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Conflicting mineral and whole-rock isochron ages from the Late-Archaean Lewisian Complex of northwestern Scotland: Implications for geochronology in polymetamorphic high-grade terrains

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Abstract—We report major and trace element (including rare-earth element, REE) geochemistry and whole-rock Sm-Nd geochronology on two suites of mafic/ultramafic inclusions in an agmatite complex within a trondhjemitic host at Gruinard Bay, in the Late Archaean Lewisian Complex of northwest Scotland. A suite of medium-grained equigranular amphibolites (main amphibolite suite, MAS) has broadly chondritic high field strength element (HFSE) ratios (Ti/Zr ca. 115, Ti/Y ca. 318, Y/Zr ca. 0.39), flat heavy REE patterns at five to fifteen times chondrites, usually with mild light REE enrichment ($La_N < 50$) and small Eu anomalies. This is consistent with an origin as basaltic liquids which may represent suitable mafic precursors for tonalite-trondhjemite-granodiorite (TTG) suite magmas, as argued previously by Rollinson and Fowler (1987). A Sm-Nd regression line (MSWD = 3.4) obtained from the MAS yields an age of 2.943 ± 0.091 Ga with an initial ϵ_{Nd} value of +2.69 ± 0.62. A suite of coarser grained hornblendites and metagabbros (hornblendite-metagabbro suite, HMS) displays significant geochemical differences from the MAS, exemplified by significantly light REE-enriched patterns in which the hornblendites have the highest total REE with $La_N/Nd_N < 1$ and large negative Eu anomalies, and the most felsic metagabbros have low total REE and large positive Eu anomalies. Samples from the HMS yield a Sm-Nd isochron (MSWD = 0.52) age of 2.846 ± 0.073 Ga with an initial ϵ_{Nd} value of +1.10 ± 0.72. The HMS geochemistry is consistent with an origin as amphibole and plagioclase cumulates to the local TTG magmas. Combining the new data with previously published data from the host trondhjemites at Gruinard Bay (Whitehouse, 1989a) yields a Sm-Nd isochron (MSWD = 1.09) age of 2.795 ± 0.028 Ga with an initial ϵ_{Nd} value of +0.66 ± 0.38. The relatively low initial ϵ_{Nd} values for the HMS and combined HMS-trondhjemite isochrons compared to contemporaneous depleted mantle ($\epsilon_{Nd}(2.8)$ ca. +2.1 to +3.3) is consistent with derivation of a parental magma from light-REE-enriched components of a mafic (basaltic) precursor which was itself derived from a depleted mantle source approximately 100–200 Ma previously. The MAS suite at Gruinard Bay may be representative of such a mafic precursor.

Our new geochronological data differ significantly from the ca. 3.3 Ga mineral Sm-Nd and Pb-Pb regression ages presented recently for two members of the HMS by Burton et al. (1994) which, if correct, require radical revision of the early geological history of the Lewisian Complex. Comparison of the relatively old mineral ages with our new whole-rock age contradicts the conventional assumption that mineral isotope systematics are less robust to later disturbance than whole-rock systematics. However, we argue that extensive retrogression, evident in all lithologies at Gruinard Bay, makes them less than ideal samples for a mineral isotopic study and it is likely that the ca. 3.3 Ga ages may be spurious, the product of disturbance of isotope systematics during the Laxfordian. On the basis of the geochronology presented here, evolution of the Lewisian closely follows generally accepted models for late-Archaean crust formation.

1. INTRODUCTION

The isochron method applied to suites of whole-rock samples and/or separated minerals can provide valuable geochronological information from terrains which have experienced complex polyphase development. The small distances required for isotopic exchange and re-equilibration in mineral sub-systems result in mineral isochron ages that most likely represent post-magmatic cooling and/or metamorphic recrystallisation, whilst whole-rock isochrons potentially preserve magmatic crystallisation and/or high-grade metamorphism ages (e.g., Wetherill et al., 1968). Reliable assignment of age information from any mineral or whole-rock isochron to geological events requires an assessment of the validity of the two critical isochron prerequisites, namely a common initial isotope ratio for all regressed samples (e.g., co-magmatic whole-rocks or equilibrium metamorphic minerals) and maintainance of a closed system to parent and/or daughter isotope migration during subsequent evolution. Such an assessment must take into account all relevant geological and geochemical observations.

We present Sm-Nd whole-rock data from a suite of amphibolites from the Lewisian Complex which yield a ca. 2.8 Ga isochron age that is ca. 500 Ma younger than published Pb-Pb and Sm-Nd isochron ages on separated minerals from

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samples from the same lithology (Burton et al., 1994). A second suite of structurally older amphibolites yields a ca. 3.0 Ga regression line that is also younger than the mineral ages in the first suite. This apparent reversal of the usual situation in which mineral isochron ages are younger than whole-rock isochron ages can be explained only if the mineral isochrons or the whole-rock isochrons or both are spurious. These possibilities will be examined in relation to the observed geology and geochronology of this region.

