

ISOTOPIC CONSTRAINTS ON THE ORIGIN OF APPALACHIAN MAFIC COMPLEXES*

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ABSTRACT. Isotopic analyses of modern oceanic basalts and ophiolites have shown that both modern and ancient oceanic crust have a characteristic Nd and Sr isotopic signature indicative of derivation from a depleted mantle reservoir. It also appears that the Nd isotopic system is not appreciably disturbed by metamorphism. These isotopic characteristics have been extended to the Pt. Sal, Kings-Kaweah, and Josephine ophiolites of California. We have used these characteristics in an attempt to identify pieces of proto-Atlantic oceanic crust among the mafic and ultramafic rocks of the Appalachians. Sm-Nd mineral isochrons for the Baltimore Mafic Complex, Maryland (BMC) yield an age of 490 ± 20 my which we interpret as the igneous crystallization age. BMC whole rock samples do not define isochrons and have initial isotopic compositions of $-6.4 < \epsilon_{Nd}(T) < -2.2$, $+51 < \epsilon_{Sr}(T) < +115$. $\epsilon_{Nd}(T)$ and $\epsilon_{Sr}(T)$ are anti-correlated. This is not the signature of depleted mantle and oceanic crust but is similar to old continental crust. We propose that the BMC is a mafic continental intrusion, possibly subduction related, which was contaminated with old continental crust during emplacement. Whole rock samples from the Thetford Mines Complex, Quebec (TMC), do not define isochrons and have $-1.5 < \epsilon_{Nd}(T) < +4.2$, $+2.6 < \epsilon_{Sr}(T) < +114$. These data do not in any way reflect the signature of normal oceanic crust. These results are in contrast with geologic relationships that show the TMC to have the characteristics of an ophiolite complex. The TMC is chemically and isotopically similar to a class of other ophiolites that have affinities to modern boninites. The Chunky Gal Amphibolite, North Carolina, Lake Chatuge complex, North Carolina, and Hazen's Notch Amphibolite, Vermont were found to have a depleted mantle signature with $+5 < \epsilon_{Nd}(T) < +8$ and may be fragments of oceanic crust. The Webster-Addie body, North Carolina, has $\epsilon_{Nd}(T) \sim -1$, $\epsilon_{Sr}(T) \sim +30$ and is not isotopically similar to oceanic crust or the other North Carolina mafic bodies analyzed. From these isotopic results we infer that Appalachian mafic rocks have diverse origins: some are continental intrusives (BMC), others are probably fragments of oceanic crust (Vermont and North Carolina amphibolites). Future models for the development of the Appalachians must allow for these various origins. The possibility that some ophiolites are not normal oceanic crust but have an origin in a partially continental setting or as anomalous oceanic crust may require further attention.

I. INTRODUCTION

The pre-Mesozoic mafic and ultramafic bodies of the Appalachians are largely restricted to the latest Precambrian to Ordovician age eugeosyncli-

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nal rocks lying to the east of the chain of anticlinoria cored by Grenville age (~ 1.0 - 1.3 AE) Precambrian basement. In some places, particularly in the Southern Appalachians, ultramafic bodies also lie enclosed within basement complex rocks. The mafic and ultramafic bodies thus define a belt or series of parallel belts, in places narrow and well-defined, elsewhere broad and diffuse, which extends the length of the Appalachians. The majority of the bodies making up the belt are relatively small (<1 km), apparently rootless pods of peridotite which have been serpentinized to varying degrees. Also found within this belt are larger complexes of mafic and ultramafic rocks which range from relatively complete allochthonous sheets such as the Bay of Islands ophiolite in Newfoundland, through bodies in which metamorphism and deformation have partially obscured an ophiolite stratigraphy, to bodies that have little if any recognizable stratigraphy but, because of their location in the belt and mafic to ultramafic composition, have recently been described as dismembered or fragmented ophiolites.

Historically, the mafic and ultramafic rocks of the Appalachians have been interpreted as intrusive bodies, emplaced into an ensialic eugeosynclinal pile (Pratt and Lewis, 1905; Hess, 1939, 1955; Chidester and Cady, 1972), and which reached their present rootless stratigraphic position by diapiric rise of serpentinized peridotite. More recently, in light of the plate tectonic interpretation of ophiolites as fragments of obducted oceanic crust, the mafic and ultramafic rocks of the Appalachians have been interpreted as large and small fragments of proto-Atlantic oceanic crust and mantle that were incorporated into the Appalachian Orogen during the closure phase of a Wilson cycle (Church and Stevens, 1971; Williams, 1971; Dewey and Bird, 1971; Upadhyay, Dewey, and Neale, 1971; St. Julien, 1972; Laurent, 1975, 1977; Crowley, 1976; Morgan, 1977; Williams and Talkington, 1977; Malpas, 1977). The objective of this study is to use the Sm-Nd and Rb-Sr isotopic systems to place constraints on the origin of several Appalachian mafic bodies in which either extensive metamorphism, or tectonism, or both have made their origin uncertain.

The naturally occurring radioactive isotopes ^{87}Rb and ^{147}Sm decay with halflives $\tau_{1/2}^{\text{Rb}} = 49$ AE and $\tau_{1/2}^{\text{Sm}} = 106$ AE to the respective daughter isotopes ^{87}Sr and ^{143}Nd . Closed systems with differing parent-to-daughter ratios will evolve over time and develop measurable differences in the daughter element isotopic abundances. For Nd, these differences are conveniently represented using epsilon notation (DePaolo and Wasserburg, 1976a,b, 1977) in which $\epsilon_{\text{Nd}}(\text{T})$ is the deviation in parts in 10^4 of the sample isotopic composition from that of a nominal chondritic reservoir at time T:

$$\epsilon_{\text{Nd}}(\text{T}) = \left[\left\{ \left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{SAMPLE}}(\text{T}) / I_{\text{CHUR}}(\text{T}) \right\} - 1 \right] \times 10^4$$

where $I_{\text{CHUR}}(\text{T}) = I_{\text{CHUR}}(0) - (^{147}\text{Sm}/^{144}\text{Nd})_{\text{CHUR}} \cdot [\exp(\lambda^{\text{Sm}} \cdot \text{T}) - 1] \lambda^{\text{Sm}}$
 $= 6.54 \times 10^{-12} \text{ yr}^{-1}$, $I_{\text{CHUR}}(0) = 0.511847$, $(^{147}\text{Sm}/^{144}\text{Nd})_{\text{CHUR}} = 0.1967$,
 and CHUR stands for Chondritic Uniform Reservoir (DePaolo and Wasserburg, 1976b; Jacobsen and Wasserburg, 1981; Wasserburg and others,

1981). One can also reference the Sm/Nd ratios to this reservoir by defining the enrichment factor

$$f^{\text{Sm/Nd}} \equiv \left\{ \left(\frac{{}^{147}\text{Sm}/{}^{144}\text{Nd}}{\text{SAMPLE}} / \frac{{}^{147}\text{Sm}/{}^{144}\text{Nd}}{\text{CHUR}} \right) - 1 \right\}$$

Positive values of $f^{\text{Sm/Nd}}$ thus indicate light rare earth element (LREE) depletion; negative values indicate LREE enrichment. One can similarly define ϵ_{Sr} with respect to a reservoir corresponding to the undifferentiated mantle (UR) with nominal values $I_{\text{UR}}(0) = 0.7045$, $({}^{87}\text{Rb}/{}^{86}\text{Sr})_{\text{UR}} = 0.0827$, $\lambda^{\text{Rb}} = 1.42 \times 10^{-11} \text{ yr}^{-1}$:

$$\begin{aligned} \epsilon_{\text{Sr}}(\text{T}) &= \left[\left\{ \left(\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}} \right)_{\text{SAMPLE}}(\text{T}) / I_{\text{UR}}(\text{T}) \right\} - 1 \right] \times 10^4; \\ I_{\text{UR}}(\text{T}) &= I_{\text{UR}}(0) - ({}^{87}\text{Rb}/{}^{86}\text{Sr})_{\text{UR}} \cdot [\exp(\lambda^{\text{Rb}}\text{T}) - 1]; \\ f^{\text{Rb/Sr}} &\equiv \left\{ \left(\frac{{}^{87}\text{Rb}/{}^{86}\text{Sr}}{\text{SAMPLE}} / \frac{{}^{87}\text{Rb}/{}^{86}\text{Sr}}{\text{UR}} \right) - 1 \right\} \end{aligned}$$

(DePaolo and Wasserburg, 1976b; O'Nions, Hamilton, and Evensen, 1977; Jacobsen and Wasserburg, 1981). There is good reason to believe that the Earth has approximately chondritic rare earth ratios. Although these ratios are somewhat variable between different chondrites, the Nd model system parameters are reasonably well established. The Sr model system, on the other hand, is based on a correlation between Nd and Sr isotopic compositions in mantle-derived rocks and on the inference that the bulk Earth has a chondritic Sm/Nd ratio. The Sr parameters are therefore less well established. Nevertheless, the current estimate of $({}^{87}\text{Sr}/{}^{86}\text{Sr})_{\text{UR}} = 0.7045$ appears to be a reasonably close estimate of the bulk earth value.

Isotopic analyses of young rocks thought to have been derived directly from the mantle in oceanic regions have shown that the mantle is isotopically heterogeneous. The largest volume of rock is represented by mid-ocean ridge basalts (MORB) which have low values of $\epsilon_{\text{Sr}}(0)$ and high values of $\epsilon_{\text{Nd}}(0)$ (fig. 1), implying that these rocks are derived from a large ion lithophile (LIL) and LREE depleted part of the mantle which has existed for a significant fraction of the Earth's history. Other oceanic volcanics such as found on oceanic islands show a wide range of isotopic composition. In particular, it appears that there are *at least* two isotopically distinct, ancient mantle reservoirs capable of yielding basaltic magmas and which give rise to the general inverse correlation between $\epsilon_{\text{Nd}}(0)$ and $\epsilon_{\text{Sr}}(0)$ known as the mantle array: (1) a LREE and LIL element depleted reservoir characterized by negative values of $\epsilon_{\text{Sr}}(0)$ (~ -20 to -30) and positive values of $\epsilon_{\text{Nd}}(0)$ ($\sim +8$ to $+13$) which is the source for present-day basaltic oceanic crust, and (2) a reservoir variously referred to as primitive or undepleted mantle with $\epsilon_{\text{Nd}} \sim \epsilon_{\text{Sr}} \sim 0$ (DePaolo and Wasserburg, 1976a,b, 1979; DePaolo, 1983) or enriched mantle with $\epsilon_{\text{Sr}} > 0$, $\epsilon_{\text{Nd}} < 0$ (White and Hofmann, 1982; McCulloch and others, 1983). Mixtures of magma derived from these two sources appear to be responsible for the isotopic variation of rocks erupted on oceanic islands, in continental volcanic provinces, and

