

## Isotopic and geochemical systematics in Tertiary-Recent basalts from southeastern Australia and implications for the evolution of the sub-continental lithosphere

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**Abstract**—Tertiary-Recent Tasmanian and Newer (Victoria/South Australia) basalts range from quartz tholeiite to olivine melilitite and show systematic increases in their incompatible element abundances with increasing degree of silica undersaturation. These two basalt provinces show similar relative abundances of rare earth elements (REE), differences in the relative concentrations of Rb, Ba, Th, K and Nb, and distinct, restricted isotopic compositions. The Tasmanian basalts have  $^{87}\text{Sr}/^{86}\text{Sr}$  from 0.7026 to 0.7034, and  $\epsilon_{\text{Nd}}$  from +7.5 to +5.8; the Newer basalts have higher  $^{87}\text{Sr}/^{86}\text{Sr}$  from 0.7038 to 0.7045, and lower  $\epsilon_{\text{Nd}}$  from +4.2 to +1.7. The range in Sr and Nd isotope compositions can be defined by primary magma compositions for both provinces, using Mg-values, Ni content and the presence of spinel lherzolite nodules. Major and trace element and Sr, Nd and Pb isotope compositions are uniform on a scale of up to 50 km for four separate Newer basanite centers. The chemical and isotopic data are consistent with a model whereby tholeiitic basalts are derived by large degrees of partial melting from a chemically uniform but isotopically variable source, and generation of undersaturated, alkaline basalts by smaller degrees of partial melting of the same source. No isotopic or geochemical evidence was found which would suggest that the more evolved basalts have been contaminated by continental crust.

In contrast to tholeiitic and alkalic basalts from Hawaii, there is a continuous spectrum of isotope compositions for the Newer tholeiitic to alkalic basalts. A model is proposed for the generation of these basalts involving mixtures of hotspot mantle plume-derived melt and lithospheric mantle-derived melt, where observed differences between ocean island and continental alkaline basalts are attributed to differences between the sub-oceanic and sub-continental lithospheric mantles. Isotopic differences between tholeiitic and alkalic basalts are interpreted to be due to varying degrees of exchange and mixing between the hotspot plume and lithospheric mantle melt components. The model is consistent with the generation of these basalts from a source which has been recently enriched in the LREE.

### INTRODUCTION

INTEGRATED PETROLOGIC, major and trace element and isotopic studies of primitive basalts can provide much needed insight into the composition and evolution of their mantle source regions. In contrast to ocean island basalts (OIB) the study of continental intraplate basalts has been relatively neglected due to the possibility of contamination by the continental crust during their ascent and emplacement. However, ALLEGRE *et al.* (1981), in a study of continental intraplate basalts which showed minimal effects of crustal contamination, suggested that continental alkaline basalts are derived from source regions similar to that of OIB. Their work was reconnaissance in nature and there now exists a need for more detailed investigations into the similarities and potential differences between the source regions of continental alkaline basalts *versus* OIB. From detailed studies which integrate major and trace element and isotope data we may begin to evaluate the role played by the convective (asthenospheric) mantle and that of the non-convective lithospheric mantle in contributing their geochemical and isotopic signatures to tholeiitic and alkalic basalts from continental and oceanic intraplate settings. Ultimately we hope to resolve how differences between the sub-oceanic and sub-continental lithospheric mantles are

manifested in the basalt types erupted from these regions. Finally, when modelling contamination processes, what role is played by the crust and by the lithospheric mantle?

Chemical and isotopic data are presented for southeastern Australian tholeiitic and alkalic basalts with primary, or near primary, major and trace element compositions. We have also analyzed more evolved basalts to determine if they have been contaminated by crustal rocks and to document such effects, if present. Comparisons are also made between the isotopic compositions of spinel lherzolite nodules and their host basalts in order to evaluate the possible role of lithospheric mantle contamination. Many samples have been previously analyzed for major (IRVING and GREEN, 1976) and trace element concentrations (FREY *et al.*, 1978). These earlier studies placed constraints on depth of melt separation and degree of partial melting involved in the generation of these basalts. In this study we present Sr and Nd isotopic data for all of these basalts and provide a broader major and trace element and isotopic data base by which to evaluate the conclusions of these earlier studies. Comparisons are also made between the chemical and isotopic composition of these continental, tholeiitic to alkalic basalts and similar basalts found in the Hawaiian islands.

### Geological background

The samples come from two Tertiary-Recent basalt provinces in southeastern Australia: the Tertiary basalts of Tasmania and the Pliocene-Recent Newer basalts of western Victoria and southeastern South Australia. Figure 1 shows the distribution and localities of the Newer basalt samples. The Newer basalt field is a monogenetic volcanic field covering some 25,000 Km<sup>2</sup>, while the Tertiary Tasmanian basalts occur as scattered flows and plugs in eastern Tasmania. WELLMAN and MCDUGALL (1974) and SUTHERLAND (1981) propose that the Tertiary-Recent volcanism along eastern Australia is related to the migration of mantle plumes beneath the continent. Basalts from southeastern Australia have erupted through predominantly Ordovician geosynclinal sediments of the Ballarat and Melbourne troughs and Silurian intrusives.

Thin section examination of the basalts show little alteration apart from minor hematite, especially around the edges of olivine crystals. Detailed location and petrographic information for these samples can be found in IRVING and GREEN (1976) and FREY *et al.* (1978).

### ANALYTICAL METHODS

Methods used for preparation of rock powders and determination of major element composition for these basalts is described in IRVING and GREEN (1976). Trace element concentrations and isotope ratios were determined on splits of whole rock powders used in the earlier studies of IRVING and GREEN (1976) and FREY *et al.* (1978). Trace element concentrations reported in Table 1, except for Nd, were determined by B. Chappell using the XRF technique of NORRISH and CHAPPELL (1977). Nd concentrations were determined by isotope dilution. Sr and Rb concentrations were determined by both XRF and isotope dilution; agreement between the two methods was within 2% absolute or better. Trace element concentrations reported in Table 2 were determined by spark-source mass spectrometry (TAYLOR and GORTON, 1977). There is generally good agreement with the Rb, Sr and Nd concentrations reported in FREY *et al.* (1978) and those re-determined in this study, except that isotope dilution yields 93 ppm Nd for sample 2860 compared to 77 ppm Nd reported in FREY *et al.* (1978).

Samples were totally spiked with <sup>150</sup>Nd and <sup>147</sup>Sm tracers and decomposed in open beakers using HF, HF-HClO<sub>4</sub> solutions and then converted to chlorides using 6 N HCl. The REE, Sr and Rb were initially separated on 5 gram cation exchange columns using 1 N, 2.5 N and 6 N HCl as the elutriants. Rb was purified on 2 gram columns using 1 N HCl, while Sr was purified on 2.5 gram columns using 1 N and 2.5 N HCl. Nd and Sm were separated from each other and the remaining REE on a separate column using 0.2 M 2-methylactic acid with a pH of 4.6. During the analyses of these basalts the blank, using our stock solutions, was 0.5 ng for Nd and 3.5 ng for Sr. No blank corrections were necessary

for any of the measured isotope ratios. Nd and Sr isotope ratios were measured as metal using a triple Re filament in the MSZ mass spectrometer (CLEMENT and COMPTON, 1972). Sr isotope ratios were measured by peak switching <sup>86</sup>Sr-<sup>86</sup>Sr and <sup>87</sup>Sr-<sup>87</sup>Sr, and then normalized to <sup>86</sup>Sr/<sup>88</sup>Sr = 0.1194. Nd isotope ratios were measured by peak switching between <sup>142</sup>Nd-<sup>144</sup>Nd, <sup>143</sup>Nd-<sup>144</sup>Nd and <sup>146</sup>Nd-<sup>144</sup>Nd, and these ratios were normalized to <sup>146</sup>Nd/<sup>142</sup>Nd = 0.636151 to correct for the effects of mass fractionation. Precision for isotopic measurements are reported as the  $2\sigma_{\text{mean}}$  for the analysis. Rb, Sr, Sm and Nd concentrations (ppm) and <sup>87</sup>Sr/<sup>86</sup>Sr and <sup>143</sup>Nd/<sup>144</sup>Nd isotopic ratios determined for BCR-1 during the measurement of the basalt samples are: Rb = 46.9 ± 1, Sr = 326 ± 3, Sm = 6.59 ± 0.1, and Nd = 28.74 ± 0.05, and BCR-1 isotope ratios are <sup>87</sup>Sr/<sup>86</sup>Sr = .70494 ± 1 (n = 3) and <sup>143</sup>Nd/<sup>144</sup>Nd = 0.511833 ± 10 (n = 7). This is -0.2ε<sub>Nd</sub> units lower than the value previously reported in this laboratory (*e.g.*, MCCULLOCH and CHAPPELL, 1982) and is a result of the installation of a new electrometer and data acquisition system. The ratio obtained for Ndα (WASSERBURG *et al.*, 1981) during the same period is 0.511101 ± 8 (n = 8), which compares with WASSERBURG *et al.*'s (1981) reported value of 0.511121 ± 8. We measured <sup>143</sup>Nd/<sup>144</sup>Nd = 0.511040 ± 7 (n = 4) for the La Jolla standard during the final stages of this study. There is excellent agreement in relative values of the <sup>143</sup>Nd/<sup>144</sup>Nd ratio between our data and that reported in WASSERBURG *et al.* (1981) for these three Nd isotope standards (BCR-1, Ndα, and La Jolla). Finally, we measured <sup>87</sup>Sr/<sup>86</sup>Sr = 0.71022 ± 1 (n = 7) for NBS - 987 during the course of this study, and <sup>87</sup>Sr/<sup>86</sup>Sr = 0.70800 ± 3 (n = 2) for the E & A SrCO<sub>3</sub> standard during the final stages of this study.

A number of repeat Sr and Nd isotopic analyses were duplicated during the period of data collection. The duplicate analyses for samples 2177, 69-1006, 69-1018, 2679, 2128, 2854 and 2860 are reported in Table 3; all repeats fall within the reported  $2\sigma_{\text{mean}}$  uncertainty limits.

### RESULTS

The major and trace element compositions of Newer basalts are reported in Tables 1 and 2. Sr and Nd isotope compositions of these and other Newer basalts and Tasmanian basalts are given in Table 3. Major and trace element concentrations for samples reported in Table 3 and not found in Table 1 are given in FREY *et al.* (1978).