The Lewisian Complex of northwest Scotland is one of the most intensively studied late Archaean high-grade terrains. Early workers divided the mainland complex into three regions based largely upon the presence and/or deformational state of a suite of mafic (Scourie) dykes (Fig. 1a; Peach et al., 1907; Sutton and Watson, 1951). Subsequent structural and geochronological studies have led to a generally accepted geological history (summarised by Park et al., 1994) involving late-Archaean (Scourian) ca. 3.0–2.7 Ga crust generation and high-grade metamorphism, Scourie dyke intrusion at 2.4 Ga and 2.0 Ga (Heaman and Tarney, 1989), and localised reworking during a mid-Proterozoic event at ca. 1.8–1.6 Ga (Laxfordian). The central region has been largely unaffected by Laxfordian events, locally preserving granulite facies assemblages, Scourian structures and, more rarely, intrusive relationships. At Gruinard Bay in the southern part of the central region, an area of relatively low strain during Scourian high-grade and subsequent metamorphism has been preserved (Fig. 1). Spectacular agmatites are exposed in which amphibolitic enclaves were clearly disrupted and intruded by a trondhjemitic magma. Rollinson and Fowler (1987) have shown that some of the amphibolite enclaves at Gruinard Bay have appropriate geochemistry for derivation as low pressure basaltic rocks, and thus might represent potential tonalite-trondhjemite-granodiorite (TTG) precursor material. Whole-rock Sm-Nd and Pb-Pb isotopic data (Whitehouse, 1989a,b) have been used to argue that the TTG magmas at Gruinard Bay were formed during the widespread late-Archaean (Scourian) episode of crustal generation at ca. 2.9–2.7 Ga. Although the precision on these ages is relatively poor because of the small range of compositions measured, this interpretation is consistent with previous observations that the Gruinard Bay area shares a common history with the rest of the Lewisian Complex. Evidence for Badcallian high-grade metamorphism (at Gruinard Bay hornblende-granulite facies, Fowler, 1986) is preserved, which elsewhere has been estimated at ca. 2.7 Ga (Chapman and Moorbath 1977; Whitehouse 1988). Zircon dating from the Scourie area suggests an important event at ca. 2.5 Ga (Corfu et al., 1994; Kinny and Friend, 1994; Friend and Kinny, 1995) which might obscure any earlier events.

It has been argued recently, on the basis of mineral Sm-Nd and Pb-Pb regression ages, that amphibolites from Gruinard Bay are ca. 3.3 Ga old (Burton et al., 1994) and thus represent "the oldest rocks in Europe" (Galer, 1994). Burton et al. (1994) used the isotopic data from these amphibolites, together with ca. 2.4 Ga mineral regression ages obtained from host trondhjemites and previous literature data, to argue strongly that "many of the TTG lithologies in the Lewisian may have been formed by partial melting of the older amphibolitic crust." If correct, the Burton et al. (1994) observations require that "the history of crustal development in the Lewisian Complex will have to be radically rethought" (Galer, 1994).

2. FIELD RELATIONS AND PETROGRAPHY

The field relations of the rocks in the Gruinard Bay area have been extensively described (e.g., Bowes et al., 1964; Davies, 1977; Crane, 1978; Field, 1978; Rollinson and Fowler, 1987) and the observations made during this study do not change any of the major interpretations. The area is dominated by trondhjemite hosting spectacular agmatite trains which contain a range of disoriented lithologies. Amphibolite enclaves occur as disrupted blocks up to 300 m long, together with fragments of foliated tonalite which represent an earlier (i.e., pre-trondhjemite) phase of TTG magmatism. An older, rarely seen, generation of amphibolite enclaves within these foliated tonalites described by Rollinson and Fowler (1987) and Rollinson (1987) are not considered in this study. The agmatite trains were formed in pre-Scourie dyke times (Davies, 1977), and have experienced hornblende-granulite facies metamorphism, equated with the Badcallian event (Park, 1970). The rocks are now dominantly retrogressed to amphibolite facies assemblages, with the former presence of orthopyroxene and/or clinopyroxene indicated by amphibole ± quartz pseudomorphs. Relict orthopyroxene-bearing assemblages are found in quartzofeldspathic gneisses about 1.5 km north of the study area (Fig. 1b; Field, 1978). Totally pseudomorphed, but undeformed, areas of retrogressed granulites occur about 3 km to the southeast (Fig. 1b; Crane, 1978). Despite the pervasive retrogression, cryptic evidence of the hornblende-granulite facies event can be detected in the form of large ion lithophile element (LILE) depletion (Fowler, 1986), in common with other parts of the central region (Sheraton et al, 1973). Retrogression from granulite facies could have occurred during any of several later events as the fabrics are known to be composite (e.g., Crane, 1978). In the Gruinard Bay area Crane (1978) detailed development of Inverian fabrics over earlier Scourian fabrics, consistent with the mode of occurrence of kyanite in metasedimentary rocks within 2 km to the east-southeast of our sample localities (Fig. 1b; described in this paper for the first time; see Appendix for detailed geological description). Further, as the local mid-Proterozoic Scourie dykes have been converted to amphibolite, there was clearly a phase of post-dyke amphibolite facies metamorphism, probably during Laxfordian reworking (Crane, 1978; Corfu et al., 1994).

