Because of the higher metamorphic grades reported for the AGC, it is considered that these characteristics may be the result of partial melting of eclogite, rather than amphibolite, at deeper levels in the crust. If the composition of a basaltic komatiite protolith is considered, the low alumina contents in the latter imply that an isochemically derived eclogitic assemblage will be dominated by clinopyroxene rather than garnet. Thus, differences in the melting of eclogite as opposed to an amphibolite will, in this case, lie in the presence of clinopyroxene, rather than hornblende, in the melt residue. This mineralogical difference will affect the relative distribution of trace elements such that Sr, for example, will be retained to a greater degree in the clinopyroxenitic residue, resulting in the generation of magmas that are depleted in this element. Similarly, because hornblende is considered to have higher distribution coefficients for the heavy rare-earth elements than an equivalent amount of clinopyroxene, melts that have equilibrated with the latter mineral will also contain higher HREE contents and lower Ce_N/Yb_N ratios than equivalent magmas that formed in the presence of hornblende. Similar petrogenetic processes occurring at different levels in the crust could, therefore, result in rocks with broadly similar bulk compositions but significantly different trace element characteristics. It is important to stress, however, that it is not feasible to derive a siliceous partial melt from a predominantly clinopyroxene-bearing eclogitic assemblage as, in the absence of garnet, the residues from such a process are unlikely to be sufficiently silica-depleted. As such, the origin of tonalitic-trondhjemitic magmas in terms of the above considerations could probably only occur in an environment intermediate between amphibolite and granulite facies conditions, where hornblende and clinopyroxene are likely to coexist and a silica-depleted residual phase is still present. A constraint on the generation of tonalite and trondhjemite magmas is imposed, therefore, by the tendency for hornblende to become progressively unstable with increasing crustal depth and the concomitant inability of anhydrous, clinopyroxene-dominated, eclogites either, to fuse at all, or, to form melts with the required composition.

The above discussion has certain implications for the correlation that is implied in the scheme of magmatic cycle presented earlier, namely that between the AGC in Swaziland and the migmatite gneiss terrane southwest of the Barberton greenstone belt. The above scheme places the various components of the Ancient Gneiss Complex into the first magmatic cycle, a correlation that was initially carried out because of gross similarities in the broad characteristics of the two regions. This interpretation is not likely to remain unchallenged, however, by proponents who view the AGC as a discrete basement to the Barberton greenstone belt and who have pointed to chemical differences (i.e. such as higher Rb/Sr and lower Ce_N/Yb_N ratios) between the former and the migmatite-gneiss terrane in the Barberton region. The considerations arising from the present study suggest that these chemical differences can, however, be viewed in terms of a petrogenetic scheme that is common to both terranes, but is governed by the depth in the crust at which the processes took place. Hence, in the absence of direct evidence proving that the AGC is older than the base of the Barberton greenstone belt, it appears unnecessary to distinguish the AGC purely on the basis of its geochemical characteristics. The present study, therefore, views the AGC as being largely correlatable with the granite-greenstone terrane underlying those portions of the Barberton Mountain Land that are characterized by rocks of the first magmatic cycle. The relatively minor differences that undoubtedly occur are considered to be a function of crustal depth and accompanying mineral phase changes and not necessarily to the presence of a primordial basement. If such rocks do occur in the AGC they are equally likely to be found in the Barberton Mountain Land as well.

3.3. The mode of emplacement of tonalite-trondhjemite plutons

Although certain of the migmatite outcrops described previously point to the fact that tonalite-trondhjemite gneiss protoliths have intruded and rafted off remnants of pre-existing greenstone material, the evidence for similar magmatic styles of emplacement in the tonalite and trondhjemite plutons is not common. Geological relationships in the latter suggest that emplacement of these bodies into their present positions was accomplished structurally, at some stage subsequent to their total, or near-total, solidification. For example, where examined, the contact between the Barberton greenstone belt and the Stolzburg pluton (illustrated in Figure 3) is highly sheared and fractured indicating a rigid emplacement of the latter into what was probably, an originally overlying, volcanic succession. No evidence for the magmatic intrusion of the trondhjemite into adjacent greenstones (such as veins of trondhjemitic composition) was observed. Similarly, the exposed contact between the Kaap Valley tonalite pluton and the Barberton greenstone belt is generally sheared, again suggestive of a rigid, diapir-like emplacement (see Plate 7E). The best example of this mode of emplacement, however, is provided by the small Doornhoek pluton (Chapter 5) which is completely enveloped by rocks of the Theespruit Formation in the Onverwacht Group (Figures 2 and 79). The contact between the pluton and the adjacent greenstones is relatively well-exposed and nowhere is there evidence for the magmatic intrusion of trondhjemite into the latter. Instead, the extreme margins of the body tend to be relatively well-foliated, by comparison with the remainder of the pluton which is characterized by a very weak, or non existent fabric. These relationships suggest that the greenstones originally overlay the Doornhoek pluton and that the latter was rigidly emplaced into the Theespruit Formation by gravitationally-induced inversion typical of classical diapiric or piercement structures. The presence of a weak fabric over most of the pluton is in accordance with available experimental data on diapirism which shows that a zone of neutral finite strain can usually be found close to the interface between the rising, buoyant mass (undergoing vertical extension) and the denser overburden which is experiencing horizontal extension.

The presentation of a comprehensive set of Rb-Sr whole rock age determinations for a number of tonalite-trondhjemite plutons in the Barberton region illustrates a range in age for these bodies of between 3,49 - 2,75 b.y. In view of the suggested structural emplacement of these bodies it is considered likely that the younger ages in this range reflect the rehomogenization of Sr-isotopes in response to these tectonic upheavals. Thus, although components of the Batavia cell, for example, have been dated at approximately 2,8 b.y. old (see Chapter 6), this age is younger than those obtained for units from the Mpuluzi batholith, which intrudes the latter. This indicates that the age of the Batavia cell does not reflect the original formation of the trondhjemite magma and that the Sr isotopes were homogenized by a later tectonic event. Similar consideration can be given to the recently obtained age of 3,18 b.y. for the Doornhoek pluton (Barton et al., 1981). This pluton is characterized by an anomalously high initial $^{87}Sr/^{86}Sr$ ratio (i.e. 0,7169) with respect to other tonalite-trondhjemite plutons in the region, again suggesting that this is not the original age of the pluton but that the Rb-Sr isotope system has been reset by a later event.

The geochronological data that is presently available in the Barberton region presents difficulties in terms of the detailed understanding of tectonic evolution in the area, particularly when viewed in the light of the suggested structural emplacement of the early gneiss plutons. Although the older tonalite-trondhjemite bodies yield ages that are consistent with their geological relationship (i.e. younger and, therefore, intrusive into the adjacent greenstones), these bodies nevertheless appear to have been structurally emplaced in the same way as their apparently younger equivalents. No definitive criteria is available, therefore, to decide whether isotopic age determinations in components of the first magmatic cycle reflect either, the origin of the body or, a younger event responsible for homogenization of the isotopic systems. Thus, the *primary* relationships between tonalite-trondhjemite gneiss plutons and the lowermost successions of the Barberton greenstone belt are unclear with the possible exception of the magmatic features identified in various migmatite exposures. It may, therefore, be feasible, in terms of the above discussion, to interpret the Doornhoek pluton as having once represented the floor upon which the Theespruit Formation was deposited and that it was subsequently structurally emplaced into its presently intrusive position. The single most important factor that militates against an interpretation of this nature, are the indications suggesting that Archaean tonalites and trondhjemites appear to have been derived by partial melting of mafic precursors. Such material may, however, have pre-dated the Barberton greenstone belt although no evidence exists for its presence.

