Rare earth element geochemistry of the Betts Cove ophiolite, Newfoundland: complexities in ophiolite formation

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Abstract—The Betts Cove ophiolite includes the components of typical ocean crust: pillow lavas, sheeted dikes, gabbros, and ultramafics. However, the trace element geochemistry of basaltic rocks is unusual. Three geochemical units are recognized within the lava and dike members. Within the pillow lavas, the geochemical units correspond to stratigraphic units. Upper lavas have 'normal' (*i.e.,* typical for ocean floor basalts) TiO_2 contents (0.75 to 2.0 wt%), heavy rare earth elements (HREE) values in the range 6–20× chondrites and chondrite-normalized REE patterns with relative LREE depletion. Intermediate lavas have TiO_2 contents between 0.30 and 0.50 wt%, HREE contents from 4–7× chondrites and extreme relative LREE depletion. Lower lavas have anomalously low TiO_2 contents (<0.30 wt%) and unusual convex-downwards REE patterns with REE abundances around 2–5× chondrite. These geochemical differences can be explained if the three groups were derived from different mantle sources. Independent mantle sources for the three units are consistent with their different $^{143}Nd/^{144}Nd$ ratios varying at 480 m.y.B.P. from 0.51222 in a lower lava to 0.51238 in an upper lava. The upper lavas may be partial melts of a source similar in composition to that of modern MORB, the intermediate lavas may be from a very depleted oceanic mantle (second stage melt), and the lower lavas may have formed by melting an extremely depleted mantle that had been invaded by a LREE-enriched fluid. A possible tectonic environment where these different sources could be juxtaposed is a back-arc or inter-arc basin.

INTRODUCTION

Since the interpretation of ophiolites as pieces of oceanic crust and mantle (Moores and Vine, 1971; Church and Stevens, 1971; Dewey and Bird, 1971; Davies, 1971), a multitude of papers on the internal stratigraphy, structure and geochemistry of these complexes has shown that each ophiolite has distinctive characteristics; in particular, there are marked variations in stratigraphy and basalt composition. Whereas some ophiolites have features typical of modern ocean crust (*e.g.,* Bay of Islands, Newfoundland and Oman) (Malpas, 1978; Suen *et al.,* 1979; McCulloch *et al.,* 1980; Hopson *et al.,* 1981), other ophiolites exhibit features, such as low TiO_2 in basalts, which have not yet been described from modern ocean floors (Sun and Nesbitt, 1978a; Coish and Church, 1979). The basaltic portions of ophiolites range widely in TiO_2 content (Church, 1977) from low TiO_2 ophiolites (*e.g.,* TiO_2 < 0.2% in the lower basalts at Betts Cove) to high TiO_2 ophiolites (*e.g.,* TiO_2 > 0.62% at Bay of Islands). The origin of such compositional variations is enigmatic. Miyashiro (1973) suggested the low TiO_2, high SiO_2 contents of some volcanic rocks in the Troodos ophiolite indicated that the complex formed at the base of an island arc rather than at a mid-ocean spreading center. Similarly, Sun and Nesbitt (1978a) noted unusual compositions of volcanics in the Troodos and Betts Cove ophiolites and suggested that they could not have formed at a 'normal' mid-ocean ridge but equivocated as to whether they represented crust formed at the base of an island arc or in a back-arc extensional

basin. Coish and Church (1978, 1979) described the Betts Cove and Blow-Me-Down ophiolites, Newfoundland and stressed the importance of different mantle source regions in producing different basaltic compositions in ophiolites, but the link between source regions in the mantle and tectonic environments was not discussed.

The Betts Cove ophiolite, Newfoundland, is an example of a low-TiO_2 ophiolite. Its stratigraphy and basalt chemistry can be described as 'unusual' in that these features are unlike those of 'classic' ophiolites such as Oman and the Bay of Islands. The 'unusual' chemical characteristics include extremely low contents of TiO_2 (<0.20 wt.%), Zr, Y and REE in basalts (Coish and Church, 1979). In order to provide a better understanding of the complex processes that formed the Betts Cove ophiolite, we present a detailed account of the rare earth element (REE) geochemistry of mafic rocks and correlate trends in REE abundances with bulk rock and mineral compositions. Mantle sources for the different basalt types are discussed and finally, a tectonic scheme for the formation of the ophiolite is described.

GEOLOGIC SETTING AND AGE OF THE BETTS COVE OPHIOLITE

The Betts Cove ophiolite is located on the eastern side of the Burlington Peninsula, northeastern Newfoundland on the western limit of the Central Mobile Belt (Dunnage Zone of Williams, 1979) (Fig. 1). It is 100 km east of the Bay of Islands ophiolite suite which lies on Ordovician continental shelf-slope sediments. The base of the ophiolite abuts, *via* a high-angle fault on its NW margin, a geologically complex terrane, part of which formed the eastern margin
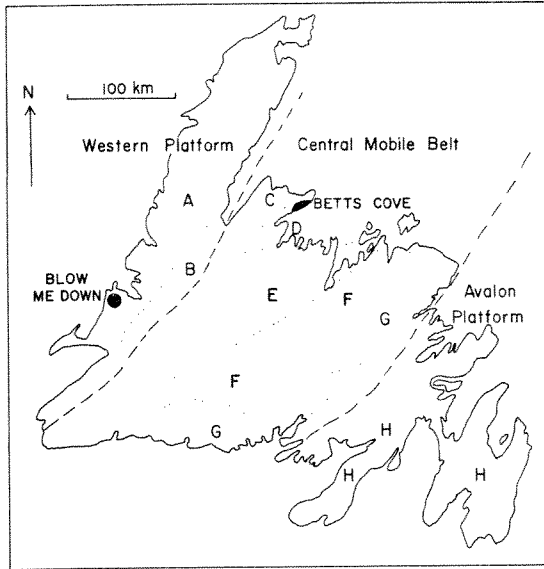
FIG. 1. Map of the island of Newfoundland showing the location of the Betts Cove ophiolite and Blow-Me-Down massif of the Bay of Islands ophiolite. A to G are tectono-stratigraphic zones after Williams et al. (1972).

of the Cambrian-Ordovician continent (Williams and Stevens, 1974). Unlike interpretations of the Bay of Islands ophiolite to the west, there is no direct evidence at the base of the Betts Cove ophiolite to indicate thrusting of the massif over great distances. Either late high-angle faulting has obliterated evidence of earlier low-angle thrusts or the ophiolite is essentially *in situ*. There is some indirect evidence to sup-

port the former interpretation in that amphibolite thought to be associated with a dynamothermal contact aureole has been reported from the base of an ultramafic complex a few kilometers to the west of the Betts Cove area (Riccio, personal communication, 1974; Church, 1977). The ophiolite is conformably overlain by pillow lavas, pyroclastic and argillaceous sediments of the upper Snooks Arm Group (Upadhyay et al., 1971; DeGrace et al., 1976; Jenner and Fryer, 1980).

The age of the Betts Cove ophiolite is not precisely constrained. It is certainly pre-Arenig (lower Ordovician) since it lies stratigraphically below sediments which contain Arenig graptolites (Snelgrove, 1931). Mattinson (1975) reported an age of 463 m.y., based on U-Pb relationships in zircons separated from a trondhjemite body supposedly associated with the gabbro member of the ophiolite. However, as pointed out by Church (1976) and acknowledged in Mattinson (1977), the dated trondhjemite may be younger than the ophiolite. Thus, 463 m.y.B.P. is a minimum age for the Betts Cove complex. The Bay of Islands ophiolite is dated at 508 m.y. (Mattinson, 1976; Jacobsen and Wasserburg, 1979). The Betts Cove ophiolite probably formed sometime between 508 and 463 m.y.B.P. assuming that it formed no earlier than the Bay of Islands complex.

INTERNAL STRATIGRAPHY AND PETROGRAPHY

The Betts Cove ophiolite comprises pillow lavas, sheeted dikes, gabbros and ultramafic cumulates (Fig. 2). In most places, the lavas and dikes are in fault contact. However, in rare instances, there is a stratigraphic contact exposed where dike abundances decrease rather abruptly from 100% in the sheeted member to around 30% in the overlying pillows. Nowhere is there a gradual change from 100% to 0% dikes. However, most of the dikes in the pillow lavas are near the base of the unit. Many of the dikes in the pillows are different from those in the sheeted complex in that they
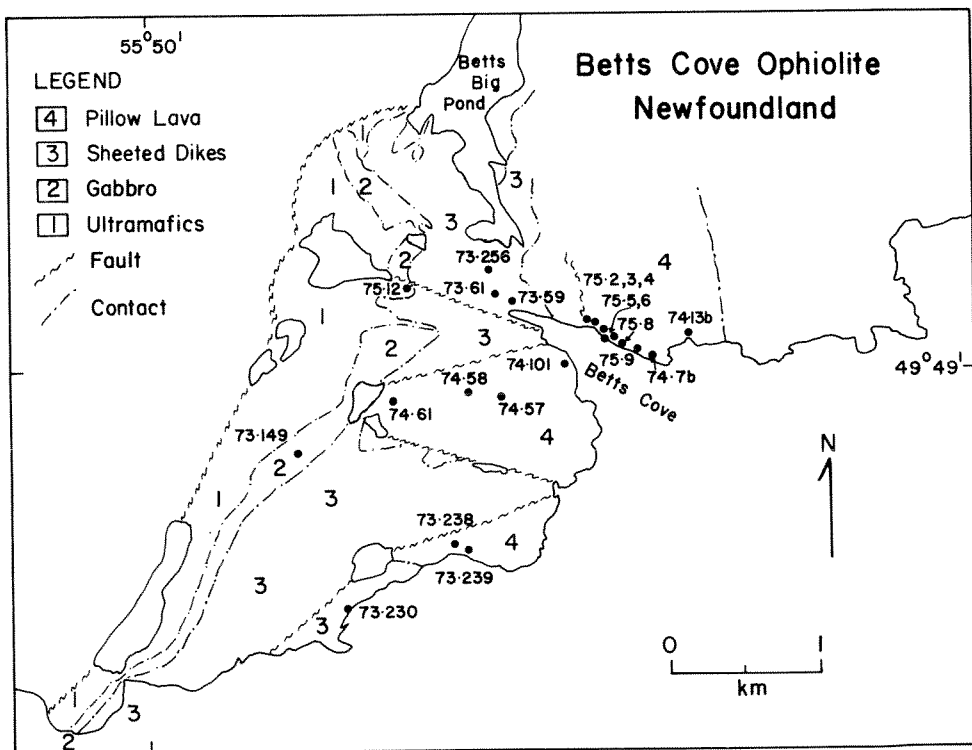

FIG. 2. Geologic map of the Betts Cove ophiolite. Sample locations are shown for the analyzed rocks in this paper.