Several of the components of the agmatites have relict igneous layering, preserved in metamorphic minerals (Bowes et al., 1964; Davies, 1977). Otherwise, field relationships provide no conclusive evidence for their origin as all textures are thoroughly metamorphic. Few field relationships that clarify the relative timing of protolith materials are preserved. Fragmentation and reorientation within the agmatite trains has so confused the structural relationships that the relative timing of protolith materials must be constrained by other methods.

The basic rocks considered here have been divided into

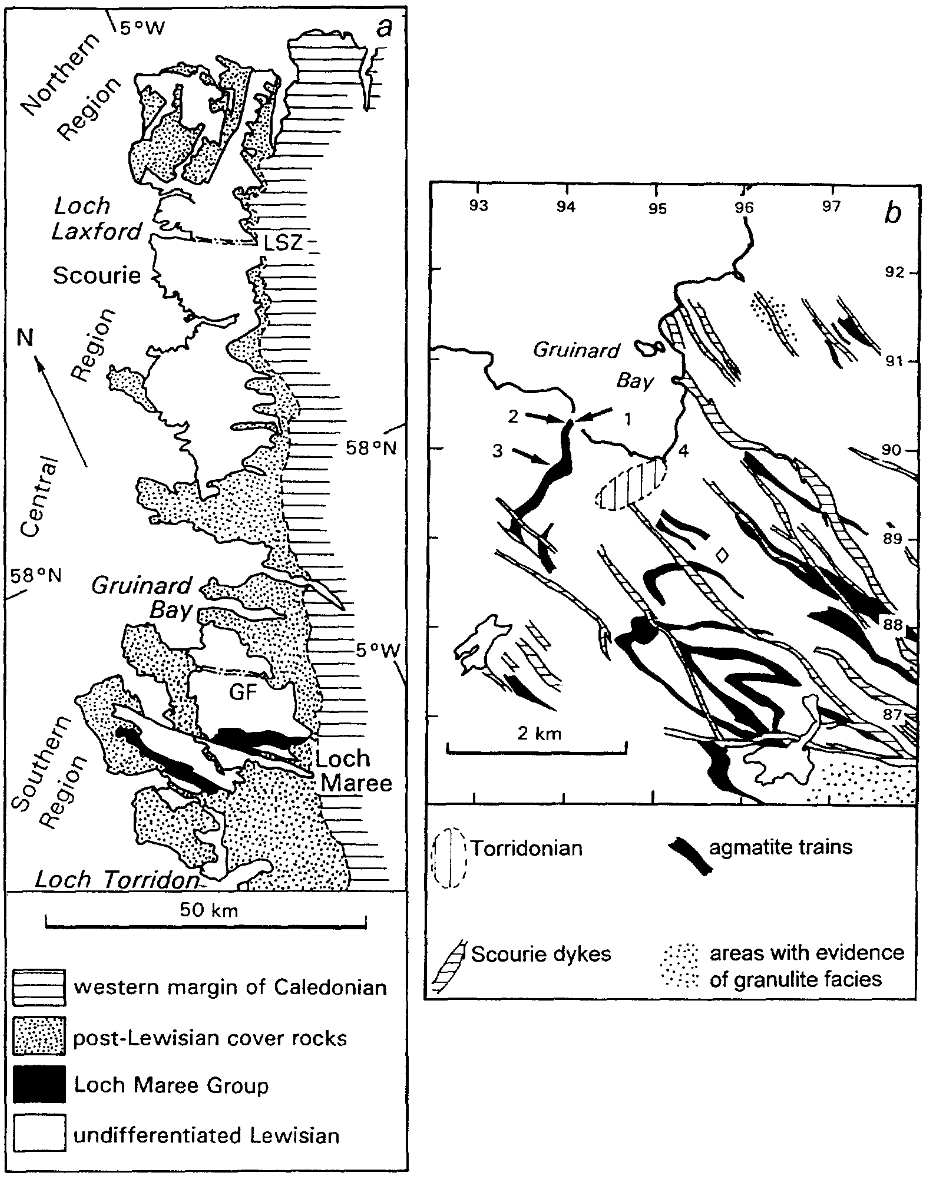

FIG. 1. (a) Simplified geological map of northwest Scotland showing subdivision of the mainland Lewisian Complex into northern, central, and southern regions. The central region is bounded to the north by the Laxford Shear Zone (LSZ) and to the south by the Gruinard Front (GF; Park and Tarney, 1987), located south of Gruinard Bay. (b) Detailed sketch map of the area immediately to the south-east of Gruinard Bay showing localities discussed in the text (after Davies, 1977). Locality 1 represents the sample site of Burton et al. (1994) and is the location for Fig. 2. The new samples of the MAS and HMS came from localities 2, 3, and 4. The diamond indicates the location of the kyanite-bearing metasediments described in the Appendix.