#### Major elements, Ni and Cr contents

The Pliocene-Recent Newer basalts consist of a broad compositional spectrum of tholeiitic and alkalic series basalts including quartz tholeiite, olivine tholeiite, hawaiite, nepheline hawaiite, nepheline mugearite, Fe-rich olivine basalt, olivine

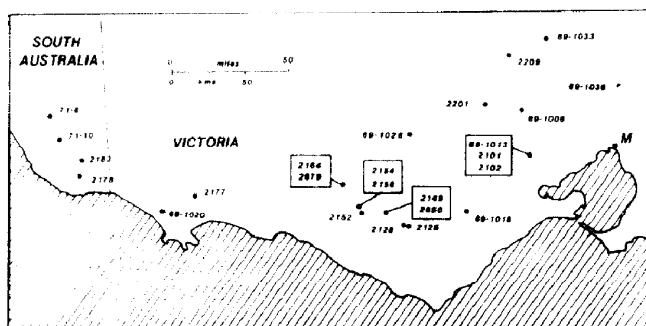


FIG. 1. Location map for the Newer basalts studied here. The sampled area covers the areal distribution of Newer basalts. (M = Melbourne)

TABLE 1. Major and Trace Element Data for Victorian Newer Basalts

	2125 Olivine Tholeiite	69-1006 Fe-rich Basalt	69-1020 Olivine Basalt	69-1033 Olivine Basalt	2201 Alkali Ol Basalt	2149 Olivine Basalt	69-1013 Hawaiite	2154 Neph.	2183 Hawaiite	2178
SiO <sub>2</sub>	50.11	49.09	49.23	49.58	47.37	46.79	48.74	46.26	48.34	47.46
TiO <sub>2</sub>	2.06	3.40	1.88	2.65	2.43	2.62	2.36	2.72	2.35	3.11
Al <sub>2</sub> O <sub>3</sub>	13.89	17.20	14.37	13.86	14.80	13.60	14.59	13.15	13.41	14.43
Fe <sub>2</sub> O <sub>3</sub>	1.76	2.38	1.87	1.85	1.83	2.06	1.89	2.06	1.89	2.25
FeO	9.55	9.53	9.37	9.25	9.13	10.28	9.44	10.30	9.44	9.00
MnO	0.16	0.14	0.16	0.15	0.17	0.18	0.18	0.18	0.17	0.16
MgO	8.47	3.76	9.19	8.45	9.46	9.78	8.31	9.67	8.88	7.35
CaO	8.60	7.17	9.10	8.20	8.47	9.00	8.07	8.65	9.58	7.51
Na <sub>2</sub> O	3.55	4.14	3.20	3.45	3.54	3.34	4.50	4.18	3.69	4.63
K <sub>2</sub> O	1.17	2.32	1.26	1.88	1.96	1.60	1.20	1.83	1.68	3.02
P <sub>2</sub> O <sub>5</sub>	0.41	0.87	0.37	0.68	0.84	0.76	0.72	1.00	0.57	1.08
CO <sub>2</sub>	0.06	0.03	0.03	0.05	0.08	0.07	0.06	0.06	0.18	0.09
H <sub>2</sub> O <sup>+</sup>					1.29	0.35		0.48		
H <sub>2</sub> O <sup>-</sup>	0.06	1.08	0.75	0.87	0.70	0.23	1.17	0.18	0.17	0.18
Mg#	62.3	41.3	63.6	62.0	64.9	62.9	61.1	62.6	62.6	59.3
V	140.	173.	144.	167.	143.	163.	130.	129.	173.	133.
Cr	240.	7.	253.	229.	222.	217.	346.	232.	249.	100.
Ni	163.	17.	174.	189.	189.	181.	313.	214.	185.	99.
Cu	29.	14.	42.	49.	44.	51.	39.	40.	56.	37.
Zn	103.	127.	103.	120.	125.	118.	116.	125.	115.	122.
Sr	467.	995.	482.	625.	890.	713.	775.	953.	642.	1090.
Rb	25.	39.	28.	39.	40.	32.	58.	39.	34.	62.
Ba	245.	485.	370.	395.	500.	400.	485.	550.	455.	820.
Pb	3.	3.	2.	4.	3.	4.	5.	6.	3.	4.
Y	20.	31.	20.	24.	24.	22.	22.	25.	21.	27.
Nb	27.	52.	28.	42.	59.	45.	58.	69.	44.	79.
Zr	168.	330.	144.	279.	287.	241.	294.	362.	202.	385.
La	22.	39.	20.	33.	44.	33.	44.	56.	34.	62.
Ce	48.	89.	41.	73.	94.	78.	97.	129.	70.	135.
Nd	23.7	44.9	21.2	35.9	40.5	--	40.3	54.0	32.6	57.2

	2101 Neph.	2102 Mugearite	69-1031 Mugearite	2209	2156	71-17 Basanite	71-26 Mugearite	2164	71-10 Olivine	70-1174 Analcimite	71-8 Analcimite
SiO <sub>2</sub>	49.63	48.88	48.87	48.78	45.59	45.29	45.40	45.16	43.85	43.75	44.09
TiO <sub>2</sub>	2.18	2.19	2.26	2.26	2.52	2.43	2.72	2.81	3.64	3.68	3.87
Al <sub>2</sub> O <sub>3</sub>	15.45	15.41	15.71	15.77	12.28	12.00	12.64	13.55	9.17	9.37	10.56
Fe <sub>2</sub> O <sub>3</sub>	3.47	3.45	3.64	3.66	1.85	1.94	2.01	1.99	2.39	2.50	2.42
FeO	6.94	6.90	7.28	7.32	10.00	9.70	10.04	9.97	9.57	10.00	9.69
MnO	0.16	0.16	0.16	0.16	0.18	0.17	0.17	0.18	0.17	0.18	0.18
MgO	6.41	7.20	6.48	6.55	12.65	13.76	11.59	10.53	15.54	15.18	12.03
CaO	6.78	6.52	6.82	6.84	8.14	7.96	8.33	8.96	10.33	10.66	11.88
Na <sub>2</sub> O	5.43	5.72	5.01	4.90	3.86	3.95	4.13	3.79	4.14	3.40	3.89
K <sub>2</sub> O	2.91	2.87	2.89	2.87	2.01	1.95	2.08	2.06	0.35	0.39	0.36
P <sub>2</sub> O <sub>5</sub>	0.64	0.64	0.88	0.89	0.91	0.85	0.90	0.99	0.84	0.88	1.02
CO <sub>2</sub>	0.54	0.04	0.08	0.08	0.06	0.07	0.08	0.19	0.05	0.11	0.05
H <sub>2</sub> O <sup>+</sup>	0.39			1.19	0.11			0.86	1.33	1.85	1.75
H <sub>2</sub> O <sup>-</sup>	0.28	0.08	1.30	0.40	0.04	0.20	0.21	0.26	0.88	1.17	1.46
Mg#	62.2	65.1	61.3	61.4	69.3	71.7	67.3	65.3	74.3	73.0	68.9
V	101.	113.	104.	104.	133.	132.	142.	145.	264.	264.	263.
Cr	144.	170.	99.	103.	412.	471.	317.	203.	535.	540.	292.
Ni	131.	149.	94.	92.	388.	439.	298.	221.	460.	496.	250.
Cu	28.	33.	35.	33.	43.	43.	46.	64.	57.	53.	68.
Zn	144.	155.	147.	148.	140.	135.	132.	134.	107.	108.	102.
Sr	995.	825.	1140.	1141.	835.	835.	845.	925.	804.	1200.	1021.
Rb	67.	72.	68.	64.	40.	40.	40.	40.	40.	29.	10.
Ba	860.	870.	720.	725.	450.	440.	465.	490.	820.	875.	855.
Pb	5.	7.	3.	6.	4.	4.	3.	4.	5.	5.	5.
Y	21.	22.	24.	24.	24.	23.	23.	25.	22.	21.	24.
Nb	89.	93.	70.	71.	66.	64.	63.	72.	76.	76.	86.
Zr	480.	493.	364.	361.	331.	319.	318.	346.	327.	327.	354.
La	56.	55.	60.	54.	49.	60.	47.	58.	63.	65.	69.
Ce	117.	114.	109.	111.	105.	100.	105.	120.	138.	137.	153.
Nd	47.1	45.4	--	47.7	47.6	--	--	52.4	64.9	--	68.0

Mg# = 100[Mg/(Mg+Fe<sup>2+</sup>)]. Sample localities are given in Table 3, except for 69-1031 (Mt. Franklin), 71-17 and 71-26 (Mt. Noorat), and 70-1174 (Mt. Watch). Major and minor elements in wt%, trace elements in ppm.

basalt, alkali olivine basalt, nepheline basanite and olivine analcinite. The Fe-rich olivine basalts are the most chemically evolved rocks of the Newer suite studied by IRVING and GREEN (1976). A comprehensive description of the general geology, petrography and major element geochemistry of the Newer basalts is presented by IRVING and GREEN (1976).

The Tertiary Tasmanian basalts are older than the Newer basalts and are composed of tholeiitic and alkalic series basalts. These basalts were erupted some 20 to 30 Ma ago and exhibit a similar compositional range similar to that of the Newer basalts. Tasmanian basalts measured for Sr and Nd isotopic composition include olivine tholeiite, basanite, olivine nephelinite and an olivine melilitite.

Both the Newer and Tasmanian basalts have normative mineralogical composition ranging from nepheline to olivine normative, and one Newer basalt is a quartz normative tho-

leite. Most nepheline normative basalts reported here, together with a Newer olivine basalt and the Tasmanian Andover olivine tholeiite contain Cr-diopside spinel ilherzolite nodules. Whole rock Mg-values for most of these basalts range from 59 to 74, with many between 68 and 75. Basalts with Mg-values > 68 are considered to be primary mantle liquids derived from a refractory upper mantle peridotite with a Mg-value of ≥88-89 (IRVING and GREEN, 1976; FREY *et al.*, 1978). The Newer Fe-rich olivine basalt (69-1006) with a Mg-value of 41 and the evolved Scottsdale, Tasmanian olivine nephelinite (2860) with a Mg-value of 54 are the most chemically evolved lavas in each suite. These two samples were analyzed in order to investigate the isotopic effects of crustal level fractionation.