Table 1. Analytical error in major and trace element analyses

	G-2 mv	G-2 rv	Major Elements AGV-1 mv	AGV-1 rv	JB-1 mv	JB-1 rv
SiO2	68.77	69.19	58.92	59.72	52.88	52.62
TiO2	0.50	0.50	1.07	1.05	1.35	1.34
Al2O3	15.28	15.35	17.03	17.22	14.81	14.62
Fe2O3	2.68	2.62	6.92	6.88	9.11	9.04
MgO	0.82	0.77	1.62	1.55	7.91	7.76
CaO	1.90	1.98	4.89	5.00	9.35	9.35
K2O	4.53	4.52	2.88	2.93	1.45	1.42
P2O5	0.11	0.14	0.41	0.50	0.24	0.26

Trace Elements

	BCR-1 mv	BCR-1 rv	JB-1 mv	JB-1 rv		BCR-1 RNAA	BCR-1 INAA	ID	(SD)
Ni	14	13	145	135	La	25.5	24.5	25.1	(.95)
Cr	10	16	375	400	Ce	53.0	56.0	54.1	(.50)
Zr	177	188	104	122	Nd	28.2	27.9	28.7	(.50)
Y	43	37	27	26	Sm	6.70	6.54	6.63	(.13)
Sr	352	330	454	440	Eu	2.03	2.05	1.97	(.05)
Rb	50	47	45	41	Tb	1.05	1.20	-	-
Ba	661	680	480	490	Yb	3.33	3.51	3.46	(.13)
					Lu	0.48	0.53	0.54	(.05)
					Sc	-	32.5	32.0	(1.0)

mv — measured value from this study
rv — recommended value (Abbey, 1977)
RNAA — Radiochemical neutron activation (Hickey and Frey, 1981)
INAA — Instrumental neutron activation values (this study)
ID — Average of isotope dilution analyses compiled from Arth and Hanson (1972), Gast et al. (1970), Hooker et al. (1975), Nakamura (1974), Philpotts and Schnetzler (1970), Shimizu (1975), Sun and Hanson (1976), Sun and Nesbitt (1978b)
SD — one standard deviation

have sinuous contacts whereas sheeted dikes have straight boundaries. The contact between the sheeted dikes and underlying gabbro is intrusive; the dikes intrude the gabbro. The gabbro member is thin (0.5 km) and consists of mainly isotropic gabbro with minor layered gabbro. Small bodies of trondhjemite, and dikes and layers of clinopyroxenite are associated with the gabbro. Some of the clinopyroxenite is pegmatitic; clinopyroxene grains reach 15 cm in length. The clinopyroxenite is in contact with the underlying ultramafic cumulates. There has been disagreement whether pyroxenite/ultramafic contact is intrusive or gradational (Upadhyay et al., 1971; Riccio, 1972; Church, 1977); apparently the contact varies along strike. In some places, e.g., near Kitty Pond, ultramafic layers grade upward through concordant clinopyroxenite into gabbro; elsewhere, as at Candlemas Head, clinopyroxenite 'dikes' intrude, and contain inclusions of, ultramafic cumulates.

The petrography of the mafic members in the ophiolite has been described elsewhere (Coish, 1977a,b; Coish and Church, 1979) so only salient features are mentioned here. All mafic rocks have greenschist facies mineralogy—albite, actinolite, chlorite, epidote, calcite, quartz. However, there are relict minerals, and igneous textures are preserved. These features, along with pseudomorph forms, can be used to interpret igneous relationships.

The pillow lava member is divided into three petrological sequences: 1) upper volcanics 2) intermediate volcanics 3) lower volcanics. The tripartite petrologic subdivision is apparently stratigraphic. Unquestionably, the upper lavas are at the highest stratigraphic level. However, the relationship between the lower and intermediate lavas is not simple. Just north of Betts Cove, along the shoreline (Fig. 2), the intermediate lavas indeed are stratigraphically between the upper and lower lavas whereas in other places they are interfingered with the lower lavas.

The upper lavas include dark pillows and occasional massive flows that contain abundant plagioclase and some clinopyroxene phenocrysts. Titanomagnetite (altered to hematite) is prominent. The groundmass is usually quenched with feathery plagioclase and pyroxene. The intermediate volcanics include reddish (hematitic) as well as green pillows. Phenocrysts include clinopyroxene, some plagioclase

and altered olivine(?). Quench textures occur in the groundmass. The lower volcanics are green pillows that have olivine and orthopyroxene(?) pseudomorphs, and clinopyroxene phenocrysts. Plagioclase occurs as a groundmass mineral. The lower lavas are also distinguished by extremely low contents of Fe-Ti oxides.

The sheeted dike complex is composed of aphyric diabase dikes, picritic dikes and perknite bodies (Coish and Church, 1979). Aphyric diabase dikes were primarily clinopyroxene-plagioclase rocks with plagioclase occurring interstitial to clinopyroxene grains. Picritic dikes are generally coarser grained than diabase dikes. They contain chlorite pseudomorphs after olivine and orthopyroxene. Relict clinopyroxene phenocrysts also occur. Aphyric diabase dikes are considered to be intrusive equivalents of both the intermediate and some lower lavas whereas the picritic dikes are represented in the lower lavas only. Perknite bodies are coarse-grained rocks concentrated near the base of the sheeted complex. They were probably originally composed of clinopyroxene, olivine and minor plagioclase. They are not chilled against other dikes but cut ultramafics in a few places (W. R. Church, pers. comm., 1980). Based on chilled margin relationships, the aphyric dikes are the youngest intrusions, the perknites the oldest, and the picritic dikes have an intermediate age.

GEOCHEMISTRY

Analytical methods and results

Major element and Ni, Cr, Zr, Y, Sr, Rb, and Ba concentrations were determined on a Phillips 1450 X-Ray Fluorescence machine following procedures of Norrish and Hutton (1969) as outlined in Coish and Church (1979). Rare earth elements and Sc were analyzed by neutron activation techniques at M.I.T. REE concentrations in samples BC-75-5, BC74-13b, BC74-7b, NH-71-1, NH-71-2, BC-75-8, BC-75-9, BC-75-7 and BC-75-6 were determined by instrumental neutron activation analysis (Frey et al., 1974); all other samples were analyzed by a radiochemical neutron activation procedure described in Hickey and Frey (1981). An indication of analytical error is shown by comparison

Table 2a: Major and Trace Elements in Betts Cove Ophiolite

	Upper Volcanics								Intermediate Volcanics						
Field #	75-5	75-7	75-6	75-9	74-13b	75-8	74-7b	73238	75-4	73239	75-3	75-2	74-57	7458a	7458b
Anal #	2	7	6	9	5298	8	5297	8	3	29	4	5	12	14	13
SiO_2	47.17	50.75	47.78	52.64	50.54	45.88	50.25	55.30	50.16	53.57	49.34	53.08	52.93	53.37	47.14
TiO_2	0.78	1.38	1.07	1.08	1.45	1.89	2.22	0.31	0.31	0.35	0.48	0.49	0.50	0.51	0.53
Al_2O_3	17.83	14.73	17.78	15.21	15.39	15.31	14.18	16.14	15.06	14.35	16.96	14.42	15.47	16.26	15.69
Fe_2O_3	3.42	4.33	5.94	3.92	11.31	2.20	1.99	1.74	3.64	1.82	2.96	4.02	1.36	1.14	2.77
FeO	4.46	8.61	6.09	7.34	-	8.65	8.13	3.60	4.46	5.50	5.54	3.73	5.49	5.31	5.98
MnO	0.00	0.14	0.00	0.00	0.00	0.00	0.00	0.12	0.16	0.14	0.16	0.17	0.16	0.16	0.17
MgO	10.24	6.03	7.17	5.71	6.20	6.60	6.51	6.56	8.84	8.04	10.28	8.30	9.08	7.91	10.62
CaO	2.29	6.31	3.38	3.29	9.54	12.67	11.51	8.38	8.22	9.07	3.95	4.94	6.80	8.11	12.05
Na_2O	0.97	4.74	3.90	6.48	4.89	2.85	3.67	4.63	2.13	4.27	4.16	4.74	4.94	3.92	1.88
K_2O	6.13	0.01	2.89	0.26	0.16	0.12	0.17	0.03	2.00	0.03	0.67	0.35	0.08	0.15	0.05
P_2O_5	0.07	0.13	0.04	0.04	0.12	0.08	0.24	0.03	0.01	0.03	0.02	0.04	0.03	0.03	0.04
LOI	6.67	2.40	4.29	3.22	2.53	3.40	2.54	2.52	4.34	2.75	4.80	4.78	2.91	3.22	4.08
Total	100.03	99.56	100.33	99.19	102.13	99.65	101.41	99.36	99.33	99.92	99.32	99.06	99.75	100.09	101.00
Trace Elements (ppm)															
Sc	49	41	0	39	42	43	45	31	34	35	41.5	42	-	34	37
Ni	136	-	41	25	89	86	70	128	106	128	100	110	156	224	244
Cr	415	73	75	30	208	245	211	301	376	323	404	423	339	436	432
Rb	57	-	30	-	2	3	1	2	19	-	10	4	1	2	2
Sr	107	-	103	86	210	222	165	81	223	41	125	77	169	202	146
Y	14	-	23	-	28	31	41	12	9	14	11	13	14	13	14
Zr	39	61	61	70	90	138	187	14	20	20	27	30	24	20	20
Ba	173	-	144	58	0	26	0	27	156	25	41	32	25	28	8
La	1.43	2.01	-	3.27	3.43	5.93	7.65	0.78	0.89	0.99	1.64	1.78	0.8	0.58	0.56
Ce	4.84	7.5	-	10.13	11.3	18.22	21.8	2.25	2.22	3.4	4.05	4.61	2.86	2.29	2.2
Nd	3.92	7.14	-	7.7	9.4	13.3	16.29	2.15	1.5	-	3.01	3.65	2.73	2.16	2.16
Sm	1.39	2.3	-	2.45	3.24	4.23	5.19	0.73	0.53	0.75	1.01	1.23	0.92	1.04	0.96
Eu	0.41	0.94	-	0.78	1.2	1.56	2.00	0.24	0.32	0.29	0.36	0.41	0.29	0.34	0.40
Tb	0.44	0.8	-	0.6	0.9	1.17	1.16	0.25	0.17	-	0.24	0.34	0.3	0.33	0.26
Yb	1.42	2.77	-	2.44	2.88	3.24	4.00	1.19	0.92	1.2	0.91	1.25	1.22	1.44	1.34
Lu	0.22	0.044	-	0.37	0.45	0.48	0.62	0.18	0.13	0.21	0.12	0.16	0.17	0.2	0.2