In conclusion, this section appeals for a cautionary approach to the understanding and significance of the geological relationships between early tonalite and trondhjemite gneiss plutons and the lowermost successions in the greenstone belt (i.e. the Onverwacht Group), particularly in view of the apparent tendency in the former for structural reactivation. Detailed assessment of the structural characteristics of these plutons and their relationship to their isotope systematics as a whole, is considered a prerequisite in understanding the tectonic evolution of the various components of the first magmatic cycle. The data presented in this thesis serve only to emphasize the complexity in this particular topic, detailed research into which would seem to require combined structural and isotopic studies of key units. It is clear, however, that the first magmatic cycle reflects an unstable crustal environment where potent tectonic forces may have been generated by acute gravitational instabilities prevailing in rock types with significantly different densities. The existing geological scenario presents only the final increment in this structurally deformed sequence, the unravelling of which requires a complete understanding of the tectonic evolution in this suite of rocks.

3.4. The significance of tonalites and trondhjemites in the generation of younger granitic phases

Geological, chemical and isotopic relationships indicate that the areally extensive, sheet-like, potash-rich batholiths which constitute the second magmatic cycle were probably derived by reworking the tonalite and trondhjemite gneisses of the first cycle. The multi-component batholiths are enriched in large-ion lithophile elements, depleted in Sr and generally have slightly higher initial $^{87}Sr/^{86}Sr$ ratios than their tonalitic precursors, suggesting an origin by partial melting of the latter. In addition, the batholiths structurally and topographically overlie the tonalite-trondhjemite basement suggesting that melt phases may have coalesced into sheet-like bodies of considerable areal extent, but with volumes that are compatible with a partial melt of the pre-existing rock types.

These considerations imply that the various components of the first and second magmatic cycles are genetically related. Furthermore, the genetic relationship that exists between the tonalites and trondhjemites of the first cycle and pre-existing mafic (possibly amphibolitic) precursors, suggests an even broader interrelationship in terms of a secularly-dependent sequence comprising greenstone-tonalite-potash batholith. This notion implies a crustal evolutionary sequence which envisages the progressive transformation of ensimatic crust into continental sial and, although oversimplified in terms of the previous disucssion on the tectonic emplacement of gneiss bodies, nevertheless provides a framework for understanding the generation of the various granitic components in the area. If the components of both magmatic cycles were consecutively derived from the same ultimate source rock, this would have interesting implications for the configuration of the early Archaean crust. As seen in the regional maps of the study area (e.g. Figures 1 and 97), the existing volumes of granite (i.e. more specifically the combined volumes of the first and second magmatic cycles) far outweigh the relatively insignificant remnants of the lower volcanic greenstone successions currently exposed. If, therefore, this granitic (*sensu lato*) material is to be considered as having been ultimately derived from the latter, then large volumes of this mafic precursor must once have underlain the Archaean crust prior to the development of these granites. Whether or not this mafic crust was primordial and global in extent would be dependent on whether all Archaean granitic rocks (i.e. at least those > 3,0 b.y. old) were themselves genetically interrelated and ultimately derived from such material. It is pertinent to note that the enormous sizes of certain greenstone belts (e.g. the Abitibi greenstone belt in Canada) indicate that they must originally have underlain significant portions of the Archaean crust. However, in terms of the Archaean controversy (see Chapter 2), currently differing views in the interpretation and significance of factors such as the initial $^{87}Sr/^{86}Sr$ ratio, for example, mean that proponents of a primordial sialic crust are equally able to consider the latter as a precursor to subsequently generated granitic phases. A single uniformitarian model for crustal evolution may not therefore be applicable to Archaean terranes from different parts of the globe.

The previous discussions present the tonalite and trondhjemite gneisses of the Barberton region in the context of a key component in the geological evolution of this Archaean granite-greenstone terrane. In the study area they represent the earliest vestiges of sialic material that have interacted with, and probably formed from, greenstones generated in an oceanic, ensimatic crustal setting. During periods of tectonic instability and possibly associated with crustal degassing, these rocks themselves underwent partial melting to form large sheet-like, potash-rich bodies, the emplacement of which took place concurrently with the stabilization and cratonization of the developing continental masses.