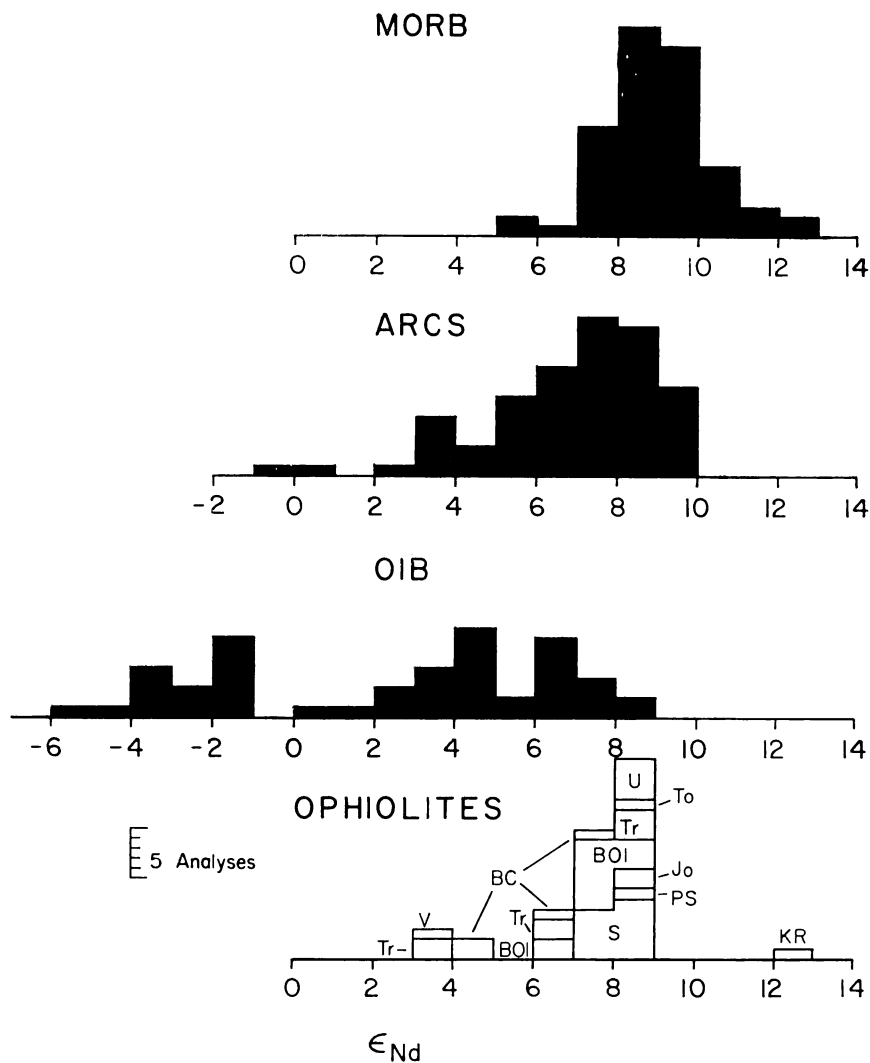


Fig. 1. Histogram of $\epsilon_{Nd}(T)$ values comparing mid-ocean ridge basalts (MORB), oceanic island arcs, and ocean island basalts (Carlson, MacDougall, and Lugmair, 1978; Cohen and others, 1980; DePaolo, ms; DePaolo and Wasserburg, 1976b, 1977; Dosso and Murthy, 1980; Hawkesworth and others, 1977; 1979a,b; Jahn and others, 1980; McCulloch and Perfit, 1981; Menzies and Murthy, 1980; O'Nions, Hamilton, and Evensen, 1977; 1978; Richard, Shimizu, and Allegre, 1976; Zindler and others, 1979) and ophiolites (key to ophiolites: Tr—Troodos, V—Vourinos, BC—Bett's Cove, BOI—Bay of Islands, S—Samail, PS—Pt. Sal, Jo—Josephine, KR—Kings-Kaweah, U—Kempersai, Urals, To—Toba, Japan) (Chen and Shaw, 1982; Coish, Hickey, and Frey, 1982; Edwards and Wasserburg, 1983; Hannah and Futa, 1982; Jacobsen and Wasserburg, 1979; McCulloch and others, 1981; Richard and Allegre, 1980).

in kimberlites. The concept of endmember reservoirs is somewhat of an idealization; in actuality it is likely that there is a continuum of sources in the mantle that are either undepleted or were depleted or enriched at a variety of times and that today give rise to the spectrum of isotopic compositions observed in modern mantle-derived rocks (compare with Jacobsen and Wasserburg, 1979). The conclusions of the present study depend on the fact that the largest volume of basaltic oceanic crust, both modern and ancient, is in general characterized by a depleted mantle signature.

Old, LREE, and LIL-enriched continental crust forms a complement to the depleted mantle reservoir and has negative values of $\epsilon_{Nd}(0)$ and large positive values of $\epsilon_{Sr}(0)$. Mixing of depleted mantle with crustal rocks has also been proposed as a mechanism for producing the parts of the mantle array populated by some continental flood basalts. The latter point of view has recently been emphasized by Carlson, Lugmair, and MacDougall (1981, 1983). Some of these mixing relationships are shown in figure 2 in which $\epsilon_{Nd}(0)$ is plotted against $\epsilon_{Sr}(0)$ for rocks from various settings. In this figure, oceanic crust, represented by mid-ocean ridge basalts and island arcs, is seen to have a distinctive isotopic signature of $\epsilon_{Nd}(0) \sim +8$ to $+13$ and relatively non-radiogenic values of $\epsilon_{Sr}(0)$.

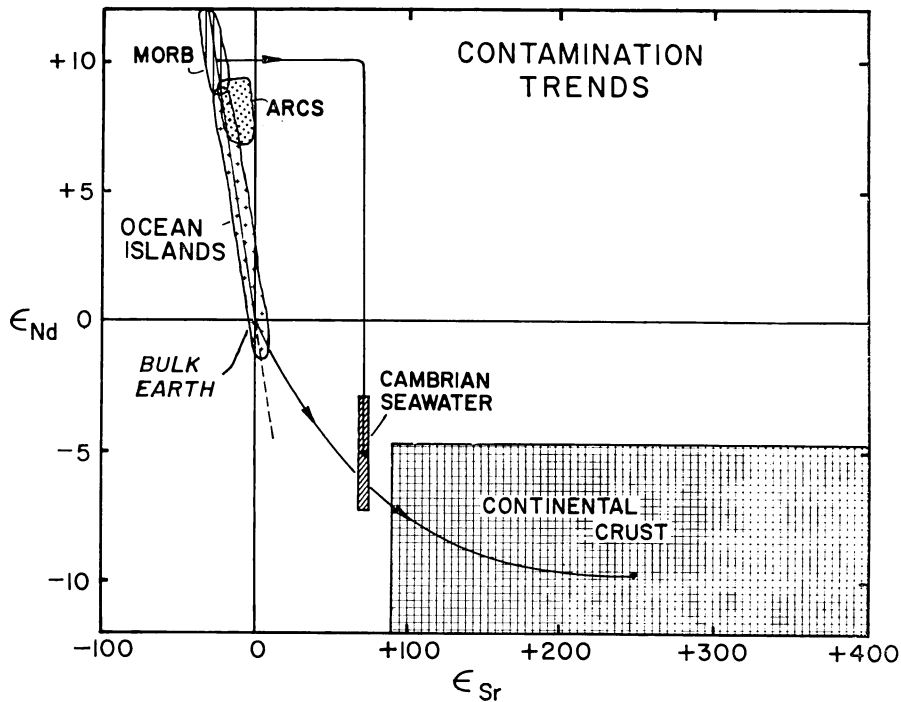


Fig. 2. ϵ_{Nd} versus ϵ_{Sr} diagram showing contamination trends and possible endmembers for mixing relationships. Isotopic composition of Cambro-Ordovician seawater from Peterman, Hedge, and Tourtelot (1970), Burke and others (1982) and Hooker, Hamilton, and O'Nions (1981).

Well-characterized ophiolite complexes such as the Samail Ophiolite, Oman, and Bay of Islands Ophiolite, Newfoundland, which are thought to represent sections of fossil oceanic crust, also have a depleted mantle isotopic signature (fig. 1). It should be noted, however, that there are some ophiolites that do not have the isotopic signature of modern oceanic crust. These occurrences are discussed later in light of our results.

In order to extend the isotopic observations on geologically well-characterized ophiolites, we have analyzed samples from three California ophiolites of Mesozoic age: the Josephine Ophiolite (157 my, Harper, 1980; Harper and Saleeby, 1980), the Point Sal Ophiolite (160 my, Hopson and Frano, 1977), and the Kings-Kaweah Ophiolite (200 my, Saleeby, 1982). The Josephine Ophiolite and Kings-Kaweah Ophiolite have metamorphic histories comparable to that inferred for many of the Appalachian mafic complexes; both were metamorphosed to greenschist or amphibolite facies during regional metamorphic events after their emplacement onto the continental margin. Our data for the California ophiolites are given in table 1 and figure 1. These rocks clearly have the Nd isotopic signature of the depleted mantle source characteristic of modern oceanic crust or oceanic island arcs, especially when allowance is made for the evolution of the source. For a depleted mantle with $f^{\text{Sm}/\text{Nd}} \sim +0.2$ one would expect this reservoir to evolve by $\sim +0.5 \epsilon_{\text{Nd}}$ units every 100 my. The present-day range of $+8$ to $+13 \epsilon_{\text{Nd}}$ for the MORB derived from the depleted oceanic mantle would thus correspond to a range of $+5.5$ to $+10.5$ at ~ 500 my ago. The value of $\epsilon_{\text{Nd}}(\text{T}) = +12.4$ for the Kings-Kaweah sample deserves some comment. This value is among the highest reported for samples from a depleted mantle reservoir and implies that the Kings-Kaweah Ophiolite was derived from mantle that was either more highly depleted than average or was depleted at a much earlier-than-average time. Ophiolite localities showing such high ϵ_{Nd} values will require more attention in the future.

The Sr isotopic compositions of ophiolites range from typical depleted mantle values to relatively radiogenic values. This dispersion is due to the relative ease with which the Sr isotopic system is disturbed by exchange during both regional metamorphism and metamorphism on the seafloor (Jacobsen and Wasserburg, 1979; McCulloch and others, 1981). This fact has limited the utility of the Sr isotopic system for constraining the origin of mafic rocks in complex orogenic terrains. In contrast, the Nd isotopic system appears to be relatively immune to such disturbances, suggesting that the distinctive Nd isotopic signature of oceanic crust is preserved during hydrothermal alteration on the sea floor and during obduction and regional metamorphism. Although the susceptibility of Nd and the REE in general to disturbance in response to different types of metamorphism is not well known, the available data suggest that the REE are quite robust. This feature could be of great importance in the Appalachians where metamorphism has obliterated the original mineralogy and modified the major element chemistry of most of the mafic and ultramafic rocks.

