on the upper basalts would help to clarify the issue.

A further possibility is that, in spite of the continuous nature of the rhyolite outcrop between Swaziland and Nuanetsi, the Nuanetsi rhyolites have a greater crustal contribution than those of the southern Lebombo. There would then be no reason to expect similarities between the rhyolites and basalts of Nuanetsi or between the rhyolites of the two areas. The parallel enrichment in K_2O in the basalts and rhyolites of the two provinces may then be coincidental.

The absence of data from the area immediately to the south of Nuanetsi prevents any conclusions being drawn on the nature of the boundary between the two provinces. The data available indicate a discontinuous variation between the provinces.

VIII.17 Conclusions

Analysis of the chemical variations observed in the rhyolites shows that most of the samples belong to a single trend which can be explained by

crystal fractionation of an original dacitic magma. The more basic, interbedded rhyolites do not appear to belong to the same trend, or relate to it by fractional crystallisation, and are probably produced by differing degrees of partial melting of the original parental material. The most acid of the rhyolites are enriched in several of the trace elements to an extent not possible by crystal fractionation of the original dacitic magma; they are probably produced by partial melting of a previously, and hence enriched, source region. The uniformity of most of the flows may be due to similarities in the source region chemistry and in the processes of melt extraction; or, given the presence of a magma reservoir, the compositions of the melt decrements may be buffered, the compositional variations smoothed out by large-scale homogenisation within the magma chamber.

The primary dacitic magma is probably formed by differentiation of basaltic material, either magmatic or previously solidified; other acid flows, within the basalts have a very different chemistry and are probably derived from acid crustal materials. Trace element considerations indicate that the source, for the former flows, cannot have contained large amounts of amphibole or garnet; these and the major element data are compatible with an origin from a clinopyroxene-plagioclase source. The process of differentiation may be either fractional crystallisation or partial melting; both are compatible with the data.

One of the most striking features of the Lebombo-Nuanetsi Province is the pronounced bimodality of rock-types with a scarcity of rocks with silica contents between 55 and 60 percent (see Cox, 1972). This silica gap is a prominent feature of many volcanic terrains and has yet to be satisfactorily explained. If the acid rocks are derived by partial melting of either the basic material or acid crustal materials, a discontinuity in the distribution of magma types will be expected. If the acid magmas have

evolved by fractional crystallisation of the basic magma however, a continuum of compositions will be available for sampling and eruption and the problem becomes one of suppression of the intermediate magmas. For example, in the Aden volcano, Hill (1974) postulated eruptive cycles characterised by initial periods of energetic basic volcanism succeeded by periods of quiet crystal fractionation. When the magmas were largely solidified, the volatile pressure had increased to the point where explosive eruption of the acid magmas could take place; intermediate magmas, although present, were not erupted except in unusual circumstances until the later stages of Caldera formation.

This is a plausible example of an eruptive mechanism which discriminate strongly against intermediate magmas. It is however debatable whether such a mechanism can be plausibly extended to a structure as large as the Lebombo.

The Swaziland data by themselves do not support any simple explanation. They can more readily be explained by the partial melting model as it can produce the variations observed in the more basic rhyolites and, given a fractionated source, result in the enrichments present in the acid rhyolite. It may be possible to elaborate further on this point when the data from the whole Karroo province have been synthesised, but it is outside the scope of the present work.

The geographical variation, resulting in the identification of a northern and a southern province, does not seem to be simply explicable in terms of regular variation in source chemistry. The rhyolites of the two areas have differences which are not mirrored in the basalts and their genetic relationship is uncertain.

The geophysical, tectonic and thermal problems of generating large volumes of acid and basic magmas are discussed in the next chapter.

CHAPTER IX

SYNTHESIS

IX SYNTHESIS

The Lebombo contains large and approximately equal thicknesses of rhyolite and basalt, the rhyolite overlying the basalt with a modicum of interbedding. Apart from a few isolated flows which have characteristics of an origin by crustal anatexis, the chemistry of the rhyolites seems to be intimately related to that of the basalts; the more basic interbedded acid flows and some of the most acid rocks have compositions suggestive of an origin by partial melting of basaltic material; the rest of the rhyolites have a composition compatible with fractional crystallisation of a dacitic magma which may itself have been produced by partial melting of solidified basalt or fractional crystallisation of basaltic magma. An origin for the main group of rhyolites by crustal fusion is probably ruled out by the relationships in the granite system (Figure IX.1) and by the low initial strontium isotope ratios of the rhyolites.

The eruption of the rhyolites as ignimbrites required the presence of a substantial fluid phase, if only at the time of eruption. The absence of hydrous phases and the agreement of the rhyolite compositions with the dry granite minimum (Figure IX.1) indicate that water cannot have played a significant part in the evolution of the rhyolite magma. The effect of large amounts of carbon dioxide in the fluid phase would be to reduce the activity of water whilst maintaining the total fluid pressure. Magmas formed under conditions of a mixed H_2O - CO_2 vapour phase however will be enriched in H_2O relative to CO_2 because of the former's greater solubility in silicate melts.

The presence of a dry, or CO_2 -rich, vapour phase is also compatible with the inferred mineralogy for the source. Trace element data indicate that the rhyolites were derived from a basaltic source with residual clinopyroxene and plagioclase; they are incompatible with a source containing amphibole or

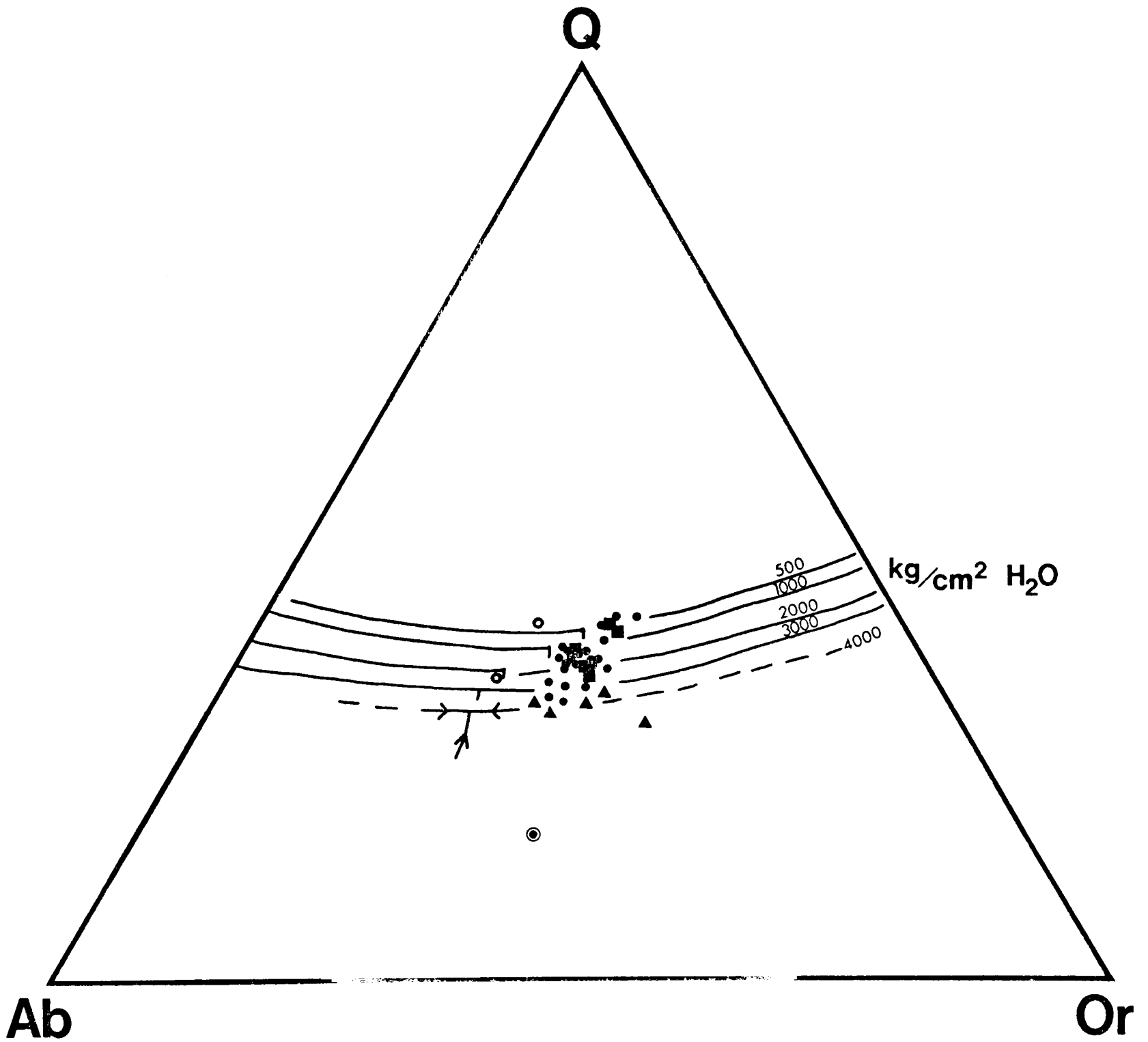


Fig. IX.1. Q-Ab-Or diagram for the Swaziland data. Minima taken from Tuttle and Bowen (1958). The rhyolites only plot in the albite field for low values of water pressure. The rhyolitic interbeds plot significantly closer to the granite minima.