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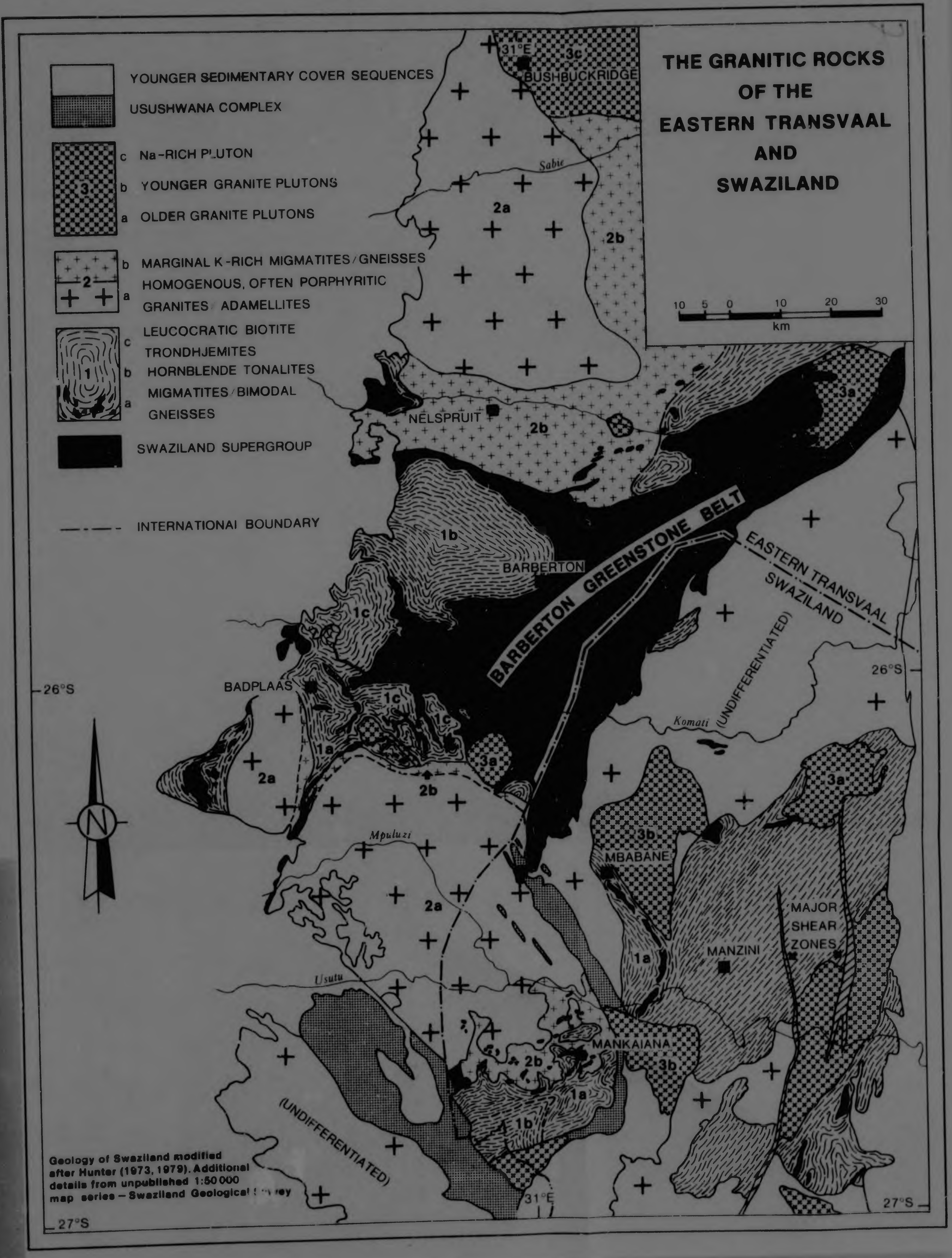
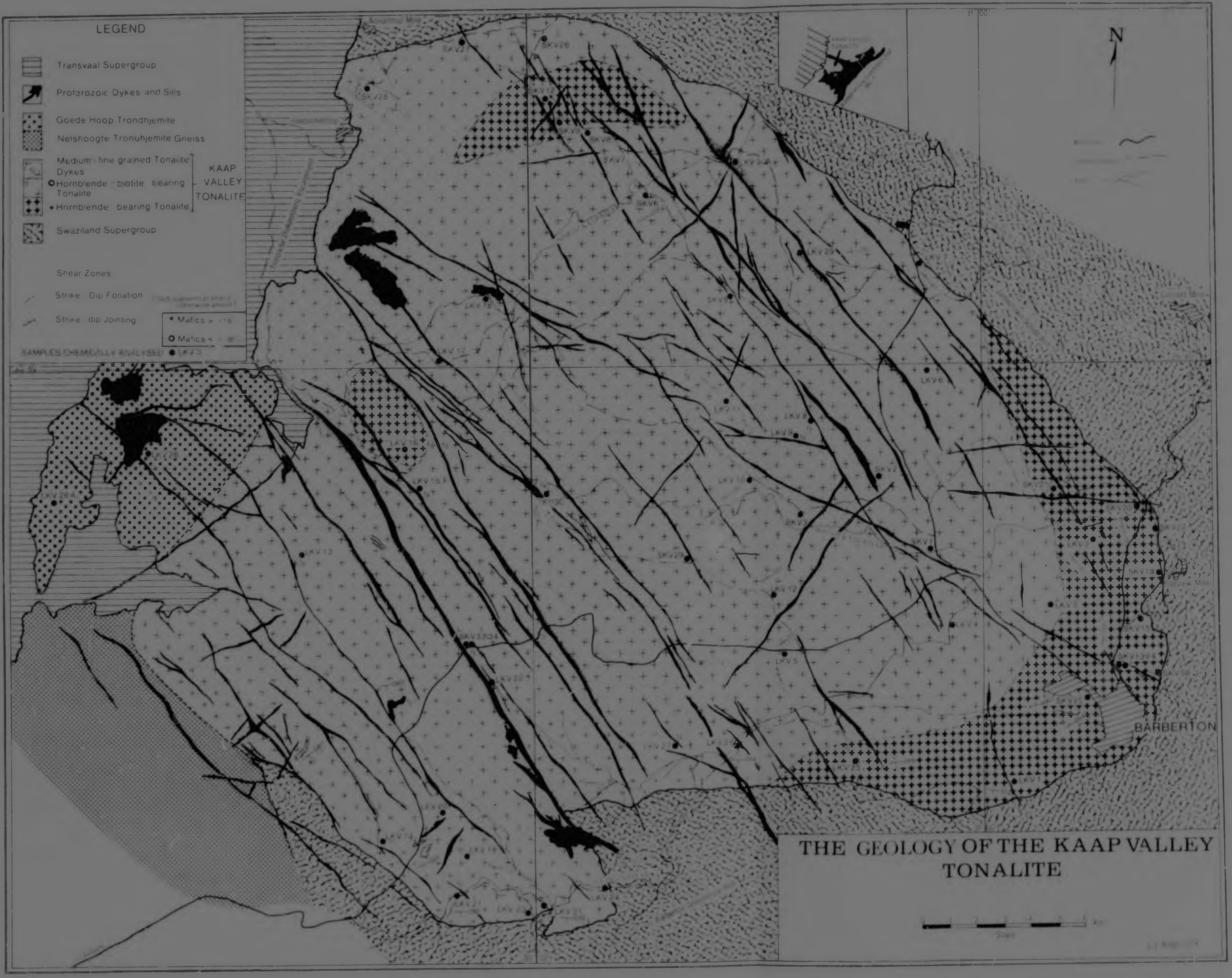


THE GRANITIC ROCKS OF THE EASTERN TRANSVAAL AND SWAZILAND

Geology of Swaziland modified after Hunter (1973, 1979). Additional details from unpublished 1:50 000 map series – Swaziland Geological Survey



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APPENDICES

APPENDIX 1 - ANALYTICAL PROCEDURES

1. MAJOR AND TRACE ELEMENTS ANALYSED USING X-RAY FLUORESCENCE SPECTROMETRY

All the major and trace element analyses presented in this thesis, unless otherwise specified, were carried out by the writer, using the Philips PW1140 X-ray fluorescence spectrometer in the Department of Geology, University of the Witwatersrand.

Considerable care was exercised in sample collection and preparation; individual samples were invariably fresh and any weathered rims were removed prior to processing. Samples were crushed in a jaw-crusher and a representative aliquot eventually pulverized to -300 mesh using a Siebtechnik pulverizer. Analyses were carried out on the XRF spectrometer using a fusion disc for all major elements except sodium (using the methods of Norrish and Hutton, 1969), and a whole rock pressed pellet (bakelite-boric acid based) for the latter element and Rb, Sr and Ba. Mass absorption corrections were applied to Ba using the tables of Birks (196[illegible]) and to Rb and Sr using the Compton peak method described in Reynolds (1967).

The accuracy and precision of the analytical methods used in the present study have been reported in McCarthy (1976) and similar values can be expected in the data presented herein. A close check was maintained on the analytical procedure by sending duplicate powder samples to an independent analytical laboratory. Comparative analyses in the present study were carried out by Bergström and akker (Johannesburg) and details of the duplicated analyses are presented in Table I. Examination of this table shows that, for the most part, differences between the two sets of data do not exceed ± 5% for either major or trace elements. In some instances, however, discrepancies may be fairly large, as shown for element pairs demarcated by an asterisk in Table I. In some cases these differences may be attributable to machine wander which is possibly related to fluctuations in the temperature of the instruments. It is noticeable that most of the asterisks are related to differences in Na_2O and K_2O between the two laboratories and this is undoubtedly the result, at least in the case of sodium, of using a pressed pellet instead of a fusion disc for analysis of the latter. Whatever the reason for inter-laboratory fluctuations, it is unlikely that significant differences in the relative concentrations of any one element from sample to sample, as analysed in the same laboratory and using standardized techniques, will occur.