2. SAMPLES AND PROCEDURES

We have analyzed samples of two of the largest mafic/ultramafic complexes in the Appalachians south of Newfoundland: the Thetford Mines Complex (TMC) of southern Quebec (Laurent, 1975, 1977) and the Baltimore Mafic Complex (BMC) of Maryland (Hertz, 1951; Hopson, 1964; Southwick, 1970; Morgan, 1977; Hanan, ms). Both these complexes have been called ophiolites in the literature. The TMC has a complete ophiolite stratigraphy, although it is strongly deformed. The BMC on the other hand consists of only a cumulate dunite-chromitite base overlain by cumulate pyroxenites which grade into cumulate and non-cumulate two pyroxene gabbros and norites. The BMC does not include tectonized harzburgite, a dike complex, pillow basalts, or overlying marine sediments. A variety of lithologies of varying degrees of metamorphism from both Thetford and Baltimore have been analyzed. For the BMC, mineral separates were made for internal isochrons on three samples having well-preserved igneous mineralogy. In addition, we have analyzed a variety of samples of mafic/ultramafic rocks from diverse localities which have on occasion been suggested to be obducted oceanic crust. These are comprised of one sample of the Hazen's Notch Amphibolite associated with the serpentinitized dunite of the Belvidere Mountain ultramafic body of northern Vermont (Laird, ms; Chidester, Albee, and Cady, 1978; Laird and Albee, 1981), two samples of Chunky Gal Mountain amphibolite from the Buck Creek peridotite body of North Carolina (Hadley, 1949; Sailor and Kuntz, 1973; McElhaney and McSween, 1983), an amphibolite from the nearby Lake Chatuge complex (Hartley, 1973; Jones, Hartley, and Walker, 1973), and a sample of websterite from the type locality of Webster-Addie, North Carolina (Miller, 1953; Steuber, 1969). We have also analyzed two samples of the Grenville age basement gneiss from exposures in the Woodstock and Towsend Domes, Maryland (Tilton and others, 1958; Tilton, Doe, and Hopson, 1970) and one sample of Wissahickon schist as possible representatives of the Appalachian basement and metasedimentary cover through which ensialic mafic rocks may have intruded. Sample locations and brief petrographic descriptions are given in app. 1.

For the total rock analyses, 10 to 25 g interior pieces were crushed in a stainless steel mortar and pestle and a ~ 0.5 g split taken for HF + HClO₄ dissolution. With the exception of Md-11, an augen gneiss, all the samples analyzed were relatively fine-grained, and the powdered material adequately represents the whole rock. No special attempt was made to homogenize a large volume of sample Md-11. Nevertheless, the Nd isotopic results for this sample are in good agreement with those of fine-grained gneiss Md-7. Similarly, no attempt was made at dissolving the zircons present in Md-7, Md-11, Md-32, and RL80-7, although the HF + HClO₄ dissolution use does substantially attack those zircons that are metamict. The lack of dissolution of zircons is unlikely to have any effect on our Rb-Sr results due to the low concentrations of these elements in zircon. One may calculate the expected contribution of the zircons to the REE budget of the whole rock. For the rocks we analyzed, the worst case is for

TABLE 1
Rb-Sr and Sm-Nd isotopic results

Sample	$\frac{^{143}\text{Nd}^*}{^{144}\text{Nd}}$	ppm Nd	$\frac{^{147}\text{Sm}\ddagger}{^{144}\text{Nd}}$	$f^{\text{Sm}/\text{Nd}}$	$\epsilon_{\text{Nd}}(\text{T})^{\text{a}}$	$\frac{^{87}\text{Sr}^*}{^{86}\text{Sr}}$	ppm Sr	$\frac{^{87}\text{Rb}^+}{^{86}\text{Sr}}$	$f^{\text{Rb}/\text{Sr}}$	$\epsilon_{\text{Sr}}(\text{T})^{\text{a}}$
Baltimore Mafic Complex										
Md-32	0.51132	16.1	0.131	-0.334	-6.2	0.71209	302.9	0.0104#	-0.874	+114.9
Norite	± 2				± 0.5	± 4				± 0.6
	0.51131	16.3	0.130	-0.337	-6.4	0.71214	302.9	—	—	—
	± 3				± 0.5	± 4				
Md-32	0.51120	6.7	0.093	-0.526	—	0.71254	494.4	0.0042#	0.949	—
Plagioclase	± 2					± 4				
Md-32	0.51140	4.0	0.152	-0.228	—	0.71272	8.55	0.1161#	0.404	—
Orthopyroxene	± 2					± 4				
Md-32	0.51135	796.9	0.135	-0.314	—	0.71214	205.6	0.0100#	0.879	—
Apatite	± 2					± 3				
Md-3	0.51183	0.68	0.227	0.516	-2.2	0.70756	93.6	0.0089	-0.893	+50.8
Feldspathic websterite	± 2				± 0.4	± 2				± 0.3
Md-3	0.51130	0.12	0.068	-0.656	—	—	—	—	—	—
Plagioclase	± 2									
Md-3	0.51188	0.21	0.253	0.285	—	—	—	—	—	—
Orthopyroxene	± 2									
Md-3	0.51186	1.86	0.253	0.285	—	—	—	—	—	—
Clinopyroxene	± 3									
Md-44	0.51152	1.16	0.190	-0.035	-6.0	0.71052	131.0	0.0125	-0.849	+92.5
Gabbro	± 2				± 0.4	± 8				± 1.1
Md-44	0.51160	1.55	0.220	0.119	—	—	—	—	—	—
Pyroxene	± 2									
Md-44	0.51114	0.38	0.076	-0.612	—	—	—	—	—	—
Plagioclase	± 2									
Md-40	0.51152	5.36	0.168	-0.143	-4.5	0.70970	85.1	0.0133	-0.839	+80.7
Metagabbro	± 3				± 0.6	± 3				± 0.4
Md-30	0.51147	2.48	0.168	-0.144	-5.5	0.70950	180.0	0.0146	-0.823	+77.7
Gabbro	± 1				± 0.3	± 3				± 0.4
Md-15	0.51171	0.51	0.250	0.273	-6.1	0.71047	25.8	0.0193	-0.767	+91.0
Orthopyroxenite	± 2				± 0.4	± 3				± 0.4
Md-39	0.51156	4.84	0.159	-0.194	-3.1	0.70916	142.5	0.0230#	-0.721	+72.1
Hornblende Gabbro	± 3				± 0.7	± 4				± 0.6

Thetford Mines Samples										
Qe-18	0.51194	0.86	0.178	-0.097	+3.0	0.70763	53.5	0.1009	-0.220	+42.5
Lower Pillow Basalt	±2				±0.4	±3				±0.4
Qe-13A #3	0.51181	1.94	0.178	-0.097	+0.4	0.70761	72.3	0.0101	-0.878	+51.5
Lower Pillow Basalt	±3				±0.6	±4				±0.6
Qe-40	0.51197	1.05	0.201	0.022	+2.2	—	—	—	—	—
Upper Pillow Basalt	±2				±0.4					
Qe-8B	0.51209	1.43	0.205	0.044	+4.7	0.71287	98.6	0.3957	3.784	+87.2
Metadiabase	±2				±0.5	±4				±0.6
Qe-12A	0.51205	0.96	0.211	0.072	+3.2	0.71202	94.5	0.0106	-0.872	+114.0
Metagabbro	±1				±0.3	±3				±0.4
Qe-30A	0.51172	0.11	0.181	-0.078	-1.5	0.70412	88.4	0.0039	-0.953	+2.6
Metagabbro	±2				±0.4	±4				±0.6
Qe-24	0.51242	0.11	0.326	0.655	+2.9	0.70936	6.58	0.2457	1.971	+52.6
Clinopyroxenite	±2				±0.4	±5				±0.7
Vermont Sample										
VJL-360C	0.51232	6.9	0.217	0.101	+8.0	0.70412	63.1	0.1509#	0.824	-12.3
Hazens Notch Amphibolite	±3				±0.5	±3				±0.4
North Carolina Samples										
CG 12-1	0.51241	3.42	0.248	0.259	+7.7	0.70543	95.2	0.0143#	-0.827	+20.1
Chunky Gal Amphibolite	±2				±0.4	±6				±0.8
CG 2-18	0.51210	12.7	0.182	-0.073	+6.0	0.70680	340.4	0.0290	-0.649	+38.1
Chunky Gal Amphibolite	±3				±0.5	±5				±0.7
C-1	0.51220	2.09	0.226	0.137	+5.1	0.70320	162.0	0.0258#	-0.688	-12.7
Lake Chatuge Amphibolite	±2				±0.5	±4				±0.6
XTC-127	0.51190	0.57	0.231	0.176	-1.1	0.70691	8.13	0.1323#	0.599	+29.2
Webster-Addie websterite	±2				±0.4	±4				±0.6
XTC-127 leached residue	0.51207	0.31	0.278	0.414	-0.9	0.70654	7.30	—	—	—
	±1				±0.3	±4				

TABLE 1 (continued)

Sample	$\frac{^{143}\text{Nd}^*}{^{144}\text{Nd}}$	ppm Nd	$\frac{^{147}\text{Sm}^\ddagger}{^{144}\text{Nd}}$	$f^{\text{Sm}/\text{Nd}}$	$\epsilon_{\text{Nd}}(\text{T})^*$	$\frac{^{87}\text{Sr}^*}{^{86}\text{Sr}}$	ppm Sr	$\frac{^{87}\text{Rb}^\#}{^{86}\text{Sr}}$	$f^{\text{Rb}/\text{Sr}}$	$\epsilon_{\text{Sr}}(\text{T})^*$
Baltimore Gneiss, Wissahickon Schist Samples										
Md-7	0.51119	24.6	0.117	-0.407	-7.8†	0.72947	199.0	1.363#	15.48	+227.4†
Biotite Gneiss	±2				±0.4	±3				±0.4
Md-11	0.51117	69.8	0.115	-0.415	-8.2†	0.75406	128.8	3.225#	37.99	+393.2†
Augen Gneiss	±2				±0.4	±6				±0.9
RL 80-7	0.51120	22.4	0.102	-0.484	-6.6†	0.71773	123.6	0.6253#	6.561	+134.2†
Wissahickon schist	±2				±0.3	±3				±0.4
California Ophiolites ^b										
P-59	0.51229	1.25	0.223	0.134	+8.2	0.70333	298.7	0.0061	-0.926	-14.2
Pt. Sal Gabbro	±4				±0.9	±4				±0.6
JP-042	0.51238	0.74	0.275	0.396	+8.8	0.70349	88.4	0.0686	-0.171	-13.9
Josephine Gabbro	±2				±0.4	±3				±0.4
Y-13b	0.51238	1.14	0.257	0.305	+8.4	0.70342	214.7	0.0470#	-0.431	-14.2
Josephine Gabbro	±2				±0.4	±3				±0.4
Ca-10	0.51254	1.57	0.243	0.237	+12.4	0.70232	131.6	0.0061	-0.926	-27.9
Kings R. Gabbro	±3				±0.5	±3				±0.4

* Errors are 2σ of the mean; $^{143}\text{Nd}/^{144}\text{Nd}$ normalized to $^{146}\text{Nd}/^{142}\text{Nd} = 0.636151$; $^{87}\text{Sr}/^{86}\text{Sr}$ normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$.

† Uncertainty = 0.2 percent.

* Uncertainty = 1.0 percent. Rb measured on aliquot only—except as noted.