Table 2a Cont'd

Field #	Lower Volcanics								Perknites		Trondjhemites	
	73149	74101b	74101a	73256	7359	7461b	7461a	73230	7361	7512	NH711	NH712
Anal #	52	10	9	31	25	19	18	45	68	58	71-1	71-2
SiO2	53.29	60.39	42.92	50.87	55.79	50.99	46.66	53.54	52.07	52.13	75.75	80.44
TiO2	0.11	0.12	0.14	0.14	0.16	0.17	0.19	0.27	-	0.08	0.09	0.02
Al2O3	11.38	11.68	11.04	12.64	12.85	11.29	12.66	15.44	5.30	10.97	13.50	10.95
Fe2O3	2.91	1.44	3.34	8.56	0.73	1.07	3.07	3.13	8.99	1.35	1.07	0.44
FeO	4.96	4.16	8.06	-	6.62	6.60	5.77	4.23	-	7.45	-	-
MnO	0.17	0.15	0.20	-	0.14	0.14	0.16	0.13	0.21	0.17	0.02	0.03
MgO	11.82	7.60	15.57	13.40	9.31	12.89	14.36	6.53	19.35	13.71	1.32	0.45
CaO	8.74	6.71	11.67	6.11	5.27	11.86	13.17	9.01	10.64	6.12	5.54	3.73
Na2O	2.52	5.03	0.52	1.63	3.16	0.72	0.17	4.14	-	3.01	2.32	2.61
K2O	0.12	0.01	0.01	0.36	0.04	0.08	0.04	0.04	-	0.16	0.17	0.09
P2O5	0.02	-	0.03	0.01	-	0.01	0.01	0.03	-	0.01	-	-
LOI	3.06	2.19	6.09	6.26	5.37	3.27	3.97	2.23	-	3.88	-	-
Total	99.10	99.48	99.59	99.98	99.44	99.09	100.23	98.72	96.56	99.04	99.78	98.76
Trace Elements (ppm)												
Sc	46	-	-	41	41	36	34	36	44	45	12	7.5
Ni	155	268	231	238	95	335	387	61	-	279	-	-
Cr	778	756	730	670	274	787	918	107	-	867	57	52
Rb	4	-	1	4	5	2	5	-	1	1	-	-
Sr	196	60	74	71	102	237	160	52	2	121	-	-
Y	8	8	5	9	8	1	1	11	2	5	-	-
Zr	8	16	14	13	15	10	13	21	5	5	-	-
Ba	66	28	12	30	16	11	5	11	2	85	-	-
La	0.48	0.79	0.92	0.96	0.66	1.59	1.73	1.25	0.29	0.58	3.4	7.07
Ce	1.2	2.16	2.48	2.04	1.84	3.88	4.2	2.98	1.11	-	7.4	13.1
Nd	0.32	1.1	1.29	0.99	0.98	1.65	1.69	1.68	0.25	-	1.86	3.49
Sm	0.15	0.23	0.37	0.28	0.23	0.44	0.48	0.48	0.05	0.13	0.45	0.62
Eu	0.063	0.074	0.15	0.071	0.047	0.236	0.274	0.21	0.024	0.11	0.09	0.112
Tb	0.069	0.073	0.12	0.117	0.101	0.149	0.149	0.179	0.014	-	-	-
Yb	0.68	0.58	0.95	0.97	0.77	0.81	0.77	1.18	0.21	0.66	1.18	1.47
Lu	0.2	0.092	0.21	0.21	0.124	0.145	0.19	0.18	0.035	0.15	0.195	0.225

of values obtained in this study with published values of standard rock samples (Table 1). $^{143}Nd/^{144}Nd$ ratios and Sm/Nd ratios were determined at M.I.T. using mass spectrometric techniques described in Zindler et. al. (1979).

Table 2a presents complete analyses of only those samples for which REE concentrations have been determined. A larger array of samples (95) has been analyzed for major and trace elements, excluding the rare earths (Coish, 1977a). Figures 4 and 5 include the complete data array. Ratios of $^{143}Nd/^{144}Nd$ for four samples are in Table 2b.

Effects of alteration on bulk rock chemistry

In order to interpret the igneous petrogenesis of the metamorphosed basic rocks in the Betts Cove ophiolite, the effects of metamorphism and alteration on bulk rock compositions must be evaluated. It is well established that some elements can be mobilized during low-temperature alteration of mafic rocks (Bischoff and Dickson, 1975; Hajash, 1975; Coish, 1977b; Mottl and Holland, 1978). The difficult question is which elements move under what conditions. Theoretical and experimental studies have shown that Ti, Zr, Y, Al2O3, Ni and the REEs are relatively immobile under greenschist facies metamorphic conditions (Pearce and Cann, 1973; Humphris and Thompson, 1978; Menzies and Seyfried, 1979). Hellman and Henderson (1977) and Hellman et al. (1979) have criticized some of the methods used in the aforementioned studies and they suggested that under some metamorphic conditions, Ti and the REEs are mobile. They caution that the REEs and Ti should not be indiscriminantly used to describe the igneous evolution of low-grade metamorphic rocks. It seems that the best approach is to show, for individual areas, that elements are indeed immobile.

There is no doubt that in the Betts Cove volcanics, the abundances of major elements, such as Ca and Na, and trace elements, such as Sr and Ba have been strongly affected by metamorphism (Coish, 1977b). Examples of such mobility are samples BC74-58a and b, BC74-101a and b and BC74-61a and b (Table 2) which are rim and core samples from three pillows.

In a study of metamorphic effects at Betts Cove, Coish (1977b) suggested that Ti, Zr, Y, Ni and Cr contents reflect igneous values. To test the suggestion that Ti is immobile, the Ti content of relict clinopyroxene was determined. Table 3 presents data on Ti in clinopyroxene, the predicted Ti content in the liquid using a K_D (Cpx/Liq) = 0.3 (Pearce and Norry, 1979), and the Ti content of the whole rocks. Although the correlation is not perfect, there is no doubt that the Ti content of the rock varies with the Ti content of constituent clinopyroxene. Unless Ti can be mobilized in clinopyroxene without visible alteration, the Ti content of clinopyroxene reflects the Ti

Table 2b. Nd isotopic data for Betts Cove lavas.

	1	2	3	4
Upper lava:				
BC-74-13b	0.513024 ±18	0.2054	0.512378	+7.17
Inter. lava:				
BC-74-57	0.513140 ±17	0.2475	0.512362	+6.86
Lower lavas:				
BC-74-101a	0.512791 ±16	0.1749	0.512241	+4.49
BC-74-101b	0.512728 ±19	0.1624	0.512217	+4.06

NOTES: 1) Col 1 - $(^{143}Nd/^{144}Nd)$; Col 2 - $(^{147}Sm/^{144}Nd)$; Col 3 - $(^{143}Nd/^{144}Nd)o$; Col 4 - ϵ_o^{CHUR}.

2) $^{143}Nd/^{144}Nd$ normalized to $^{146}Nd/^{144}Nd$ = 0.7219 and BCR-1 $^{143}Nd/^{144}Nd$ = 0.51263.

3) Initial ratios calculated assuming an age of 480 m.y..

4) ϵ_o^{CHUR} calculated assuming CHUR present day $^{143}Nd/^{144}Nd$ is 0.51262 and $^{147}Sm/^{144}Nd$ = 0.1936.

Table 3. Titanium contents in clinopyroxene and equilibrium
liquid compared to observed whole rock values (a)

Sample	1	2	3
BC73-141	.02	.07	.08
BC73-61	.02	.07	.02
BC73-57	.04	.15	.19
BC74-101	.03	.09	.12
BC74-90	.22	.73	.65
BC74-13b	.90	3.00	2.22

1 - TiO2 in clinopyroxene
2 - TiO2 in equilibrium liquid (K_D = 0.3, Pearce & Norry, 1979)
3 - TiO2 content in whole rock

(a) Possible reasons for discrepancies between columns 2 and 3
    are: 1) inappropriate choice of partition coefficients for
    the wide range of TiO2, 2) the difficulty of measuring Ti
    content by microprobe below levels of 0.05 wt.% and 3) cpx
    may not be a liquidus phase in BC74-13b, and thus the
    predicted Ti content is that of a liquid more fractionated
    than the whole rock.

content in the magma. Thus, the correlation seen in Table 3 implies that gross differences in bulk rock TiO$_2$ at Betts Cove are due to igneous processes.