TABLE 1

COMPARISONS OF DUPLICATE ANALYSES CARRIED OUT BY XRF TECHNIQUES AT THE BERGSTRÖM AND BAKKER AND GEOLOGY DEPARTMENT LABORATORIES

		G	B	G	B	G	B	G	B	G	B	G	B	G	B	G	B	G	B
		B15	B15	B103	B103	B124	B124	C3	C3	C4	C4	D1	D1	D4	D4	LKV16	LKV16	SKV2B	SKV2B
	SiO_2	73,43	73,50	71,52	76,60	-	71,60	74,28	75,20	71,39 *	74,83	73,35	73,59	67,96	67,20	64,45	62,90	[illegible],55	65,20
	TiO_2	0,12	0,09	0,06	0,04	-	0,15	0,28	0,28	0,26	0,23	0,32	0,31	0,64	0,68	0,53	0,53	,46	0,46
	Al_2O_3	14,96	15,40	14,19	14,40	-	15,68	11,86	12,90	14,31 *	13,12	15,03	15,06	16,52	15,50	16,17	16,85	15,53	15,86
	Fe_2O_3*	0,88	0,88	0,16	0,38	-	1,37	2,19	1,34	2,90	2,27	2,35	2,44	4,50	4,19	4,73	4,47	3,85	3,82
	MnO	0,01	0,08	<0,01	0,01	-	0,01	0,01	0,01	0,18	0,25	0,01	0,03	0,06	0,07	0,06	0,09	0,10	0,05
Wt.%	MgO	0,44	0,30	0,35	0,20	-	0,48	0,78	0,80	0,27	0,28	1,09	0,80	1,23	1,30	2,32	2,43	2,14	2,13
	CaO	1,12	1,51	1,54	1,28	-	1,78	0,87	0,86	1,04	1,12	2,67	2,61	3,29	3,16	4,77	4,81	3,59	3,61
	Na_2O	4,69 *	3,80	6,46 *	9,50	-	5,90	2,77	3,20	2,73	2,85	4,37	5,00	4,39	4,80	4,87	4,40	6,17 *	5,20
	K_2O	4,10 *	5,47	5,53	0,44	-	2,72	4,96	3,81	4,71 *	3,39	2,21	1,69	1,49	1,44	0,91	1,15	1,03	1,09
	P_2O_5	0,07	0,04	0,02	0,02	-	0,04	0,06	0,03	0,07	0,04	0,17	0,20	0,41	0,22	0,23	0,26	0,16	0,19
	L.O.I.	1,63	1,18	0,39	0,36	-	0,42	0,49	0,27	1,12	0,74	1,03	0,71	1,15	1,18	1,63	1,20	2,32 *	1,27
	TOTALS	100,10	99,38	100,96	99,53	-	99,21	99,17	99,28	100,23	99,29	100,88	99,43	100,94	99,47	100,26	99,30	100,86	99,93
	Rb	97	93	21	12	46	45	93	91	128	140	10 *	45	59	52	41	36	49	37
ppm	Sr	139 *	490	84	135	663	740	95	105	160	101	825	880	333	360	734	790	635	693
	Ba	555	597	8	<10	611	750	817	880	1193 *	750	787	710	220	220	457	480	214	320

* - Total iron as Fe_2O_3.

G - Analyses carried out by the author at the Department of Geology, University of the Witwatersrand.

B - Analyses carried out by Bergström and Bakker, Johannesburg.

* - Elements where comparisons between G and B are questionable.

TABLE I

COMPARISONS OF DUPLICATE ANALYSES CARRIED OUT BY XRF TECHNIQUES AT THE BERGSTROM AND BAKKER AND GEOLOGY DEPARTMENT LABORATORIES

		G	B	G	B	G	B	G	B	G	B	G	B	G	B	G	B	G	B
		B15	B15	B10A	B10A	B12A	B12A	C3	C3	C4	C4	D1	D1	D4	D4	LKV16	LKV16	SKV28	SKV28
	SiO_2	73,43	73,50	77,52	76,60	-	71,50	74,29	75,20	71,39 *	74,80	70,39	70,50	67,06	67,20	64,45	62,90	65,55	66,20
	TiO_2	0,12	0,09	0,06	0,04	-	0,15	0,20	0,20	0,26	0,23	0,32	0,31	0,64	0,68	0,53	0,53	0,46	0,46
	Al_2O_3	14,96	15,40	14,19	14,40	-	15,[illegible]0	12,66	12,90	14,91 *	13,[illegible]0	15,83	15,00	16,52	15,50	16,19	16,80	15,50	15,80
	Fe_2O_3+	0,88	0,89	0,19	0,38	-	1,30	2,18	1,98	2,99	2,22	2,90	2,44	4,50	4,19	4,70	4,47	3,85	3,82
	MnO	0,01	0,04	<0,01	0,01	-	0,01	0,01	0,01	0,08	0,05	0,01	0,03	0,06	0,07	0,06	0,09	0,10	0,05
Wt.%	MgO	0,44	0,30	0,06	0,20	-	0,40	0,78	0,80	0,87	0,88	1,00	0,80	1,23	1,00	2,92	2,60	2,14	2,10
	CaO	1,12	1,53	1,54	1,58	-	1,79	0,87	0,88	1,34	1,12	2,61	2,61	3,39	3,16	4,77	4,81	3,58	3,61
	Na_2O	4,59 *	3,80	6,46 *	5,50	-	5,90	3,77	3,20	2,73	2,80	4,37	5,00	4,59	4,80	4,87	4,40	6,17 *	5,20
	K_2O	4,10 *	3,47	0,53	0,44	-	2,12	3,86	3,81	4,77 *	3,39	2,21	1,69	1,49	1,44	0,91	1,15	1,03	1,26
	P_2O_5	0,07	0,04	0,02	0,02	-	0,04	0,06	0,03	0,07	0,04	0,17	0,20	0,21	0,22	0,23	0,26	0,16	0,16
	L.O.I.	0,60	0,32	0,39	0,36	-	0,42	0,49	0,27	1,12	0,74	1,03	0,71	1,13	1,16	1,63	1,29	2,32 *	1,27
	TOTALS	100,32	99,38	100,96	99,53	-	99,23	99,17	99,28	100,23	99,29	100,84	99,49	100,84	99,42	100,26	99,30	100,86	99,93
	Rb	97	95	21	12	46	45	83	93	128	140	10 *	45	59	52	41	36	49	31
ppm	Cr	339 *	490	84	105	663	740	95	105	180	190	426	480	333	380	734	790	635	690
	Ba	655	550	5	< 0	620	750	817	580	1193 *	750	387	310	220	220	457	480	218	220

+ - Total iron as Fe_2O_3.

G - Analyses carried out by the author at the Department of Geology, University of the Witwatersrand.

B - Analyses carried out by Bergstrom and Bakker, Johannesburg.

* - Elements where comparisons between G and B are questionable.

2. RARE EARTH ELEMENTS ANALYSED USING INSTRUMENTAL NEUTRON ACTIVATION TECHNIQUES

Each of the different rock types discussed in this thesis was representatively analysed for the rare-earth elements (REE) using the instrumental neutron activation (INA) techniques available from the Activation Analysis Research Group at the Nuclear Physics Research Unit, University of the Witwatersrand. In this technique samples are pulverized to less than -200 mesh and loaded into sealed glass tubes. The sample aliquots are then subjected to an epithermal neutron flux at the poolside irradiation position of the SAFARI I reactor at Pelindaba. The samples are protected from a thermal neutron flux by cadmium shielding and the total irradiation time is six hours. Technical details of the subsequent analytical procedures utilized by the Activation Analysis Research Group are presented in Fesq et al. (1973) but certain points relating to the procedures for REE analysis are presented below.