Uncertainty = 0.4 percent. Rb measured on separated sample.

† Initial ϵ values calculated for 490 my.

* Errors on initial ϵ values reflect only the analytical uncertainty and do not include errors due to age estimates.

^b Initial ϵ values calculated using the ages: Pt. Sal = 160 my; Josephine Ophiolite = 157 my; Kings R. Ophiolite = 200 my.

Md-7 which had ~ 2 mg insoluble residue from ~ 700 mg rock. If the residue were entirely zircon at 100 ppm Nd, then the undissolved material would comprise ~ 1 percent of the total rock Nd. Assuming an improbably large Sm/Nd fractionation between zircon and whole rock leading to a 20 ϵ -unit difference in isotopic composition between the zircon and rock, the shift in the whole rock ϵ_{Nd} value due to the lack of zircon dissolution would be only ~ 0.2 ϵ -units. We, therefore, believe the undissolved zircons to be an insignificant problem for these rocks.

Mineral separates were made using a combination of magnetic separation, heavy liquids (STBE and MI), and hand-picking. The purity of the mineral separates was ~ 99 percent for plagioclase and apatite and > 95 percent for pyroxene with fine-grained magnetite largely making up the remaining < 5 percent. The mineral separates were rinsed in acetone and distilled water and air dried before weighing.

In order to assess the possibility of contamination along grain boundaries during serpentinization and metamorphism, a split of the powdered Webster-Addie sample was leached briefly in cold 0.1M HCl, rinsed in water, then leached for 1 hr in sub-boiling 1.5M HCl. The insoluble residue was dried, weighed, and dissolved in HF + HClO₄.

In order to spike the samples properly, solution aliquots containing 1 to 2 mg of sample were spiked and approximate Rb, Sr, Sm, and Nd concentrations determined by isotope dilution. Using these concentrations, the remainder of the sample was spiked and chemically separated to determine both isotopic compositions and concentrations precisely. On many samples, only aliquot data were collected for Rb concentrations. The errors on these data are consequently higher but due to the low Rb/Sr ratios involved, the error in the calculated initial ratios is insignificant. Details of the chemical separation and mass spectrometric procedures are given in Eugster and others (1970), Papanastassiou and Wasserburg (1973), Papanastassiou, DePaolo, and Wasserburg, (1977), and DePaolo (ms). Blanks for the complete analytical procedure are < 50 pg Nd and < 200 pg Sr and are negligible.

3. RESULTS

Our results for the Baltimore Mafic Complex are presented in table 1 and figures 3 to 6. Figure 3, A to C, shows our Sm-Nd mineral data on three samples of the BMC. We report 2σ errors for the age and initial values calculated using the regression routine of Williamson (1968). The large error on the age for the isochron for Md-32 is due to the relatively small spread in Sm/Nd ratios which in turn is probably caused by small amounts of apatite in the plagioclase and orthopyroxene mineral separates. On the basis of these three concordant isochrons, we assign an igneous crystallization age of 490 ± 20 my to the BMC. It is important to note that the initial $\epsilon_{Nd}(T)$ values for these isochrons are different from one another, and all are negative.

The results of our Rb-Sr analyses on minerals from sample Md-32 are plotted in figure 3D. In contrast to the Nd isotopic system, the mineral data fail to define an Rb-Sr isochron with several different mineral

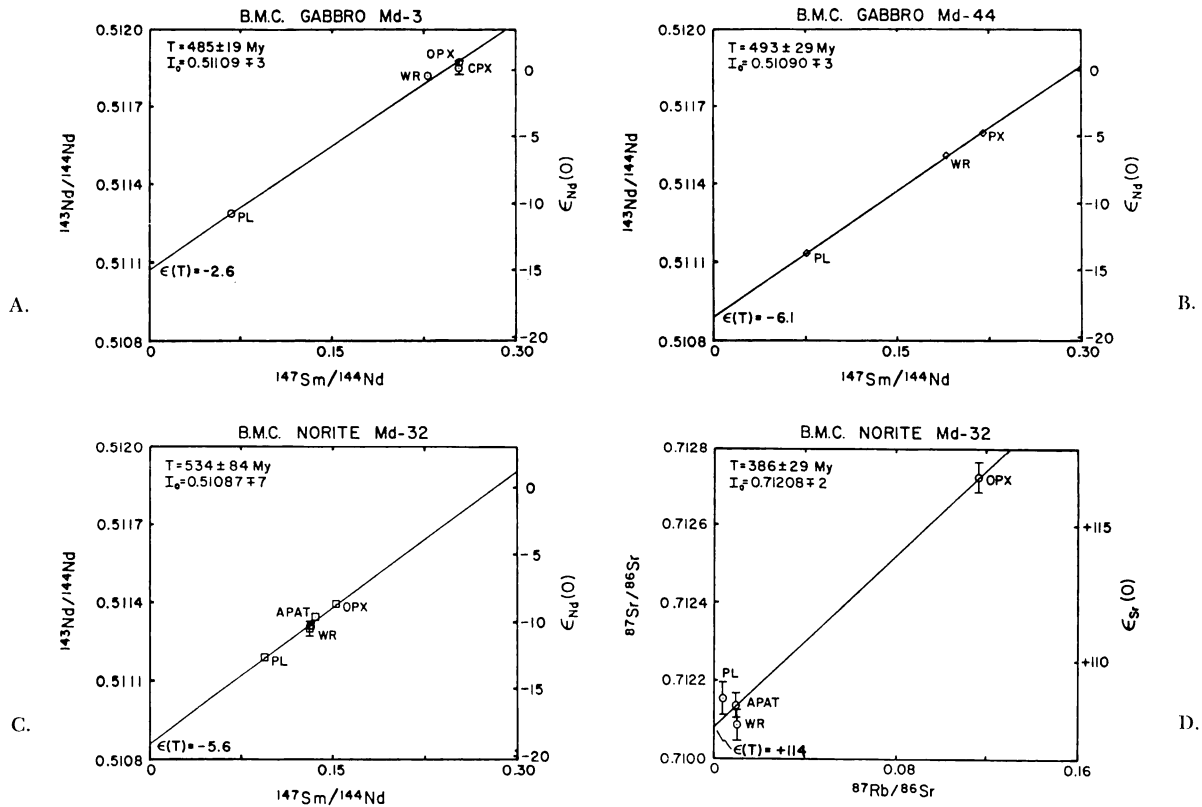


Fig. 3. (A) Sm-Nd evolution diagram showing internal isochron relationship for Baltimore Mafic Complex feldspathic websterite sample Md-3; (B) As above, two-pyroxene gabbro sample Md-44; (C) As above, norite sample Md-32; (D) Rb-Sr diagram showing mineral data for Baltimore Mafic Complex norite sample Md-32. Note the lack of an isochron relationship in contrast to the Sm-Nd data.

phases. The slope is defined only by the orthopyroxene point and the cluster of points with low Rb/Sr ratio. A linear fit to the data yields an age of 386 ± 28 my and a high initial $\epsilon_{\text{Sr}}(\text{T}) = +114$ (0.7121). The results of our Sm–Nd analyses of whole rocks from the BMC are plotted in figure 4A, and our Rb–Sr results for the same samples in figure 4B. As can be seen from these figures, the whole rock data define neither Sm–Nd nor Rb–Sr isochrons, as could have been anticipated from the variation in $\epsilon_{\text{Nd}}(\text{T})$ among the mineral isochrons. Instead, the Nd data vary from $\epsilon_{\text{Nd}}(\text{T}) = -6.4$ to -2.2 and the Sr data from $\epsilon_{\text{Sr}}(\text{T}) = +115$ to $+51$ (0.7120 to 0.7075) for an age of 490 my. The Nd data show a general correlation between $\epsilon_{\text{Nd}}(0)$ and Sm/Nd about the 490 my reference isochron (fig. 4A). There is no obvious correlation in the Sr data (fig. 4B). In terms of trace element chemistry, all the Baltimore samples, with the exception of obvious cumulate rocks, are moderately to strongly LREE-enriched relative to chondrites, as can be seen by the negative values of $f^{\text{Sm}/\text{Nd}}$ in table 1. These values lie between typical continental crustal values of $f^{\text{Sm}/\text{Nd}} \sim -0.4$ and a chondritic value of $f^{\text{Sm}/\text{Nd}} \sim 0$. The whole rock Rb/Sr ratios have a restricted range and are low relative to nominal undepleted mantle (UR).

Our results for the Thetford Mine Complex samples are presented in table 1 and figure 5. We were unable to obtain a mineral isochron for the TMC due to the pervasive low-grade metamorphic overprint throughout the Thetford locality which has destroyed the igneous mineralogy. For the purpose of calculating initial isotopic ratios we have assumed an age of 500 my. With the exception of Qe 24 (a cumulate clinopyroxenite), all these samples have approximately chondritic Sm/Nd ratios. The calculation of $\epsilon_{\text{Nd}}(\text{T})$ values is therefore nearly independent of the assumed age. As can be seen in figure 5, the whole rock samples from Thetford do not define an isochron in either the Sm–Nd or Rb–Sr evolution diagrams. Using the assumed age, we calculate very variable initial isotopic compositions with $\epsilon_{\text{Nd}}(\text{T}) = +4.2$ to -1.5 , $\epsilon_{\text{Sr}}(\text{T}) = -2.6$ to $+114$ (0.7041 to 0.7119). The samples have a wide range of Rb/Sr ratios, ranging from values much less than bulk mantle values to moderately LIL-enriched values. These samples are also characterized by low concentrations of Sm and Nd.

Our isotopic results for the North Carolina and Vermont samples are presented in table 1. We again assume an age of 500 my for the purpose of calculating $\epsilon(\text{T})$ values for these samples. Although this may be in error by as much as ± 100 my, in the worst case (XTC-127, leached residue) this corresponds to a shift of only ∓ 1 ϵ -unit in $\epsilon_{\text{Nd}}(\text{T})$ and does not affect our conclusions.