The concentrations of the total REEs (La + Sm + Yb) vary sympathetically with TiO$_2$ content (Fig. 3). Because TiO$_2$ contents in these rocks are due to igneous processes, it seems likely that the gross differences in REE contents of the Betts Cove mafic rocks also reflect igneous processes. Moreover, in pillow rim and core samples 74-61a and b and samples 74-58a and b (Fig. 9 and Table 2), the chondrite-normalized REE patterns are very similar. REE abundances in pillow rim-core samples (BC74-101a,b) differ by factors of 1.1 to 2.2 (Table 2); nevertheless, both samples are characterized by convex-downwards, chondrite-normalized REE patterns (Fig. 9). The differences in absolute abundances could be related either to the concentration of varioles in the center of the pillow (Upadhyay, 1978; 1981) or to differences in alteration processes between rim and core (Coish, 1977b).

Major and trace elements (excluding REEs)

The three-fold division of the mafic rocks into lower, intermediate and upper groups is best defined by TiO$_2$ content (e.g., Fig. 5 of Coish and Church, 1979 and Figs. 4 and 5 of this paper). Note that both lava and dike rocks fall into the lower and intermediate groups but only lavas form the upper group.

Within the lower, low TiO$_2$, group, samples of picritic dikes and perknites are low in Al$_2$O$_3$ content (<10%) and have high Ni abundances (>280 ppm). These samples contain 10 to 40% large chlorite pseudomorphs after olivine and they are in part accumulative in origin; thus their bulk rock analyses do not represent liquid compositions. The compositional cutoff between rocks that have liquid compositions and those that are accumulative is somewhat arbitrary, although some insight into the problem can be gained by examining dikes in which phenocrysts are observed to be concentrated towards the center. Analyses of three samples collected at
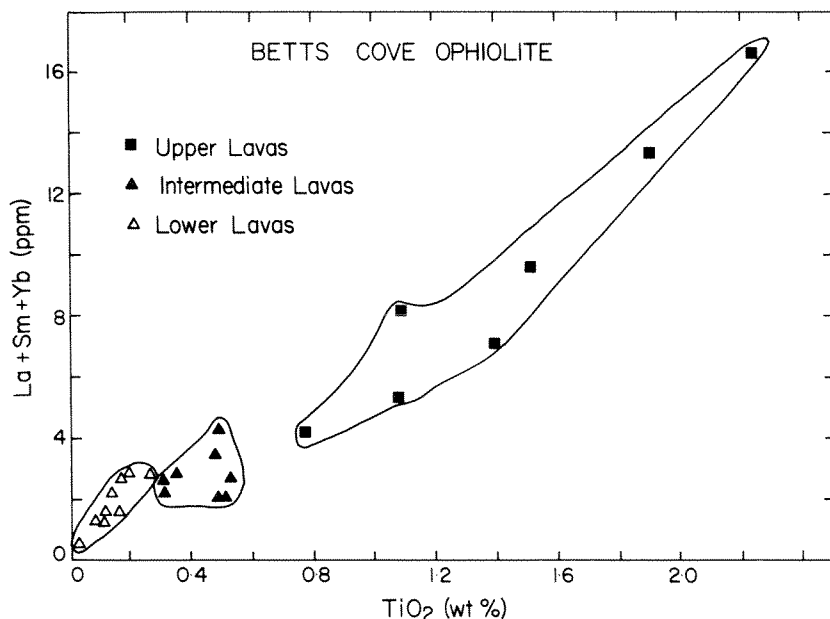

FIG. 3. Variation of TiO$_2$ content with total rare earth element content (La + Sm + Yb) in basalts from Betts Cove ophiolite.
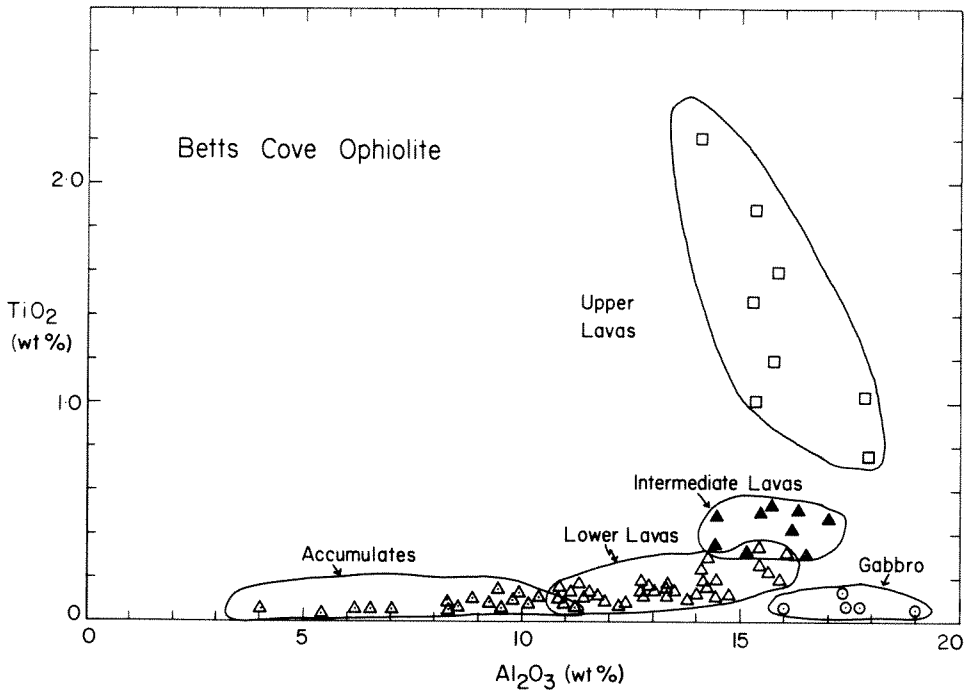
FIG. 4. Variation of TiO_2 vs Al_2O_3 for mafic rocks in the ophiolite.

equal intervals from the margin to the center of one such dike are in Table 4. The increase in Ni and MgO towards the center can be accounted for by mechanical accumulation of olivine and clinopyroxene phenocrysts. If it is assumed that the three samples are representative of the whole dike, then an average composition (on a volatile-free basis) gives *maximum* contents of MgO (17.4 wt.%), Ni (363 ppm) and Cr

(1252 ppm), and *minimum* values for Al_2O_3 (10.6 wt.%) and TiO_2 (.07 wt.%) in the *liquid*. (These values are maxima and minima because olivine and clinopyroxene phenocrysts may have undergone mechanical accumulation in the liquid prior to the intrusion of the liquid as a dike. The above calculations deal only with accumulation that took place during dike emplacement. Any effect of an earlier accumulation
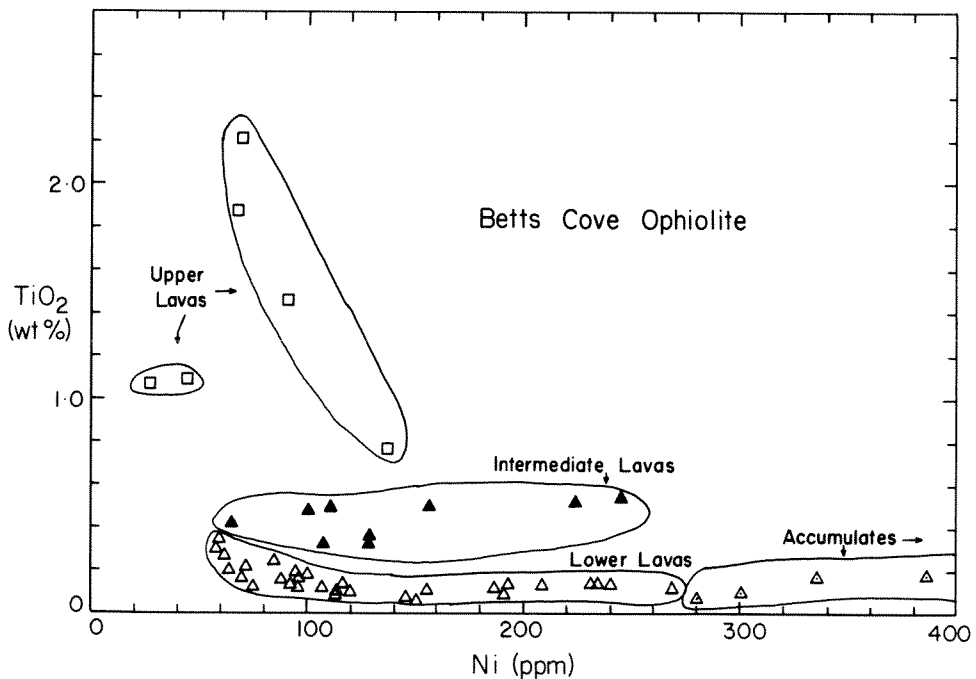
FIG. 5. Variation of TiO_2 vs Ni contents.

Table 4. Chemical Variations within a Sheeted Dike

	Distance from margin		
	7 cm	13 cm	18 cm
SiO2	55.44	47.11	49.02
TiO2	0.08	0.06	0.06
Al2O3	12.30	9.48	8.47
Fe2O3	0.78	1.68	1.57
FeO	5.97	8.42	7.54
MnO	0.16	0.19	0.18
MgO	11.60	19.31	18.60
CaO	4.88	6.47	6.88
Na2O	4.35	0.87	0.95
K2O	nd	nd	nd
P2O5	0.01	0.01	0.02
LOI	3.34	5.96	4.97
Total	98.91	99.55	98.55
FeO^t/MgO	.57	.51	.48
Trace Elements			
Ni	112	397	526
Cr	521	1452	1589
Zr	7	–	4
Y	7	4	7
Sr	90	27	26
Rb	nd	nd	1
Ba	15	19	17

is masked by the later process). The cutoff between liquids and accumulative rocks is chosen at about 280 ppm Ni (Fig. 5) mainly because rocks with higher Ni concentrations tend to contain large phenocrysts. The cutoff value is consistent with calculations for the picritic dike that showed flowage differentiation. Accumulative rocks will not be used to discuss magmatic evolution of the Betts Cove ophiolite.