Subsequent to neutron bombardment, the sample aliquots themselves emit a characteristic radiation, the intensity of which varies depending on the half-lives of the various isotopes contained in the sample. It is necessary, therefore, to count certain of the elements with shorter half-lives prior to those with longer ones. With respect to the REE, the present analytical procedure counted La and Sm approximately seven days after irradiation, Nd, Tb, Yb and Lu approximately 15 days after irradiation and Ce and Eu approximately 30-35 days after irradiation. Another important factor in relation to this analytical procedure is associated with variations in the neutron flux that is incident on each of the sample aliquots as a result of their different positions in the poolside location of the reactor. This necessitates the existence of an internal neutron flux monitor which operates by comparing the concentrations of iron (Fe) as determined by INA analysis with known concentrations of the same element. In the present study all irradiated samples had previously been analysed for major elements by XRF techniques, so that the magnitude of the Fe peak, as determined by neutron activation analysis, could be compared with Fe concentrations obtained using a constant-flux technique. In this way small corrections in the magnitudes of emitted radiation (i.e. proportional to the incident neutron flux), which sometimes amounted to as much as 10% of the radiation counts, were made. Calibration procedures for the neutron activation analysis technique were carried out using international standards such as NIM-G and GSP-1, as well as a local standard OG-382. The precision and

Rhodonite - Mn in plagioclase and garnet

Spessartine garnet - Si, Al, Fe in garnet

The maximum variation in replicated samples with respect to each of these standards is less than 2% (G. Davies, personal communication). It should be stressed, however, that in certain samples, mineral alteration may result in stoichiometrically imperfect analyses (see Table 20 and Figure 46) but this is solely a function of the altered structural state of the individual minerals in question.

3.2. Analytical procedure - Geological Survey

All the analyses obtained using the Geological Survey facilities were carried out by R.C. Wallace on a JEOL 50A electron microprobe. The running voltage for this instrument is also 15 kilovolts with a sample current of 255 milliamps. Data processing was carried out with respect to in-house standards represented by the following series of minerals, each pertaining to the undermentioned elements.

Si, Ca - wollastonite
Mg - pyrope
Na - albite
K - sanidene
Ti - rutile
Fe - specularite
Al - corundum

Data indicating the maximum percentage variation in replicated samples with respect to these standards are not available but are assumed to be of the same order of magnitude as the University of the Witwatersrand facility.

* * *

APPENDIX 2 - PETROGENETIC MODELLING TECHNIQUES

1. PARTITION COEFFICIENTS USED IN PETROGENETIC MODEL CALCULATIONS

1.1 Granitic *(sensu lato)* rocks:-

	Garnet	Clino-pyroxene	Hornblende	Biotite	K-felspar	Plagioclase
Rb *	0,01	0,03	0,015	3,26	0,66	0,05
Sr *	0,015	0,52	0,22	0,12	3,87	2,84
Ba *	0,017	0,13	0,04	6,36	6,12	0,36
La + (Δ)	0,30	0,4	(0,4)	0,35	0,05	0,3
Ce + (Δ)	0,35	0.5	(0,9)	0,32	0,05	0,3
Nd + (Δ)	0,53	1,1	(2,8)	0,29	0,03	0,2
Sm + (Δ)	2,7	1,7	(4,0)	0,26	0,02	0,1
Eu + (Δ)	1,5	1,6	(3,4)	0,24	1,10	2,15
Tb + (Δ)	12	1,6	(4,1)	0,30	0,01	0,06
Yb + (Δ)	40	1,6	(4,9)	0,44	0,01	0,05
Lu + (Δ)	30	1,5	(4,5)	0,33	0,01	0,05

1.2 Basaltic rocks:-

	Olivine	Ortho-pyroxene	Clino-pyroxene	Hornblende	Plagioclase	Garnet
Rb + (¢)	0,01	0,02	0,02 (0,04)	0,29 (0,02)	0,07	0,04
Sr + (¢)	0,01	0,02	0,12 (0,50)	0,46 (0,02)	1,83	0,01
Ba + (¢)	0,01	0,01	0,01 (0,05)	0,42 (0,05)	0,23	0,02
La + (Δ)	0,01	0,01	0,10	0,20 (0,4)	0,15	0,03
Ce + (Δ)	0,01	0,02	0,10	0,20 (0,9)	0,12	0,03
Nd + (Δ)	0,01	0,03	0,12	0,33 (2,8)	0,08	0,07
Sm + (Δ)	0,01	0,05	0,18	0,52 (4,0)	0,07	0,29
Eu + (Δ)	0,01	0,05	0,18	0,59 (3,4)	0,34	0,49
Tb + (Δ)	0,01	0,12	0,17	0,54 (4,1)	0,06	2,0
Yb + (Δ)	0,01	0,34	0,16	0,49 (4,9)	0,07	11,5
Lu + (Δ)	0,02	0,42	0,13	0,43 (4,5)	0,06	11,9

Sources:-
* - Compilation by Hanson (1978)
+ - Compilation by Arth and Hanson (1975)
¢ - Compilation by Condie and Hunter (1976)
Δ - Arth and Barker (1976)

2. CHONDRITIC RARE-EARTH ELEMENT CONTENTS USED FOR NORMALIZING RARE-EARTH ELEMENT PLOTS

	ppm		ppm
La	0,2446	Tb	0,0374
Ce	0,6379	Dy	0,2541
Pr	0,0964	Ho	0,0567
Nd	0,4738	Er	0,1660
Sm	0,1540	Tm	0,0256
Eu	0,0580	Yb	0,1650
Gd	0,2043	Lu	0,0254

(These data represent the regressed means of numerous analyses of CI chondrites after Evensen et al., 1978)

3. EQUATIONS USED IN PETROGENETIC MODEL CALCULATIONS

3.1. Magmatic Crystallization

The distribution of trace elements between crystallizing mineral phases and residual magma can be considered in terms of equilibrium between either, the surface of the crystallizing phase and the liquid, or, the total solid and the liquid. If the latter situation pertains to the entire solidification history of the rock, then *equilibrium crystallization* prevails and the trace element distribution is determined by :-

$$Cs = Co.K_D/F + K_D (1-F) \quad .... \; 1$$

(where the various symbols used are defined below)

and $$Cl = Co/F + K_D (1-F) \quad .... \; 1a \quad \text{(after Arth, 1976)}$$

If, on the other hand, equilibrium is only a surface phenomenon then the crystallization procedure is analagous to instantaneously removing the crystallizing phases from the liquid from which they formed (i.e. *fractional crystallization*) and trace element distribution is determined by :-

$$Cs = Co.K_D.F^{(K_D - 1)} \quad .... \; 2$$

and $$Cl = Co.F^{(K_D - 1)} \quad .... \; 2a \quad \text{(after Greenland, 1970)}$$

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$$Cs = Co.K_D/F + K_D\ (1-F) \ \ldots\ldots\ldots\ldots \ 1$$

(where the various symbols used are defined below)

and $$Cl = Co/F + K_D\ (1-F) \ \ldots\ldots\ldots\ldots \ 1a \quad \text{(after Arth, 1976)}$$

If, on the other hand, equilibrium is only a surface phenomenon then the crystallization procedure is analagous to instantaneously removing the crystallizing phases from the liquid from which they formed (i.e. *fractional crystallization*) and trace element distribution is determined by :-

$$Cs = Co.K_D.F^{(K_D - 1)} \ \ldots\ldots\ldots\ldots \ 2$$

and $$Cl = Co.F^{(K_D - 1)} \ \ldots\ldots\ldots\ldots\ldots\ldots \ 2a \quad \text{(after Greenland, 1970)}$$

3.2 [illegible]

In the case of partial melting, the present study considered only one situation, namely that where continuous equilibrium is maintained between the liquid (melt) phase and residual solid until such time as the melt is bodily removed from the solid. This is commonly referred to as batch melting and takes into account the fact that the parental material is likely to consist of a significantly different assemblage to that reflected in the partial melt. The distribution of trace elements during batch melting is given by :-

$$Cl \;=\; Co/K_P + F\,(1 - K_M) \;..........\; 3$$

and

$$Cs \;=\; Co.K_D/K_P + F\,(1 - K_M) \;.......\; 3a$$

(after Shaw, 1970).