The samples of amphibolite from the Lake Chatuge district and Chunky Gal amphibolite have initial $\epsilon_{\text{Nd}}(\text{T})$ in the range $+5.1$ to $+7.7$. The $\epsilon_{\text{Sr}}(\text{T})$ values for the Chunky Gal amphibolites are moderately radiogenic at $+20$ and $+38$ (0.7053 and 0.7066) while the Lake Chatuge sample has $\epsilon_{\text{Sr}}(\text{T}) = -13$ (0.7030). These samples are slightly LREE-enriched to moderately LREE-depleted and have Rb/Sr ratios lower than bulk earth values (UR) (table 1). The North Carolina amphibolite whole rock samples do not define Rb–Sr or Sm–Nd isochrons. The unleached websterite from the Webster-Addie body has $\epsilon_{\text{Nd}}(\text{T}) = -1.1$, $\epsilon_{\text{Sr}}(\text{T}) = +29.2$ (0.7060)

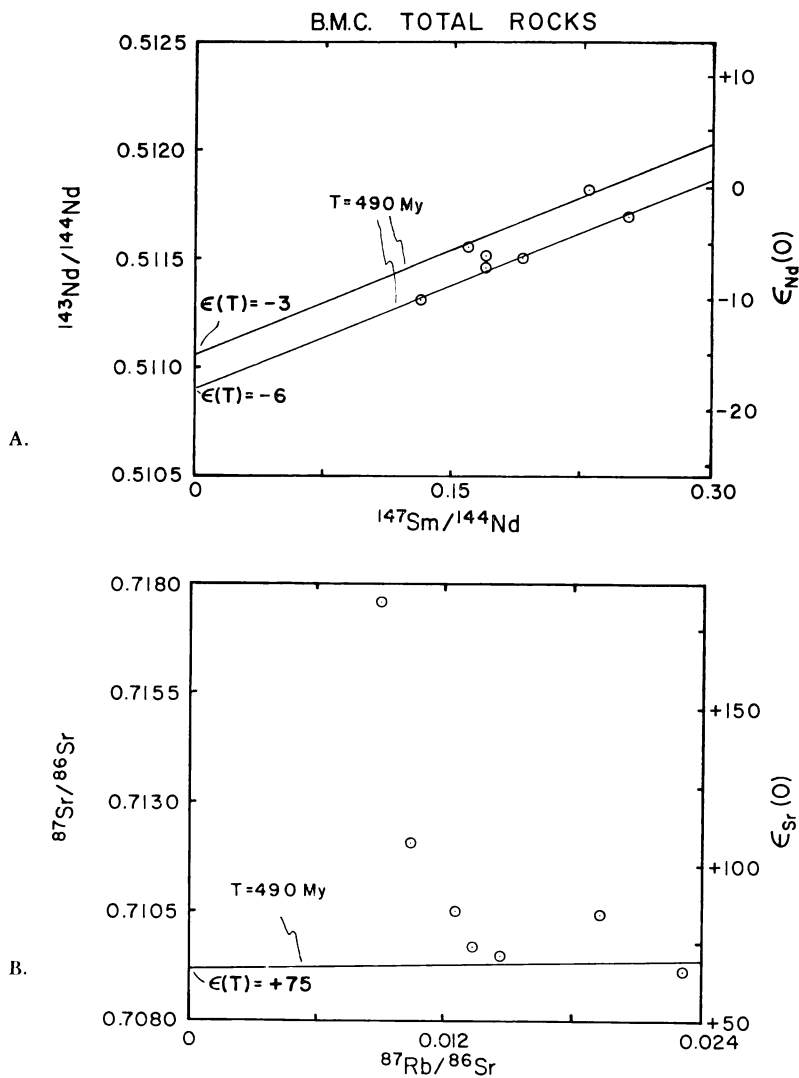


Fig. 4 (A). Sm-Nd evolution diagram with data for whole rock samples of the Baltimore Mafic Complex. Reference isochrons are drawn for 490 my and $\epsilon_{\text{Nd}}(T) = -3$ and -6 . Note the implied range in $\epsilon_{\text{Nd}}(T)$ and general correlation about the 490 my isochron. (B). Rb-Sr evolution diagram with the data for whole-rock samples of the Baltimore Mafic Complex. Reference isochron is drawn for 490 my and initial $\epsilon_{\text{Sr}}(T) = +75$. Note the implied range in $\epsilon_{\text{Sr}}(T)$ and limited range of Rb/Sr ratios.

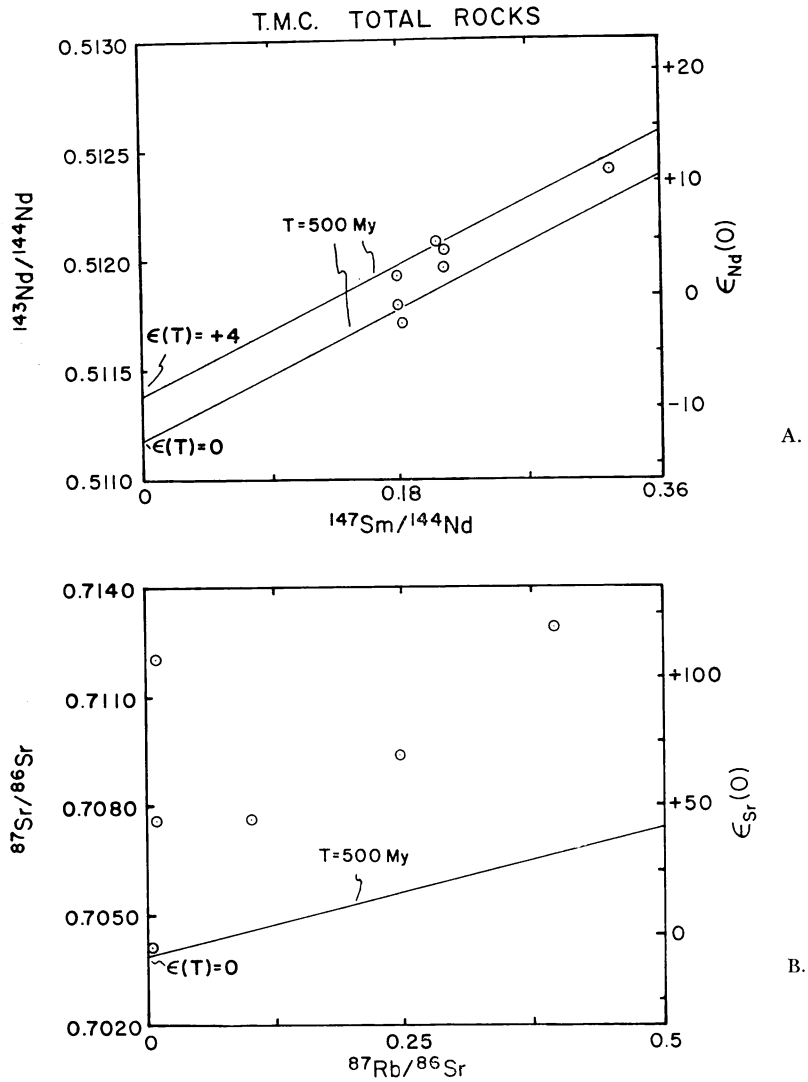


Fig. 5 (A). Sm-Nd evolution diagram with data for whole rock samples of the Thetford Mines Complex. Reference isochrons are drawn for 500 my and $\epsilon_{Nd}(T) = +4$ and 0. Note the implied range in $\epsilon_{Nd}(T)$.
 (B). Rb-Sr evolution diagram with data for whole rock samples of the Thetford Mines Complex. Reference isochron is drawn for 500 my and $\epsilon_{Sr}(T) = 0$. Note the implied range of $\epsilon_{Sr}(T)$ and wide variation in Rb/Sr ratios.

and is slightly LREE depleted. Approximately half the REE were removed by leaching the sample, and the leached residue was strongly LREE depleted (table 1). The measured $^{143}\text{Nd}/^{144}\text{Nd}$ value, however, increased from the unleached to the leached sample so that the calculated $\epsilon_{\text{Nd}}(\text{T})$ is identical within error for both samples. By leaching the sample we have apparently selectively dissolved a relatively LREE enriched phase. No evidence was found for a component with a high value of $\epsilon_{\text{Nd}}(\text{T})$.

The samples of basement gneiss Md-7 and Md-11 (table 1) are typical of old upper continental crust and have negative $\epsilon_{\text{Nd}}(0)$ and large positive $\epsilon_{\text{Sr}}(0)$ reflecting the upper crust's LREE- and LIL-enriched nature (McCulloch and Wasserburg, 1978). At the time of crystallization of the BMC (~ 490 my), these samples would have had $\epsilon_{\text{Nd}}(\text{T}) = -7.8$ and -8.2 and $\epsilon_{\text{Sr}}(\text{T}) = +227$ and $+393$ (0.7200 and 0.7315). The T_{CHUR} model ages for these samples are 1.25 and 1.27 AE which are quite reasonable for Grenville age crust (Tilton and others, 1958; Tilton, Doe, and Hopson, 1970; Grauert, 1974). T_{UR} ages (at 1.37 AE and 1.11 AE) are generally less reliable but in this case are close to the T_{CHUR} ages. The sample of Wissahickon schist has $\epsilon_{\text{Nd}}(490) = -6.6$, $\epsilon_{\text{Sr}}(490) = +134$, and $\text{T}_{\text{CHUR}} = 1.04$ AE, $\text{T}_{\text{UR}} = 1.72$ AE. The Nd model age implies that the Wissahickon sediments were dominantly derived from Grenville age basement such as the Baltimore Gneiss. The Wissahickon schist is known to have been metamorphosed at ~ 330 my from K-Ar studies (Lapham and Basset, 1964).

4. DISCUSSION

An Sm-Nd age of $\sim 490 \pm 20$ my was found for three different rocks from the Baltimore Mafic Complex. This is in agreement with zircon U-Pb ages (Sinha, personal commun.). We therefore interpret these results as the igneous crystallization age of the BMC. This places the time of magmatism prior to or at the earliest stages of the Taconic Orogeny. Samples from both the Susquehanna block (Md-32, Md-44) and the Baltimore block (Md-3) have ages that are identical within error, confirming the close relationship of these two exposures. This age is similar to the age of 505 ± 10 my obtained by Jacobsen and Wasserburg (1979) for the Bay of Islands complex. In contrast to their isotopic results which demonstrated an oceanic signature for the Bay of Islands complex ($\epsilon_{\text{Nd}}(\text{T}) = +6.5$ to $+8.1$, $\epsilon_{\text{Sr}}(\text{T}) = -19.6$ to $+32.5$), our isotopic data clearly indicate that the BMC was not derived from depleted oceanic mantle. In particular, the negative values of $\epsilon_{\text{Nd}}(\text{T})$ and positive values of $\epsilon_{\text{Sr}}(\text{T})$ are most like the isotopic signature of old continental crust. The variation in initial isotopic composition of both Nd and Sr does not appear to correlate with stratigraphic position in the complex, rock type, or degree of metamorphism. Hanan (1980, 1981) has shown that $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are shifted to higher values in metamorphic minerals along late stage fractures in the gabbro. These fractured rocks were avoided in the present study. It is clear from our attempt at an Rb-Sr mineral isochron (fig. 3D) that there has been redistribution of Rb or Sr on a mineral scale during post-crystallization metamorphism. Nevertheless, for reasons that will be shown below, we believe that the variation in whole rock initial Sr isotopic composition