The three groups of basic rocks have a similar wide range in Ni content but the Ni-TiO_2 trends (Fig. 5) distinctly define the three groups. The wide range in Ni content probably reflects fractionation of mafic phases, but it is evident that such fractionation can not account for the TiO_2 abundance variations between the groups.

The upper group of lavas are also distinguished by a trend of decreasing Al_2O_3 with increasing TiO_2 whereas within and between the lower and intermediate groups there is a positive correlation between TiO_2 and Al_2O_3 contents (Fig. 4). The distinctive trend for the upper lavas reflects the importance of plagioclase as a fractionating phase which is consistent with the high abundance of plagioclase in these rocks.

Rare earth elements: Upper volcanic group (>0.75 wt.% TiO_2)

The upper group of lavas has the highest total abundances of the rare earths (4–25× chondrites for LREE and 7–18× for HREE) of the three groups (Fig. 3). The chondrite-normalized patterns are depleted in LREE relative to the middle and heavy rare earths. Also, there is a decrease in abundance from Gd to Lu giving rise to a convex-upwards pattern similar to that seen in many mid-ocean ridge basalts (Frey *et al.*, 1968; Staudigel, 1979; Sun *et al.*, 1979) and basalts from other ophiolites (*e.g.*, Troodos, Kay and Senechal, 1976). With the exception of sample BC-75-9, the patterns are sub-parallel (Fig. 6). The (La/Sm)$_N$ ratio is nearly constant whereas the (La/

Yb)$_N$ ratio and total REEs increase with increasing TiO_2 content (Figs. 3 and 7).

Some of the samples in the upper volcanic group have negative Eu anomalies that may have been produced by alteration processes. Two observations support this idea: 1) the most primitive sample (BC-75-5) has the largest negative Eu anomaly—a situation not in accord with an igneous origin for the anomaly, and 2) whenever a negative Eu anomaly occurs, the rock usually has an anomalously low CaO content (*e.g.*, BC-75-5 and 75-9, Table 2). If the low CaO values were inherited from a melt, then the plagioclase content of the rock should be low. However, plagioclase phenocrysts are abundant in all the upper lavas. To account for the combination of low CaO and high plagioclase content, it is speculated that CaO and Eu were removed from the rock during albitization of the plagioclase phenocrysts. It is interesting that all samples have been albitized but only some have negative Eu anomalies. This may be in part related to the distance of transport of Eu after it has been mobilized. Although CaO and Eu may be always mobilized during albitization, if the CaO and Eu are reprecipitated by the formation of epidote and calcite on or around the plagioclase grain, then the whole rock CaO and Eu are unchanged. On the other hand, if precipitation does not occur on the scale of a hand specimen, then redistribution of CaO and Eu results. Eu anomalies have been attributed to alteration in other basic rocks (*e.g.*, Sun and Nesbitt, 1978a).

Rare earth elements: Intermediate volcanic group (0.30 to 0.50 wt.% TiO_2)

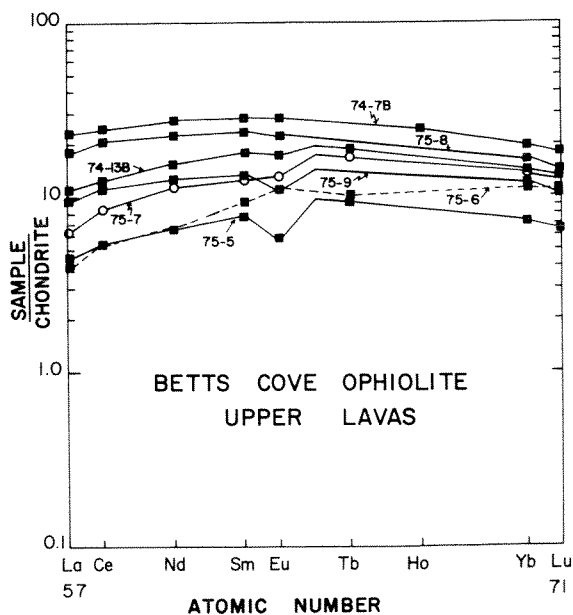Most of the volcanic rocks with intermediate TiO_2 contents have chondrite-normalized REE patterns

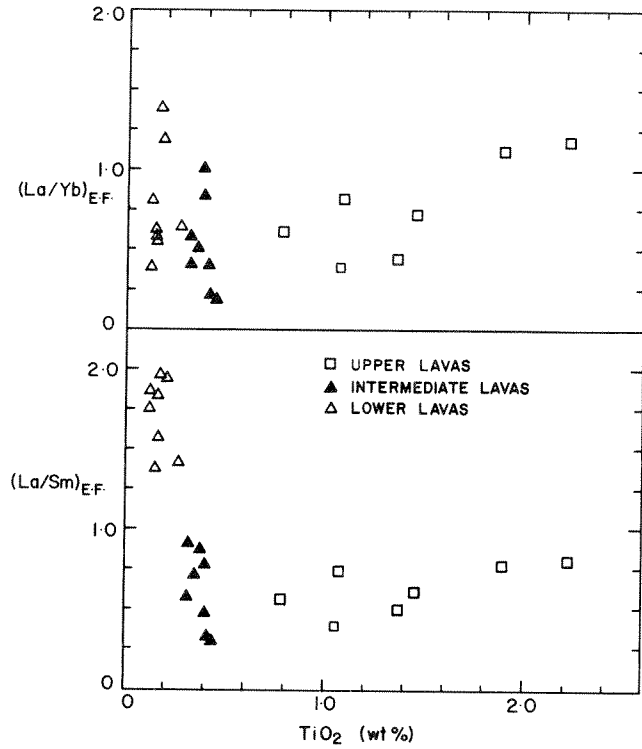FIG. 6. Rare earth element patterns in upper lavas.

FIG. 7. Variation of $(La/Yb)_N$ and $(La/Sm)_N$ ratios with TiO_2 content for three lava groups. N denotes chondrite normalized values.

with relative depletion of LREE (Fig. 8); $(La/Yb)_N$ ratios vary between 0.25 and 1.09. The normalized middle to heavy REE are relatively flat without the maxima at intermediate REE seen in the upper lavas. Three samples (BC-75-2,3,4) have relatively high $(La/$ $Sm)_N$ ratios (0.79 to 0.92) compared to other samples (Fig. 8). It is suspected that the LREEs of these samples may be enriched due to alteration. Among the intermediate lavas, these samples have the highest contents of K_2O, lowest CaO contents and the largest
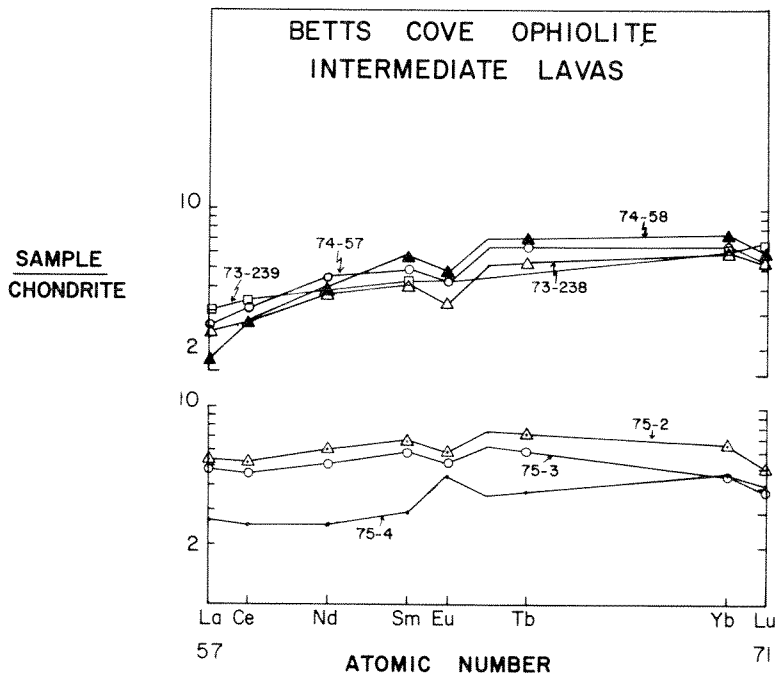
FIG. 8. Rare earth element patterns in intermediate lavas.

losses on ignition (Table 2), and they are closely associated with a zone of high oxidation (red pillows). If these flat REE patterns are indeed produced by alteration (there is no way to *absolutely* disprove an igneous origin), then the characteristic igneous pattern of the intermediate lavas has low REE abundances (<10× chondrites) with relative LREE-depletion compared to chondrites. Similar patterns are found in the most depleted lavas from the Troodos ophiolite (Kay and Senechal, 1976; Smewing and Potts, 1976) and in some very primitive ocean floor basalts (Frey *et al.*, 1974).

Rare earth elements: Lower volcanic group (<0.30 wt. % TiO_2)

The REE abundances in rocks with the lowest TiO_2 contents are low (2–7× chondrites) and the chondrite-normalized patterns have unusual convex-downwards distributions, *i.e.*, the light and heavy rare earths have greater abundances than the middle rare earths (Fig. 9). Most of the patterns have $(La/Yb)_N$ < 1.0; but sample BC-74-61 is an exception as it exhibits a mild enrichment in La relative to Yb. Negative Eu anomalies occur in BC73-59, 73-256, and 74-101b; again, these samples have anomalously low CaO values (see Table 2), which may be due to intense albitization. There is a positive correlation between TiO_2 and total REE content within the lower volcanic group (Fig. 3).