The symbols used in equations 1-3 above are explained as follows :-

Cs - concentration of a given trace element in the solid phases.

Cl - concentration of a given trace element in the liquid phases.

F - in equation [illegible] and 2 refers to the fraction of liquid remaining
- in equation [illegible] refers to the fraction of melting that has taken place.

Co - in equations 1 and 2 is the concentration of a given trace element in the original melt.
- in equation 3 is the initial concentration of a given trace element in the parent.

K_D - is the bulk partition coefficient for a given trace element and is given by :-

$$K_D \;=\; D1.F1 + D2.F2 + \;..........\; + DN.FN$$

where D1,D2 ...DN are the distribution (or partition) coefficients of the given trace element into mineral phases 1, 2 ... N respectively (as presented in section 1) and F1,F2 ... FN are the weight fractions of the same mineral phases.

K_P, K_M - are the bulk partition coefficients for the parent or source material and for the liquid (melt) phase during the batch melting situation. In the unusual case where simple modal melting takes place (i.e. $K_P = K_M$) then equation 3 reduces to :-

$$Cl \;=\; Co/K_P + F\,(1 - K_P)$$

It is important to realize, however, that at the time of removal of a liquid from a residue, the ratio Cl/Co is dependent only on the K_D of the residue and naturally, the degree of melting F. If none of the mineral phases in the original parent have been consumed during melting (e.g. a very small degree of melting) then Kp refers to the bulk partition coefficient of the entire source material; if any one phase has been consumed, however, then Kp refers only to the bulk partition coefficient of the residue *at the time of removal* of the melt (i.e. after a specified percentage of partial melting).

4. WORKED EXAMPLE OF A TRACE ELEMENT MODEL CALCULATION

The distribution of trace elements between a crystallizing mineral assemblage and residual magma or, between a partial melt phase and a restite fraction, is calculated using the equations provided in section 3 above. In the worked example that follows, an example of how these calculations are undertaken is provided. The case that is worked through is one involving partial melting and equilibrium crystallization, although the same principles can be applied to fractional crystallization processes as long as the relevant equations are used.

Prior to examining the case example, it is necessary to stress that the object of using trace element modelling techniques in this thesis was *not* to exactly replicate the empirical trace element concentrations in a particular suite of rocks. Uncertainties in analytical data and partition coefficients as well as the constraints imposed by the mobility of elements in rocks and a restricted sample coverage do not allow for such precise modelling techniques. The trace element models described in the text were used simply to place constraints on the possible origins of certain of the rock types examined. In some instances, for example, it is almost impossible to differentiate between a rock formed by a small degree of partial melting and one that results from an advanced degree of fractional crystallization. Thus, in the present study the petrogenesis of a particular rock type was not exhaustively tested for in terms of a number of possible model origins, but was, rather, gauged in terms of the most likely situation prevailing within the framework of the geological relationships existing with respect to the rock in question.

The worked example provided below is an attempt to model the REE characteristics of the anatectites described in section 3 of Chapter 3. In terms of their relationship to other rock types associated with the migmatites, these rocks appear to have been derived by anatexis or partial melting of tonalitic or trondhjemitic gneisses and this suggestion is partly confirmed by considerations regarding the mesonormative compositions of the anatectites. In terms of their mesonormative compositions the anatectites can be subdivided into two categories, namely, those with cotectic or minimum melt compositions and those with Or-depleted compositions. The REE characteristics of the two sub-types are also different so that :

(i) the cotectic anatectites are relatively LREE enriched ($Ce_N/Yb_N > 20$) with either no Eu anomaly or a slight negative anomaly; and

(ii) the Or-depleted anatectites are not as LREE enriched ($Ce_N/Yb_N < 6$) with a positive Eu anomaly (see Figure 26, Chapter 3).

These differences imply two sets of conditions for each of the anatectite sub-types and these are considered in turn below.

4.1 Cotectic anatectites

The model for the origin of these rocks is one that envisages a 30% partial melt of a trondhjemitic gneiss. The figure of 30% is an arbitary one and represents a probable upper limit for the melt fraction envisaged for the cotectic anatectites.

The REE content of a typical trondhjemite gneiss is given below,

i.e

	La	Ce	Nd	Sm	Eu	Tb	Yb	Lu
ppm	13	22	8	1,30	0,40	0,18	0,28	0,03

The bulk partition coefficients (K_D) for the REE into a trondhjemitic assemblage comprising 55% plagioclase, 35% quartz, 5% K-felspar and 5% biotite are calculated using the formula provided in section 3 above :-

i.e. $K_D = D1.F1 + D2.F2 + ... + DN.FN$

$\therefore$ K_D for La can be calculated knowing the partition coefficients of La into plagioclase, quartz, K-felspar and biotite (as provided in section 1 above):-

$$K_D = (0,55 \times 0,3) + (0,35 \times 0,00^*) + (0,05 \times 0,05) + (0,05 \times 0,35) = 0,185$$

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* - Note that quartz is considered to have a partition coefficient of zero for all elements.

The same procedure will provide the K_D's for each of the REE under consideration.

Batch melt equations require the bulk partition coefficient of the melt phase. The K_D's for the cotectic anatectites can be calculated for an assemblage comprising 35% plagioclase, 35% quartz and 30% K-felspar :-

i.e. K_D for La = (0,35 x 0,3) + (0,35 x 0,00) + (0,3 x 0,05) = 0,12

and likewise for the remainder of the REE.

Knowing the REE concentration in the parent and the K_D's for both the parent and melt assemblages. it is possible to calculate the trace element concentrations in the melt phase using the batch melt equations provided in section 3 above :-

i.e. $Cl = Co/Kp + F\ (1 - K_M)$

$\therefore \quad Cl_{La} = 13/0,18 + 0,3(1 - 0,12) = 29,3$

Thus, the concentration of La in a 30% partial melt of a trondhjemite will be approximately 2,5 x that in the parental rock (i.e. a LREE enriched pattern). As seen in Figure 27, when the above calculation is repeated for all the REE and plotted on a chondrite normalized diagram, the resultant pattern is not dissimilar to the empirical data for the cotectic anatectites. The model is compatible, therefore, with the preferred interpretation of these rocks, although it is nevertheless possible that they may have been derived by a different process and from a different parental rock type. The suggested model is, however, considered to best suit the observed geological relationships. Finally, it may be possible to improve the model fit by a trial and error process involving considerations of, for example, different parental assemblages, partition coefficients and degrees of partial melting. As previously mentioned, this approach is regarded as being relatively futile in view of the numerous other discrepancies involved in the modelling process.

4.2 Or-depleted anatectites

The Or-depleted anatectites differ fundamentally from their cotectic equivalents in that they have pronounced positive Eu anomalies. The most feasible process whereby Eu^{2+} is concentrated into a particular assemblage in preference to the other REE (in a trivalent state) is by the crystallization of plagioclase, for which Eu^{++} has a high partition coefficient (see section 1 above). Thus, the melts from which the Or-depleted anatectites were derived must contain a high proportion of "cumulate" plagioclase that is responsible for the positive Eu anomaly. This situation is completely different to the anatectites discussed above which were characterized by a cotectic composition (i.e. all three phases, namely plagioclase, quartz and K-feispar, crystallized simultaneously and in equilibrium) so that the resultant solids entirely reflect the liquids from which they were derived. The Or-depleted anatectites, whose compositions have moved away from the granite cotectic (i.e. possibly by higher degrees of partial melting), appear to have crystallized plagioclase + quartz only, in a process that is akin to fractional crystallization. The resultant solids need not, therefore, reflect the composition of the liquids from which they were derived.