Ni and Cr contents of most of these basalts vary from 92 ppm to 496 ppm and 99 ppm to 540 ppm respectively. The

TABLE 2. Trace Element Data for Newer Basalts (in ppm)

	2177 Quartz Tholeiite	2152 Olivine Tholeiite	69-1026 Olivine Basalt	69-1036 Alkali Ol. Basalt	71-10 Olivine Analcimite	71-8
Y	91.	49.	53.	27.	22.	28.
Zr	158.	172.	162.	173.	171.	171.
Nb	28.	32.	28.	38.	38.	38.
Ba	281.	477.	623.	309.	894.	893.
La	47.0	46.8	45.8	32.3	62.	68.
Ce	97.	103.	125.	47.	125.	127.
Pr	12.0	11.1	10.3	5.7	16.0	18.2
Nd	50.2	49.0	40.8	24.6	65.2	67.8
Sm	11.7	9.68	9.04	5.83	11.6	11.7
Eu	3.92	2.92	2.64	1.90	3.41	3.33
Gd	12.3	9.1	9.2	5.2	6.8	6.8
Tb	1.91	1.46	1.44	0.89	1.23	1.21
Dy	11.5	9.3	8.2	5.2	5.2	5.0
Ho	2.40	1.48	1.60	0.84	0.91	0.89
Er	5.88	3.52	3.91	0.18	2.14	1.91
Yb	4.03	2.62	2.89	1.61	1.79	1.62
Hf	3.08	3.69	3.37	3.67	6.44	6.71
Cs	0.39	0.20	0.06	0.43	1.12	1.15
Pb	2.52	2.45	2.50	3.27	4.55	5.82
Th	2.29	2.95	2.37	2.63	7.25	7.14
U	0.43	0.60	0.23	0.63	1.67	1.34

two most evolved lavas have low Ni and Cr contents: Fe-rich olivine basalt (69-1006) has 17 ppm Ni and 7 ppm Cr and olivine nephelinite (2860) has 64 ppm Ni and 11 ppm Cr. Primary basalts have a higher and more restricted range in Ni and Cr content, ranging from 298 to 496 ppm Ni and 310 to 540 ppm Cr. The high Mg-values and Ni and Cr concentrations, and the presence of Cr-diopside lherzolite nodules (implying rapid uninterrupted ascent through the crust) are consistent with the derivation of these basalts by partial melting of a peridotitic mantle with no subsequent chemical modifi-

cation (*i.e.* crystal fractionation, crustal contamination, etc.). With primary basalts as a focusing point we can look at the trace element and isotopic composition of these and other, more evolved, basalts from these fields.

#### Incompatible elements

Figures 2a and 2b show the incompatible element compositions for a variety of Newer basalt types normalized to values for the primitive mantle. There are progressive increases in the incompatible element concentrations with increasing degree of silica undersaturation (Fig. 2a). Incompatible element abundances in the nepheline mugearites are similar to those in an alkali olivine basalt (Fig. 2b), except that the nepheline mugearites have lower Ti contents relative to the alkali olivine basalt. This suggests that a Ti-bearing phase is involved in the evolution of these magmas. Relative abundances of the incompatible elements Nb through Lu (Fig. 2b) in the hawaiite and nepheline hawaiites are similar to those in the alkali olivine basalt, though significant differences exist in the relative abundances of the most incompatible elements (Rb, Ba, Th, K and Nb). These differences may reflect variations in the initial source composition, different degrees of partial melting, and/or differences in the composition and relative amounts of fractionating phases involved in their generation (*e.g.* IRVING and PRICE, 1981).

The olivine analcimites display a large relative depletion in K, variable depletions in Rb, and a minor depletion in Sr compared to a Newer basalt (Fig. 2a). Less pronounced Sr depletions are evident in the many Newer basalts suggesting either Sr is behaving more compatibly during partial melting

TABLE 3. MEASURED Sr AND Nd ISOTOPE COMPOSITIONS OF SE AUSTRALIAN BASALTS

Sample	Locality	Rock Type	<sup>87</sup> Sr/ <sup>86</sup> Sr	<sup>143</sup> Nd/ <sup>144</sup> Nd	$\epsilon_{Nd}$
VICTORIAN AND SOUTH AUSTRALIAN NEWER BASALTS					
69-1013	Anakies (W)	hawaiite*	.70386	.512004 ± 20	+3.2
2154	Noorat	ne. hawaiite*	.70390	.512030 ± 20	+3.6
2183	Mt. Gambier	ne. hawaiite*	.70408	.511984 ± 14	+2.9
2178	Mt. Schank	K-rich ne. hawaiite	.70413	.512007 ± 26	+3.3
2101	Anakies (E)	ne. mugearite*	.70384	.512006 ± 24	+3.3
2102	Anakies (E)	ne. mugearite*	.70384	.512004 ± 24	+3.3
2209	Mt. Franklin	ne. mugearite*	.70377	.512053 ± 18	+4.0
				.512049 ± 29	+4.0
2177	Mt. Eekersley	qz. tholeiite	.70491	.511765 ± 24	+1.9
			.70489		
2125	Stonyford	ol. tholeiite	.70427	.511945 ± 22	+2.1
69-1018	Mt. Gellibrand	ol. tholeiite	.70412	.511990 ± 23	+3.0
				.511970 ± 20	+2.6
2152	Marida Yallock	ol. tholeiite	.70442	.511865 ± 24	+2.6
69-1006	Mt. Gorong	Fe-rich ol. basalt	.70426	.512008 ± 22	+3.2
			.70423	.512016 ± 18	+3.0
69-1026	Mt. Widdenin	ol. basalt	.70436	.511775 ± 31	+1.7
69-1020	Mt. Kincaid	ol. basalt	.70446	.511925 ± 26	+1.7
69-1033	Green Hill	ol. basalt*	.70407	.511970 ± 24	+2.6
2201	Mt. Warrenheip	alkali ol. basalt*	.70398	.512018 ± 18	+3.2
2149	Camperdown	alkali ol. basalt*	.70404	.512025 ± 18	+3.7
69-1036	Mt. Frazer	alkali ol. basalt*	.70384	.511992 ± 26	+3.0
2156	Mt. Noorat	basanite*	.70379	.512007 ± 22	+3.3
2184	Mt. Shadwell	basanite*	.70389	.512021 ± 29	+3.6
2679	Mt. Shadwell	basanite*	.70381	.512031 ± 20	+3.5
				.512016 ± 27	+3.3
2128	Mt. Porndon	basanite*	.70386	.512000 ± 24	+3.2
			.70389	.512002 ± 14	+3.2
2650	Mt. Leura	basanite*	.70381	.511997 ± 18	+3.1
71-10	Mt. Watch	ol. analcimite*	.70450	.511930 ± 21	+1.8
71-10		(residue')	.70434		
71-8	Mt. McIntyre	ol. analcimite	.70434	.511971 ± 18	+2.1
TASMANIAN BASALTS					
T-14	Andover	ol. tholeiite*	.70324	.512162 ± 22	+6.4
2896	Lugnata	basanite*	.70310	.512200 ± 28	+7.1
2854	Scottsdale	ol. nephelinite*	.70265	.512212 ± 24	+7.3
			.70268	.512224 ± 16	+7.6
2860	Scottsdale	ol. nephelinite	.70260	.512215 ± 24	+7.4
			.70263		
2927	L. J. Marsh	ol. melilitite*	.70335	.512134 ± 22	+5.6

\*Basalts containing Cr-diopside spinel lherzolite xenoliths. ol. = olivine ne = nepheline. All <sup>87</sup>Sr/<sup>86</sup>Sr values have an uncertainty of  $\leq \pm 0.00005$ .

'leach experiment with 6N HCl, see text for details.

$$\epsilon_{Nd} = \left[ \frac{(^{143}\text{Nd}/^{144}\text{Nd})_{\text{Sample}}}{(^{143}\text{Nd}/^{144}\text{Nd})_{\text{Chur}}} - 1 \right] 10^4, \text{ where } (^{143}\text{Nd}/^{144}\text{Nd})_{\text{Chur}} = 0.511838$$

than predicted by its relative position in Fig. 2, or the source region for these basalts is heterogeneous and relatively more depleted in Sr compared to the LREE. Depletions in K and Rb in the olivine analcimites are suggested to be relatively recent, since these basalts have the lowest Rb/Sr ratios but high  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios (0.7043) compared to most Newer basalts. This depletion K and Rb suggests that a residual K and Rb-rich phase is involved in the generation of the olivine analcimites.

IRVING and GREEN (1976) noted the dominant groundmass mineral of the olivine analcimites is analcime and argued that the analcime is not a replacement of leucite. In particular they pointed out that this would not be in accord with the low  $\text{K}_2\text{O}$  content (0.4%) and high  $\text{Na}_2\text{O}$  contents (3.4% to 3.9%) of these lavas. In thin section the olivine analcimites show some secondary alteration, characterized by hematite rims on phenocrystic olivines. The chemical composition of these basalts is unlikely to have been significantly affected by surface alteration since they do not show depletions in Ba, Th, Cs, U, or the LREE (Fig. 2a, Table 2), and have Th/U ratios of 4.3 and 5.3, similar to other Newer basalts (FREY *et al.*, 1978; and this study, Table 2). A leaching experiment was carried out on one olivine analcinite (71-10) where 100 mg of sample was leached for 18 hours in 6 N HCl in 170°C. After rinsing several times with cold 6 N HCl the residue was dissolved in HF and  $\text{HClO}_4$ . The leached residue has a slightly lower  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio ( $0.70434 \pm 4$ ) than the unleached sample ( $0.70450 \pm 4$ ), although these ratios overlap at their  $3\sigma$  uncertainty. This indicates that surface alteration has at most only minor effects on the Sr isotope composition. The Mt. Watch olivine analcinite contains Cr-diopside bearing, spinel lherzolite nodules, and both analcimites have Mg-values and Ni contents characteristic of primary mantle derived magmas, implying that their chemical and isotopic compositions are inherited from their mantle source regions.

FREY *et al.* (1978) reported anomalously high REE and Y contents for three Newer basalts (see their Figs. 7 and 9). We have redetermined the trace element contents of these samples and confirm the anomalous enrichments (Table 2). The absolute concentrations of REE and Y are increased 2 to 3 fold compared to other tholeiitic and transitional Newer basalts, while the remaining incompatible element concentrations are similar to those in other Newer basalts (Fig. 3). Identified REE and Y-enriched Newer basalts include a quartz tholeiite, an olivine tholeiite and an olivine basalt. The REE and Y-enriched basalts have the lowest  $\epsilon_{\text{Nd}}$  values (+0.6 and -1.4) and the highest  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios (0.7044 to 0.7049) compared to all other Newer basalts (Fig. 5a). Apart from the REE and Y-enriched Newer basalts, two other localities, both ocean islands (Hawaii and Norfolk Island, Tasman Sea), have been reported with basalts having similar REE and Y enrichments (GREEN, 1978; RODEN, M. F. *et al.*, 1984). A Hawaiian Koolau tholeiite (69TAN-2), with a similar REE and Y enrichment, has a higher  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio and a lower  $\epsilon_{\text{Nd}}$  value compared to other Koolau tholeiites (Fig. 5a) (RODEN, M. F. *et al.*, 1984); this difference in Sr and Nd isotope composition is similar to that found in the Newer REE and Y-enriched basalts. The characteristic offset in Sr and Nd isotopic compositions reflects a time-integrated enrichment in these two source regions, and documents that this process can operate in both the sub-oceanic and sub-continental mantle. At present no adequate geochemical or physical model has been proposed which can explain the observed features in the REE-Y enriched lavas (FREY *et al.*, 1978; GREEN, 1978; RODEN, M. F. *et al.*, 1984). A geochemical and isotopic study of normal and REE and Y-enriched Norfolk Island basalts is underway, and a more detailed discussion of this feature will be presented elsewhere (MCDONOUGH *et al.*, in prep.).