two groups; (1) medium-grained amphibolites and (2) medium- to coarse-grained hornblendites and metagabbros. The medium-grained amphibolites (main amphibolite suite, MAS) show considerable variation in amphibole-plagioclase proportions from >80% to ca. 30% amphibole and are sometimes banded as a result. These correspond to the amphibolites described by Rollinson and Fowler (1987) and Rollinson (1987) as potential source material for TTG magmatism. Their mineralogy is dominated by granoblastic hornblende and plagioclase with occasional opaque grains. Clinopyroxene is locally preserved and may also be represented by hornblende commonly sieved with quartz. Otherwise, the amphibole is dusted with exsolved opaques. Retrograde epidote is abundant within the plagioclase, and titanite, scapolite, calcite, and chlorite are all common. Sericitisation of plagioclase is also widespread.

The second group exposed within the agmatite trains and engulfed by trondhjemite is a suite of medium- to coarse-grained hornblendites and metagabbros (hornblendite-metagabbro suite, HMS). These have a similar amphibolite facies assemblage with edenitic hornblende dusted densely with oxides and exsolved ilmenite or titanite, clearly representing an early, higher-Ti amphibole (cf. Burton et al., 1994). Symplectic hornblende intergrowths with quartz, as in the MAS, again probably represent former orthopyroxene and/or clinopyroxene (Fig. 2a). Discrete quartz aggregates are locally present. Plagioclase contains abundant retrograde epidote (Fig. 2b), and late biotite overgrows the early fabric. Large apatites are abundant and small grains of titanite and pyrite are common. Scapolite and calcite have also been observed.

The relative age of the MAS and HMS can be observed at one critical locality, close to map reference NG940902 (locality 1 in Fig. 1b; low outcrop to north of road, approximately 50 m west of parking area). Here, a fragment of foliated, medium-grained amphibolite, equated with our MAS, is enclosed within coarse-grained unfoliated hornblendite, our HMS (Fig. 2c), the material analysed by Burton et al. (1994).

3. SAMPLING

Samples were collected from Gruinard Bay around and including the locality of Burton et al. (1994; Fig. 1b), and were chosen particularly to represent the range of amphibolites exposed. Specifically, one set of both MAS and HMS amphibolites was collected from either side of the road cut at NG941902 (the precise locality of Burton et al., 1994; Fig. 1b, locality 1). Another set of samples, including members of our MAS and HMS amphibolites, was taken from the inland extension of the agmatite train, at Meall Buidhe, where they are exposed in a ca. 50 m outcrop at NG937897 (Fig. 1b, localities 2, 3). A set of HMS samples was collected from the roadcut at NG953900 (Fig. 1b, locality 4). Additional samples of the MAS were provided by H. R. Rollinson, and these localities are documented by Rollinson (1987) and Rollinson and Fowler (1987).

4. ANALYTICAL METHODS

Elemental data were obtained by ICPAES at Oxford Brookes University, using an ARL 3510 sequential instrument, following rock dissolution by standard fusion and $HF/HClO_4$ procedures. REE were preconcentrated before analysis by standard cation exchange procedures. Natural rock standards, including certified reference materials, were used for all calibrations. Accuracy and precision of the data are estimated at 2–3% rsd for major elements and 5–10% rsd for trace elements. Chondrite normalised REE values are calculated using the reference data of Nakamura (1974).

Isotopic data were acquired at the University of Oxford. Samarium and neodymium were separated from whole-rock powders using standard dissolution and ion-exchange chromatography methods (modified after Eugster et al., 1970). Element concentrations were determined by isotope dilution using a mixed ^{149}Sm-^{150}Nd enriched tracer added prior to dissolution. Isotope ratios were determined using a VG54E single collector thermal ionisation mass spectrometer, operated by the ANALYST software of Ludwig (1992). Neodymium isotopic ratios were corrected for within-run mass fractionation by normalisation to a ^{146}Nd/^{144}Nd ratio of 0.7219; replicate analyses of La Jolla standard yielded a mean ^{143}Nd/^{144}Nd ratio of 0.511851 $\pm$ 0.000025 (0.005%, 2σ); error on ^{147}Sm/^{144}Nd ratios is ca. $\pm$0.1%. All regression calculations in this paper use a ^{147}Sm decay constant of 6.54×10^{-12} a^{-1} (Lugmair and Marti, 1978) and have been performed using ISOPLOT (Ludwig, 1991); all quoted errors are at the 95% confidence level. ϵ_{Nd} parameters are calculated relative to CHUR (^{143}Nd/^{144}Nd = 0.512638; ^{147}Sm/^{144}Nd = 0.1966; Jacobsen and Wasserburg, 1984). Depleted mantle models use the parameters of DePaolo (1981a) and DePaolo et al. (1991).