is a magmatic characteristic, and that on the scale of several centimeters, the whole rocks have remained largely closed systems since crystallization. That the variation in initial Nd isotopic composition is a magmatic characteristic is clearly demonstrated by the different initial values obtained for our Sm–Nd mineral isochrons. This interpretation hinges on the assumption that the Sm–Nd mineral isochrons record both the time of crystallization from a melt and the melt's isotopic composition. It is possible that the mineral isochrons record a metamorphic event and that the initial isotopic compositions are a result of large scale introduction of Nd and Sr from the surrounding rocks. Such a metamorphic event would have had to have occurred shortly after crystallization in order to account for the agreement between the Sm–Nd mineral isochron and U–Pb zircon ages. In addition, the samples chosen for isochron work have anhydrous, apparently igneous mineralogy (see app. 1) implying granulite grade metamorphism. Without a hydrous fluid phase to mediate the process, we believe it is unlikely that isotopic exchange can occur over a scale of kilometers. Furthermore, as will be shown below, amphibolite grade metamorphism, presumably in the presence of abundant fluid, did not appreciably disturb the Nd isotopic signature of mafic rocks from North Carolina and Vermont. Therefore, we do not believe that the Nd systematics of the BMC samples record a metamorphic event. In the discussion that follows, we shall assume that this is the case. In figure 6 we have plotted our results for the BMC in an $\epsilon_{Nd}(T)$ versus $\epsilon_{Sr}(T)$ diagram. The data define an array lying to higher $\epsilon_{Sr}(T)$ than the mantle array with $\epsilon_{Nd}(T)$ inversely correlated with $\epsilon_{Sr}(T)$. It is this coherence, together with the assumption that the $\epsilon_{Nd}(T)$ values are magmatic, that leads us to believe that the variation in $\epsilon_{Sr}(T)$ is not due to later metamorphism. The trend in the ϵ_{Nd} – ϵ_{Sr} diagram is exactly what one would expect for contamination of a mantle-derived magma with high ϵ_{Sr} and low ϵ_{Nd} , old continental crust. Figure 2 schematically illustrates this along with other contamination trends and the isotopic characteristics of endmember rock types. Hanan (ms) suggested that the radiogenic Sr isotopic compositions of the BMC are due to interaction with seawater. This is almost certainly not the case for two reasons: (1) Cambro-Ordovician seawater had $\epsilon_{Sr} \sim +72$ (0.709) (Peterman, Hedge, and Tourtelot, 1970; Burke and others, 1982), a value *lower* than the initial isotopic composition of several of our samples; (2) interaction with seawater is not expected to shift the Nd isotopic composition by a significant amount until extremely high water/rock ratios are reached ($\sim 10^5$) due to the extraordinarily low concentration of Nd in seawater ($\sim 2 \times 10^{-12}$ g/g) (Goldberg and others, 1963; Høgdahl, Melson, and Bowen, 1968; DePaolo and Wasserburg, 1977). The mixing line between oceanic crust and seawater is shown in figure 2. This process is thought to account for the spread of $\epsilon_{Sr}(T)$ at essentially constant $\epsilon_{Nd}(T)$ seen in figure 6 for the Bay of Islands data (Jacobsen and Wasserburg, 1979). Our data for the BMC rocks in figure 6 do not show this relationship. We suggest instead that the Baltimore Complex magma was contaminated by or partially derived from older crustal material. Possible candidates for the contaminant are the Grenville age basement rocks which are exposed in several mantled gneiss domes in

the Baltimore area and the overlying metasediments. Our data for the Baltimore gneiss and Wissahickon schist samples are plotted on figure 6 at the isotopic composition they would have had 490 my ago. It is clear from this figure that the gneisses are suitable candidates for the contaminant in terms of their isotopic composition, as would be sediments or metasediments largely derived from Grenville age basement. The particular sample of Wissahickon schist analyzed appears to have too low $\epsilon_{Sr}(T)$ to be a reasonable bulk contaminant. Undoubtedly there exists a wide variation in both Rb/Sr and ϵ_{Sr} within the Wissahickon schist, and it is likely that assimilation of portions of this formation could lead to the observed isotopic pattern. Several other lines of evidence support the idea that the BMC magma was contaminated with older crustal material. Brown primary amphibole is common in the gabbro, suggesting a relatively high f_{H_2O} as might result from the assimilation of hydrous phases, and partially assimilated mafic xenoliths of uncertain origin are present in the upper portions of the gabbro (Hanan, ms). Further evidence is the observation originally

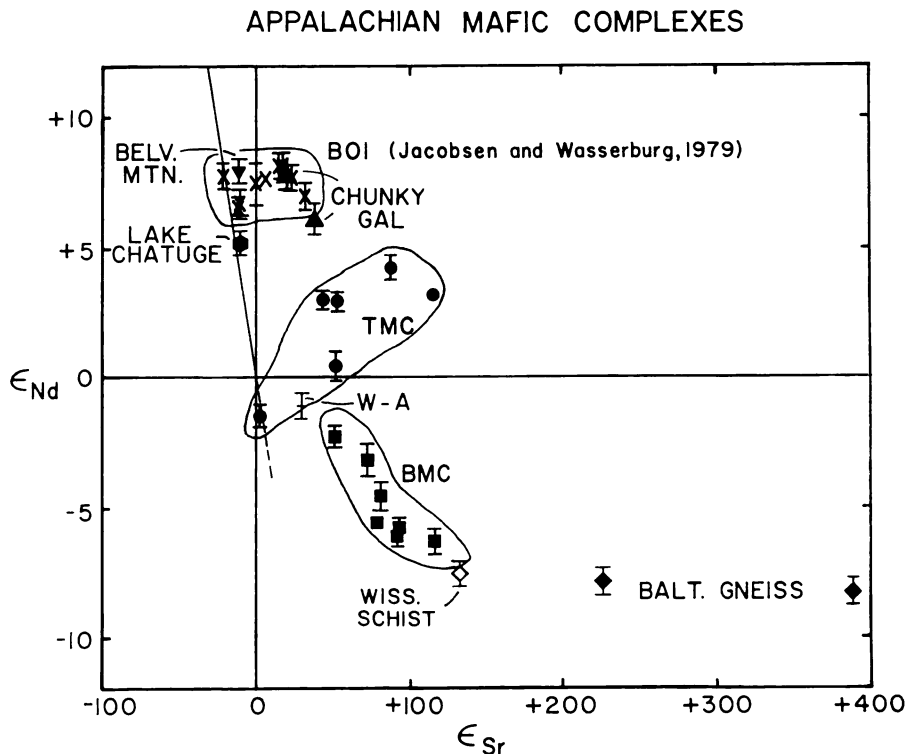
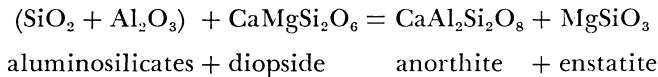


Fig. 6. $\epsilon_{Nd}(T)$ versus $\epsilon_{Sr}(T)$ diagram with data for Vermont, North Carolina, Thetford, Baltimore, and Bay of Islands (BOI) (Jacobsen and Wasserburg, 1979). Vermont, North Carolina, and Bay of Islands samples have the isotopic signature of oceanic crust. The Baltimore (BMC) samples fall on a trend suggesting contamination with continental crust. The Thetford Mines samples (TMC) do not show any obvious trend. Compare with figure 2.

due to Bowen (1928) that assimilation of aluminous rocks by basaltic magma should bring orthopyroxene onto the liquidus and increase the anorthite content of the crystallizing plagioclase at the expense of clinopyroxene crystallization according to the generalized reaction:



More recent experimental work by Irvine (1975) has substantiated Bowen's observation. Both the early appearance of orthopyroxene in the crystallization sequence and high anorthite content of the plagioclase are characteristics of the BMC. Taken to its extreme, this mechanism should produce norites. We note that Md-32, a norite sample, is also the most enriched in incompatible elements of the samples analyzed, as evidenced by the presence of zircon and apatite. This sample also has the highest $\epsilon_{\text{Sr}}(\text{T})$ and lowest $\epsilon_{\text{Nd}}(\text{T})$ and would thus represent the most contaminated sample we have analyzed. The T_{CHUR} model age for this sample is 1.2 AE, close to that calculated for the samples of Baltimore Gneiss (~ 1.26 AE), suggesting that assimilation of the Baltimore Gneiss or a rock with similar isotopic characteristics caused the variation in initial isotopic composition of the BMC. A calculation for simple mixing between a magma with $\epsilon_{\text{Nd}} = \epsilon_{\text{Sr}} = 0$, and 5 ppm Nd, 100 ppm Sr and a contaminant with $\epsilon_{\text{Nd}} = -8$, $\epsilon_{\text{Sr}} = +300$, 50 ppm Nd and 200 ppm Sr shows that > 25 percent bulk assimilation is necessary to account for the most contaminated samples. This is a rather large amount. This calculation is clearly an oversimplification but serves to give a sense of the magnitude of the effect. A more realistic calculation would include the effects of combined assimilation/fractionation and selective melting in the contaminant as well as incorporating other isotopic and trace element constraints. In the absence of information on liquid compositions during the evolution of the BMC magma, however, such a detailed calculation is not warranted. Furthermore, the lack of correlation between isotopic composition and rock type or position in the crystallization sequence suggests that the BMC magma chamber was chemically and isotopically heterogeneous and that the magma was therefore not well-mixed, thus this basic assumption of most mixing models cannot be justified in the case of the BMC.

Having documented the probable role of continental crustal material in the genesis of the BMC, we now ask in what tectonic settings such a process might operate. From the preceding arguments, it appears that the complex did not form at an oceanic spreading center and is therefore not an ophiolite, if that term is considered to mean obducted oceanic crust. The remaining possibilities include oceanic island volcanism, subduction-zone related volcanism, and continental mafic intrusion of an unspecified nature (not related to a subduction zone). Oceanic island volcanism is an unlikely candidate on the basis of our isotopic data. Although oceanic island basalts are often derived from less depleted mantle sources than normal MORB or a mixture of depleted and enriched mantle, the large negative $\epsilon_{\text{Nd}}(\text{T})$ together with large positive $\epsilon_{\text{Sr}}(\text{T})$ characteristic of the

BMC have not been found among oceanic island rocks. The most enriched oceanic island alkali basalts measured to date have $\epsilon_{\text{Nd}} = -5.7$, $\epsilon_{\text{Sr}} = +17$ (Dosso and Murthy, 1980), while the BMC has $\epsilon_{\text{Sr}}(\text{T}) \sim +100$ for $\epsilon_{\text{Nd}} \sim -6$. Subduction related magmas have two mechanisms whereby interaction with continental crustal material may occur: by subduction of continent-derived sediments and by intrusion into continental crust. Although subduction and subsequent melting of sediments have often been called upon to explain some features of andesitic magmas and are probably an important mechanism in their genesis (Brown and others, 1982), it is unlikely that sufficient sediment could be carried down with the subducting plate to produce the strong continental signature seen in the BMC. This possibility cannot be ruled out, however, as the work of Whitford (1975) and Whitford and others (1977, 1979) has shown that some andesites of the Sunda-Banda Arc, Indonesia, have strongly positive values of $\epsilon_{\text{Sr}}(0)$ and negative values of $\epsilon_{\text{Nd}}(0)$, probably due to large amounts of continental detritus from Australia being subducted beneath the arc. In general, however, only those arcs built upon thick continental crust, such as the Andes, have negative ϵ_{Nd} and positive ϵ_{Sr} similar to the Baltimore samples. Studies by DePaolo and Wasserburg (1977) and James (1982) of Andean volcanics and Nohda and Wasserburg (1981) of Japanese volcanics have also appealed to contamination by continental material during intrusion in order to explain the observed isotopic and trace element compositions. The last case, continental mafic intrusion of an unspecified nature, is no different in contamination mechanism from the case of a continental arc. Studies of the Stillwater Complex of Montana (DePaolo and Wasserburg, 1979), the Shabogamo Intrusive Suite of Labrador (Zindler, Hart, and Brooks, 1981), the Kalka Intrusion of Central Australia (Gray, Cliff, and Goode, 1981) and the Cortlandt Complex of New York (Domenick and Basu, 1982) have similarly suggested on the basis of Nd isotopic evidence that crustal contamination played a role in the genesis of these mafic intrusions. We emphasize that the negative values of $\epsilon_{\text{Nd}}(\text{T})$ for the BMC samples rule out a depleted mantle source for the complex, and that the complex is therefore not a fragment of typical oceanic crust. We suggest, instead, on the basis of our isotopic data, that the BMC was intruded into continental crust, possibly as part of a Cambro-Ordovician arc complex, and was contaminated by old LREE enriched crustal material with $\epsilon_{\text{Nd}} \ll 0$ during emplacement. Its present allochthonous nature (Crowley, 1976) must be due to post-crystallization tectonic movements not related to obduction of oceanic crust.