The Tasmanian basalts (Fig. 2c) have incompatible element characteristics similar to those of the Newer basalts; there are progressive increases in incompatible element concentrations with increasing degree of silica undersaturation. Compared with a typical Newer primary basanite (*e.g.*, 2679), Tasmanian

primary basalts (Fig. 2c) have a more fractionated pattern (steeper slope) for Rb, Ba and Th; in addition K contents are more depleted with respect to Th and Nb. These features reflect compositional differences between the source regions of the Newer basalts and Tasmanian basalts.

#### $\text{P}_2\text{O}_5$ versus Nd

$\text{P}_2\text{O}_5$  abundances strongly correlate with Nd concentrations for the Newer and Tasmanian basalts (insert, Fig. 4). Most of the Newer basalts show a constant  $\text{P}_2\text{O}_5/\text{Nd}$  ratio of  $189 \pm 6$  ( $2\sigma_{\text{mean}}$  uncertainty). Three groups of Newer basalts show deviations from this ratio: (1) the olivine analcimites ( $\text{P}_2\text{O}_5/\text{Nd} = 129$  and 150) which also has significant depletions in K and Rb, (2) the Anakies (east) nepheline mugearites (136 and 141), and (3) the REE and Y-enriched basalts (62, 74 and 83). Differences in the  $\text{P}_2\text{O}_5/\text{Nd}$  ratio of the olivine analcimites and the REE and Y-enriched basalts may reflect compositional heterogeneities in their source regions. Apatite megacrysts are common in the Anakies nepheline mugearites (IRVING, 1974) and small amounts of these megacrysts will significantly change this ratio. The  $\text{P}_2\text{O}_5/\text{Nd}$  ratio for the Tasmanian basalts is  $201 \pm 16$  and overlaps with that of the Newer basalts. There is no difference in the  $\text{P}_2\text{O}_5/\text{Nd}$  ratio between tholeiitic and alkalic basalts from either province indicating that the magma source region beneath southeastern Australia has a  $\text{P}_2\text{O}_5/\text{Nd}$  of 190 to 200. The  $\text{P}_2\text{O}_5/\text{Nd}$  ratio for continental alkaline basalts and OIB shows considerable overlap, with a total range from 140 to 200, suggesting that an average  $\text{P}_2\text{O}_5/\text{Nd}$  for the mantle source regions for hotspot magmatism is about  $170 \pm 30$  ( $\text{P}/\text{Nd} = 74 \pm 13$ ). This value agrees with SUN's (1982) primitive mantle estimate for  $\text{P}_2\text{O}_5/\text{Nd} = 155$  ( $\text{P}/\text{Nd} = 67$ ), but is 35% greater than the primitive earth mantle estimate of 110 ( $\text{P}/\text{Nd} = 50$ ) proposed by WECKWERTH *et al.* (1983). The positive correlation between P and Nd is similar to that previously observed for P and Ce (SUN and HANSON, 1975; FREY *et al.*, 1978).

#### Sr and Nd isotopic compositions

The  $^{87}\text{Sr}/^{86}\text{Sr}$  values of the Newer basalts are distinctly higher and  $\epsilon_{\text{Nd}}$  values lower than those of the Tasmanian basalts (Fig. 5a). Excluding the anomalous REE and Y-enriched basalts, the Newer basalts have a range of  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios from 0.7038 to 0.7045, and  $\epsilon_{\text{Nd}}$  values from +3.8 to +1.7. The  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios for Tasmanian basalts range from 0.7026 to 0.7034, and  $\epsilon_{\text{Nd}}$  varies from +5.8 to +7.6. The range in  $^{87}\text{Sr}/^{86}\text{Sr}$  for the Newer basalts is similar to that previously reported (DASCH and GREEN, 1975; STUCKLESS and IRVING, 1976). There are some significant discrepancies between the isotopic composition of whole rock powder splits of the same samples measured in the earlier studies and those reported here. For example, we determined  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios of 0.70386 and 0.70389 for the Mt. Porndon basanite (2128), whereas STUCKLESS and IRVING (1976) report a value of 0.7044 and DASCH and GREEN (1975) report a value of 0.7045. We obtained  $\epsilon_{\text{Nd}}$  values of +3.2 and +3.3 for samples 2128 and 2102, whereas, DEPAOLO (1978) reports values of  $+4.0 \pm 0.4$  and  $+4.5 \pm 0.3$  respectively, for these same samples (DEPAOLO (1978) mislabelled sample 2101 and 2120). DEPAOLO's (1978)  $\epsilon_{\text{Nd}}$  values are higher, but allowing for laboratory bias these values agree at the limit of analytical uncertainties.

Victorian hawaiites, nepheline hawaiites, nepheline mugearites, alkali olivine basalts and basanites have a mean Sr and Nd isotopic composition of 0.70385 and +3.3, with relatively minor variation (Fig. 5a). Primary magmas, as defined above, cover the measured range in Sr and Nd isotope compositions for the Newer basalts. The Mt. Shadwell basanite has a  $\epsilon_{\text{Nd}}$  value of +3.7 and a  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio of 0.7038 and the Mt. Watch olivine analcinite (from South Australia) has a  $\epsilon_{\text{Nd}}$  value of +1.8 and a  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio of 0.7043; both of these basalts satisfy the criteria for primary magmas. Two exceptions to this are the Mt. Franklin nepheline mugearite which has a



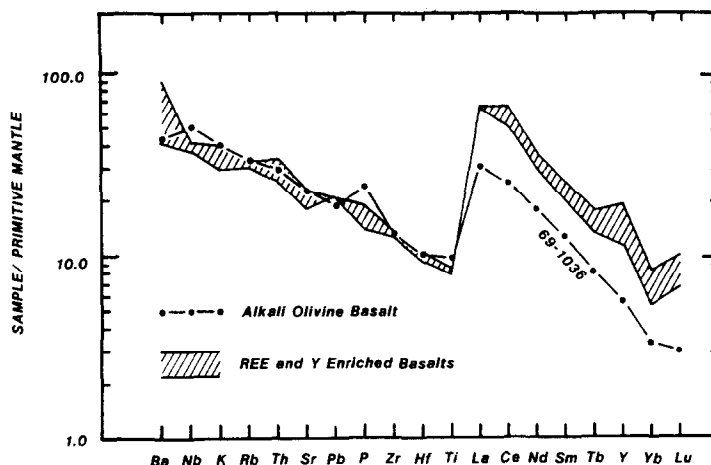


FIG. 3. Primitive mantle normalized incompatible element abundances of Newer REE and Y-enriched basalts compared to a typical Newer alkali olivine basalt. Normalizing values are as reported in Fig. 2, except for Pb which is 0.12. Trace element data from Table 2 and FREY *et al.* (1978).

slightly higher  $\epsilon_{\text{Nd}}$  value of +4.2 and a similar  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio of 0.7038 and the REE-Y enriched Newer basalts which have much lower  $\epsilon_{\text{Nd}}$  values and generally higher  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios. Victorian tholeiites and olivine basalts show a greater range in their Sr and Nd isotope compositions than that found in the Victorian alkaline basalts, with all but the Fe-rich olivine basalt plotting at higher  $^{87}\text{Sr}/^{86}\text{Sr}$  and lower  $\epsilon_{\text{Nd}}$  values (Fig. 5a inset). The four South Australian Newer basalts, a nepheline mugearite, a nepheline hawaiite and two olivine analcimites, have a large spread in isotope compositions from  $\epsilon_{\text{Nd}}$  of +3.3 to +1.8 and  $^{87}\text{Sr}/^{86}\text{Sr}$  from 0.7039 to 0.7045. This data indicates lower  $\epsilon_{\text{Nd}}$  values and higher  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios in the source region beneath the far western part of the field (South Australia) and are in contrast with those measured for the Victorian Newer alkaline basalts. Finally, the transitional to tholeiitic Victorian basalts plot in an array projecting away from the field of the alkaline Victorian basalts.

Parent/daughter ratios for the Sr and Nd isotope systems in both the Newer and Tasmanian basalts show different patterns. The Sm/Nd ratios of the Tasmanian basalts vary from 0.19 to 0.23 and overlap with Sm/Nd ratios for the Newer basalts (0.17 to 0.26), whereas, Rb/Sr ratios for the two regions are distinctly different. Tasmanian basalts have Rb/Sr ratios of 0.010 to 0.029; Newer basalts have Rb/Sr ratios of 0.039 to 0.087, except for the two olivine analcimites with ratios of 0.009 and 0.050. Although the Rb-Sr systematics of the two regions can be accounted for by a single stage fractionation event, this is not consistent with the Sm-Nd isotope systematics.

The Newer basalts have higher  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios and relatively flat normalized patterns for the most incompatible elements (Rb, Ba, Th, K and Nb) (Fig. 2a and 2c). In comparison, the Tasmanian basalts have lower  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios and a more fractionated, depleted pattern for these elements (*cf.* the pattern of 2679 in Fig. 2c). The low  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios and the concave downward pattern for these elements in the Tasmanian basalts suggests that their mantle source region has had a long-term depleted character in contrast to that for the Newer basalts. Depletions in incompatible elements and low  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios are also features which characterize the MORB source region.

The entire eastern Australian borderland is peppered with large Cenozoic subalkaline to alkaline basalt provinces; Sr and Nd isotope compositions have been measured in only the Southern Highlands and nearby Mesozoic Kiama alkaline basalts (MENZIES and WASS, 1983). The Sr and Nd isotopic fields of the Tasmanian, Newer and REE-Y enriched Newer basalts are more restricted compared to the field for the Southern Highland basalts (Fig. 5b); collectively these regions exhibit a large range in Sr and Nd isotope compositions for the whole of the eastern Australian Cenozoic basalt province. This range in Sr and Nd isotope compositions is similar to that observed in other Cenozoic continental intraplate basalt provinces, including west Antarctica (FUTA and LE MASURIER, 1983), Kenya (NORRY *et al.*, 1980), the Basin and Range, and Colorado Plateau, (MENZIES *et al.*, 1983; ALIBERT and ALBARÈDE, 1984), western Europe (ALIBERT *et al.*, 1983; CHAUVEL and JAHN, 1984), China (ZHOU and CARLSON, 1982) and New Zealand (BARREIRO, 1983; McDONOUGH and McCULLOCH, unpublished data). All of these Cenozoic continental intraplate basalt provinces have Sr and Nd isotope compositions that generally plot on the lower left side of the oceanic mantle array, with  $\epsilon_{\text{Nd}}$  values ranging from greater than +8.0 to about -1.0.