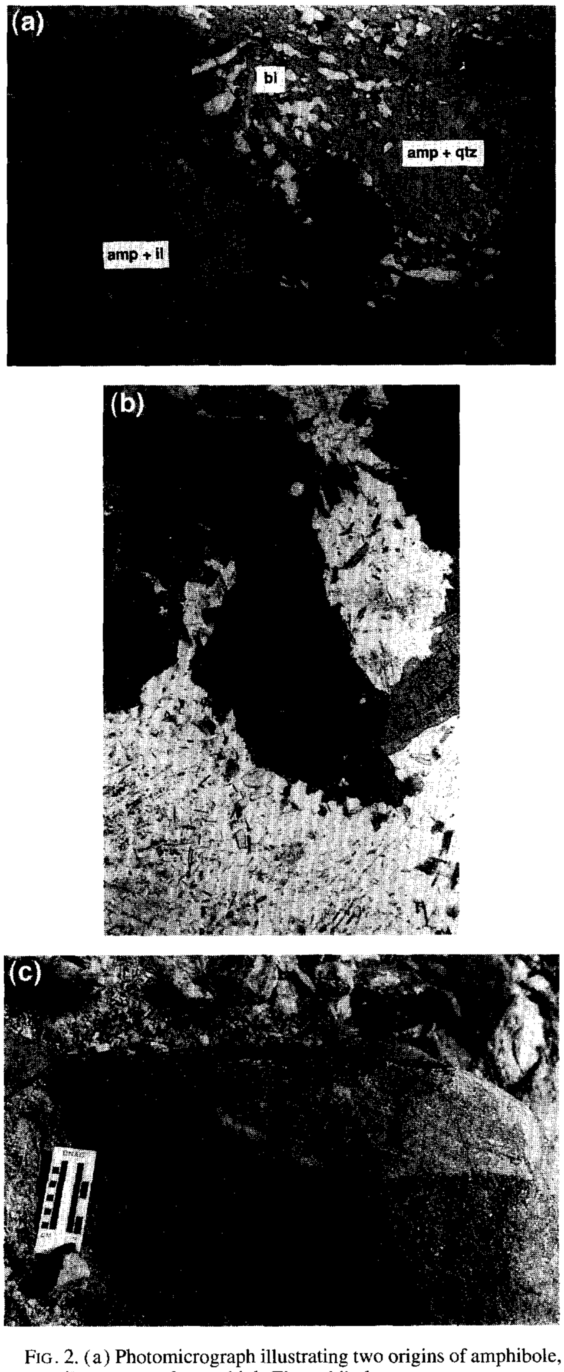

FIG. 2. (a) Photomicrograph illustrating two origins of amphibole, amp + il represents former high-Ti amphibole, amp + qtz represents amphibole after pyroxene. (b) Photomicrograph illustrating retrogressive development of epidote in plagioclase. Long axis of both photomicrographs is 4 mm. (c) Field relationship showing the relative age of the HMS (host) and MAS (inclusion). See text for description and field location.