(Dots = rhyolites; squares = granophyres; triangles = interbedded rhyolites; open circles = rhyolitic interbeds; dot with circle = trachyandesite).

garnet. The assemblage clinopyroxene plus plagioclase is stable over a wide range of pressure and temperature if dry; garnet is developed at high pressures, and amphibole in the presence of water. If the basalt is under-saturated with water, a three-phase assemblage, clinopyroxene, plagioclase and amphibole, may be developed. Under these conditions amphibole will melt incongruently to form clinopyroxene plus liquid and therefore will not be a residual phase after appreciable degrees of melting.

Approximately ten per cent rhyolite can be generated from a basaltic source. The rhyolite succession in Swaziland is 5 km thick and at least 50 km in width, making a volume within Swaziland of about 30,000 cubic kilometres; the volume of parental basalt must be therefore approximately 300,000 cu. km., implying, if exactly underlying the rhyolites, a vertical thickness of some 50 km. No fissures or other potential sources for the ignimbrites have been identified and it is probable that they lie to the east of the present rhyolite outcrop, buried by the monoclinal flexuring. The residual basalt in the source would therefore also lie to the east of the Lebombo.

The mechanisms for generating acidic magma from previously solidified basalt can be parsimoniously restricted to two principal modes: the melting of previously erupted material, and the melting of basic material intruded at low crustal levels.

The melting of previously erupted material is a potential mechanism for producing continuous evolution of magma: as the superincumbent loading increases, the base of the volcanic pile is depressed into the zone of melting; the ensuing magma erupts on to the surface, and aids the continuing depression. For rhyolite magmas however it is not a plausible model because of the need to process ten units of source material for one of rhyolite. The process is only likely to be significant for the uppermost lavas of the suc-

cession where the base has been depressed by some 15 km.

The melting of deep-level basalt is energetically inefficient, requiring the production, intrusion, solidification and remelting of basalt; it cannot be envisaged to be responsible for the whole rhyolite succession but may produce the few, volumetrically unimportant, flows which have the characteristics of partial melting.

The production of acidic magmas from basaltic ones by fractional crystallisation is more efficient, though the problem of the suppression of the basic and intermediate members of the continuous process remains. All the rhyolite flows seem to be the results of explosive eruptions; the critical factor determining the eruption of magmas may be the volatile pressure; only after extensive crystallisation is the volatile pressure sufficient for eruption to take place.

There is a correlation of chemistry with height in the succession, with the most basic rhyolites in the middle. The percentage of fractionation may be proportional to the rate of eruption; the most basic rhyolites corresponding to the peak of magmatic activity.

The problems remaining are those of energy and space: the source of the heat required and the disposal of the third of a million cubic kilometres of basalt left over from the production of rhyolitic magma. The Lebombo originated as a flexure along a continental margin related to the break-up of Gondwanaland in the late-Karoo. Crustal extension has implications of crustal thinning (Cox, 1970) and gravity data suggest that the continental crust beneath the Lebombo may have been thinned (Darracott, 1974). Extension and upwarping of the base of the crust can produce steep thermal gradients and induced melting. The separation of two continental masses

provides a space for the large volume of residual basalt magma.

The environment of crustal attenuation and extension can provide the energy required for large amounts of magma. The initial phase of the Karroo magmatism is a basaltic one, with high-magnesia lavas found in Nuanetsi and in the northern part of the Kruger Park. The Lesotho and Lebombo basalts seem to be contemporaneous and probably represent part of the same initial magmatic cycle. The sudden production of acid magmas after an outburst of basic magma, discounting the interbedded flows which are of minor volumetric importance, requires a change in tectonics in order to stop or slow down the rate of basalt extrusion and to allow acid magmas to develop. The argument about rate of eruption may also apply here; by slowing the crustal extension the rate of eruption may be reduced sufficiently for more extensive fractional crystallisation to take place. The rhyolites are broadly compatible with evolution from a dacitic magma; a high level magma chamber may develop to store the rising basic magma, fractionate and erupt when the volatile pressure reaches a critical value.

A further constraint is that of the strontium isotopes: the rhyolites have a low and fairly constant $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (0.704); the basalts have generally higher and more varied values (0.704-0.708) (Erlank, unpub. data). This isotopic variation suggests the rhyolites evolved from a relatively unradiogenic, but uniform source; the basalts from one similar but more variable. It may be compatible with the tectonic model discussed, if the later magmatism, producing the acid volcanics was produced from a more primitive part of the mantle; the implication being that the initial basic magmas were derived from the upper portion of the mantle, with relatively high Rb/Sr and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, upwelling of more primitive low Rb/Sr mantle producing magmas with less radiogenic isotopic chemistry.

In summary, a model is proposed to account for the major volcanic and tectonic features of the Lebombo volcanics. It is broadly similar to the volcanic cycle of Cox (1972) though his main tenet was that the rhyolites were derived from the crust. A crustal origin is unlikely for the majority of the rhyolites in Swaziland but still may be possible for those in Nuanetsi which are all interbedded with basalt.

It is proposed that the rise of a large magma body beneath Nuanetsi began an initial phase of volcanism with the eruption of nephelinites and high-magnesia basalts. Volcanism subsequently spread to other areas, producing large volumes of uniform basalt, with occasional acid flows produced by crustal anatexis. The peak or culmination of this phase occurred soon after the beginning, succeeded by a slow decline in the basalt liquidus temperatures (Cox, 1972). The next phase of volcanic activity produced rhyolitic magma exclusively, mainly as differentiates from basic magma, with a small percentage of melts derived from previously solidified basalt or gabbro. The rate of eruption reached a maximum towards the middle of this phase with the eruption of the most basic differentiates. The end of this period of acid volcanism was marked by the intrusion of granophyre bodies, some of which are very differentiated. The cycle is repeated on a smaller scale with the extrusion of the upper basalts and their intercalated rhyolites (Little Lebombo). The Cooyana and Bumbeni complexes of Zululand may also be related to this final episode of Karroo volcanism.

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APPENDIX A ANALYTICAL TECHNIQUESA 1 SAMPLE PREPARATION

Care was taken to ensure that the samples collected were as fresh as possible and free from major veins or fracture surfaces. Weathered surfaces were trimmed in the field. The samples were split to match-box size in a hydraulic splitter, the weathered material being discarded. After jaw-crushing the material was ground for a few seconds in a tungsten-carbide mill and reduced to a fine powder after 10 minutes in an agate mill.

A 2 MAJOR ELEMENT ANALYSISPreparation of beads

Major elements, with the exception of sodium and ferrous iron, were determined by X-ray fluorescence spectrometry of fusion beads. The method has been fully described by Carter (1976) and Cooper (1975) and is based on the methods of Norrish and Hutton (1969), and Harvey et al. (1973).

Fusion beads were prepared using Spectroflux 105 (Johnson Matthey Ltd.), a mixture of 47% Lithium tetraborate, 37% Lithium carbonate and 16% Lanthanum oxide. 2.5g flux and 0.48g rock powder were used to give a flux:rock ratio of 5.3:1. The bead was pressed in an aluminium platten. The weight loss on fusion was determined.

X.R.F. analysis

Machine settings for the analysis of fusion beads are listed in Table A 1. Calibration lines for each element were constructed using the U.S.G.S. standards: G-2, GSP-1, AGV-1, BCR-1, W-1, PCC-1 and DTS-1, with the values of Flanagan (1973). A monitor standard was run alternately with the samples to correct for drift.

The data were reduced using a computer program written by S.W. Richardson (XRFOCT76 or XRFNOVA) which corrected the data for dead time, weight loss on fusion, interelement effects and mass absorption effects using the coefficients of Norrish and Hutton (1969).