The model for the origin of these rocks, therefore, envisages 50% solidification of a plagioclase-dominated assemblage cyrstallizing from an advanced partial melt (80%) of the same trondhjemite gneiss considered previously.

The REE content of an 80% partial melt of a trondhjemite gneiss is calculated by substituting 0,8 for 0,3 in the batch melt equation considered in section 4.1 above :-

i.e.

	La	Ce	Nd	Sm	Eu	Tb	Yb	La
Cl	14,7	24,8	9,1	1,50	0,32	0,22	0,34	0,04

The REE bulk partition coefficients for an Or-depleted anatectite comprising 60% plagioclase and 40% quartz are calculated in the same way as that outlined above :-

i.e. K_D for La = (0,6 x 0,3) + (0,4 x 0,00) = 0,18

etc.

Knowing the REE content in the suggested parent magma (i.e. the 80% partial melt above) it is possible to calculate the trace element concentrations in the crystallizing phases. The concentrations in the solid phase after an arbitary 50% crystallization are given by :-

$$Cs = Co.K_D/F + K_D\ (1 - F)$$

$$\therefore\ Cs_{La} = 14,7.0,18/0,5 + 0,18\ (1 - 0,5) = 4,58$$

When the above calculation is repeated for all the REE and plotted on a chondrite normalized diagram (Figure 28), the resultant pattern is similar to that for the Or-depleted anatectites. The positive Eu anomaly, in particular, is adequately accounted for as the bulk partition coefficient for Eu is greater than unity so that this element is strongly distributed into the crystallizing assemblage. The arbitrary figure of 50% crystallization in the above model could be modified to represent a higher degree of solidification; however, the rock cannot represent 100% crystallization of its suggested parental magma as, in an equilibrium crystallization mode, such material will be identical to the parental liquid and, hence, not characterized by a positive Eu anomaly. In real terms, it is possible that portions of this liquid that had not yet solidified may have been filter pressed out of the magma chamber so that they could no longer equilibrate with the previously formed crystal assemblage. In this way it is possible for an assemblage to solidify by equilibrium crystallization and not fully reflect the composition of the liquid from which it was derived.

* * *

APPENDIX 3 - GARNET-BIOTITE GEOTHERMOMETRY

This section describes an attempt that was made at assessing the equilibration temperature of coexisting garnet and biotite from a sample of tonalitic gneiss (E2, see Table 3 and Figure 10) from a migmatite outcrop in the Batavia section (Figure 2). The data presented below (Table II) represents microprobe analyses of the relevant coexisting mineral phases from sample E2.

TABLE II

ANALYSES OF COEXISTING BIOTITE AND GARNET FROM A TONALITIC GNEISS - SAMPLE E2, MIGMATITE OUTCROP E

		1	2	3	4	5
		Biotite	Biotite	Garnet	Garnet	Garnet
	SiO_2	36,22	35,49	38,27	38,62	38,68
	TiO_2	2,31	2,08	0,08	0,09	0,08
	Al_2O_3	15,50	16,04	19,95	19,85	20,08
	FeO (*)	23,55	23,07	26,96	26,95	26,69
Wt.%	MnO	0,31	0,28	3,95	4,06	3,91
	MgO	8,67	8,71	2,84	2,80	2,93
	CaO	-	0,01	8,21	8,17	8,44
	Na_2O	0,04	0,05	0,01	0,01	0,01
	K_2O	9,77	9,50	-	-	-
	TOTALS	96,37	95,23	100,27	100,55	100,82
	STRUCTURE (+)	22	22	24	24	24
	Si	5,[illegible]93	5,534	6,088	6,128	6,108
	Ti	0,269	0,207	0,010	0,011	0,008
	Al	2,821	2,948	3,741	3,709	3,736
	Fe^{3+} (ϕ)	3,041	3,008	0,063	0,013	0,032
	Fe^{2+}			3,524	3,560	3,492
	Mn	0,040	0,037	0,532	0,545	0,524
	Mg	1,996	2,024	0,673	0,662	0,668
	Ca	-	0,002	1,399	1,388	1,422
	Na	0,012	0,015	0,003	0,003	0,004
	K	1,925	1,889	-	-	-
	TOTALS	15,697	15,664	16,033	16,014	16,020

(i) [illegible]
Column [illegible] grain core

(ii) Analy[illegible]d [illegible] Davies, Department of Geology, University of the Witwatersrand, using an ARL Electron Microprobe.

(iii) Total Iron as FeO

(iv) Structure of biotite calculated in terms of 22 oxygens to compensate for $(OH)^-$ not determined by the probe; garnet structure calculated in terms of 24 oxygens.

(v) Ferric/Ferrous ratio calculated for garnet using the stoichiometric relationship of Ryburn et al. (1976) :-

Fe^{3+}(garnet) = 16 - 4Si - 4Ti - 2Al - 2Cr

Fe^{2+}(garnet) = Fe^{Total}(garnet) - Fe^{3+}(garnet)

Examination of the above data shows a consistency in the composition of the [illegible] rim of the analysed garnets, indicating that the effects of [illegible]nation [illegible] mineral are minimal. Nevertheless, during the calculation of the [illegible] geothermometer, temperatures were obtained using the average Mg/Fe [illegible] with respect to Mg/Fe^{2+} for each of the garnet analyses (Table [illegible]) below.

Calibration of the mineral composition with temperature was carried out in [illegible], using the following two, experimentally derived, regression equations :-

lnK = -2730/T(°K) + 1,557 (after Thompson, 1976)

lnK = -2109/T(°K) + 0,782 (after Ferry and Spear, 1978)

where:

K = Mg/Fe^{2+}(garnet)/Mg/Fe(biotite)

T = Temperature in degrees Kelvin.

Substituting the relevant data into the above equations provides the following :-

Notes :- (i) Columns 1 - 4 : Grain rims
Column 5 : Grain core

(ii) Analyses carried out by G. [illegible], Department of Geology, University of the Witwatersrand, using an ARL Electron Microprobe.

(*) Total Iron as FeO

(†) Structure of biotite calculated in terms of 22 oxygens to compensate for $(OH)^-$ not determined by the probe; garnet structure calculated in terms of 24 oxygens.

(‡) Ferric/Ferrous ratio calculated for garnet using the stoichiometric relationships of Ryburn et al. (1976) :-

$$Fe^{3+}(\text{garnet}) = 16 - 4Si - 4Ti - 2Al - 2Cr$$

$$Fe^{2+}(\text{garnet}) - Fe^{Total}(\text{garnet}) - Fe^{3+}(\text{garnet})$$

Examination of the above data shows a consistency in the composition of the core and rim of the analysed garnets, indicating that the effects of zonation in this mineral are minimal. Nevertheless, during the calculation of the garnet-biotite geothermometer, temperatures were obtained using the average Mg/Fe (biotite) with respect to Mg/Fe^{2+} for each of the garnet analyses (Table III, below).