#### Pb isotopic compositions

Pb isotope data for 5 South Australian and Victorian Newer basalts have been previously reported (COOPER and GREEN, 1969), and we have recently measured the Pb isotopic composition of the Mt. Porndon basanite (2128) which has a value of  $18.58 \pm 0.02$   $^{206}\text{Pb}/^{204}\text{Pb}$ ,  $15.57 \pm 0.02$   $^{207}\text{Pb}/^{204}\text{Pb}$  and  $38.55 \pm 0.04$   $^{208}\text{Pb}/^{204}\text{Pb}$  (analytical methods reported in NELSON *et al.*, 1985). There is reasonably good agreement between the two data sets. The Pb isotope compositions for the 4 Newer basanites are similar, consistent with their similar Sr and Nd isotopic and major and trace element compositions. The  $^{206}\text{Pb}/^{204}\text{Pb}$  compositions of the South Australian basalts are slightly lower than those of the Victorian basalts. There are distinct differences in the Sr, Nd and Pb isotopic compositions of the South Australian Newer alkaline basalts when compared to the Victorian Newer alkaline basalts, indicating a compositional source difference from east to west across the

a Newer basanite which is considered to be a primary magma. Primitive mantle normalizing values (in ppm) are given above the element labels and are from SUN (1982), modified to the C1 abundances reported by ANDERS and EBIHARA (1982). In order to obtain a smooth pattern the positions of Sm and Hf have been reversed. A Zr/Hf ratio of 31.0 (value for C1 chondrite Orgueil) was used for normalization, whereas a Zr/Hf ratio of  $38 \pm 2$  for the upper mantle and Bulk Earth may be more appropriate.

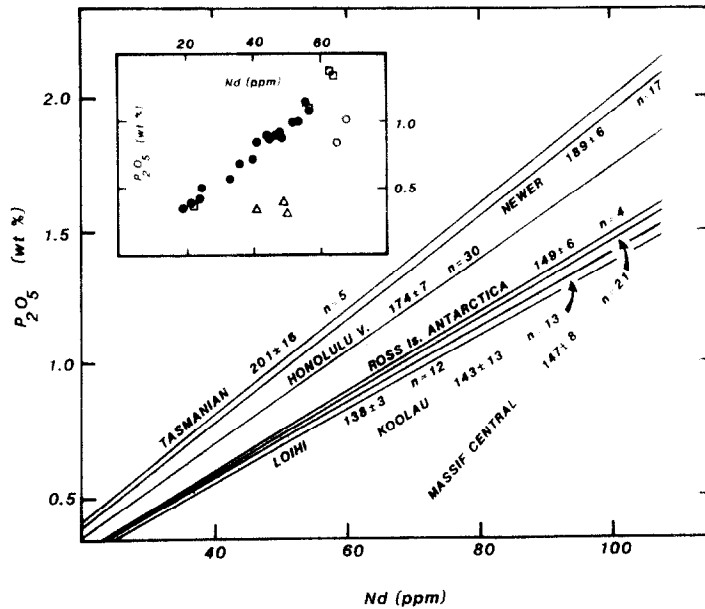


FIG. 4.  $P_2O_5$  (wt%) versus Nd (ppm) for continental alkaline basalts and OIB. The location and  $P_2O_5$ /Nd ratio for different areas are given with a reported  $2\sigma_{\text{mean}}$  uncertainty and  $n$  = the number of analyses used in determining the ratio. Data sources for the different localities are SUN and HANSON (1975), FREY *et al.* (1978), CLAGUE and FREY (1982), FREY and CLAGUE (1983), RODEN, M. F. *et al.* (1984) and CHAUVEL and JAHN (1984). Inset diagram shows a detailed  $P_2O_5$  (wt%) versus Nd (ppm) plot for Newer (filled circles) and Tasmanian basalts (open squares). The REE and Y enriched basalts (open triangles) and olivine analcimites (open circles) are shown separately, the Anakies (east) nepheline mugearites samples are not plotted.

Newer volcanic field. On a  $^{207}\text{Pb}/^{204}\text{Pb}$  versus  $^{206}\text{Pb}/^{204}\text{Pb}$  and a  $^{208}\text{Pb}/^{204}\text{Pb}$  versus  $^{206}\text{Pb}/^{204}\text{Pb}$  diagram (Fig. 6) the data for the Newer basalts plot in a broad vertical array between the

fields of Gough Island and Hawaiian Islands (SUN, 1980; COHEN and O'NIONS, 1982b), indicating their Pb isotopic compositions are similar to those observed in OIB.

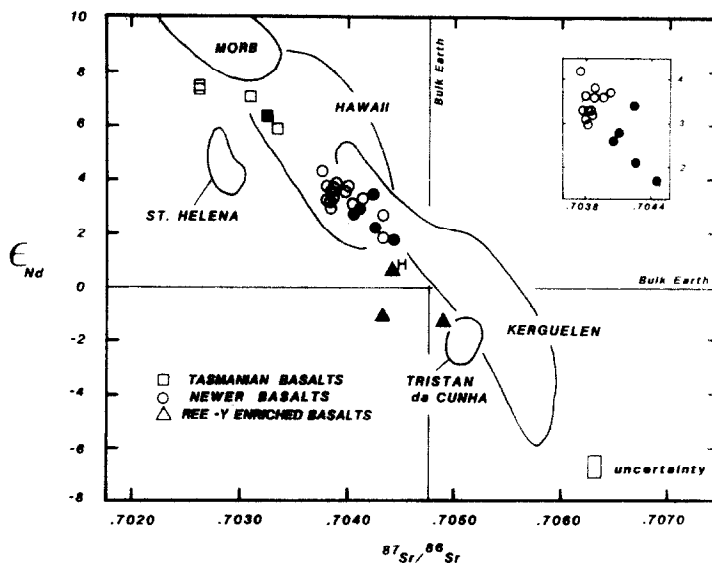


FIG. 5a.  $^{87}\text{Sr}/^{86}\text{Sr}$  versus  $\epsilon_{\text{Nd}}$  variation for the Tasmanian, Newer and REE and Y enriched Newer basalts. Filled symbols indicate the tholeiitic basalts. Also shown are the Sr-Nd isotope fields for MORB, and the ocean islands of St. Helena, Hawaii, Kerguelen and Tristan da Cunha (DEPAOLO and WASSERBURG, 1976; O'NIONS *et al.*, 1977; DOSSO and MURTHY, 1980; WHITE and HOFMANN, 1982; COHEN and O'NIONS, 1982a,b; STILLE *et al.*, 1983; RODEN, M. F. *et al.*, 1984; STAUDIGEL *et al.*, 1984). The 'H' data point is for the REE and Y enriched Koolau tholeiite, Ohau from RODEN, M. F. *et al.* (1984). Inset diagram shows the  $^{87}\text{Sr}/^{86}\text{Sr}$  versus  $\epsilon_{\text{Nd}}$  variation for only the Victorian tholeiitic (filled circles) and alkalic Newer basalts (open circles), and does not include the REE and Y-enriched, and South Australian, Newer basalts.



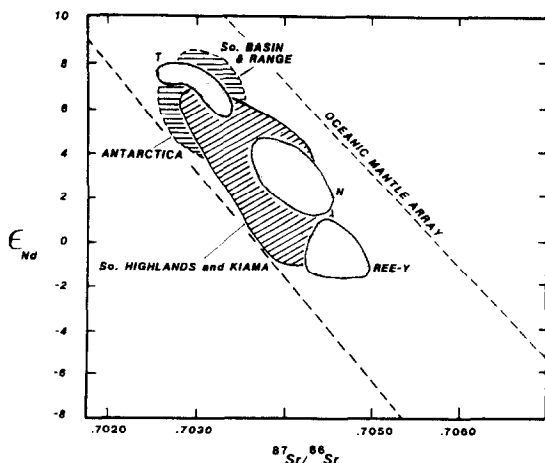


FIG. 5b.  $^{87}\text{Sr}/^{86}\text{Sr}$  versus  $\epsilon_{\text{Nd}}$  variation for the Tasmanian basalts (T), Newer basalts (N) and REE and Y enriched Newer basalts (REE - Y). Also shown are the Sr-Nd isotope fields for continental alkaline basalt provinces: South Basin and Range (MENZIES *et al.*, 1984), Antarctica (FUTA and LE MASURIER, 1983), Southern Highlands and Kiama (MENZIES and WASS, 1983). It should be noted that many other continental alkaline basalt provinces plot in this same region and have been omitted for the sake of clarity and legibility, these include: Kenya (NORRY *et al.*, 1980), Colorado Plateau (ALIBERT and ALBARÈDE, 1984), western Europe (ALIBERT *et al.*, 1983; CHAUVEL and JAHN, 1984), China (ZHOU and CARLSON, 1982) and New Zealand (BARREIRO, 1983; McDONOUGH and McCULLOCH, unpublished data). The Oceanic Mantle Array has been drawn to encompass the data for MORB and the ocean islands including St. Helena, the Austral islands and the Comores islands.

## DISCUSSION

### 1) Evaluating lithospheric contamination

CARLSON *et al.* (1981) and MAHONEY *et al.* (1982) have suggested that some continental basalts with greater than 5 wt% MgO have been contaminated by continental crust. They proposed that the observed spread in Sr and Nd isotope compositions are in part due to a crustal contamination process. MAHONEY *et al.* (1982) considered that one of the contaminating components may be enriched mantle. Others have argued that the observed chemical and isotopic variations found in continental intraplate basalts may reflect heterogeneities in the mantle or varying amounts of mixing between two or more mantle components, and there is no need to call upon crustal contamination (*e.g.*, DEPAOLO, 1983; MENZIES *et al.*, 1984). It is therefore necessary to evaluate whether the southeastern Australian basalts have had their chemical and isotopic composition affected by crustal contamination. In addition it is necessary to consider separately the roles of the crustal lithosphere and the lithospheric mantle when evaluating the effects of contamination. We define the crust as that portion of lithosphere above the Moho, and the lithospheric mantle is sandwiched between the Moho and the low velocity zone (LVZ). In the absence of a seismically defined LVZ we consider the base of the lithosphere to be the lowermost portion

of the lithospheric mantle that is mechanically coupled to the continental crust.