Results

The average relative deviation is about 1.3% for most elements, with TiO_2 , MnO , CaO and K_2O significantly better. Approximately one quarter of the samples were run in duplicate. Inter laboratory comparison with the Department of Geochemistry, Cape Town, is shown in Table A 2.

Analysis of sodium and iron

Sodium was analysed by Atomic Absorption Spectrometry on a Perkin-Elmer 306. Sample solutions were prepared by dissolving rock powders in HF and HClO_4 . Standard solutions were prepared from B.D.H. 1000 ppm aqueous standards with additional Sr to suppress the preferential ionisation of Na . The reproducibility of analyses was poor; the analyses are only accurate to $\pm 0.2\%$ Na absolute.

Ferrous iron was determined by the Ammonium metavanadate method of Wilson (1955). Duplicates gave a relative deviation of 2%.

TABLE A 1

Major elements : X.R.F. settings

Element	Tube	Analysing Crystal	KV	mA	2 θ (peak)	Background	Counter	Attenuation	Counting Time (s)	Collimator
Si	Cr	PET	60	24	109.20	111.18	Flow	1	20	Coarse
Al	Cr	PET	60	28	145.02	147.32	Flow	1	10	Coarse
Ti	Cr	LiF ₂₂₀	60	16	86.12	88.00	Flow	1	10	Fine
Fe	W	LiF ₂₂₀	60	12	57.49	55.50	Scint	2	20	Fine
Mg	Cr	RAP	60	28	44.51	46.50	Flow	0	100	Coarse
Ca	Cr	PET	30	6	45.14	48.00	Flow	1	10	Fine
K	Cr	PET	60	12	50.70	48.08	Flow	1	10	Fine
Mn	W	LiF ₂₂₀	60	28	69.98	55.50	Flow	1	20	Fine
P	Cr	PET	60	32	89.52	85.00	Flow	1	40	Coarse

All elements analysed with vacuum and spinner on.

TABLE A 2

Comparison of major element data with those obtained from the Department of Geochemistry, University of Cape Town

	RC 36 Oxford	RC 36 U.C.T.	RC 73 Oxford	RC 73 U.C.T.	RC 131 Oxford	RC 131 U.C.T.	RC 113 Oxford	RC 113 U.C.T.
SiO ₂	67.13	69.23	75.60	77.28	69.54	70.87	50.14	50.90
Al ₂ O ₃	13.45	13.24	12.05	11.89	12.55	12.71	12.43	12.50
TiO ₂	0.78	0.77	0.17	0.16	0.56	0.54	2.42	2.42
Fe ₂ O ₃	3.93	6.31*	1.61	1.864*	2.05	5.71*	3.45	17.00*
MgO	0.64	0.59	0.09	0.133	0.49	0.43	5.12	5.06
CaO	2.46	2.50	0.36	0.41	2.02	2.06	9.37	9.31
K ₂ O	4.73	4.68	5.30	5.18	4.70	4.66	0.67	0.65
MnO	0.14	0.13	0.03	0.03	0.12	0.12	0.24	0.24
P ₂ O ₅	0.21	0.23	0.03	0.03	0.12	0.13	0.35	0.36
Na ₂ O	3.40	3.18	3.16	3.07	3.30	3.05	2.39	2.22
FeO	2.32	0.00*	0.31	0.0*	3.43	0.0*	12.50	0.0*
Vol	1.28	0.97	1.10	0.913	1.08	0.52	1.61	0.45
Total	100.49	101.84	99.81	100.96	99.97	100.8	100.58	101.12

TABLE A 2 (Continued)

	RC 121 Oxford	RC 121 U.C.T.
SiO ₂	50.11	50.40
Al ₂ O ₃	14.86	14.80
TiO ₂	1.53	1.52
Fe ₂ O ₃	3.31	13.37*
MgO	6.47	6.27
CaO	11.62	11.71
K ₂ O	0.22	0.24
MnO	0.20	0.20
P ₂ O ₅	00.16	0.17
Na ₂ O	2.43	2.18
FeO	9.39	0.0*
Vol	1.54	0.379
Total	101.83	101.23

*All iron as Fe₂O₃

A 3 TRACE ELEMENT ANALYSIS

Two groups of trace elements have been analysed: Nb, Zr, Y, Sr and Rb, and Ba, La, Ce and Nd.

(1) Nb-Rb

The five elements Nb, Zr, Y, Sr and Rb have adjacent Ka peaks and after corrections for background and interferences can be analysed precisely and accurately. A computer program, XRFNB, was modified from one written by A. Duncan to process the data. The machine settings are listed in Table A 3.

Background correction

The five peaks fall on a gently curving, near-linear background. Background blanks (the U.S.G.S. peridotites PCC-1 and DTS-1 and specpure SiO_2) were used to define the curvature and background factors have been calculated from them. The background is not completely resolved between the Nb and Y peaks, two backgrounds were measured on the low-angle side of Nb Ka and three others: between Y and Sr, and Sr and Rb, and beyond Rb. The final background is taken 3° beyond Rb to avoid a number of U, Th and Pb peaks.

Because the first three peaks are not bracketed by background measurements, three background factors are calculated for each of the peaks Nb, Zr and Y. The background factors are calculated using B1 and B3, B2 and B3, and B1 and B4 respectively. These background factors and those for Rb and Sr are calculated from the simple relationship

$$F = P/B_a + B_b$$

where: F = background factor;

P = counts from the blank at the peak of the element concerned;

B_a , B_b = the two relevant background measurements.

Interference factors

To correct for the overlap of RbK β on YK α , SrK β on ZrK α and YK β on NbK α pure dunites spiked with approximately 2000 ppm Rb, Sr and Y respectively were used as interference standards. The contributions of the interfering K β peaks to the K α peaks were calculated as a function of the interfering elements K α peak; the SrK β contribution to the ZrK α peak was approximately 7% of the SrK α peak.

Calibration

The data from Flanagan (1973) for the U.S.G.S. standards were not found to produce reasonable calibration lines; the values quoted by Willis et al. (1972) for Nb, Zr and Y and those from Pankhurst and O'Nions (1973) for Rb and Sr were used (Table A 4).

TABLE A 4

	Nb	Zr	Y	Sr	Rb
G-1	21.7	21.9	11.9	249	210
W-1	6.8	95.1	20.3	192	21.3
GSP-1	25.4	543	25.4	233	254
BCR-1	11.3	192	33.4	332	47.3

Table A 5 shows the inter-lab companion between the Oxford and Cape Town analyses.

TABLE A 3

Machine conditions for Nb-Rb run

2 θ	Element	Counting time(s).	Interference	Total counts on GSP-1
29.30	Bg	100		73908
29.80	Bg	100		68491
30.36	Nb $K\alpha$	200	Y $K\beta$	163869
32.01	Zr $K\alpha$	100	Sr $K\beta$	414352
33.81	Y $K\alpha$	100	Rb $K\beta$	91047
34.78	Bg	100		35250
35.77	Sr $K\alpha$	100	None	158743
37.03	Bg	100		26284
37.91	Rb $K\alpha$	100	None	143563
41.08	Bg	100		16670

W tube; 90 kv, 20 mA; LiF_{210} crystal; Scintillation counter; Fine collimator
Full - peak P.H.A. window.

Precision and accuracy

	Relative standard deviation (%)	Counting error (ppm)	Det. limit (ppm)	Estimated accuracy %
Nb	0.7	0.35	1	2
Zr	0.2	1.2	1	2
Y	0.5	0.5	1	2
Sr	0.5	0.6	1	2
Rb	0.5	0.6	1	2

Values for precision above calculated from 24 analyses of RC 12 (Nb = 75;
Zr = 930; Y = 130; Sr = 160; Rb = 130).