Calibration of the mineral composition with temperature was carried out in duplicate, using the following two, experimentally derived, regression equations :-

$$\ln K = -2730/T(^\circ K) + 1,557 \quad \text{(after Thompson, 1976)}$$

and

$$\ln K = -2109/T(^\circ K) + 0,782 \quad \text{(after Ferry and Spear, 1978)}$$

where;

$$K = Mg/Fe^{2+}(\text{garnet})/Mg/Fe(\text{biotite})$$
$$T = \text{Temperature in degrees Kelvin.}$$

Substituting the relevant data into the above equations provides the following :-

TABLE III

ESTIMATION OF GARNET-BIOTITE EQUILIBRATION TEMPERATURE USING THE AVERAGE Mg/Fe RATIO IN BIOTITE WITH RESPECT TO THE Mg/Fe^{2+} RATIO IN EACH OF THE ANALYSED GARNETS FROM TABLE II

Average Mg/Fe (biotite) = 0,664

	Garnet (3)	Garnet (4)	Garnet (5)
Mg/Fe^{2+}(garnet)	0,191	0,186	0,191
K	0,288	0,280	0,288
T (Ferry & Spear, 1978)	767°C	753°C	767°C
T (Thompson, 1976)	704°C	694°C	714°C

AVERAGE TEMPERATURE = 733°C

Thus, geothermometric considerations indicate that garnet and biotite in the tonalitic gneiss from the Batavia section, equilibrated in the range ≅ 700 - 770°C. Such a temperature is above the water-saturated tonalite solidus for all pressures in excess of approximately 1 - 2 Kb (Stern et al., 1975).

* * *

APPENDIX 4 - PHYSICAL CONDITIONS WHEREBY ZONE MELTING AND BATCH MELTING ARE SIMILAR

1. ZONE MELTING

The general form of the zone melting equation (i.e. Equation 1, Table 28) is :-

$$\frac{Cl}{Co} = \frac{1}{K_D} - (E) \quad \text{where E is the exponential function}$$

From the equation 1 in Table 28, if ℓ is a large number, then $e^{-\ell}$ is small and E tends to zero. Hence, for a large zone length factor (ℓ), the general form of the above equation approximates to :-

$$\frac{Cl}{Co} \cong \frac{1}{K_D} \quad \text{..........} \quad (1)$$

2. BATCH MELTING

In the case of batch melting :-

$$\frac{Cl}{Co} = \frac{1}{(K_D + F\ (1 - K_M)} \quad \text{(after Shaw, 1970)}$$

If F tends to zero (i.e. for a very small degree of partial melt), then the above equation approximates to :-

$$\frac{Cl}{Co} \cong \frac{1}{K_D} \quad \text{..........} \quad (2)$$

Thus, zone melting, under conditions of a large zone length factor, tends to be similar to small degrees of partial (batch) melting. Whereas the latter may be unrealistic in geological terms because of the insignificant volumes of magma generated, this constraint does not apply to zone melting because the zone length factor (i.e. height of rock column/height of zone melt) is independent of the horizontal dimension. It is conceivable, therefore, that significant volumes of magma may be generated in terms of these considerations.

APPENDIX 5 - PROCEDURES FOR THE CALCULATION OF
^{87}Sr/[illegible] H[illegible]PS

The decay of ^{87}Rb to ^{87}Sr is expressed quantitatively in terms of the following relationship :-

$$^{87}Sr_{t2} = {}^{87}Sr_{o} + {}^{87}Rb\,(e^{\lambda t1} - e^{\lambda t2})$$

where o = original quantity
λ = decay constant
$t1, t2$ = time

If $t2$ = present day, then the expression $(e^{\lambda t1} - e^{\lambda t2})$ reads as $(e^{\lambda t} - 1)$

On dividing the above expression by ^{86}Sr, i.e.

$$\frac{^{87}Sr}{^{86}Sr} = \left\{\frac{^{87}Sr}{^{86}Sr}\right\}_{o} + \frac{^{87}Rb}{^{86}Sr}\,(e^{\lambda t} - 1) \;.......... \;(1)$$

one obtains the familiar equation expressing the relationship in the Rb-Sr isochron diagram.

The above equation (1) can be simplified as follows :-

as $\frac{^{87}Rb}{^{86}Sr} \propto \frac{Rb}{Sr}$ then $\frac{^{87}Rb}{^{86}Sr} = K\frac{Rb}{Sr}$ (where K is a constant)

and, for very small values of λt,

$$e^{\lambda t} \approx 1 + \lambda t$$

Thus, $$\frac{^{87}Sr}{^{86}Sr} = \left\{\frac{^{87}Sr}{^{86}Sr}\right\}_{o} + \left\{\frac{Rb}{Sr}\right\} K\lambda t \;.......... \;(2)$$ (after Faure and Powell, 1972)

The value of the constant K, is derived as follows :-

$$^{87}Rb = Rb\,(0{,}27385)$$

and $$^{86}Sr = Sr/9{,}24286 + (^{87}Sr/^{86}Sr.\ 0{,}98850)$$

$$\therefore \quad \frac{^{87}Rb}{^{86}Sr} = \frac{Rb(0{,}27385)\,(9{,}24286 + {}^{87}Sr/^{86}Sr.\ 0{,}98850)}{Sr}$$

$$\therefore \quad K \approx (2{,}5312 + (0{,}2707\ {}^{87}Sr/^{86}Sr)) \;.......... \;(3)$$

(J.M. Barton, personal communication).

On a plot of Age v initial $^{87}Sr/^{86}Sr$, the development of the $^{87}Sr/^{86}Sr$ ratio with time is proportional to the Rb/Sr ratio of the system. In Figure [illegible], two growth vectors, one corresponding to a Rb/Sr ratio of [illegible], the other to a Rb/Sr ratio of 0,35, are plotted. The slopes of these vectors are calculated by deriving a value for K for a $^{87}Sr/^{86}Sr$ ratio equal to 0,7000. Thus, from equation (3) above :-

$$K = 2,5312 + (0,27 \text{[illegible]} \; 0,7000)$$

$$\therefore \quad K = \underline{2,7207}$$

The growth vectors in Figure 94 are calculated using the following expression, which is derived from the above discussion :-

$$\left\{\frac{^{87}Sr}{^{86}Sr}\right\}_{t2} = \left\{\frac{^{87}Sr}{^{86}Sr}\right\}_{o} + \left\{\frac{Rb}{Sr}\right\} K.\lambda(t1 - t2)$$

where the interval (t1 - t2) corresponds to 500 m.y. (i.e. t1 - t2 = 3500 - 3000 m.y.), where the values of $\frac{Rb}{Sr}$ are 0,08 and 0,35 respectively and where $\left\{\frac{^{87}Sr}{^{86}Sr}\right\}_{o}$ is taken to be 0,7000.

$$\text{Thus,} \quad \left\{\frac{^{87}Sr}{^{86}Sr}\right\}_{3,0 \text{ b.y.}} = 0,7000 + (0,35)\ 2,7207.\ 1,42 \times 10^{-11}\ (0,5 \times 10^{9})$$

$$= \underline{0,7068} \quad (\text{for} \quad Rb/Sr = 0,35)$$

$$\text{and} \quad \left\{\frac{^{87}Sr}{^{86}Sr}\right\}_{3,0 \text{ b.y.}} = 0,7000 + (0,08)\ 2,7207.\ 1,42 \times 10^{-11}\ (0,5 \times 10^{9})$$

$$= \underline{0,7015} \quad (\text{for} \quad Rb/Sr = 0,08)$$

The above data enables one to plot the $^{87}Sr/^{86}Sr$ growth vectors for rocks characterized by Rb/Sr ratios of 0,08 and 0,35 respectively, in the diagram in Figure 94.

* * *

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