*i) crustal contamination*—The southeastern Australian primary basalts and many slightly more evolved (Mg-values > 59) lavas carry spinel lherzolite nodules and a limited number of felsic granulite facies nodules. The lherzolite xenoliths are angular to subrounded and range up to 25 cm in diameter. The composition, angularity and size of these lherzolite xenoliths attest to 1) lithospheric fracturing during ascent of these magmas, 2) the rapid uninterrupted ascent (hours to days) of these magmas through the lithosphere, and 3) the mantle origin of these magmas. The high magma volatility (most are erupted as cinder cones) indicates a positive volatile pressure for these magmas during ascent, and suggests that these magmas would principally contaminate the surrounding environment without any (or very limited) uptake of crustal derived components.

The primitive nature of the samples chosen (*i.e.*, high Mg-value > 67, Ni > 280 ppm, Cr > 300 ppm, and presence of Cr-diopside lherzolite nodules) should mean that they had little opportunity to be contaminated by crustal rocks. The primary basalts cover the total range in Sr and Nd isotope composition for the Newer and Tasmanian fields. Evolved basalts, analyzed to specifically evaluate the possible role of crustal contamination, have relatively low Mg-values (59 to 41) which are attributed to crystal fractionation at crustal pressures (IRVING and GREEN, 1976; FREY *et al.*, 1978). These basalts are the most likely candidates to show

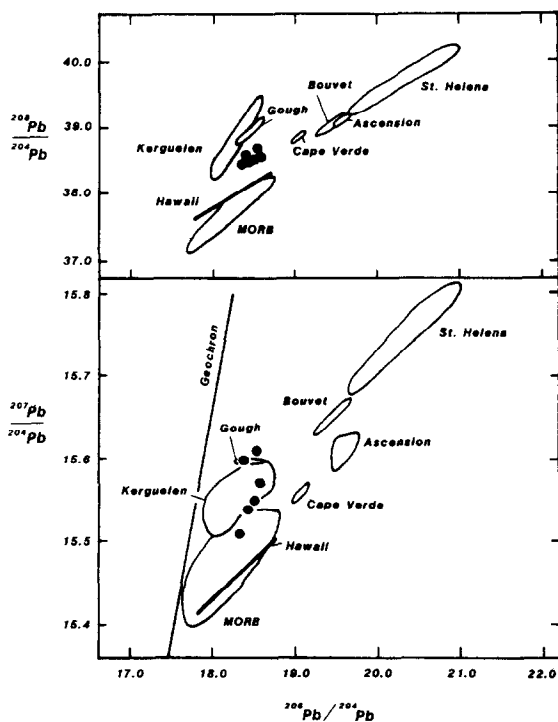


FIG. 6.  $^{207}\text{Pb}/^{204}\text{Pb}$  versus  $^{206}\text{Pb}/^{204}\text{Pb}$  and  $^{208}\text{Pb}/^{204}\text{Pb}$  versus  $^{206}\text{Pb}/^{204}\text{Pb}$  for Newer basalts (solid circles) (COOPER and GREEN, 1969; and this study) and OIB and MORB (data from SUN, 1980; COHEN and O'NIONS, 1982a,b).

any effects of crustal contamination. The isotopic compositions of the evolved basalts are not distinctive from the primary basalts, implying crustal contamination is not significant. In addition, the Fe-rich olivine basalt (69-1006) with the lowest Mg-value plots on the high  $\epsilon_{\text{Nd}}$  side of the Sr-Nd isotopic array. This feature is opposite to what would be expected if a fractionated lava was contaminated by crust similar to that exposed in the region. From these observations we conclude that the Tasmanian and Newer basalts have not been contaminated by the crust, or at least not to any detectable extent.

Internal melts in the few felsic, granulite facies xenoliths that were found may be derived from the melting of these xenoliths, or they may represent the infiltration of host basalt melt, or some combination of these two processes. Preliminary microprobe analyses of some internal melt pockets suggests that they are derived by decompression melting of the felsic xenolith. Some internal melts in felsic granulite xenoliths from the Snake River Plains basalts are derived from the melting of the xenolith (LEEMAN *et al.*, 1985). The lherzolite nodules show only limited interaction with the host basaltic melt at the margins of the xenoliths with a few showing limited invasion of basaltic melt several millimeters into the xenolith. Such relationships demonstrate the limited ability of the basaltic magmas to infiltrate and contaminate their xenoliths, and presumably their surrounding conduit system, throughout the entire lithosphere. This process could conceivably result in some minor crustal contamination which would be most pronounced in the fractionated lavas, but as stated, there is no evidence for this.

There is no evidence to suggest that the REE and Y-enriched Newer basalts inherited their unusual chemical and isotopic characteristics as a result of crustal contamination. Because the relative concentrations of the non-REE trace elements in these basalts are similar to those in other Newer basalts (Fig. 3) this restricts the composition of the enriched phase, or processes which may have generated this signature. Also, since this 2 to 3 fold enrichment in the REE and Y has been identified in ocean island basalts, this suggests that the source of this enrichment is in the mantle. This chemical signature may be a result of hotspot magmatism.

*ii) lithospheric mantle contamination*—The extent to which Cr-diopside lherzolite nodules, presumably fragments of the lithospheric mantle, might have modified the Sr and Nd isotope compositions of Newer and Tasmanian basalts must also be considered. Sr and Nd isotope compositions of spinel lherzolites from a number of Newer volcanic centers show a wide variation in composition from  $^{87}\text{Sr}/^{86}\text{Sr}$  0.7037 to 0.7085 and  $\epsilon_{\text{Nd}} = +11.4$  to  $-7.5$  (CHEN and FREY, 1981; MCDONOUGH *et al.*, unpublished data). This is in contrast with the limited range in isotopic composition for their host basalts, most particularly the four lherzolite nodule-bearing basanites with identical Sr, Nd and Pb isotope compositions (Mt. Shadwell, Mt. Leura, Mt. Noorat

and Mt. Porndon). There is no means of constraining the average isotopic composition of this part of the sub-continental lithospheric mantle, but the uniform isotopic compositions of these basalts would not be anticipated if lherzolite nodules with such a diversity of isotopic compositions had contaminated these lavas. Also there is no textural evidence for the melting of lherzolite nodules by their host basalt. This suggests that the basalts have not been significantly contaminated by the entrained lherzolite nodules.

While the spinel lherzolite xenoliths are unlikely to have modified their host basaltic magmas, partial melts derived from similar lherzolitic material at greater depths may be important in modifying the compositions of these basalts. Partial melts derived from garnet lherzolites in the lower parts of the sub-continental lithospheric mantle may be added to basaltic magmas prior to xenolith entrainment and eruption. Such material may represent a mantle component which is common to continental intraplate basalts. Since the Newer basalts are isotopically relatively uniform, and the southeastern Australian sub-continental lithospheric mantle is isotopically diverse (as inferred from Victorian spinel lherzolites), then an added lithospheric-mantle melt component must be homogenized with an ascending mantle diapir or its basaltic magmas prior to eruption. In a later section we consider possible mechanisms by which melt material derived from the lithospheric mantle may be added to these basaltic magmas.

## 2) Scales of isotopic homogeneity in the mantle beneath SE Australia

Several studies have identified different scales of isotopic homogeneity in the mantle. Sr and Nd isotope studies of MORB have shown that some ridge segments possess a uniform isotopic composition on a scale of 10 to 50 kilometers (*e.g.*, DUPRÉ *et al.*, 1981), and in an extreme example MACDOUGALL and LUGMAIR (1985) found uniform Sr and Nd isotopic compositions for MORBs along a 550 kilometer section of the East Pacific Rise. On a much larger scale HART (1984) recently suggested the existence of a distinctive globe-encircling Sr and Pb isotope signature in the southern hemisphere region of the mantle. Hart suggested that this portion of the mantle has had a long term history which is in general compositionally different from other parts of the mantle.

Chemical and isotopic homogeneity over a relatively small scale has been identified in the Newer volcanic field. Four different basanite centers (Mt. Shadwell, Mt. Leura, Mt. Noorat and Mt. Porndon), separated by more than 50 km, have uniform chemical and isotopic compositions (FREY *et al.*, 1978; and this study). Abundant upper mantle nodules and very few crustal nodules are found at these localities. Mt. Shadwell, Mt. Noorat and Mt. Porndon basanites are primary, mantle-derived magmas. The isotopic compositions of these basanites are all within analytical uncertainties:  $^{87}\text{Sr}/^{86}\text{Sr}$  of .70385,  $\epsilon_{\text{Nd}}$  of +3.4, and  $^{206}\text{Pb}/^{204}\text{Pb}$  of 18.5,

$^{207}\text{Pb}/^{204}\text{Pb}$  of 15.55 and  $^{208}\text{Pb}/^{204}\text{Pb}$  of 38.5. This data indicates that there possibly exist regions in the upper mantle beneath southeastern Australia which are isotopically homogeneous over a scale of about 50 kilometers.

On a larger scale, the variation in Sr and Nd isotope composition for the Newer basalts is about half of that found for tholeiitic and alkalic basalts from Oahu, Hawaii (STILLE *et al.*, 1983; RODEN, M. F. *et al.*, 1984), which covers most of the known variation in Sr and Nd isotope compositions for Hawaiian basalts (Fig. 5a). The range in Sr and Nd isotope data for the Tasmanian basalts is similar to that of the Newer basalts, but because of the limited sampling this data may not adequately represent the total range for these basalts.

The presence of low  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios ( $<0.7040$ ) in the Newer and Tasmanian basalts (see also COMPSTON *et al.*, 1968), and for basalts from the Austral islands (VIDAL *et al.*, 1984; MCDONOUGH *et al.*, unpublished data), Rodrigues island (BAXTER *et al.*, 1985), and Comores islands (WHITE and DUPRÉ, 1984), all of which are southern hemisphere localities centered on the conjectured DUPALL anomaly (HART, 1984), is not consistent with the proposed large scale mantle anomaly in this region.

### 3) Geochemical characteristics of the mantle source region

Both the Newer and Tasmanian basalts show progressive increases in their incompatible element concentrations with increasing degree of silica undersaturation at a relatively constant Mg-value (Figs. 2a–c). This suggests that different degrees of partial melting of relatively similar mantle sources could possibly explain the chemical variation found in the olivine tholeiite to olivine nephelinite/melilitite spectrum in both regions.