TABLE A 5

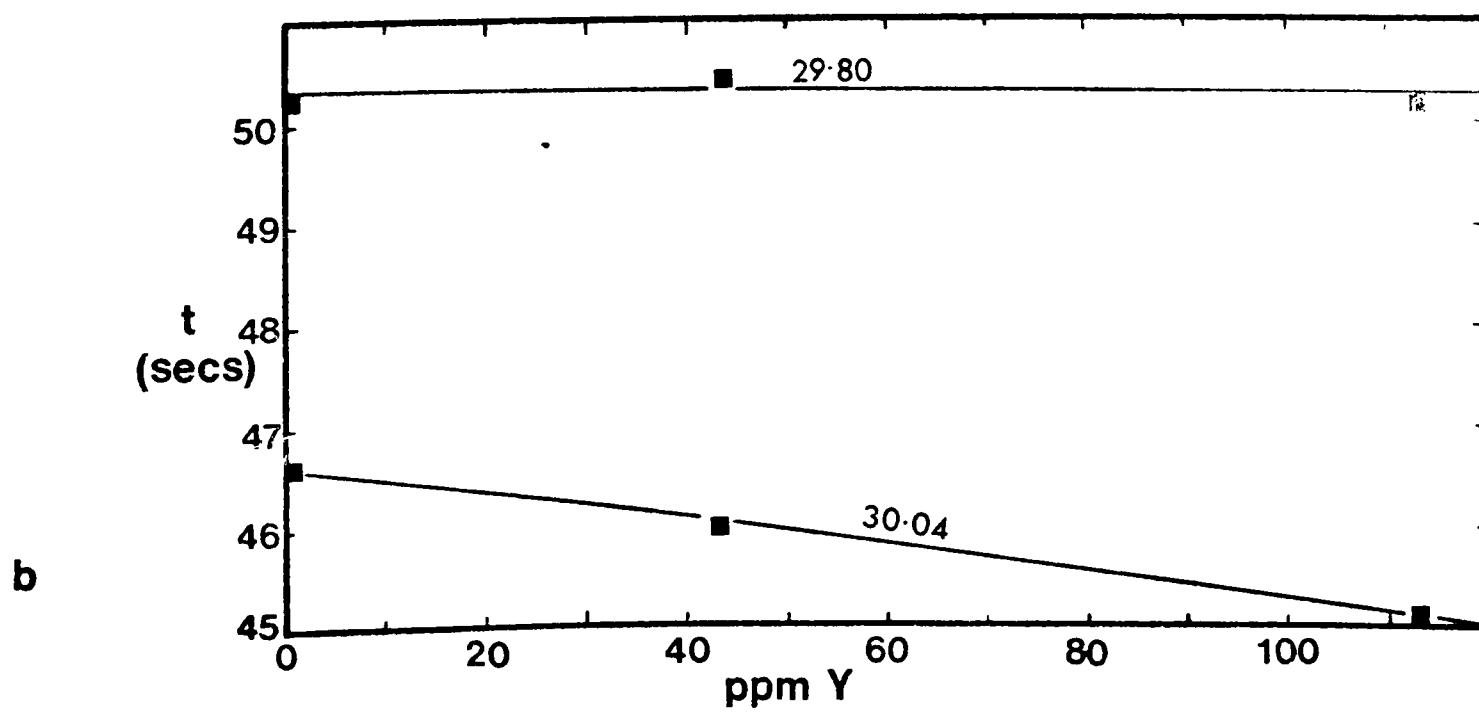
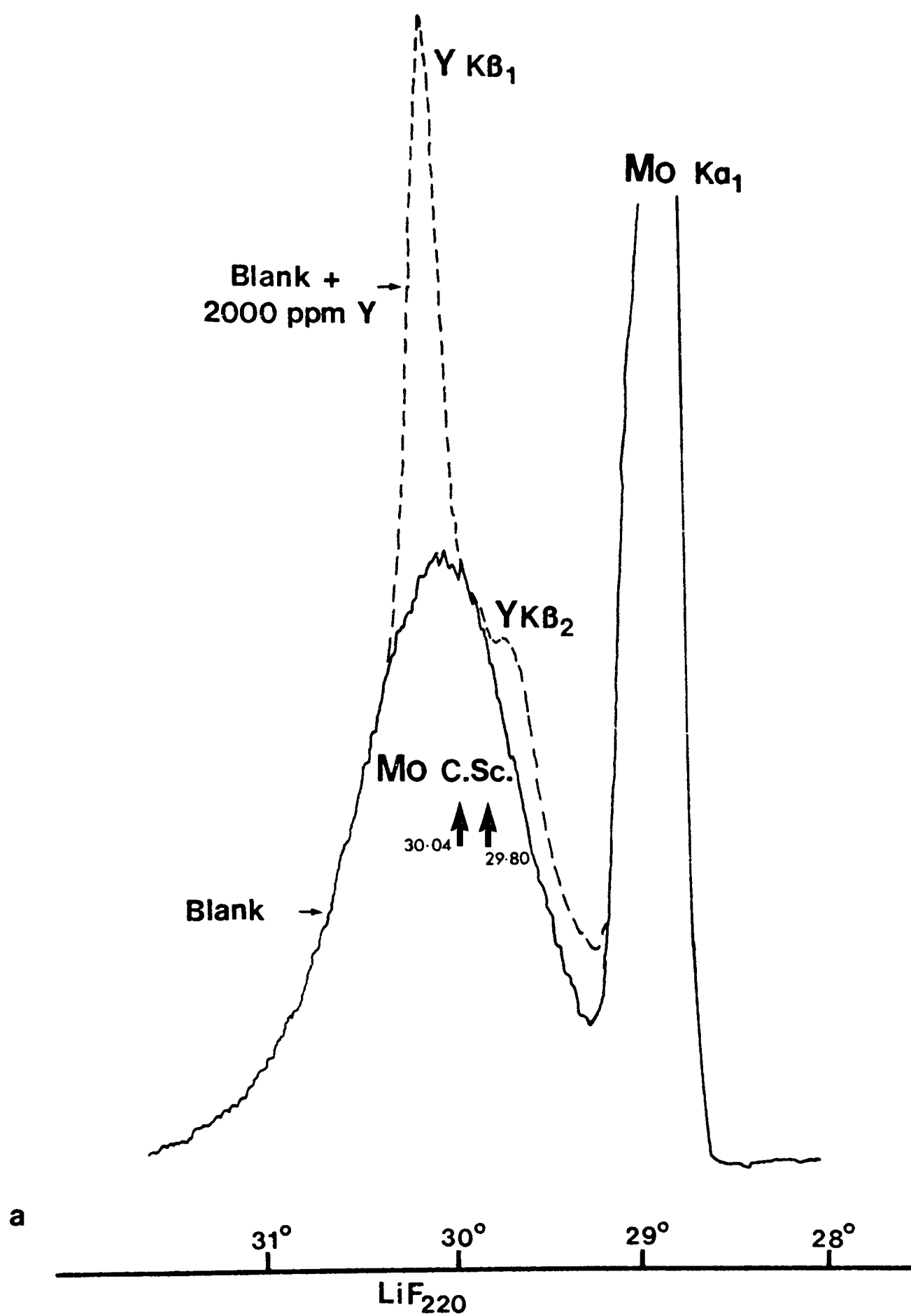
Comparison of Oxford and U.C.T. analyses (U.C.T. determinations first)

	Nb	Zr	Y	Sr	Rb
RC 36	80	1025	89	188	121
	78	1059	95	193	127
RC 73	77	237	82	35	229
	74	244	88	34	235
RC113	10	197	42	221	15
	10	202	44	234	15
RC121	4	97	23	201	4
	5	98	24	203	3
RC131	74	819	109	122	139
	73	830	117	124	144

Mass absorption correction

The molybdenum Compton scatter peak was used as an estimate of the mass absorption coefficient. There is significant interference of Y on the scatter peak (Figure A 1); this can be almost completely avoided by measuring on the side of the peak. The addition of 2000 ppm Y is sufficient to double the Compton peak and hence halve the measured mass absorption coefficient. Calculated values for the U.S.G.S. standards were used for calibration. The peak was counted till 400,000 counts had been accumulated, thus obviating the need for a dead-time correction.

Fig. A. 1 The effect of Y on the Mo Compton Scatter peak. (a) shows the relative positions of the interfering peaks; (b) shows how the count rate (or time taken to accumulate 400,000 counts) is dependent on the Y content at the two 2θ positions. By measuring the Compton peak at 29.80 the Y interference can be minimised.



Computation

A Fortran program, XRFNB, was used to reduce the raw data. The input data consisted of four groups of data: count data on the background blanks; count data on interference standards; count data, mass absorption coefficients and concentration values for the calibration standards; and finally count and mass absorption coefficient data on the unknown. No correction for drift was made; one sample was run after every five unknowns as a check for drift. After three days the count rate on the Zr peak of RC 12 had changed by 0.2%; no drift of 2θ angles was detected.

Ba, La, Ce and Nd

The elements Ba, La, Ce and Nd were analysed by X.R.F. using a blank peridotite for background correction and the background as an estimate of mass absorption. The analyses have an accuracy of 10% relative, but a precision, measured from duplicates, of about 2% relative. Machine conditions are listed in Table A 5. The data were reduced on a hand calculator.