Basaltic rocks with similarly low TiO_2 (<0.3%) contents are rare. Laurent and Hebert (1977) report

analyses of very low TiO_2 basalts from lower lavas in the Thetford Mines ophiolite which may be an extension of the Betts Cove ophiolite belt south into the Quebec Appalachians (Church, 1977). Other very low TiO_2 volcanic rocks are found in western Mongolian ophiolites (Zonenshain and Kuzmin, 1978), Adamsfield ultramafic complex, Tasmania (Varne and Brown, 1978), Arakapas fault zone, Cyprus (Simonian and Gass, 1978) and the Marianas Trench-Bonin Islands region of the western Pacific (Dietrich *et al.*, 1978; Sun and Nesbitt, 1978a; Hickey and Frey, 1982). Low TiO_2 volcanics from the Mariana Trench-Bonin Island region, which are of Tertiary age, are high-MgO andesites or "boninites", whereas the older ophiolitic, low-TiO_2 volcanics are basaltic.

Rare earth element abundance data are available for only the Arakapas fault zone and the Marianas Trench-Bonin Islands region. Basalts from the Arakapas fault zone have very low concentrations of the REEs, similar to the abundances in lavas at Betts Cove; however, they do not exhibit the convex-downwards normalized REE patterns but have a straight LREE-depleted profile (Simonian and Gass, 1978). The very low TiO_2 volcanics from the Marianas Trench-Bonin Islands also have low REE abundances and have convex-downwards normalized REE patterns similar to those in the Betts Cove lower lavas. This type of REE pattern is in fact common in boninites, which are also characterized by lower TiO_2 contents (0.1–0.2% TiO_2) than most other island-arc volcanics (Hickey and Frey, 1982).
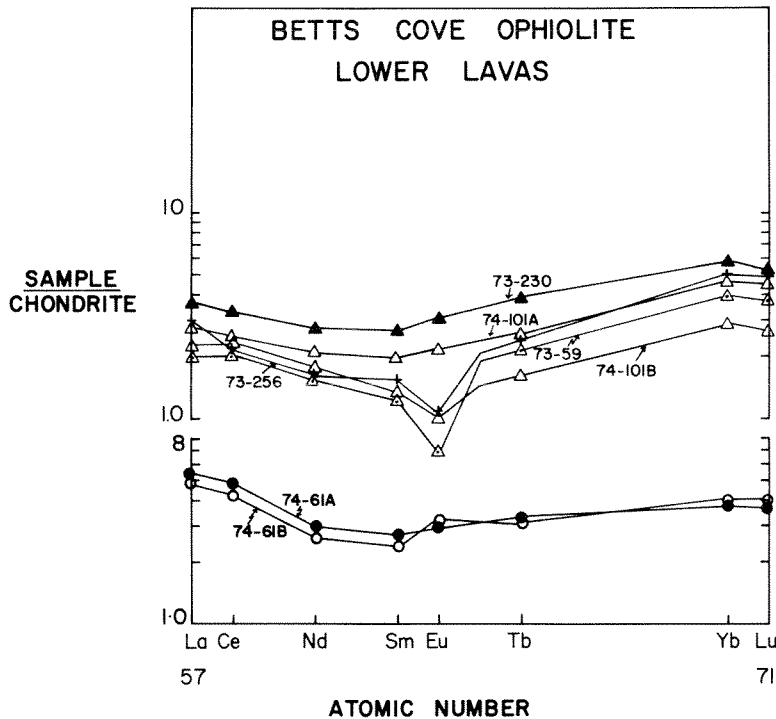

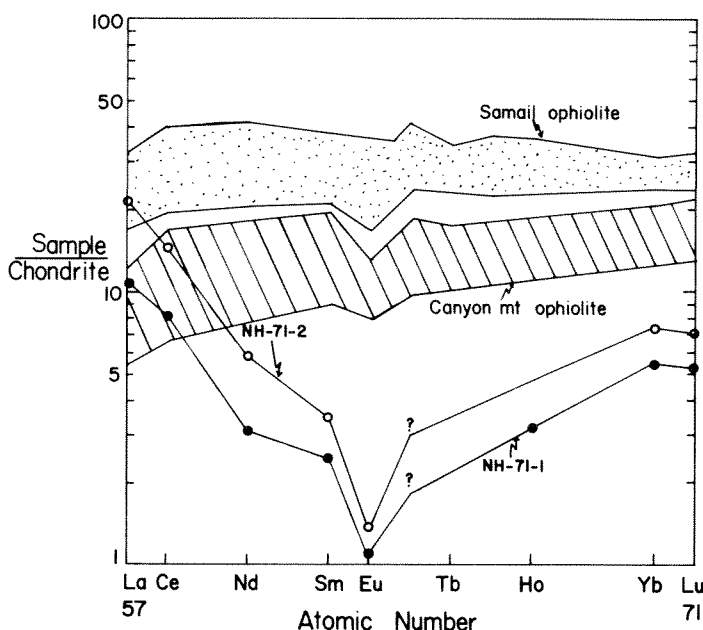FIG. 9. Rare earth element patterns in lower lavas.

FIG. 10. Rare earth element patterns in two trondhjemite samples at Betts Cove compared to REE contents in plagiogranites from other ophiolites. Fields for Canyon Mountain and Samail ophiolites from Fig. 4, Gerlach et al., 1981.

Nd Isotopes

A pillow rim and core pair (BC74-101a and b) from the lower lava unit had similar (within analytical precision) $^{143}Nd/^{144}Nd$ ratios at 480 mybp (Table 2b). These rim and core samples have very different major element compositions (Table 2a) which establishes that the significant changes in major element contents resulting from metamorphism, alteration or different variole concentrations were *not* accompanied by changes in $^{143}Nd/^{144}Nd$ ratio.

Individual samples from the intermediate and upper units have significantly higher $^{143}Nd/^{144}Nd$ ratios than in the pillow from the lower unit (Table 2b). These $^{143}Nd/^{144}Nd$ values for the intermediate and upper unit are within the range reported for the Bay of Islands ophiolite (Jacobsen and Wasserburg, 1979) and are consistent with a source similar to that of modern MORB.

Trondhjemites

The Betts Cove ophiolite has minor amounts of trondhjemite that occur as dikes and small isolated bodies in the gabbro member. Table 2a shows the major, trace and rare earth element compositions of two samples. Major element concentrations are similar to those in rocks termed 'oceanic plagiogranite' (Coleman and Peterman, 1975; Dixon and Rutherford, 1979; Ishizaki and Yanagi, 1975; Kay and Senechal, 1976; Gerlach et al., 1981). Such plagiogranites are interpreted as: 1) residual liquids of extensive fractional crystallization of oceanic tholeiite magma (Coleman and Peterman, 1975), or 2) as immiscible liquids formed from late-stage, Fe-rich basaltic magmas (Dixon and Rutherford, 1979), or 3) partial melts of tholeiitic basalts under hydrous conditions (Gerlach et al., 1981).

The chondrite-normalized REE patterns of trondhjemitic rocks from Betts Cove are unusual and significantly different from the REE patterns determined for plagiogranites from other ophiolites (Fig. 10). The V-shaped patterns of the Betts Cove trondhjemites are defined by La concentrations at about 10–20× chondrites, Sm about 1–2× chondrites, and Yb about 5–7× chondrites (Fig. 10). These REE abundances are equal to or less than those in the upper lavas; consequently, it is unlikely that these trondhjemites are genetically related to the upper unit. The very low REE contents and accentuated convex-downwards chondrite-normalized pattern of the trondhjemites are most similar to the REE abundances in the lower unit of Betts Cove volcanics; thus, the trondhjemites may be genetically related to the lower lavas. The marked negative Eu anomaly in the trondhjemites (Fig. 10) probably reflects an important petrogenetic role for plagioclase.

DISCUSSION

Origin of compositional variations within the volcanic groups

The metamorphosed and altered nature of the Betts Cove volcanics precludes detailed evaluation of the compositional variations within the intermediate and lower volcanic units. In these units, despite a rough correlation between total rare earths and TiO2 in the lower lavas (Fig. 3), there is no systematic variation between La/Sm and La/Yb, and TiO2 contents
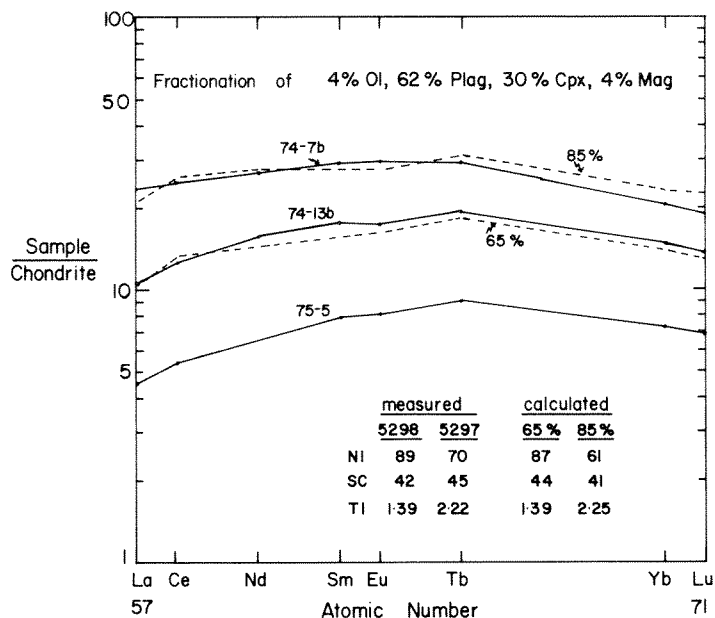
FIG. 11. Quantitative model for fractional crystallization within the upper lava group. Solid lines represent rare earth concentrations in upper lavas 75-5, 74-13b and 74-7b. Dashed lines are calculated values after 65% and 85% fractional crystallization of a liquid with the composition of 75-5. The fractionated phases are 4% olivine, 62% plagioclase, 30% clinopyroxene and 4% magnetite. Rayleigh fractionation equations and K_D values compiled by Frey *et al.* (1978) (set 1 for olivine and plagioclase and set 2 for clinopyroxene) were used in the calculations.