Table 1. Representative major and trace element analytical data

	LG1 MAS	LG2 MAS	G2 MAS	UM1 HMS	UM2 HMS	UM3 HMS	G1 HMS	5B HMS	5C HMS	5E HMS	5F HMS	5G HMS	5H HMS
SiO_2	47.79	55.35	46.36	45.18	50.58	42.03	44.95	52.62	52.16	49.71	52.52	50.87	49.84
TiO_2	0.7	1.07	0.8	1.9	1.73	1.97	1.53	0.31	0.15	0.18	0.14	0.14	0.16
Al_2O_3	14.23	16.91	13.7	10.07	10.28	11.9	14.4	16.76	19.74	21.88	18.8	20.58	21.02
Fe_2O_3	11.8	7.65	13	15.56	15.63	17.54	15.1	9.36	5.9	6.68	6.34	7.17	6.16
MnO	0.19	0.19	0.22	0.19	0.18	0.22	0.15	0.16	0.11	0.11	0.12	0.13	0.12
MgO	6.3	2.42	8.09	11.91	8.59	11.54	8.16	6.59	6.28	4.29	5.89	5.05	4.92
CaO	12.54	8.98	11.78	10.35	7.93	10	8.9	8.86	8.99	10.42	8.67	8.68	8.35
Na_2O	2.21	3.8	2.66	1.2	1.66	1.62	3.26	4.04	3.82	4.34	3.88	3.72	4.05
K_2O	1.87	1.49	1.99	0.9	1.46	1.47	1.75	1.24	2.43	1.94	2.16	2.37	2.42
P_2O_5	0.09	0.11	0.09	0.57	0.23	0.27	0.23	0.15	0.1	0.08	0.09	0.07	0.08
LOI	1.53	1.38	1.3	2.17	1.94	1.86	1.49	0.89	1.09	1.05	2.12	1.47	2.26
Total	99.25	99.35	99.99	100	100.21	100.42	99.92	100.98	100.77	100.68	100.73	100.25	99.38
Trace elements (ppm)													
Ni	170	128	139	101	72	97	71	131	199	117	183	162	127
Co			43	51	54	54	48	37	27	29	33	25	27
V	284	236	271	632	500	605	450	106	54	47	53	57	64
Zn	72	79	129	183	174	197	119	109	92		101	89	
Ba	254	156	372	193	126	448	245	378	891	942	972	662	774
Sr	307	202	197	78	222	81	499	336	439	518	414	440	479
Zr	38	56	40	51	64	59	54	53	25	20	19	23	24
Y	15	17	16	51	19	39	26	12	5	7	5	7	4
La	10.7	7.7	7.2	32.6	21.3	24.3	19.6	13.2	7.9	10.7	6.26	9.36	3.84
Ce	18.8	11.1	9.8	76.4	42.6	62.3	43.3	27.4	14.2	21.2	11.3	18.2	6.4
Nd	11.2	5.1	4.1	69.3	30	53	35.6	14.8	6.51	8.9	5.2	8.29	4.16
Sm	2.4	2	1.8	16.3	6.3	11.6	7.8	2.5	1.23	1.56	1.17	1.84	1.12
Eu	0.82	0.89	0.74	3.14	1.52	2.79	2.15	0.88	0.55	0.65	0.51	0.6	0.49
Gd	1.59	1.38	1.35	16.2	6.13	11.6	7.9	1.81	0.89	0.9	1.17	1.55	0.67
Dy								2.09	0.76	1	0.8	1.12	0.62
Er	1.25	1.29	1.25	5.21	1.66	4.15	3	1.17	0.56	0.75		0.74	
Yb	1.79	1.95	1.81	4.1	1.69	3.58	2.35	1.11	0.44	0.62	0.42	0.85	0.29

5. GEOCHEMISTRY

5.1. Elemental Data

Representative geochemical data for both groups of amphibolites defined above are presented in Table 1. The petrography illustrates that the observed mineral assemblages experienced significant and incomplete re-equilibration to lower metamorphic grade than the peak attained in the area (hornblende-granulite facies). The high-grade event must itself have overprinted the primary mineralogy of both amphibolite groups, so that the primary geochemistry may have been masked by at least two periods of potential mobility. There is good evidence that K, Rb, U, and Th abundances were severely affected by the high-grade event (Fowler, 1986), being depleted in the same way as the classic Scourie granulites but to a lesser degree, in accord with the less extreme metamorphic conditions. Many of the major element oxides are likely to have been disturbed during retrogression; for example, the abundance of quartz and the presence of late biotite in some members of the HMS suggest that to construct petrogenetic hypotheses on the basis of SiO_2 and K_2O would be unwise. Therefore, discussion concentrates on elements and oxides such as Zr, Ti, Y, Ni, MgO, Al_2O_3, and the REE, which are regarded as robust under most relevant circumstances. The REE in particular are generally considered to be one of the least mobile element groups, and are, therefore, used below to provide the clearest petrogenetic constraints.

5.1.1. Geochemistry of the main amphibolite suite (MAS)

The MAS forms part of the Rollinson (1987) main group of olivine-normative tholeiites, with broadly chondritic high field strength element (HFSE) ratios (average Ti/Zr ca. 115 cf. 110 in chondrites (Fig. 3a), average Ti/Y ca. 318 cf. 256 (Fig. 3b) and average Y/Zr ca. 0.39, same as chondrites), relatively flat heavy REE patterns between five and fifteen times chondrites with mild light REE enrichment not exceeding $La_N = 50$, and only small Eu anomalies (Fig. 3c). Such chemistry is consistent with basic magmatism derived from a chondritic mantle reservoir (Rollinson and Fowler, 1987; Rollinson, 1987). Relatively low Mg#'s (the majority fall between 50 and 70) and Ni abundances (with one exception less than 200 ppm, Table 1) indicate that significant low pressure fractionation may have occurred prior to disruption by TTG magmas and incorporation into the agmatite trains now exposed.