TABLE A 5

Analysis of Ba, La, Ce and Nd : Machine conditions

2θ	Element	Counting time(s)	Comments
109.85	Bg	100	
111.74	Ce $L\beta_1$	100	No correction made for
112.79	Nd $L\alpha_1$	100	mutual overlap of Ce and Nd
111.27	Ba $L\beta_2$	100	
117.56	Bg	100	
119.72	La $L\beta_1$	100	

W tube; 70 KV, 28 mA; Flow Counter; LiF_{220} crystal; fine collimator, vacuum on.

La, Ce and Nd values normalised to chondrite values of Nakamura (1974): 0.32, 0.265, 0.630 respectively.

A 4 POTASSIUM-ARGON

Potassium and argon-isotope analyses were done at the Institute of Geological Sciences, London, using a flame photometer and an A.E.I. MS 6 mass spectrometer respectively.

Argon analysis

A 40-80 mesh fraction was sieved from whole-rock jaw-crushed aggregate. It was washed repeatedly in deionised water in an ultrasonic bath until all adhering dust had been removed. The sample was then split, half being finely ground for K_2O determinations.

The argon samples were loaded in Mo crucibles and baked and evacuated overnight. A ^{38}Ar spike was added from a calibrated gas burette and the argon extracted by R.F. induction heating. The argon was cleaned by liquid nitrogen cold fingers, titanium sponges and a zirconia getting-pump. The argon was equilibrated into the mass spectrometer and measured by manual peak-switching. The data were reduced in a Hewlett-Packard desk calculator.

Potassium

Potassium was analysed by flame photometer with solutions prepared by the dissolution of rock powder in HF. An internal Lithium standard was used. The precision was about 2%; systematic errors were corrected by normalising the data to the accepted value for the U.S.G.S. standard BCR-1. Selected samples were checked by A.A. and reproducibility of better than 3% was obtained.

APPENDIX B

Major and trace element analyses and calculated C.I.P.W. normative minerals

TABLE B 1 : Major element analyses : R.W. Cleverly and K.G. Cox

Trace element analyses : R.W. Cleverly

TABLE B 2 : Trace element analyses : R.W. Cleverly

Rock Types

R = rhyolite

IR = interbedded rhyolite

RI = rhyolitic interbed

G = granophyre

D = dolerite

B = basalt

Major elements in weight percent.

Trace elements in parts per million.

* = duplicate analysis

** = triplicate analysis

TABLE B 1

Sample name	RC 12	RC 33	RC 36	RC 40	RC 42	RC 44	RC 45	RC 47	RC 48	RC 49	RC 50	RC 52	RC 55	RC 56	RC 57	RC 58	RC 59	RC 61	RC 62
Rock type	R	R	IR	IR	RI	G	IR	R	R	R	R	R	R	R	R	R	R	R	IR
Flow Number	XIII	XIV	-	-	-	-	-	XIV	XVIII	XVII	XVIII	XI	XI	XI	XI	XI	XI	XI	-
SiO ₂	70.25	69.87	67.13	66.78	66.67	71.72	66.54	69.26	70.00	69.52	68.98	70.98	70.62	70.62	70.78	70.48	71.62	70.92	68.18
Al ₂ O ₃	12.14	12.33	13.45	13.31	12.09	12.57	13.21	12.06	12.09	12.41	12.55	12.58	12.17	12.10	12.20	12.33	12.49	12.32	12.93
TiO ₂	0.49	0.48	0.78	0.75	1.13	0.51	0.77	0.55	0.55	0.61	0.59	0.52	0.48	0.48	0.50	0.51	0.50	0.45	0.54
Fe ₂ O ₃	2.15	3.46	3.93	4.83	6.44	3.12	4.84	3.34	4.06	3.15	2.85	3.08	2.60	3.83	3.21	3.49	2.97	2.92	4.78
MgO	0.14	0.27	0.64	0.64	0.58	0.24	0.52	0.62	0.47	0.40	0.29	0.36	0.21	0.23	0.27	0.43	0.27	0.35	0.24
CaO	1.91	1.30	2.46	2.48	2.91	1.82	1.75	1.32	1.67	1.97	1.80	2.08	1.78	1.64	1.91	1.75	1.99	1.75	1.57
K ₂ O	4.72	4.49	4.73	4.33	3.36	4.55	5.79	4.48	4.44	4.35	4.23	4.37	4.33	4.53	4.35	4.40	4.56	4.54	5.29
MnO	0.11	0.12	0.14	0.14	0.08	0.11	0.14	0.11	0.12	0.14	0.13	0.13	0.12	0.11	0.13	0.14	0.13	0.12	0.16
P ₂ O ₅	0.11	0.11	0.21	0.22	0.22	0.11	0.21	0.12	0.14	0.14	0.14	0.12	0.11	0.11	0.11	0.11	0.12	0.09	0.14
Na ₂ O	3.17	3.57	3.40	3.78	3.90	3.65	2.97	3.14	3.57	3.48	3.45	3.41	3.41	3.49	3.54	3.34	3.37	3.45	3.18
FeO	2.76	1.89	2.32	1.44	1.28	2.19	1.56	2.60	2.08	3.15	3.3	2.64	2.68	1.59	2.36	2.18	2.60	2.14	1.89
Vol.	2.77	1.49	1.28	1.88	2.11	1.05	1.37	1.43	1.59	1.29	2.75	1.02	1.03	0.93	1.29	1.31	1.32	1.72	1.20
Total	100.72	99.37	100.49	100.56	100.77	101.63	99.68	99.04	100.74*	100.61	101.04*	101.28	99.52	99.66	100.64	100.48	101.93	100.31*	100.16*
Nb	75	79	78	77	12	68	78	74	74	71	70	71	72	72	74	74	73	71	82
Zr	930	869	848	1055	242	796	1060	894	911	903	855	870	830	845	865	851	845	792	1216
Y	128	170	95	96	31	115	99	125	132	127	131	125	122	126	128	130	128	115	115
Sr	161	129	193	174	158	135	182	130	145	132	134	140	136	125	136	133	142	154	219
Rb	133	131	127	117	95	131	138	132	129	122	115	125	128	134	132	132	125	134	125
Ba	-	-	1355	1360	509	-	-	-	-	-	-	-	1236	-	-	-	-	-	1571
La	-	-	94	91	28	-	-	-	-	-	-	-	95	-	-	-	-	-	97
Ce	-	-	220	205	64	-	-	-	-	-	-	-	214	-	-	-	-	-	229
Nd	-	-	106	139	28	-	-	-	-	-	-	-	107	-	-	-	-	-	118
Q	29.37	29.32	24.33	23.50	26.27	29.58	23.53	29.86	29.11	27.83	28.99	29.73	30.02	30.16	29.77	30.38	30.19	30.26	26.41
Or	27.87	26.51	27.97	25.60	19.83	26.90	34.23	26.49	26.17	25.73	25.10	25.83	25.61	26.77	25.71	26.00	26.98	26.94	31.36
Ab	26.82	30.21	28.77	31.98	33.00	30.88	25.13	26.57	30.21	29.44	29.19	28.85	28.85	29.53	29.95	28.26	28.51	29.19	26.91
An	4.96	4.37	7.45	5.56	5.57	4.48	5.61	5.56	4.06	5.39	4.73	6.10	5.10	3.97	4.56	5.68	5.48	5.12	5.53
Di	3.32	1.10	2.64	3.37	3.10	2.82	1.32	0.16	2.51	2.96	2.86	2.91	2.62	1.25	3.50	1.84	3.13	2.51	0.94
Hy	1.31	0.21	0.38	0.02	0.0	0.0	0.69	2.79	0.10	1.93	2.26	1.10	1.38	0.0	0.0	0.72	0.81	0.55	0.0
Il	3.11	5.01	5.62	2.90	1.11	4.52	3.26	4.84	5.49	4.57	4.13	4.46	3.77	4.09	4.65	5.07	4.30	4.34	5.03
Il	0.93	0.91	1.52	1.43	2.15	0.97	1.46	1.04	1.03	1.15	1.09	0.98	0.91	0.91	0.95	0.97	0.95	0.85	1.03
Ap	0.27	0.25	0.49	0.51	0.52	0.26	0.49	0.29	0.34	0.33	0.32	0.28	0.25	0.25	0.26	0.26	0.27	0.21	0.34
Hm	0.0	0.0	0.05	2.83	5.68	0.0	2.59	0.0	0.32	0.0		0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.35
C	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0		0.0	0.0	0.0	0.0	0.0	0.0	0.0	
Wo					1.43	0.18													0.03
Ru																			