(Fig. 7). Moreover, there are no systematic correlations between La and Zr contents, or between these elements and a compatible trace element, such as Ni. It is likely that the lack of coherence results from abundance changes caused by metamorphism and alteration.

Systematic abundance trends are more apparent in the upper lavas; *e.g.,* Zr and REE contents correlate positively with TiO_2 contents (Fig. 3, and Coish and Church, 1979, Fig. 6) whereas Ni, Cr and Al_2O_3 abundances are negatively correlated with TiO_2 contents (Figs. 4, 5 and Table 2). These trends are consistent with fractionation of the observed phases, plagioclase and clinopyroxene.

In detail, among the upper lavas, La/Sm remains nearly constant whereas La/Yb ratio increases with increasing TiO_2 and total rare earth content (Fig. 7). Thus, there is an increase in the negative slope of the middle to heavy rare earth pattern with increasing fractionation. The change in La/Yb ratio is very important in modelling the fractionation process. If sample BC-75-5 is chosen as the most primitive sample of the upper lavas, and assuming that the Eu anomaly in BC-75-5 is not of igneous origin, then samples BC74-13b and BC74-7b can be derived from BC-75-5 by 65% and 85% crystallization, respectively (Fig. 11). Plagioclase, olivine, clinopyroxene and magnetite are fractionated. High clinopyroxene/melt REE partition coefficients (set 2 of Frey *et al.*, 1978) are used in order to account for La/Yb ratios in derivative samples 74-13b and 74-7b that are higher than those in the parental sample 75-5. This scheme

is not unique; an alternative model could involve small amounts of garnet along with plagioclase, clinopyroxene and magnetite. In this case, low cpx/melt REE partition coefficients could be used. However, it is unlikely that garnet would be a fractionation phase since the chamber was probably at a shallow level.

Origin of compositional differences between volcanic groups
Crystal fractionation:

Because the lower lavas have the lowest concentrations of REE, Ti and other incompatible elements, they might be considered as parental to lavas of the overlying units. However, the large differences in TiO_2 content and chondrite-normalized REE patterns between the lower and overlying units are not likely to result from crystal fractionation. This conclusion is confirmed by the distinctly lower [143]Nd/[144]Nd ratios in the lower unit.

Our limited Nd isotopic data indicate that the intermediate and upper units might be genetically related, and superficially, it appears that the differences between the REE patterns of the intermediate and upper lavas could result from crystal fractionation. However, attempts to *quantitatively* model fractionation of intermediate lavas to give upper lava compositions have failed. Various combinations of olivine, plagioclase, clinopyroxene and garnet were used in the calculations. The principal difficulties in the modelling are: 1) The change in TiO_2 content from 0.50 to 2.0 wt.% requires at least 80% crystallization;

Ni and MgO contents in the upper lavas are too high for this amount of crystallization to have occurred (Table 2). 2) Some intermediate lavas are very LREE-depleted [$(La/Yb)_N$ = 0.5] whereas the most fractionated upper lavas have $(La/Yb)_N$ around 1.0. Large amounts of clinopyroxene would have to fractionate to cause the increase in La/Yb ratio. However, higher Sc values in upper lavas relative to those in intermediate lavas would seem to preclude clinopyroxene fractionation. A similar argument can be made for the consistent differences in Sm/Yb (Fig. 7).

Complex melting of a homogeneous source:

Dynamic melting was proposed by Langmuir *et al.* (1977) as a realistic model that creates melts with a wide range of REE contents from a homogeneous source. An essential feature of this model is that later melts can be derived from the residues of earlier derived melts. In the simplest form of dynamic melting the derived melts are isotopically homogeneous, and the older lavas are more enriched in incompatible elements than the younger lavas. Neither of these criteria is fulfilled by the Betts Cove units where the stratigraphically oldest lavas are isotopically distinct and are the most depleted in incompatible elements.

More complicated versions of the dynamic melting model have been proposed (Wood, 1979, 1981). Inherent in the dynamic melting model is the large array of variables whose adjustment can produce a wide variety of trace element abundances including the variety of REE patterns (Figs. 6 to 9) in the Betts Cove volcanics. Reasonable circumstantial proof of the model requires independent evidence supporting the model; *e.g.,* Wood (1979) utilized the correlation of compositional changes with time and volume to argue for dynamic melting as the process creating lavas with diverse REE patterns on the Faeroe Islands, the Isle of Skye, Troodos ophiolite and various regions in Iceland. However, as discussed by Zindler *et al.* (1979) recent lavas from the Reykjanes Peninsula of Iceland are heterogeneous in $^{143}Nd/^{144}Nd$ ratios. Consequently, in order to explain these lavas by a dynamic melting scheme, Wood (1981) postulated dynamic melting of a heterogeneous mantle; specifically a veined mantle equilibrated in Sr isotopes but not equilibrated in Nd isotopes. The Betts Cove lavas are also heterogeneous in $^{143}Nd/^{144}Nd$ ratio and in addition the age relations of the lavas are unambiguously *not in accord* with the sequence of compositions expected from dynamic melting (Wood, 1979, 1981). It could be argued that the sequence of melting within the mantle is not always reflected in the eruption sequence as stated by Langmuir *et al.* (1977, p. 151). Since several *ad hoc* assumptions are required to explain Betts Cove volcanics by a dynamic melting process, we prefer to emphasize that the mantle source of the lower Betts Cove volcanics was compositionally different from the sources of the overlying lava units.

Different sources:

Several lines of evidence indicate that compositional differences between the three lava groups at Betts Cove are best explained by different mantle sources. The lower $^{143}Nd/^{144}Nd$ ratios of the lower lavas in addition to their relative LREE enrichment, $(La/Sm)_N$ > 1, indicates that they could not have been derived from the same source as the intermediate and upper lavas which have higher $^{143}Nd/^{144}Nd$ and $(La/Sm)_N$ < 1. The intermediate and upper lavas have similar $^{143}Nd/^{144}Nd$ initial ratios which could indicate a common source. However, the difficulty in deriving these two groups from a single source by simple partial melting or fractional crystallization processes, and reversal of sequence (*i.e.,* upper lavas less depleted in Ti and REE than intermediate lavas) from that expected in a dynamic melting process, suggests that the intermediate and upper Betts Cove lavas also originate from compositionally distinct sources.

The extremely low TiO_2, Zr, Y, and HREE contents of the lower lavas require a severely-depleted peridotite source. As noted by Coish and Church (1979, Table 5) derivation of these lavas by 5–20% partial melting requires a source depleted in Ti and HREE by factors of 2 to 20 compared to the source of normal MORB. This highly depleted source would be expected to have very low $(La/Sm)_N$ and $(La/Yb)_N$ ratios; however, the lower lavas are enriched in LREE relative to middle REE, *i.e.,* $(La/Sm)_N$ > 1, thereby, forming the convex-downwards chondrite normalized REE pattern. In order to reconcile extreme Ti-depletion with relative LREE enrichment, it can be argued that the severely depleted mantle was invaded by a small amount of LREE-enriched, but Ti-depleted material. This interpretation was proposed by Sun and Nesbitt (1978a) and Hickey and Frey (1979; 1982) to explain similar convex downward REE patterns in boninites from western Pacific island arcs. Because boninites are restricted to island arcs, it was suggested that the LREE metasomatism was related to subduction. Evidence that convex downward, chondrite-normalized patterns occur in depleted mantle is the existence of such REE patterns in refractory mantle peridotites from several tectonic regimes (Frey, 1982).

The intermediate lavas were probably derived from a depleted mantle that was not affected by LREE metasomatism. The REE patterns of these lavas can be obtained by partial melting of an ultramafic source with REE concentrations similar to those of lherzolite sample V-14 collected near the base of the tectonite unit in the Bay of Islands ophiolite (Suen *et al.,* 1979). Twenty percent continuous melting (Langmuir *et al.,* 1977) or 30% batch melting of V-14 produces a liquid with rare earth element contents similar to sample BC74-58, the most depleted intermediate lava. Under the above melting conditions, the source should have TiO_2 values of

less than 0.2 wt.%, Sc around 25 ppm and Ni about 2000 ppm. The source of the intermediate lavas was probably more depleted than the source for modern MORB (lower TiO_2 and La/Sm ratio) but in terms of TiO_2 the source was less depleted than the source of the lower lavas. Similar rocks in the Troodos ophiolite have been interpreted as melts from an ultramafic source that had already been depleted of a basaltic melt fraction (Smewing and Potts, 1976; Kay and Senechal, 1976).

The REE abundances in the most primitive upper lava (BC-75-5) could have been derived by about 20% batch melting of a source like that of normal MORB (e.g., White and Schilling, 1978). In addition, $^{143}Nd/^{144}Nd$ ratios for a Betts Cove intermediate and upper lava (Table 2b) are within the range of values reported for the Bay of Islands ophiolite ($\epsilon_{Nd} = 6.5 - 8.1$, Jacobsen and Wasserburg, 1979) and are consistent with a source similar to that for normal MORB. Consequently, the upper lavas may be direct analogues to young normal MORB, while the intermediate lavas may be derived from a peridotite residue from which a normal MORB magma has been previously extracted. In this case their similar $^{143}Nd/^{144}Nd$ ratios do not reflect a common source material, but derivation of their sources from an isotopically homogeneous mantle.

The positive ϵ_{ND}^{CHUR} value for the Betts Cove lower lava requires source materials with a time-integrated $(Sm/Nd)_N > 1$, in contrast to the relative LREE enrichment, $(Sm/Nd)_N < 1$, of the lavas. This discrepancy supports a mantle metasomatic process as the origin for the convex downward, normalized REE patterns. The lower $^{143}Nd/^{144}Nd$ value for the lower lava compared to the upper and intermediate lava suggests either that the LREE-enrichment occurred much earlier than the eruption of the lavas or that more recent enrichment occurred involving a material of low $^{143}Nd/^{144}Nd$. Hickey and Frey (1982) noted similar low $^{143}Nd/^{144}Nd$ ($\epsilon_{Nd} = -0.3$ to $+5.6$) in boninites with convex downward, normalized REE patterns from island arcs in the western Pacific. They concluded that the LREE-enrichment was recent and was derived from either subducted crustal sediment or a material similar to the source of ocean island volcanics.