5.1.2. Geochemistry of the hornblendite-metagabbro suite (HMS)

The HMS is chemically comparable with (and compositionally extends) the calc-alkaline group of Rollinson (1987). Differences from the MAS are highlighted by the REE patterns (Fig. 4a), in which the hornblendites and metagabbros form an array of subparallel, significantly light REE-enriched patterns. Within this suite of samples, two subgroups may be identified. The hornblendites have highest total REE content, $La_N/Nd_N < 1$ and a large negative Eu anomaly, the most felsic metagabbros have lowest total REE and large positive Eu anomalies, and there is a range in between these extremes. Note that the data of Burton et al. (1994) have been plotted on the diagram and are entirely consistent with hornblendites of the HMS. The consistent light and middle REE enrichment separates the HMS from the MAS, and is illustrated by Sm/Nd vs. Nd and Gd/Yb

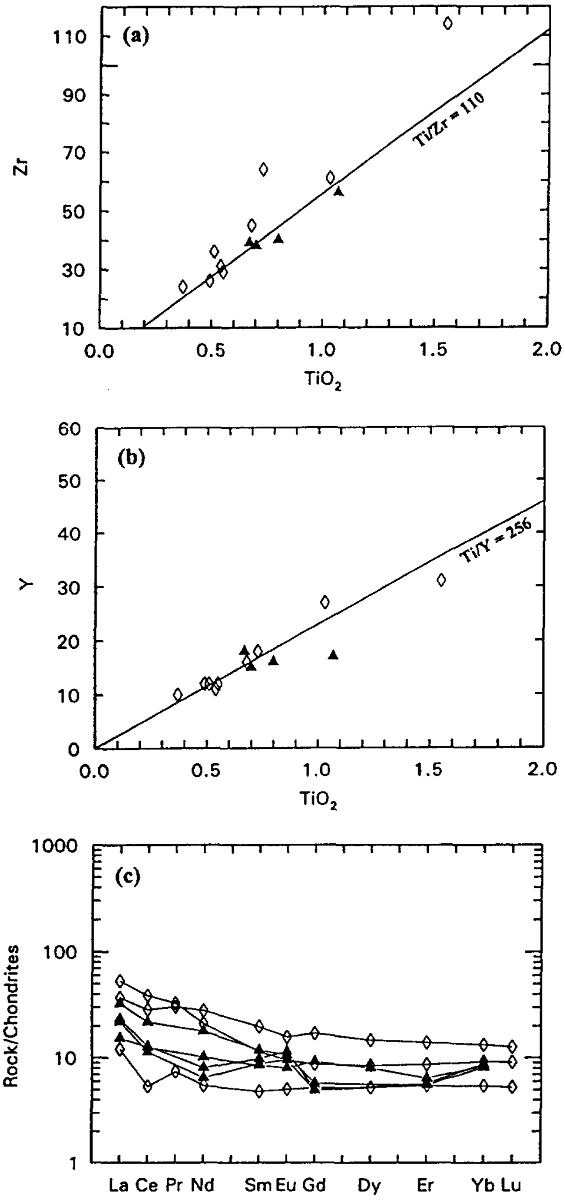

FIG. 3. Geochemistry of the MAS: (a, b) Incompatible element plots for the MAS (filled triangles) and Gruinard Bay olivine normative tholeiites (Rollinson, 1987; open diamonds). Lines represent chondritic ratios; (c) Chondrite normalised REE patterns (after Nakamura, 1974).

vs. Nd plots in Fig. 4b and 4c. Figure 4b highlights the restricted range of Sm/Nd in the HMS through a considerable range of total REE abundance, as opposed to significant variation in Sm/Nd coupled with a restricted total REE range in the MAS. The whole-rock data of Burton et al. (1994) unequivocally plot within the HMS. Figure 4c emphasises the flat HREE profile of the MAS (Gd/Yb < 2, chondritic Gd/Yb $= 1.23$), in contrast to relative middle REE enrichment in the HMS (Gd/Yb up to 4.04). These REE differences between the MAS and the HMS argue strongly for a fundamentally different origin. The overall shape of the REE patterns for the HMS resembles that of the local TTG magmas (Rollinson and Fowler, 1987), and a possible genetic link might be explored. Rollinson and Fowler (1987) dis-