Sample name	RC 65	RC 66	RC 70	RC 73	RC 74	RC 75	RC 80	RC 87	RC 90	RC 91	RC 92	RC 105	RC 119	RC 128	RC 131	RC 134	RC 136	RC 144		
	Rock		G		R		R		R		R		Kl		IR		K			
	R	Cecil Nack	R	G	R	K	Ia	XVI	XX	XXIV	XXVI	G	Kl	IR	R	R	K	R		
Flow Number	Cecil Nack	Cecil Mack	-		IX		VII		XX		XXVI		-		XIX		XXV		XIV	
SiO ₂	68.29	68.14	72.56	75.60	74.02	70.78	69.38	69.56	70.31	69.76	74.95	69.24	71.41	61.36	69.54	69.90	69.99	70.68		
Al ₂ O ₃	12.48	12.51	12.74	12.05	12.17	11.77	12.49	12.68	12.68	12.35	12.64	12.38	11.95	12.94	12.55	11.92	12.22	12.40		
TiO ₂	0.57	0.54	0.43	0.17	0.29	0.27	0.47	0.55	0.55	0.53	0.42	0.51	0.66	1.43	0.56	0.48	0.47	0.46		
Fe ₂ O ₃	4.74	4.74	1.94	1.61	3.78	3.43	5.17	3.04	2.60	3.31	2.04	2.04	2.44	4.82	2.05	3.75	3.78	2.59		
MgO	0.30	0.31	0.29	0.09	0.22	0.03	0.07	0.65	0.45	0.89	0.05	0.46	0.35	1.30	0.49	0.56	0.37	0.20		
CaO	2.07	1.91	1.70	0.36	0.16	1.47	1.42	1.68	1.78	1.62	0.70	2.07	1.70	3.74	2.02	2.17	2.24	1.15		
K ₂ O	4.41	4.33	4.87	5.30	5.13	4.80	4.80	4.68	4.58	4.68	5.27	4.34	3.67	3.48	4.70	3.98	4.42	3.21		
NaO	0.18	0.17	0.08	0.03	0.04	0.11	0.10	0.12	0.16	0.09	0.01	0.12	0.08	0.18	0.12	0.14	0.14	0.05		
P ₂ O ₅	0.13	0.11	0.08	0.03	0.03	0.06	0.11	0.13	0.12	0.11	0.07	0.11	0.13	0.41	0.12	0.12	0.10	0.09		
Na ₂ O	3.79	3.70	3.46	3.16	3.00	3.03	3.86	3.06	3.59	3.20	2.99	3.21	3.50	3.31	3.30	3.35	3.45	3.36		
FeO	2.36	1.98	2.01	0.31	0.04	1.11	1.01	2.87	3.03	1.84	0.13	3.39	2.86	3.76	3.43	0.89	1.48	0.43		
Vol.	1.10	0.97	0.69	1.10	2.00	2.64	1.54	1.45	0.98	2.11	0.90	1.45	1.85	1.63	1.08	3.17	1.93	6.22		
Total	100.38	100.39*	100.84	99.81	100.87	99.49	100.42	100.47	100.84	100.46	100.17*	99.32*	100.20*	98.36	99.97*	100.43*	100.58	100.85		
Nb	77	75	59	74	86	79	81	73	71	70	78	68	18	64	73	68	79			
Zr	1354	1321	584	244	694	894	1154	889	817	748	584	793	372	900	830	734	873			
Y	118	113	97	88	184	141	122	127	123	112	155	119	56	89	117	109	145			
Sr	177	168	100	34	102	121	163	152	141	115	149	139	129	244	124	86	132	132		
Rb	123	121	162	235	143	137	128	133	136	140	163	122	122	88	144	146	130	479		
Ba	-	1393	1104	572	1816	-	-	-	-	-	1170	-	598	1091	1261	-	-			
La	-	95	99	104	178	-	-	-	-	-	195	-	43	67	97	-	-			
Ce	-	227	183	177	305	-	-	-	-	-	260	-	93	177	211	-	-			
Nd	-	116	85	76	186	-	-	-	-	-	135	-	39	88	103	-	-			
Q	26.04	27.39	29.89	36.13	36.34	32.32	26.53	28.79	27.15	29.13	36.01	28.03	32.56	20.81	26.65	30.57	29.17	36.33		
Or	26.05	25.47	28.81	31.30	30.33	28.34	28.36	27.67	27.05	27.67	31.12	25.63	21.66	20.57	27.79	23.54	26.10	18.98		
Ab	32.07	31.31	29.27	26.74	25.38	25.64	32.66	25.89	30.37	27.07	25.30	27.16	29.61	28.01	27.92	28.34	29.19	28.43		
An	3.96	4.59	4.83	1.60	0.60	4.36	2.59	7.03	4.99	5.51	3.04	6.56	5.94	10.18	5.56	6.00	4.83	5.12		
Di	1.58	1.84	2.64	0.0	0.0	0.17	0.36	0.39	2.64	1.40	0.0	2.55	1.43	4.62	3.15	3.16	1.97	0.0		
Hy	0.0	0.0	0.91	0.22	0.56	0.0	0.0	3.49	2.58	1.56	0.13	2.75	2.61	1.98	3.50	0.17	0.0	0.51		
Mt	6.55	5.40	2.81	0.60	0.0	3.14	2.25	4.41	3.76	4.69	0.0	2.96	3.52	6.99	2.97	1.81	3.88	0.23		
Il	1.08	0.99	0.81	0.32	0.17	0.50	0.89	1.05	1.05	1.00	0.29	0.97	1.24	2.71	1.06	0.91	0.89	0.87		
Ap	0.29	0.26	0.18	0.08	0.06	0.15	0.26	0.30	0.28	0.25	0.16	0.27	0.30	0.97	0.29	0.28	0.24	0.21		
Hm	0.22	0.97	0.0	1.19	3.78	1.26	3.62	0.0	0.0	0.07	2.04	0.0	0.0	0.0	0.0	2.47	1.11	2.43		
C	0.0	0.0	0.0	0.54	1.46	0.0	0.0	0.0	0.0	0.0	0.91	0.0	0.0	0.0	0.0	0.0	0.0	1.52		
Wo.	1.44	0.76	0.0	0.0	0.0	0.97	1.37	0.0	0.0	0.0		0.0	0.0	0.0	0.0	0.0	1.29	0.0		
Ru				0.0	0.19						0.27									