A detailed major and trace element melting model has been presented by FREY *et al.* (1978) for many of the basalts reported here. FREY *et al.* (1978) used the REE to test an equilibrium melting model to explain variations in melt composition from olivine tholeiite to olivine basalt to alkali olivine basalt to basanite to olivine nephelinite/melilitite. Their model requires decreasing degrees of partial melting of a garnet lherzolite source to account for this spectrum of basalt compositions. They argued that the source could possess either a LREE-enriched or chondritic REE pattern. Their model calculations showed that to derive this spectrum of basalts from a LREE-enriched ( $\text{LREE} = 7$  to  $9 \times$  chondrites,  $\text{HREE} = 2.5$  to  $3 \times$  chondrites) source requires a range in the degree of partial melting from 20–25% for the tholeiites to 4–6% for the nephelinite/melilitite. Using a chondritic source ( $\text{LREE}_N/\text{HREE}_N = 1.0$ ), with about 2 to 3 times chondrite REE abundances, requires varying the degrees of partial melting from about 5% to about 0.5% to generate this spectrum. This type of chemical argument can be carried a step further by attempting to derive this range of basalt

compositions from a LREE-depleted source region, since all but the high Y-REE bearing basalts have positive  $\epsilon_{\text{Nd}}$  values, requiring the source region(s) for these basalts to have a time integrated LREE-depleted ( $\text{Sm}/\text{Nd} > 0.31$ ) character. To derive the olivine tholeiites in a single stage melting process requires less than 1% partial melting, and considerably less for the olivine nephelinites/melilitites. Although geophysical models cannot place strict lower limits on the percentage of partial melt which can be extracted from the upper mantle (STOLPER *et al.*, 1981; MCKENZIE, 1984), to generate tholeiitic basalts by about 1% partial melting, or less, is unlikely, as this would require a source that is very depleted chemically and mineralogically. We consider that about 5 to 25% partial melting of an upper mantle peridotite is needed for the generation of tholeiitic basalts (JAQUES and GREEN, 1980), with the absolute degree of partial melting depending on the proportion of fertile components (*i.e.*, garnet and clinopyroxene). Therefore, we suggest that in agreement with earlier findings (FREY *et al.*, 1978), the Newer and Tasmanian olivine tholeiites were generated by up to 20 to 25% partial melting and the olivine nephelinites/melilitites by 3 to 6% partial melting of a LREE-enriched garnet lherzolitic upper mantle or by about 5% partial melting for the tholeiites down to about 0.5% partial melting for the olivine nephelinites/melilitites for a garnet lherzolitic upper mantle with a chondritic REE pattern. In addition to the melting models tested by FREY *et al.* (1978), there are other more sophisticated melting models which may more accurately describe melting and magma extraction processes (LANGMUIR *et al.*, 1977; O'HARA, 1985). Although we have not tested such models, the application of these models are important to our understanding of source composition and melt volumes involved.

Though it is suggested that the calculated REE source patterns for these two regions are similar, the Newer and Tasmanian basalts come from isotopically distinct sources (Fig. 5a). Additionally, there are significant differences between the relative concentrations of K, Th, Ba and Rb in the two regions (Figs. 2a–c), which reflects differences in their upper mantle source compositions. Finally, the upper mantle source region for the Newer and Tasmanian basalts has a time integrated Sm/Nd ratio greater than its present Sm/Nd ratio, as indicated by the positive  $\epsilon_{\text{Nd}}$  values of these basalts.

Relatively recent enrichment events can produce a basaltic source region which possesses a Sm/Nd ratio that is less than the time integrated Sm/Nd ratio. For example, RODEN, M. K. *et al.* (1984) found that the St. Paul's Rocks ultramafics have a LREE-enriched pattern and a positive  $\epsilon_{\text{Nd}}$  value. They suggested that these ultramafic rocks inherited this LREE-enrichment about 155 m.y. ago, and found that a LREE-enriched alkali basalt from the massif possessed a similar  $\epsilon_{\text{Nd}}$  value, suggesting that the ultramafic material of the massif is an ideal mantle source from which the alkali basalts from the flanks of the massif can be derived by

partial melting. There exists similar ultramafic material derived from the sub-continental lithospheric mantle beneath this part of southeastern Australia (*i.e.*, Cr-diopside bearing spinel lherzolite xenoliths) which have both chondritic and LREE-enriched patterns and positive  $\epsilon_{Nd}$  values (FREY and GREEN, 1974; CHEN and FREY, 1981; MCDONOUGH *et al.*, unpublished data). By analogy with the model of RODEN, M. K. *et al.* (1984) it might be possible to derive the Newer basalts from such lherzolitic material. It is also possible that partial melts derived from such material are added to plume-derived melts in order to generate the range in Sr and Nd isotope compositions in the Newer basalts. In the following section we discuss these possibilities.

#### 4) *The nature of the source region and its components*

It has been proposed that the Tertiary basalts along the entire eastern Australian borderland resulted from several hotspot traces (WELLMAN and MCDUGALL, 1974; SUTHERLAND, 1981). It is not known whether the initiator of basaltic volcanism in the region is a deep seated mantle plume, physically similar to that envisioned for Hawaii, or results from a shallower thermal perturbation of unknown origin. In either case both would provide upwelling plumes of hot mantle material which ultimately supply the region with basaltic volcanism. Thus, there remains the question of the nature and origin of the different mantle components which leads to the range in Sr and Nd isotope compositions. To generate the observed spectrum of isotope compositions we consider a model involving mixing of two components: a plume component and a lithospheric mantle component.

A variety of mixing models have recently been proposed to explain systematic trends in chemical and isotopic data for basalts. CHEN and FREY (1983) have presented one such mixing model, in which there is mixing between two mantle components, one, the plume and the other, the surrounding mantle through which the plume is ascending. This model involves wall rock reaction between an ascending hot mantle plume and its surrounding, relatively cooler upper mantle peridotite. Chen and Frey's model involves mixing of very small degrees (<1.0%) of partial melt extracted from a cooler, depleted MORB-like source material (wall rock) with larger degrees of partial melt (1% to 5%) from a hotter mantle plume-derived material. Melt from the MORB-like source material would therefore possess alkaline affinities, a LREE-enriched pattern and a more positive  $\epsilon_{Nd}$  value than melt derived from the plume component. The plume-derived melt would have a composition that could vary from alkalic to tholeiitic depending on the degree of partial melting involved and the amount of clinopyroxene and garnet in the plume component, and its REE pattern would be determined by the source pattern and phases in the residuum.

The Chen and Frey model provides a good working model and has considerable merit, but there are some

significant questions that must be resolved. (1) Can the major element variations be explained by mixing varying proportions of these two components? (2) Can both chemical and isotopic endmember compositions be realistically constrained? It must be again stressed (as noted by these authors) that there is no *a priori* reason to call upon one endmember with a bulk earth Sr and Nd isotopic composition (*i.e.*,  $\epsilon_{Nd} = 0$  and  $^{87}Sr/^{86}Sr = 0.7047 \pm 5$ ). (3) Is their model wholly applicable to both the continental and oceanic environment? (4) What happens when a hot mantle plume intrudes the base of the lithosphere? Many of these questions cannot be rigorously answered and we do not intend to attempt this, but they must be considered when applying a model such as theirs. An important consideration of the Chen and Frey model which must be accounted for however are the consequences of a plume physically intersecting the base of the lithosphere. We predict that this will result in significant partial melting of the lithospheric mantle. With this in mind a lithospheric mantle-derived melt must be considered as an added component to any plume-derived melt and should be incorporated into any model invoking component mixing and mantle plumes, as is commonly done for intraplate volcanism.

We propose a model whereby the lithospheric mantle interacts with plume-derived melts during intrusion of an ascending mantle plume into the base of the lithosphere. The mixing model proposed here involves mixing a low degree of partial melt (alkalic in character) with a larger degree of partial melt (tholeiitic in character), as put forth in the Chen and Frey model, but that the larger degree of partial melt can represent melts of up to 20% melting. Also, in contrast to their model where the plume component mixes only with depleted mantle during the ascent of the plume, we propose that the plume component also mixes with a lithospheric mantle component, which in some cases may be a zone of partial melt (*e.g.*, LVZ?), as a result of the intersection of the hot plume with the base of the lithosphere. These mixtures of partial melts together contribute to the chemical and isotopic composition of the southeastern Australian basalts.

Figure 7 illustrates the model which we propose. An ascending mantle plume would intersect the base of the lithosphere. The base of the lithosphere is defined as the LVZ, and in the absence of a well defined LVZ we consider it to be the lowermost portion of the lithospheric mantle that is mechanically coupled to the continental crust. Intrusion of a mantle plume into this region would raise the temperature above the mantle solidus and initiate partial melting of the lithospheric mantle. Partial melting of lithospheric mantle may occur over a wide region. Lithospheric mantle-derived melts would mix with plume-derived melts and generate a spectrum of chemical and isotopic compositions as found in the southeastern Australian basalts. The mixing zone region is envisaged as the area where the plume intrudes the base of the lithosphere.

The range in plume Sr and Nd isotopic compositions

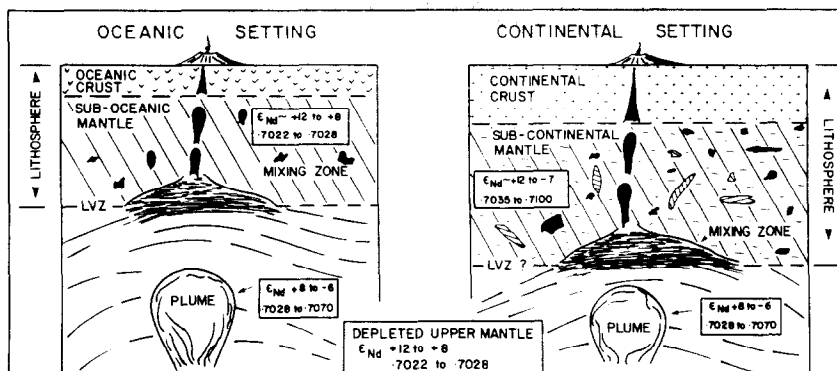


FIG. 7. A mantle model depicting the relative isotopic composition of the mixing components involved in oceanic and continental intraplate magmatism. A hot ascending mantle plume intrudes the base of the lithosphere and initiates melting of the lithosphere, magmas are mixtures of a plume derived component and a lithospheric mantle derived component. For simplicity, the Sr and Nd isotopic composition of the sub-oceanic lithospheric mantle and the depleted upper mantle is assumed to be equal to that of MORB, though we envision small, chemical and isotopic heterogeneities in these mantle reservoirs. The plume's isotopic composition is constrained by the range in Sr and Nd isotope ratios found in OIBs. The Sr and Nd isotopic composition of the sub-continental lithospheric mantle is estimated from the range in isotope ratios measured for ilherzolite nodules found in the Newer basalts (CHEN and FREY, 1981; MCDONOUGH *et al.*, unpublished data). The values assumed here are considered applicable for the Newer basalts *only*, and other continental regions would have differing isotopic compositions depending on their age and geologic history. The sub-oceanic and the sub-continental lithospheric mantles contain small scale heterogeneous regions (black, angular fragments), with the sub-continental lithospheric mantle containing more ancient chemical and isotopic heterogeneities (lined, angular fragments).