TECTONIC ENVIRONMENT

The important feature of the Betts Cove mafic rocks is their range in composition, and consequently, the wide range in composition required for their respective mantle sources. The juxtaposition of these vastly different magma types, and their sequence of appearance, are crucial factors that must be addressed in any tectonic model.

Because of the unusually low TiO_2 values and the anomalous REE patterns in the lower volcanics, the Betts Cove ophiolite has become the center of controversy concerning the tectonic environment of

ophiolite formation (Miyashiro, 1973; Church and Coish, 1976; Sun and Nesbitt, 1978a; Upadhyay and Neale, 1979; Serri, 1980, 1981; Crawford et al., 1981). Such low TiO_2 contents (<0.3%) are not found in modern ocean ridge basalts. On the other hand, some island arc-related rocks have similar low-TiO_2 contents. This has led to the suggestion that the Betts Cove, and similar ophiolites, could have formed during the early stages of island arc volcanism (Miyashiro, 1973). More recently, Serri (1980), Meijer (1980) and Crawford et al., (1981), suggested that boninites found in the Marianas trench region represent modern analogues to very-low TiO_2 basalts found in some ophiolites such as Betts Cove. Boninites are high-MgO, high-SiO_2, low-TiO_2 island-arc volcanics, which frequently exhibit U-shaped chondrite-normalized REE patterns (Crawford et al., 1981; Hickey and Frey, 1982). They are typically associated with tholeiitic island-arc volcanics (Meijer, 1980) and are found unusually close to the trench, either in a fore-arc position (Meijer, 1980; Crawford et al., 1981), or along the arc-ward trench wall (Dietrich et al., 1978; Hawkins et al., 1979). To explain the unusual position of boninites, and their chemical characteristics, Crawford et al. (1981, Fig. 3) presented the following model for their origin. They suggested that back-arc or inter-arc basins are formed by the effect of a rising mantle diapir which causes the arc to split. As the diapir rises it heats the surrounding subarc mantle, which has been previously depleted in basaltic melt components by extraction of arc tholeiite magma, and enriched in H_2O and incompatible elements derived from the subducted oceanic plate. Hydrous, remelting of this refractory peridotite leads to the generation of boninites. Continued rise of the diapir causes melting in the diapir itself and this produces tholeiitic magmas of ocean floor affinity. By this time a basin has been established and the magmatism comes from a spreading center not influenced by the subduction. This model explains the juxtaposition of boninites between island arc tholeiites below and basalts with ocean ridge characteristics above. The critical feature of this model is that it provides a mechanism by which melts can be derived from refractory peridotite source materials.

Is the above boninite model applicable to the Betts Cove ophiolite? The lower lava group at Betts Cove are chemically most similar to boninites. However, there are some important differences: 1) All boninites are silica-rich, and hence always quartz normative whereas only a few lower lavas at Betts Cove are quartz-normative (even these have probably been enriched in SiO_2 by metasomatic processes, Coish, 1977b). 2) The major phenocryst phase in boninites is low-Ca pyroxene, either orthopyroxene or clinoenstatite, compared to diopsidic clinopyroxene in the Betts Cove lower lavas. 3) Olivine appears to have played a more significant role in the evolution of the lower lavas at Betts Cove than in boninites. 4) In this model boninites are directly underlain by island arc
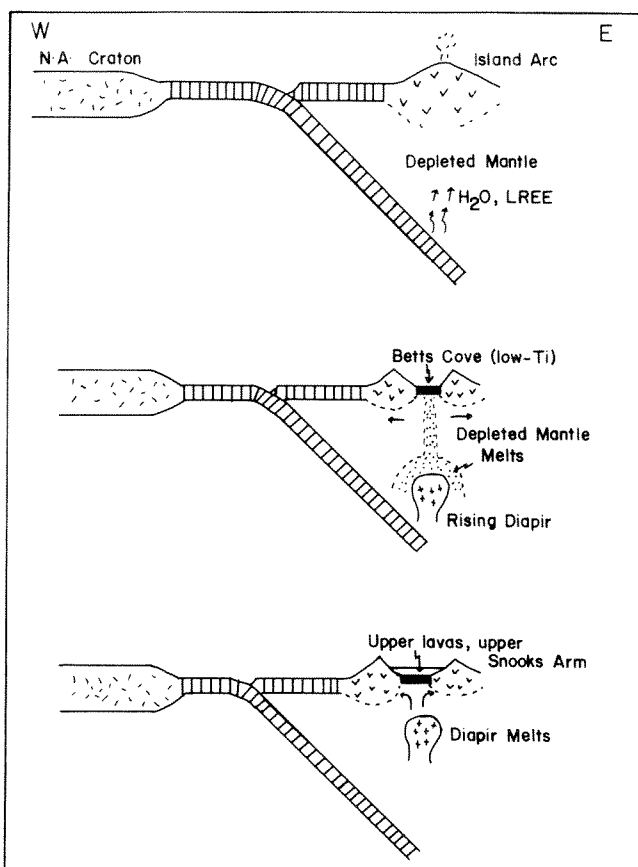
FIG. 12. Schematic diagram showing a possible tectonic model for the origin of the Betts Cove ophiolite. Model is based on ideas from Crawford *et al.* (1981) on the origin of boninites and ideas from Nelson and Casey (1979) on the tectonic evolution of western Newfoundland. The model begins with enrichment of a very depleted mantle by a fluid carrying LREE. Then a rising diapir of MORB-type mantle from below the very depleted mantle causes melting of the very depleted mantle through which it is moving. These melts erupt as low-Ti basalts in the Betts Cove ophiolite. The final stage shows melting of the MORB-type mantle diapir itself to produce the upper lavas at Betts Cove and the upper Snooks Arm basalts.

tholeiitic lavas, and in at least one area, DSDP Site 458 (Meijer, 1980), this relationship is observed; whereas sheeted dikes, gabbros and ultramafic cumulates underlie the lower lavas at Betts Cove.

These differences imply that the boninite model can not be a *direct* analogue for the formation of the Betts Cove ophiolite. However, since boninites are the only relatively young rocks that resemble the lower lavas at Betts Cove, a modified version of the boninite model may be appropriate. In the model of Serri (1980) and Crawford *et al.* (1981), boninites are extruded during the initial splitting of an island arc and thus directly overlie island arc tholeiites. This cannot be the case at Betts Cove since the sub-volcanic structure is equivalent to the intrusive part of ocean crust. Moreover, the eruption sequence of Betts Cove lavas, beginning with the most depleted source and ending with a more fertile source, requires an explanation of why the initial magmas are derived from the most depleted source.

A plausible model for the origin of the volcanic stratigraphy at Betts Cove is shown in Fig. 12. Craw-

ford *et al.* (1981) imply that diapirs rise only during the initiation of arc splitting. However, it is possible that continued spreading of the back-arc or inter-arc basin may be due to the ascent of many plumes similar to that envisioned to initiate splitting. If this is the case, then the later diapirs would also cause melting of an already-depleted peridotite source that had been enriched by fluids from the subducting plate, but now the depleted lavas so produced would extrude or intrude as dikes in a spreading center within the basin rather than within the arc volcanics. This could result in boninite-like lavas built directly upon mantle material. This model would explain the low-Ti lower lavas sitting on sheeted dikes and gabbros at Betts Cove. The intermediate lavas may also be produced at this stage and may represent melting of depleted mantle that had not been invaded by LREE-enriched fluids. This is not unreasonable since low-Ti lavas with U-shaped REE patterns are associated with low-Ti lavas with straight REE patterns in modern boninite terranes (Hickey and Frey, 1982). The upper lavas of the ophiolite and volcanics in the up-

per Snooks Arm group (Jenner and Fryer, 1980) may have formed during the latter stages in the ascent of the diapir when this MORB-type mantle began to melt. The lavas produced would thus be relatively undepleted and similar in composition to the upper volcanics. The rise of a succession of diapirs can be considered as episodic which would result in episodic spreading in the basin. Furthermore, the episodic nature of the process would result in a stratigraphic sequence, in any part of the basin, of relatively undepleted volcanics above depleted volcanics.

CONCLUSIONS

The main conclusions of this study are: 1) There are three distinct basalt compositions in the Betts Cove ophiolite which correlate with stratigraphic position. They are designated upper, intermediate and lower lavas. The groups are clearly distinguished on the basis of their TiO_2 contents. 2) The upper lavas have chondrite-normalized REE patterns that are relatively LREE-depleted at $10-20\times$ chondrite. The intermediate lavas also have LREE-depleted patterns $[(La/Sm)_N = 0.50]$ with absolute HREE abundances between 5 and $7\times$ chondrites. The lower lavas have unusual convex-downwards, chondrite normalized REE patterns with a range in absolute abundances from 2 to $7\times$ chondrite. 3) A lower lava sample has an initial $^{143}Nd/^{144}Nd$ ratio (0.51223) lower than samples from the overlying units which have initial $^{143}Nd/^{144}Nd$ ratios similar to samples from the Bay of Islands ophiolite and recent, normal MORB. 4) The compositional differences between the three lava units probably were inherited from compositionally different mantle sources. The upper lavas may have originated from a source similar to that for modern MORB. The intermediate lavas could have formed by melting a mantle source more depleted in Ti and REE than the source of normal MORBs; possibly, they represent second-stage melts. The lower lavas are probably partial melts from an extremely depleted mantle that had been secondarily enriched by a LREE-enriched, Ti-depleted fluid. 5) A possible tectonic environment where the various mantle sources may be juxtaposed is a back-arc or inter-arc basin.

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