Sample name	SK 2	SK 131	SK 138	SK 153	MZ 6	MZ 11	RC 110	RC 111	RC 113	RC 115	RC 116	RC 120	RC 121	MW 61
	Rock type			R			D	D	D	D	D	D	D	D
	Flow Number			TA			XX			VI			-	
SiO ₂	71.59	73.03	68.23	70.89	56.25	73.55	51.74	49.74	50.14	50.24	49.70	50.75	50.11	47.72
Al ₂ O ₃	12.66	12.68	12.17	12.32	14.88	12.48	13.62	12.97	12.43	13.36	14.27	13.28	14.86	14.94
TiO ₂	0.42	0.31	0.59	0.51	1.45	0.25	1.95	2.69	2.42	2.39	1.73	2.10	1.53	1.38
Fe ₂ O ₃	2.19	3.13	1.75	3.16	8.28	3.97	3.82	3.29	3.45	3.36	2.99	3.28	3.31	5.63
MgO	0.30	0.35	0.30	0.28	1.28	0.00	4.61	5.73	5.12	5.65	6.32	4.84	6.47	5.77
CaO	1.44	1.81	2.56	1.92	3.35	0.16	8.93	9.97	9.37	10.03	11.21	9.48	11.62	9.85
K ₂ O	4.93	4.80	3.12	4.59	4.24	5.20	0.86	0.62	0.67	0.55	0.32	0.66	0.22	1.10
MnO	0.08	0.12	0.14	0.11	0.21	0.05	0.22	0.23	0.24	0.22	0.20	0.22	0.20	0.17
P ₂ O ₅	0.08	0.06	0.15	0.14	0.50	0.02	0.39	0.38	0.35	0.26	0.17	0.21	0.16	0.13
Na ₂ O	3.47	2.96	3.12	2.84	3.84	3.60	2.71	2.23	2.39	2.55	2.05	2.78	2.43	3.37
FeO	1.78	1.57	4.49	2.11	2.93	0.12	10.55	11.20	12.50	11.29	10.14	11.39	9.39	6.22
Vol.	1.51	2.75	2.93	1.91	3.40	1.08	1.93	1.57	1.61	1.47	1.29	1.63	1.54	4.00
Total	100.40*	103.29*	99.54	100.78	100.87*	100.42*	101.33	100.60*	100.58*	101.36*	100.41	100.61	101.83	100.27*
Nb	62	85	73	73	64	83	12	12	10.5	8	5	12	5	3
Zr	598	842	896	821	996	646	241	196	202	150	100	210	98	85
Y	99	147	127	119	112	151	37	37	44	34	24	37	24	22
Sr	95	102	191	131	219	89	304	242	234	234	207	218	203	318
Rb	158	129	143	125	105	141	14	10.3	15.5	12.3	6.8	14.3	3.4	67
Ba	1139	-	-	-	1330	-	-	-	-	198	79	-	94	-
La	101	-	-	-	61	-	-	-	-	22	16	-	11	-
Ce	195	-	-	-	192	-	-	-	-	36	22	-	20	-
Nd	98	-	-	-	97	-	-	-	-	20	6	-	10	-
Q	29.21	33.52	30.23	32.66	10.76	32.44	5.82	4.33	4.44	1.94	2.92	3.37	1.37	0.0
Or	29.32	26.37	18.44	27.13	24.94	30.73	5.11	3.73	3.90	3.17	1.88	3.87	1.32	6.56
Ab	29.36	25.04	26.40	24.03	32.49	30.46	22.93	18.87	20.22	21.58	17.34	23.52	20.56	28.51
An	4.32	7.14	9.99	7.31	10.82	0.61	22.45	23.80	21.06	22.82	28.81	21.84	28.97	22.39
Di	1.92	1.13	1.52	1.07	1.97	0.0	16.06	19.49	19.67	20.74	21.15	19.90	22.67	20.58
Hy	0.65	0.36	6.04	0.80	2.08	0.0	16.89	18.46	19.58	18.33	18.99	17.26	17.34	1.60
Nil	3.17	4.54	2.54	4.58	5.92	0.0	5.53	4.78	4.92	4.76	4.33	4.75	4.80	7.94
Il	0.8	0.59	1.12	0.97	2.75	0.36	3.71	5.09	4.63	4.55	3.29	3.99	2.91	2.62
Ap	0.19	0.14	0.35	0.33	1.18	0.05	0.91	0.90	0.81	0.60	0.41	0.49	0.37	0.29
Hm					4.19	3.97								0.0
C						0.71								5.61
Wo														
Ru						0.06								

TABLE B 2

Trace element analyses of rhyolites and basalts

Sample name	Rock type	Nb	Zr	Y	Sr	Rb
RC 37	IR	62	908	91	229	92
RC 39	IR	75	1037	102	167	125
RC 71	B	3	118	27	282	16
RC 79	R	81	1159	123	151	126
RC 101	IR	77	1106	111	148	118
RC 126	B	7	140	29	232	11.3
RC 127	G	72	1055	117	173	100
RC 133	R	72	855	151	128	122
SK 13	R	80	1213	110	175	127
SK 14	IR	80	1017	98	127	124
SK 20	G	79	659	135	108	141
SK 25	IR	86	1103	102	155	110
SK 31	IR	76	1033	94	151	129
SK 32	IR	80	1171	101	210	116
MW 26	B	5	120	24	262	15
MW 50	R	65	886	99	132	72
MW 51	R	80	1241	139	156	114

APPENDIX CSample descriptions and localities

Latitude and longitude are given for all the samples which have been analysed. All the samples are marked on the maps except RC 65, 66 (Cecil Mack Pass), and RC 119, 120, and 121 (West of Nsoko).

Modal analyses are given for the majority of the samples analysed. These are based on point counts of a single, randomly cut thin-section. An average of 1,500 points were counted per section. In the modal analyses a cross indicates the presence of a mineral, a dash the absence of a particular mineral.

MODAL ANALYSES

Sample name	Lat°S	Long°E	Locality Details	Sample Description	Ph- eno	Qz	Fsp	Cpx	Fa	Ore	Zr	Ap
RC 12	26°37'	32°02'	Base of flow XIII; by road 2 km N of Tikuba.	Massive rhyolite.	18	0.5	14	3.3	-	0.8	x	x
RC 33	26°35'	32°02'	Base of flow XIV; 2.5 km Maphungwane store.	Fine-grained rhyolite, weathered feldspar.	x	x	x	x	-	x	x	x
RC 36	26°50'	31°59'	Interbedded rhyolite. Twin Ridge estate 5 km SE Big Bend. Easterly of two flows.	Grey matrix with eutaxitic banding.	19	0	13.4	3.4	-	2.1	-	x
RC 37	26°50'	31°59'	1 km SW RC 36. Westerly of two flows.	Similar to RC 36, no banding present.	-	-	x	x	-	x	-	x
RC 39	26°41'	32°00'	Interbedded rhyolite, 14 km N of RC 36.	Slight compaction banding. Glomeroporph. texture occasionally developed.	-	-	x	x	-	x	-	x
RC 40	26°40'	32°00'	Interbedded rhyolite; 1.5 km north of RC 39.	Slightly weathered. Similar to RC 39.	17	0	14	0.3	-	1	x	x
RC 42	26°38'	31°56'	Rhyolitic interbed. Boulders by road 2 km west of Magwan- yana.	Elongate amygdales. Very weathered feldspar. Pseudo- morphs after Fa.	-	-	x	x	x	x	-	x
RC 44	26°37'	31°59'	Granophyre dyke; 6 km WNW Tikuba.	Microgranophyric matrix, med. grained.	21	1	14	4.3	-	1.5	x	x
RC 45	26°37'	31°59'	Interbedded flow 1 km N of RC 128.	Weathered rhyolite. Glomero- porph. texture. Haematised	-	-	x	?	x	x	-	x

				Ph- eno	Qz	Fsp	Cpx	Fa	Ore	Zr	Ap
RC 47	26°37'	32°04'	Flow XIV. 1 km east of Tikuba.	20	1	14.2	1.8	-	2.3	x	x
			Fine-grained rhyolite. Quartz. Irregular banding.								
RC 48	26°36'	32°05'	Flow XVIII. Grenspan Farm.	18	4.2	8.7	3.1	-	2.2	x	x
			Grey fine-grained massive rhyolite.								
RC 49	26°35'	32°04'	Flow XVII. 1 km N. Grenspan.	23	1.4	17	4	-	1	x	x
			Massive rhyolite.								
RC 50	26°34'	32°05'	Flow XVIII 3 km N. Grenspan. Mtibhlati River.		x	x	x	-	x	x	x
			Grey massive rhyolite.								
RC 52	26°36'	32°01'	Flow XI by Maphungwane Store.	19	1.2	10.4	5	-	2.3	x	x
			Fine-grained massive rhyolite.								
RC 55	26°35'	32°00'	Flow XI. 2 km NNW Maphungwane.	17	1.5	10.2	3.7	-	1.8	x	x
			Grey massive rhyolite.								
RC 56	26°35'	32°00'	Flow XI. Adjacent to RC 55.	21	1.4	15	2.5	-	1.5	x	x
			Grey massive rhyolite, vesicular. Glomeroporph. texture.								
RC 57	26°33'	32°00'	Flow XI. 4 km NNW RC 56.	16	11	0.5	4	-	1.4	x	x
			Massive rhyolite, slight banding.								
RC 58	26°33'	32°00'	As RC 57.	26	1	14	6.5	-	2.8	x	x
			Pale grey massive rhyolite. Microgranophyric matrix.								
RC 59	26°33'	32°00'	As RC 57.	17	0.5	10	4.5	-	1.3	x	x
			Massive rhyolite, felsitic groundmass. Some pyroxene altered to amphibole.								
RC 61	26°33'	32°00'	Flow XI. 3 km S Ravelston. By school.		x	x	x	-	x	x	x
			Massive rhyolite. Microgranophyric matrix.								