We would like to draw attention to our Nd results on the Baltimore Gneiss samples. In particular, we note that the T_{CHUR} model ages (DePaolo and Wasserburg, 1976a) appear to reflect accurately the time of crustal formation. Previous isotopic work on the gneiss has pointed up the complex history of the basement complex: K–Ar and Rb–Sr mineral systems were reset by metamorphic events in the interval 300 to 500 my ago (Wasserburg, Pettijohn, and Lipson, 1957; Wetherill, Davis, and Lee-Hu, 1968), and the U–Th–Pb systematics of zircons show complex patterns of discordance due to episodic and continuous lead loss as well as possible inheri-

tance of older Pb (Tilton and others, 1958; Tilton, Doe, and Hopson, 1970; Grauert, 1974). The T_{CHUR} and, to a lesser extent, the T_{UR} model ages appear to be able to see past these later disturbances to the original time of crustal formation of the Grenville Province at ~ 1.25 to 1.30 AE.

Our Sr and Nd isotopic data for the Thetford Mines complex are plotted in figure 6. As can be seen in this figure, there is no clear correlation between $\epsilon_{\text{Nd}}(T)$ and $\epsilon_{\text{Sr}}(T)$. The data scatter from $\epsilon_{\text{Nd}}(T) = -1.5$ to $+4.2$ and $\epsilon_{\text{Sr}}(T) = +2.6$ to $+114.0$ (0.7041 to 0.7119). Like the BMC, the isotopic composition of the Thetford samples does not appear to correlate with rock type, position within the complex, or degree of metamorphism. In particular, note that the upper and lower pillow lava sequences are isotopically indistinguishable. The range in isotopic composition characteristic of the TMC is not the signature of oceanic crust (compare the data in fig. 6 with fig. 2). Neither can the variations in isotopic composition be generated by simple mixing of oceanic crust with old continental crustal or seawater Sr and Nd. More complicated processes appear to be involved. The TMC is unquestionably an ophiolite *sensu stricto* in that the complex includes all the elements of the ophiolite stratigraphy. It thus appears that the isotopic and field geologic relationships are in opposition. We do not believe that metamorphism is the primary cause of the variations in isotopic composition in the Nd system. Ophiolites with similar metamorphic histories such as the Josephine Ophiolite and the Kings-Kaweah Ophiolite of California, which have metamorphic assemblages similar to the TMC, preserve the Nd signature and, to a lesser extent, the Sr isotopic signature of oceanic crust (table 1) (Chen and Shaw, 1982). Furthermore, as will be shown below, relatively high-grade amphibolites from North Carolina and Vermont also preserve the Nd isotopic characteristics of their depleted mantle source. On the other hand, the case for the observed Sr isotopic compositions being primary is rather poor. Sr isotopic exchange during metamorphism is a well-documented process (Lanphere and others, 1963, and many later studies), and interaction with metamorphic fluids derived from dehydration reactions in metasedimentary rocks will generally increase the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio in mantle-derived rocks. Unlike the Baltimore samples, there is no apparent correlation between $\epsilon_{\text{Nd}}(T)$ and $\epsilon_{\text{Sr}}(T)$, suggesting that these systems are decoupled. Several of the samples analyzed contained secondary calcite which might be expected to have a large effect on the Sr isotopic composition of the bulk rock, although the presence of calcite does not appear to correlate with $\epsilon_{\text{Sr}}(T)$ or $\epsilon_{\text{Sr}}(0)$. In addition, the samples have quite a wide range of Rb/Sr ratios, possibly indicating rather large Rb–Sr fractionation during metamorphism. In short, the Sr isotopic system cannot be counted upon to preserve primary isotopic composition in complex metamorphic terrains, and we shall therefore concentrate our discussion on the significance of the Nd data.

Although the isotopic composition of the TMC is unusual for an ophiolite, it is not unique. Recent studies of the Vourinos Complex, Greece (Noiret, Montigny, and Allegre, 1981; Richard and Allegre, 1980), the Troodos Ophiolite, Cyprus (Hannah and Futa, 1982), and the Bett's

Cove Ophiolite, Newfoundland (Coish, Hickey, and Frey, 1982) have shown that, in part, they are also characterized by rocks with $\epsilon_{\text{Nd}}(\text{T})$ between 0 and +4 and with relatively radiogenic Sr, although the latter is usually attributed to interaction with seawater. Both the Troodos and Bett's Cove ophiolites were also found in these studies to include rocks with a more normal oceanic signature of $\epsilon_{\text{Nd}} \sim +8$, a characteristic not seen in the Thetford samples we have analyzed. The significance of these results for ophiolite genesis is not clear. The implication is that there are ophiolites of varying ages derived from a mantle source with modestly positive ϵ_{Nd} or from a blend of magmas derived from normal depleted oceanic mantle and an enriched reservoir such as enriched mantle, undepleted mantle, or continental crust. These characteristics have been attributed to the source for many oceanic island basalts (O'Nions, Hamilton, and Even- sen, 1977; Dosso and Murthy, 1980). A mixture of sources might also be expected for arcs and the magmas intruded during the initial stages of continental rifting to form an ocean basin or marginal basin. The latter cases would also allow for interaction of the mantle-derived magma with continental crust.

Church (1977) has pointed out that the TMC is chemically similar to the Bett's Cove ophiolite of Newfoundland (Coish and Church, 1979) in that they are characterized by the presence of orthopyroxene-bearing cumulates and relatively low Ti-basalts, among other features. He contrasts these complexes with ophiolites characterized by dunite-troctolite cumulate sequences and relatively Ti-rich basalts with olivine and plagioclase phenocrysts such as the Bay of Islands Ophiolite. He further identifies these two endmembers with the ophiolites of the internal and external zones, respectively, of the Tethys region (Rocci, Ohnenstetter, and Ohnenstetter, 1975). Our isotopic data together with the previously mentioned studies seem to indicate a fundamental difference in origin between the two endmembers as defined by Church, but it is unclear what this difference signifies. Clearly, more work is needed to establish the links between the variations in major and minor element geochemistry and isotopic systematics of these ophiolites and to attempt to relate these to modern analogues. Among modern rocks, these complexes have isotopic and chemical affinities to boninites, which are olivine, orthopyroxene, and clinopyroxene-phyric island arc basalts with high MgO and SiO₂ but low TiO₂ contents (Kuroda and Shiraki, 1975; Shiraki and Kuroda, 1977). Recent Nd isotopic data obtained by Hickey and Frey (1982) has shown that boninites, like the Thetford Mines samples, are characterized by a large range in $\epsilon_{\text{Nd}}(0)$ (−0.3 to +6.2) which they attribute to mixing between a strongly depleted mantle peridotite and an enriched component, either subducted sediments or metasomatized, LREE enriched mantle. We suggest that the TMC may represent the ancient intrusive and extrusive counterpart to modern boninites.

In summary, our data on the TMC are not easily reconciled with the geologic relationships. At face value, they preclude the formation of the complex at a normal mid-ocean spreading center. Nevertheless, it is possi-

ble that the metamorphism and alteration were so severe that all primary isotopic memory has been erased.

Our isotopic results for the North Carolina samples are plotted in figure 6. In contrast to the results discussed thus far, the Chunky Gal and Lake Chatuge amphibolites clearly have the isotopic signature of ancient oceanic crust. In figure 6, these data generally fall within the field defined by the Bay of Islands complex (Jacobsen and Wasserburg, 1979), although the Lake Chatuge sample has a somewhat lower value of $\epsilon_{Nd}(T)$ than the other samples. Similarly, the sample of Hazen's Notch amphibolite from Vermont also plots within this field. These samples all have characteristically oceanic values of $\epsilon_{Nd}(T) > +5$ for $T = 500$ my but extend from the mantle array to more radiogenic values of $\epsilon_{Sr}(T)$. The dispersion in $\epsilon_{Sr}(T)$ may be due to either hydrothermal interaction with seawater or continental metamorphic fluids or both. Previous Sr isotopic work by Stueber (1969) and Jones, Hartley, and Walker (1973) showed that the Sr isotopic composition of ultramafic and mafic rocks from North Carolina was in some cases considerably more radiogenic than plausible seawater values. In addition, these studies suggested that the increase in $^{87}Sr/^{86}Sr$ ratios was correlated with the degree of metamorphism; samples of unmetamorphosed troctolite and dunite had lower isotopic ratios than their metamorphic equivalents, amphibolite and serpentinite. If true, this would clearly indicate exchange of Sr during metamorphism. It is important to note, however, as has been observed in other studies of ophiolite complexes (Jacobsen and Wasserburg, 1979; McCulloch and others, 1981; Chen and Shaw, 1982), that even though the Sr system shows evidence for substantial post-crystallization isotopic exchange, the Nd system appears to preserve the primary isotopic composition, even at the relatively high grade of metamorphism represented by the Vermont and North Carolina amphibolites. This fact encourages our belief that the Nd isotopic variations observed in the BMC and TMC are primary and not due to later exchange processes.

Our results on the Vermont and North Carolina amphibolite samples are significant in that they demonstrate that there are mafic rocks in the Appalachians that are derived from a depleted mantle source and that, at least on isotopic grounds, are good candidates for being fragments of oceanic crust.

The Webster-Addie sample with $\epsilon_{Nd}(T) \sim -1$, $\epsilon_{Sr}(T) \sim +29$ is clearly different from the nearby amphibolites and does not have the signature of the depleted mantle source typical of oceanic rocks. The leaching experiment shows that the low $\epsilon_{Nd}(T)$ is not due to the presence of Nd derived from surrounding metamorphic rocks on grain boundaries. It is difficult to draw any firm conclusions about the origin of the Webster-Addie body on the basis of this single sample; however, the isotopic data do not support an origin at a mid-ocean spreading center.