is based on the observed variation in  $\epsilon_{Nd}$  values and  $^{87}Sr/^{86}Sr$  ratios found in OIB (*e.g.*, St. Helena, Iceland, Kerguelen, Society) and outside of normal MORB isotope compositions: this includes  $\epsilon_{Nd}$  values of +8 to -6 and a  $^{87}Sr/^{86}Sr$  ratios of 0.7028 to 0.7070. The Sr and Nd isotopic composition of the sub-oceanic lithospheric mantle is based on the variation found in normal MORB, and this is the same range assigned to the depleted upper mantle. Although an over-simplification of the nature of the depleted upper mantle and sub-oceanic lithospheric mantle, such a model is justified by our present knowledge. It is recognized that the depleted upper mantle may contain small scale heterogeneities in a matrix of more uniform peridotite. Support for this model is found in the variable isotopic compositions of seamounts (ZINDLER *et al.*, 1984) and MORBs (COHEN and O'NIONS, 1982a). The sub-oceanic lithospheric mantle could possess small scale heterogeneities incorporated during lithosphere forming processes at spreading centers (SLEEP, 1984). The range in Sr and Nd isotopic compositions assumed for sub-continental lithospheric mantle is based on the variation observed in spinel ilherzolites from southeastern Australia (CHEN and FREY, 1981; MCDONOUGH *et al.*, unpublished data). This range in isotope compositions is specific to this region and may well be considerably larger as the number of analyzed samples increase. The Sr and Nd isotopic composition of other regions of sub-continental lithospheric mantle would differ depending on formation age and its history of depletion and enrichment events. The range in isotopic composition for the sub-continental lithospheric mantle is larger than for any of the other mantle reservoirs,

which is to be expected, considering the variety and complexity of magmatic and tectonic processes leading to the formation of this part of the mantle. We have represented the sub-continental lithospheric mantle as possessing various heterogeneous fragments of different ages and parent/daughter isotope ratios to depict the variable chemical and isotopic nature of this region (Fig. 7).

This model predicts that for OIBs the lithospheric component possesses an isotopic character that is, in general, indistinguishable from present day MORB, while for continental intraplate basalt, such as in southeastern Australia, the lithospheric component possesses a heterogeneous isotopic character. The melt component derived from the lithospheric mantle could either be from a LVZ-derived melt, or from partial melting of the lower lithospheric mantle during the intrusion of a mantle plume into this region of the lithosphere, and/or some combination of these two sources. Thus, this model incorporates features from both the SUN (1980) and the CHEN and FREY (1983) models. NIXON *et al.* (1981) have presented a similar model for kimberlite genesis, where magmas from a deeper, ascending diapir mix with, or are contaminated by, incompatible element-rich melts derived from the base of the lithosphere.

One of our primary concerns is the chemical and isotopic characterization of the different mantle components involved in the generation of intraplate basalts. The model presented here involves at least two components: the plume and the lithospheric mantle. Clearly there is no *a priori* reason to assume that either of these components are homogeneous and, in fact, there

is chemical and isotopic evidence which demonstrates the heterogeneous nature of the sub-continental lithospheric mantle. If we turn our attention to ocean islands, where the lithospheric mantle component is chemically and isotopically less complex than beneath the continents, then it might be possible to gain an understanding of the chemical and isotopic nature of the plume component.

The chemical and isotopic composition of tholeiitic basalts from ocean islands probably provides the most direct information on the nature of the plume component. This is a consequence of tholeiitic basalts requiring higher melting temperatures than alkalic basalts, and hence are generated by larger degrees of partial melting. This suggests that large volume tholeiitic basalts are, in general, dominated by the plume component. For example, in the Hawaiian islands, Kauai, Oahu and Maui, there is a consistent trend whereby the shield-building tholeiitic lavas possess higher  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios and lower  $\epsilon_{\text{Nd}}$  values than the post caldera collapse alkalic lavas whose compositions overlap with the field of MORBs (FEIGENSON, 1984; RODEN M. F. *et al.*, 1984; CHEN and FREY, 1983). This is consistent with the inference that the tholeiitic lavas are dominated by the plume component, whereas, the post caldera collapse alkalic lavas are dominantly derived from the sub-oceanic lithosphere. There are also examples where tholeiitic basalts have overlapping Sr and Nd isotopic compositions with the associated alkalic basalts (*e.g.*, Loihi, STAUDIGEL *et al.*, 1984), and other less common examples where tholeiitic basalts have lower  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios and higher  $\epsilon_{\text{Nd}}$  values (*e.g.*, Pacific seamounts, ZINDLER *et al.*, 1984), which could be due to selective melting of a heterogeneous mantle source, either in the plume or lithospheric mantle. These examples indicate the potential problem with any single interpretive model but, we believe do not exclude the general applicability of the model presented here.

For continental intraplate basalts (*e.g.*, Newer basalts) it is more difficult to characterize the plume component because of the heterogeneous nature of the sub-continental lithospheric mantle. Development of the sub-continental lithospheric mantle involves many varied and complex geological processes which results in compositional diversity in this part of the mantle. RINGWOOD (1982) pointed out that ascending megalith diapirs (*i.e.*, plume material) of refractory peridotite are intrinsically less dense than the surrounding mantle and would rise buoyantly. Megaliths which rise beneath continents are thus permanently trapped within the sub-continental lithosphere, whereas those that rise beneath oceans are emplaced within the sub-oceanic lithosphere and later subjected to further episodes of subduction. Also, the sub-continental lithosphere evolves over relatively longer time periods and through the addition of many complex processes (*e.g.*, subduction along the continental margin) compared to the sub-oceanic lithosphere. Thus, it is anticipated that the sub-continental lithospheric mantle would be inher-

ently more complex (isotopically and chemically) than the sub-oceanic lithospheric mantle.

##### 5) Mantle sources for tholeiitic and alkali basalts

A final point for consideration is the difference in isotopic compositions sometimes observed between tholeiitic and alkalic basalts in both ocean island and continental settings. This is evident in the central portion of the Newer volcanic field where there is a slight tendency for the tholeiitic basalts to have higher  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios and lower  $\epsilon_{\text{Nd}}$  values in comparison to the alkaline basalts (Fig. 5a inset). Though this feature is only discernible where the data is restricted to a limited geographical sampling within the field, its presence may be indicating the general nature of the plume component, whereas, the isotope composition of the alkalic Newer basalts may indicate the averaged isotopic composition of the regional sub-continental lithospheric mantle component. Several Hawaiian volcanoes have tholeiitic basalts which have distinctly different isotopic compositions compared to associated alkalic basalts. Therefore, for OIBs and continental intraplate basalts the isotopic composition of the tholeiitic basalts more directly reflects the nature of the plume component, whereas the alkaline basalts indicate the nature of the underlying lithospheric mantle. An alternative model is that the plume is a heterogeneous mixture (*e.g.*, a plum pudding model) where the late-stage, post-erosional alkalic basalts (*e.g.*, Honolulu volcanics) represent the low melting temperature peridotite material in the plume and the shield-building, tholeiitic basalts represent the more refractory peridotite. This latter model may be applicable to seamounts (*e.g.*, ZINDLER *et al.*, 1984), but would not be consistent for ocean islands (*e.g.*, Hawaii), because one would expect that such low melting temperature peridotite brought up in the plume would not survive the tholeiite shield-building stage.

#### CONCLUDING REMARKS

Two Tertiary-Recent volcanic provinces of southeastern Australia contain primary and evolved tholeiitic and alkalic basalts which have been derived by partial melting of a peridotitic mantle. No evidence of crustal contamination was found in the basalts studied here. Differences are found in the relative abundances of incompatible elements and isotope composition between the two basalt provinces. Local regions within these provinces show extreme compositional homogeneity. Slight differences in Sr and Nd isotope compositions exist between tholeiitic and alkalic basalts but, unlike the tholeiitic and alkalic basalt series from Hawaii (CHEN and FREY, 1983; STILLE *et al.*, 1983; RODEN, M. F. *et al.*, 1984) the differences are not large and the compositions overlap.

Melting models involving mixtures of mantle plume-derived melt and lithospheric mantle-derived melt

components are proposed for the generation of these basalts. Lithospheric mantle-derived melts may be ponded in the low velocity zones at the base of the lithosphere. Mixing of plume-derived and lithospheric mantle-derived melt components ( $\pm$  other mantle components) occur during the intrusion of hotspot mantle plumes into the base of the lithosphere. This model may also be applicable to the generation of OIB with geochemical differences between oceanic and continental basalts resulting from compositional differences in their underlying lithospheric mantle.

From this model we predict some general features of hotspot volcanism. Firstly, one of the components involved in OIB generation is derived from the sub-oceanic lithospheric mantle and, because of the young age of the oldest exposed oceanic crust, it possesses a Sr and Nd isotopic composition that is, in general, similar to that of MORB's. Secondly, for OIBs the Sr and Nd isotopic composition of the plume component can be estimated to be equal to or greater than the most radiogenic Sr and unradiogenic Nd isotope composition in the series, and these isotopic compositions will most likely (but not always) be found in the tholeiitic basalts. In continental regions the isotopic composition of the sub-continental lithospheric mantle component may not necessarily be equivalent to that of MORB's. Finally, in general, the larger the spread in Sr and Nd isotope compositions of both continental and oceanic basalts the greater the difference in isotopic compositions between the lithospheric mantle- and mantle plume-derived melt components.

The combined isotope data for late Tertiary basalts from the Newer and Tasmanian provinces (COMPSTON *et al.*, 1968; and this study), and from the Austral islands, (VIDAL *et al.*, 1984; McDONOUGH *et al.*, unpublished data), Rodrigues island (BAXTER *et al.*, 1985), and the Comores islands (WHITE and DUPRÉ, 1984) shows that a number of regions in southern latitude between 15°S and 45°S have  $^{87}\text{Sr}/^{86}\text{Sr}$  less than 0.7040, with some as low as 0.7026 to 0.7029 (Tasmanian, and the Austral islands). HART (1984) proposed a DUPALL isotopic anomaly for this region of the southern hemisphere based upon  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios greater than 0.7040. These low  $^{87}\text{Sr}/^{86}\text{Sr}$  values within the DUPALL anomaly area and occurrence of high  $^{87}\text{Sr}/^{86}\text{Sr}$  ( $>0.7040$ ) outside of the DUPALL belt (*e.g.*, Koolau tholeiites, Oahu, Hawaii, 0.7042, and Sao Miguel, Azores, 0.7047) is not consistent with the proposed DUPALL anomaly.

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