				Ph- eno	Qz	Fsp	Cpx	Fa	Ore	Zr	Ap
RC 62	26°32'	31°58'	Interbedded rhyolite. Small outcrop at base of escarpment.	14	-	11	1.6	-	1	-	x
			Massive rhyolite.								
RC 65	27°04'	31°59'	Cecil Mark Pass. 100 m below top of hill.	18	-	12.3	2.4	1.3	1.5	x	x
			Devitrified, buff-coloured, glomeroporph. rhyolite.								
RC 66	27°04'	31°59'	20 m below RC 65.	16	-	11	2	1.4	1.4		
			Devitrified rhyolite, spherulitic spots.								
RC 70	26°25'	31°55'	Eastern Stegi granophyre. 5 km NW Stegi.		x	x	x	-	x	x	x
			Medium-grained granophyric intergrowth, relict biotite, amphibole rims to pyroxene. Fluorite.								
RC 71	26°25'	31°55'	Basalt, outcrop by road 5 km NW Stegi.			x	x	-	x	-	x
			Plagioclase phenocrysts in grey basalt. Amygdale of quartz and epidote.								
RC 73	26°25'	31°54'	Western Stegi granophyre. Lukhula hill.	-	38	58	2.2 bio	x	1.2	x	x
			Coarse biotite granophyre; trace of fayalite.								
RC 74	26°22'	32°00'	Flow IX. Oribi Farm.	27	18	6	1	x	1	x	x
			Acid rhyolite, banded. Perthitic feldspar.								
RC 75	26°23'	31°59'	Flow VII, Oribi dip.	12	0.2	8	0.8	2.5	0.6	x	x
			Massive rhyolite, glomeroporph. texture.								
RC 79	26°49'	32°01'	Flow Ia. Usutu river.		-	x	x	-	x	x	x
			Devitrified rhyolite, spotted texture.								
RC 80	26°49'	32°01'	As RC 79, 120 m east.		-	x	x	-	x	x	x
			Massive spotted rhyolite.								

					Ph- eno	Qz	Fsp	Cpx	Fa	Ore	Zr	Ap
RC 87	26°33'	32°03'	Flow XVI. 3 km west of Groenings.	Irregular banding, grey, fine-grained rhyolite, no glomeroporph. texture.	21	1.5	15	2.5	-	1.5	x	x
RC 90	26°35'	32°06'	Flow XXI. 3 km SE Groenings.	Pale rhyolite. Plentiful quartz. Fine-grained groundmass.	30	2.7	21	4	-	2	x	x
RC 91	26°34'	32°07'	Flow XXIV. By saddle in road.	Fine grey massive rhyolite.	-	x	x	x	-	x	x	x
RC 92	26°34'	32°07'	Flow XXVI. By road 0.5 km from border.	Grey rhyolite, white devitrified spots.	10	-	9.5	-	-	1	-	-
RC101	26°39'	32°00'	Interbedded rhyolite at foot of escarpment.	Fine buff rhyolite. Glomeroporph. texture, mottled banding.	-	-	x	x	-	x	-	x
RC105	26°37'	31°59'	Granophyre dyke, 4 km NE Magwanyana, by track.	Fine-grained granophyre. Microgranophyric groundmass.	21	0.5	15.5	3	-	1.4	x	x
RC110	27°01'	31°54'	Dolerite boulders by roadside 4.2 km W Nsoko.	Aphyric dolerite.	-	-	x	x	-	x	x	x
RC111	27°01'	31°54'	Dolerite 4.4 km W Nsoko.	Aphyric dolerite.	-	-	x	x	-	x	-	x
RC113	27°01'	31°53'	Dolerite 6.1 km W Nsoko.	Aphyric dolerite.	-	-	x	x	-	x	-	x
RC115	27°01'	31°53'	Dolerite 6.8 km W Nsoko.	Porphyritic dolerite, 1.5 cm plag. phenocrysts.	10	-	42	35	-	5.2	-	x
RC116	27°01'	31°52'	Dolerite 7.9 km W Nsoko.	Porph. dolerite, plagioclase phenocrysts.	2.2	-	x	x	-	x	-	x

				Ph- eno	Qz	Fsp	Cpx	Fa	Ore	Zr	Ap
RC119	27°01'	31°50'	Rhyolitic interbed, 11.6 km W Nsoko.	8	-	5.4	2.5	-	0.2	-	x
			Fine grey matrix, occasionally glomeroporphy.								
RC120	27°01'	31°47'	Dolerite k5.2 km W Nsoko.	-	0.5	50	38	-	5.5	-	x
			Aphyric dolerite.								
RC121	26°59'	31°47'	Dolerite by Craggiemore T halt.	10	-	38	40	-	4.8	-	x
			Porphyritic dolerite. Plagioclase phenocrysts.								
RC126	26°37'	32°00'	Basalt in stream.		-	x	x	?	x	-	x
			Sparsely amygdaloidal, plag. phenocrysts 1 cm.								
RC127	26°37'	32°00'	Diorite dyke, 5 km W Tikuba.		x	x	x				
			Weathered microdiorite.								
RC128	26°38'	31°59'	Basal interbedded rhyolite. 5 km W Tikuba.	20	-	11	3.4	-	2.5	-	x
			Grey massive rhyolite, pyroxene/ore very weathered.								
RC131	26°34'	32°05'	1 km E. Groenings. Flow above XVIII.	29	3.3	16	5.6	-	2.2	x	x
			Grey, massive rhyolite.								
RC133	26°34'	32°04'	Flow XVIII. ½ km S Groenings.	-	x	x	x	-	x	x	x
			Pale grey-buff massive rhyolite.								
RC134	26°41'	32°08'	Flow XXV. 9 km SE Tikuba.	25	4.2	19	-	-	1.3	x	x
			Weathered devitrified rhyolite.								
RC136	26°40'	32°04'	Flow XIV. 4 km SSE Tikuba.	17	1	12	2	x	1.5	x	x
			Grey fine-grained massive rhyolite. Perthitic feldspar. Calcite pseudomorphs after cpx, fa.								
RC144	26°32'	32°08'	Flow XXVI. Ipoye beacon.	15	1	10	3	-	1	x	x
			Flow-banded pitchstone. Perthitic cracks.								

						Qz	Ksp	Cpx	Fa	Ore	Zr	Ap
SK 2	26°25'	31°55'	Eastern Stegi Granophyre, same locality as RC 70.	Medium-grained granophyre. Slightly porphyritic.	x	x	x	-	x	x	x	x
SK 13	26°27'	31°57'	Base of rhyolite escarpment 7km N Stegi.	Pale spotty massive rhyolite.	-	x	x	-	x	-	x	x
SK 14	26°22'	31°56'	Interbedded rhyolite. 1 km S Lebombo dip.	Massive buff rhyolite.	-	x	x	-	x	-	x	x
SK 20	26°22'	31°55'	Granophyre dyke. 11 km NNW Stegi.	Greyish medium-grained microgranophyre.	x	x	x	-	x	x	x	x
SK 25	26°21'	31°50'	Interbedded rhyolite 2 km S Grand Canyon.	Grey devitrified rhyolite.	-	x	x	-	x	-	x	x
SK 31	26°20'	31°56'	Interbedded rhyolite, Grand Canyon dip.	Grey spotty massive rhyolite.	-	x	x	-	x	-	x	x
SK 32	26°18'	31°57'	Interbedded rhyolite 3 km S Lusoti.	Streaky rhyolite.	-	x	x	-	x	-	x	x
SK131	26°18'	32°01'	Flow VI. 2 km along road from Cyrildene.	Acid rhyolite, quartz-rich. Slightly contorted banding. Abundant pseudomorphs after fayalite.	x	x	x	x	x	x	x	x
SK138	26°22'	32°03'	Flow XVII. 14 km NE Stegi.	Black pitchstone, incipient spherulites, sanidine and plagioclase phenocrysts. Quartz rare.	?	x	x	-	x	x	x	x

							Qz	Fsp	Cpx	Fa	Ore	Zr	Ap
SK153	26 22	32 04	Flow XX.	15 km NE Stegi.	Black pitchstone, undevitrified glassy matrix.		x	x	x	-	x	x	x
MW 26	26 54	31 53	Basalt	14 km SSW Big Bend.8	Porphyritic basalt, amygdalae of chlorite, epidote and zeolites.		-	x	x	-	x		
MW 50	26 13	32 03	Flow III.	2 km W Blue Jay.	Black pitchstone, dark green glassy lenses.		-	x	x	-	x	x	x
MW 51	26 13	32 03	Flow III.	1.5 km W Blue Jay.	Spherulitic oxidised rhyolite.		x	x	x	-	x	x	x
MW 61	26 15	31 58	Dolerite boulders,	5 km NE Lusoti.	Dolerite, pink plagioclase ? K-feldspar.			x	x	-	x	-	x
MZ 6	26 11	32 02	Quarry	3 km NE Mlaula Station.	Reddish porphyritic rock 'trachy-andesite. Two feldspars and quartz.		x	x	x	-	x	-	x
MZ 11	26 10	32 04	Flow J (VI)	Mbuluzi River.	Quartz-rich rhyolite. Large-scale folding.		x	x	x	?	x	x	x