5. CONCLUSIONS

From Sr and Nd isotopic analyses it appears that there is no single origin for the mafic and ultramafic rocks of the Appalachians. The isoto-

pic data support the earlier notions of ensialic mafic intrusions in the case of the Baltimore Mafic Complex, while clearly supporting an oceanic origin for the Bay of Islands Ophiolite, the Belvidere Mountain Complex, and some of the North Carolina bodies. The Thetford Mines Complex appears to belong to a class of ophiolites with peculiar isotopic, and possibly chemical, characteristics, which have affinities to modern-day boninitic rocks found in oceanic arcs. The isotopic data would tend to rule out the formation of these complexes at a mid-ocean spreading center. These conclusions are critically dependent on the observation that the Nd isotopic system is relatively immune to disturbance during metamorphism and thus preserves the original igneous isotopic signature. Nevertheless, even if the observed isotopic systematics represent a metamorphic effect, we find no supportive evidence for an oceanic crustal origin for the BMC, TMC, or the Webster-Addie body. In contrast, the Sr isotopic system alone is relatively undiagnostic as to the origin of rocks in metamorphic terranes due to the relative ease with which Rb and Sr are mobilized during metamorphism. The Nd isotopic system can readily be exploited in favorable cases to obtain ages on mafic rocks. When applied to samples of the BMC, this technique yields an age of 490 ± 20 my for this body which we interpret as the time of crystallization. Old continental crust has a distinctive isotopic signature, and combined Nd-Sr isotopic studies provide a sensitive test for the presence of assimilated crust in igneous bodies. Our results on the BMC indicate that a significant fraction of the Nd and Sr in the Complex was derived from old continental material such as the Baltimore Gneiss.

In summary, we believe that studies of Appalachian geology have reached a point where the diversity in origins for the mafic and ultramafic rocks of this mountain belt must be recognized and explained. It would appear that simple models involving only obduction of ridge-generated oceanic crust are inadequate to explain the observations. Finally, we caution as others have that not every mafic and ultramafic rock in an orogenic belt is part of an ophiolite nor are all ophiolites formed at mid-ocean spreading centers.

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APPENDIX 1

Sample descriptions and locations

(Percentages are visual estimates and intended only as a guide to modal abundances)

Baltimore Mafic Complex

Md-32—Norite: Plagioclase, 80 percent; Opx, 14 percent; Magnetite, 5 percent; Sulfide, 1 percent; trace, apatite quartz, green trem-actinolite, chlorite.
Conowingo Dam Quad., Maryland-Pennsylvania; road from Oakwood to Susquehanna River in ravine of Conowingo Creek. 39°41.26'N, 76°11.35'W.

Md-3—Feldspathic websterite/two pyroxene gabbro: 40 percent Opx. with Cpx. exsolution lamellae, 40 percent; Cpx. with opx. exsolution lamellae, 35 percent plagioclase, 25 percent; trace, opaques, green trem-actinolite.
Baltimore West Quad., Maryland; Purnell Drive, 39°19.12'N, 76°47.63'W.

Md-40—Metagabbro: Pale green fibrous trem-actinolite after pyroxene 50 percent; plagioclase, 30 percent; cpx., 15 percent; chlorite, 3 percent; zoisite + epidote, 1 percent; opaques, 1 percent; trace, carbonate, chlorite.
Conowingo Dam Quad., Maryland-Pennsylvanian; railroad cut along Susquehanna River, 39°40.43'N, 76°11.13'W.

Md-15—Feldspathic orthopyroxenite: Orthopyroxene, 90 percent; plagioclase, 7 percent; cpx., 2 percent; green trem-actinolite, 1 percent; trace, opaques.
Conowingo Dam Quad. Maryland-Pennsylvania; road from Oakwood to Pilot, 39°41.23'N, 76°10.93'W.

Sample from B. Morgan, his #77-BM-65.

Md-39—Hornblende gabbro: Plagioclase, 40 percent; cpx., 20 percent; opx., 15 percent; intercumulus brown amphibole, 10 percent; fibrous green amphibole + ? after pyroxene, 10 percent; pale green trem-actinolite, 3 percent; epidote, 1 percent; opaques, 1 percent.
Conowingo Dam Quad, Maryland-Pennsylvania; railroad cut along Susquehanna River, 39°40.34'N, 76°10.97'W.

Md-30—Two-pyroxene gabbro: Plagioclase, 45 percent; opx., 20 percent; cpx., 15 percent; fibrous pale green amphibole + ? after pyroxene, 15 percent; zoisite, 2 percent; chlorite, 2 percent; opaques, 1 percent.
Conowingo Dam Quad. Maryland-Pennsylvania; road from Oakwood to Susquehanna River, 39°41.74'N, 76°11.95'W.

Md-44—Two-pyroxene gabbro: Plagioclase, 60 percent; opx., 20 percent; cpx., 20 percent; trace green trem-actinolite, opaques.
Conowingo Dam Quad. Maryland-Pennsylvania; railroad cut along Susquehanna River, 39°41.44'N, 76°12.30'W.

Baltimore gneiss, Wissahickon schist

Md-7—Layered biotite gneiss: Quartz, 40 percent; plagioclase, 30 percent; microcline, 20 percent; biotite, 10 percent; trace, apatite, zircon, sphene.
Ellicott City Quad, Maryland; bluff above railroad tracks along Patapsco River west of Woodstock, 39°19.87'N, 76°52.45'W. Near Woodstock Dome locality sampled by Tilton, Doe, and Hopson (1970).

Md-11—Augen gneiss: Quartz, 30 percent; microcline, 28 percent; plagioclase, 25 percent; biotite, 15 percent; muscovite, 1 percent; sphene, 1 percent; trace, epidote, allanite, zircon, apatite, tourmaline.
White Marsh Quad. Maryland, Hartley Mill Road in Long Green Creek near Crowley's (1976) type locality. Townsend Dome. 39°27.02'N, 76°28.54'W.

RL 80-7—Wissahickon schist: Quartz, 40 percent; plagioclase, 35 percent; biotite, 10 percent; epidote + zoisite, 10 percent; muscovite, 5 percent; chlorite, 1 percent; trace, opaques, apatite, zircon.
Conowingo Dam Quad. Maryland-Pennsylvania; route 299 northwest of Harmony Chapel, 39°39.0'N, 76°8.3'W. Sample from K. Sinha.

Thetford Mines Complex

Qe-40—Upper Pillow lava (greenschist): Fine-grained chlorite + amphibole, 40 percent; calcite, 30 percent; epidote, 20 percent; quartz + plagioclase, 9 percent; coarse euhedral green amphibole, 1 percent; trace, opaques.
Disraeli Sheet, Quebec; route 34 north of Lac Coulombe, 45°50.70'N, 71°27.63'W.

Qe-13A #3—Lower Pillow lava (greenschist): Chlorite, 60 percent; amphibole after cpx. (?) microphenocrysts, 39 percent; opaques, 1 percent; trace, calcite, zoisite. Thetford Mines Sheet, Quebec; 0.27 km southeast of Colline Poudrier, 46°0.72'N, 71°14.14'W.

Qe-18—Lower Pillow lava (greenschist): Fine-grained chlorite + amphibole, 55 percent; quartz + plagioclase, 40 percent; calcite (in vesicles), 5 percent; trace, zoisite, opaques.

Thetford Mines Sheet, Quebec; on road to Mt. Adstock, 46°1.10'N, 71°12.8'W.

Qe-12A—Metagabbro, screen in dike complex: Amphibole, 55 percent; Chlorite, 20 percent; zoisite + ? after plagioclase (?), 15 percent; quartz, 10 percent; trace, opaques, cpx. (relict).

Thetford Mines Sheet, Quebec; dike and sill complex at Colline Poudrier, 46°0.83'N, 71°14.65'W.

Qe-8B—Metadiabase: Amphibole, 50 percent; plagioclase, 45 percent; chlorite, 2 percent; cpx. phenocrysts (relict), 2 percent; zoisite, 1 percent.

Same location as Qe 12A.

Qe-30A—Metagabbro: Dark, fine-grained alteration products of plagioclase [= hydrogrossular (?) + zoisite + clays (?)], 75 percent; amphibole after cpx., 20 percent; chlorite, 5 percent; trace, zoisite, opaques, cpx. (relict).

Disraeli Sheet, Quebec; road along northeast end of the northwest arm of Breeches Lake, 45°55.28'N, 71°27.92'W.

Qe-24—Clinopyroxenite: cpx., 75 percent; amphibole after cpx., 15 percent; serpentine after olivine (?), 7 percent; opaques, 3 percent.

Disraeli Sheet, Quebec; between Lac de l'Est and Colline Lemay, 45°56.89'N, 71°24.36'W.

North Carolina Samples

CG-12-1—Chunky Gal Amphibolite: Amphibole, 50 percent; plagioclase, 48 percent; opaques, 2 percent; trace, zircon, tourmaline.

Rainbow Springs Quad., North Carolina; 35°4.8'N, 83°35.4'W, see McElhaney and McSween (1983). Sample from H. McSween.

CG-2-18—Chunky Gal Amphibolite: Amphibole, 75 percent; plagioclase, 20 percent; zoisite, 5 percent; trace, zircon.

Shooting Creek Quadrangle, North Carolina; 35°5.0'N, 83°38.0'W. See McElhaney and McSween (1983). Sample from H. McSween.

C-1—Amphibolite from Lake Chatuge ultramafic body: Amphibole, 60 percent; plagioclase, 40 percent; trace, epidote, zircon, opaques.

Sample from J. Paque.

XTC-127—Websterite, webster-addie ultramafic body: cpx., 85 percent; opx., 15 percent; trace, green amphibole, spinel, opaques.

Sample from K. Misra.

Vermont Sample

VJL-360C—Hazen's Notch Amphibolite: Blue-green barroisitic amphibole, 50 percent; epidote, 25 percent; quartz, 20 percent; garnet + chlorite after garnet, 5 percent; trace, rutile, zircon, apatite, plagioclase.

Jay Peak Quad., Vermont. Near lookout tower on Belvidere Mountain, 44°46.43'N, 72°33.06'W. See Laird (ms), Laird and Albee (1981). Sample from J. Laird.

California Samples [Chen and Shaw, 1982]

P-59—Cumulate gabbro: Plagioclase, 80 percent; cpx., 10 percent; fibrous trem.-actinolite, 5 percent; serpentine, 3 percent; brown amphibole, 2 percent; trace, chlorite, opaques.

Point Sal Ophiolite. Sample from S. Jacobsen.

JP-042—Cumulate olivine gabbro: Sericitized plagioclase, 65 percent; olivine, 10 percent; cpx., 10 percent; serpentine + magnetite after olivine, 10 percent; fibrous trem.-actinolite, 5 percent; trace, green spinel.

Oregon Mountain, Josephine Ophiolite. Sample from S. Jacobsen.

Y-13b—Cumulate gabbro: Sericitized plagioclase, 70 percent; cpx., 25 percent; chlorite + serpentine, 5 percent; trace, amphibole, opaques.

Josephine Ophiolite. Sample from G. Harper.

Ca-10—Metagabbro: Amphibole after cpx., 70 percent; plagioclase, 25 percent; cpx., 5 percent; trace, opaques, serpentine, chlorite, zoisite.
Kings-Kaweah Ophiolite, 45°4.6'N, 119°24.5'W.

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