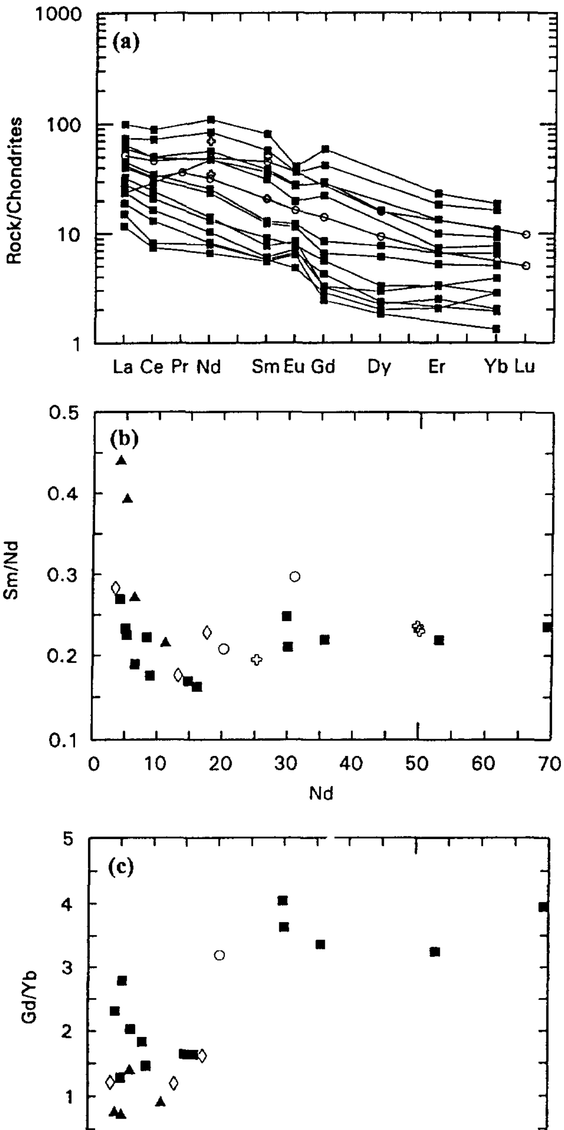

FIG. 4. Geochemistry of the HMS: (a) Chondrite normalised REE patterns for the HMS (filled squares) compared with patterns for Gruinard Bay calc-alkaline amphibolites (Rollinson, 1987; open circles) and the amphibolites of Burton et al. (1994; cross symbols); (b) Sm/Nd vs. Nd and (c) Gd/Yb vs. Nd illustrating contrasts in light and middle REE enrichment that separate MAS and HMS. Symbols as Fig. 2 and above.

cussed the geochemistry of all the TTG gneisses in the area and concluded that the major petrogenetic mechanism was wet melting of a basaltic parent, possibly represented by the MAS. They noted, however, that local amphibole and plagioclase fractionation was likely during TTG magma evolution. REE distribution coefficient data for amphibole and plagioclase in equilibrium with appropriate melt compositions are well known (Arth and Barker, 1976; Drake and Weill, 1975). Accordingly, the REE patterns of both minerals in equilibrium with examples of the local TTG magmas (data from Rollinson and Fowler, 1987) can be calculated and are plotted on Fig. 5a. The similarity of calculated patterns with measured patterns strongly suggests that the HMS represents cumulates from local TTG magmas. The Al_2O_3-rich nature of the felsic metagabbros and the abundances of transition metals and HFSE in the hornblendites are consistent with this hypothesis (Table 1). On a plot of Al_2O_3 against MgO (Fig. 5b), the HMS data scatter about a tie-line between reasonable igneous amphibole (hornblende) and plagioclase (andesine) compositions. Data for the proposed parent TTG-suite magmas (Rollinson and Fowler, 1987) are also plotted, and indicate that few samples retain large proportions of trapped melt but that the HMS may represent rather pure cumulates related to the TTG magmatism. High Sr concentrations associated with the felsic metagabbros (Fig. 5c) are consistent with this hypothesis, given the well-known substitution of Sr in the Ca^{2+} site in plagioclase.

5.2. Sm-Nd Whole-Rock Ages for the Amphibolite Groups

The two groups of amphibolites described above have been analysed for whole-rock Sm-Nd isotopic compositions. Analytical data are presented in Table 2 and in $^{143}Nd/^{144}Nd$ vs. $^{147}Sm/^{144}Nd$ diagrams (Fig. 6). The MAS yields a regression line (MSWD = 3.4) corresponding to an age of 2.943 $\pm$ 0.091 Ga with an initial ϵ_{Nd} value of +2.69 $\pm$ 0.62 (Fig. 6a; age calculated using model 3 of Ludwig (1991) in which scatter indicated by an elevated MSWD value is attributed to variation in initial ratio rather than analytical errors only). The HMS yields an isochron age (MSWD = 0.52) of 2.846 $\pm$ 0.073 Ga with an initial ϵ_{Nd} value of +1.10 $\pm$ 0.72 (Fig. 6b).

An important corrollary of our hypothesis, based upon REE geochemical evidence, that the HMS originated as cumulates from TTG magmas, is that both these suites should have a common age and initial isotopic ratio. On Fig. 6b we show the Gruinard Bay trondhjemite data from which Whitehouse (1989a) reported a regression age of 2.955 $\pm$ 0.170 Ga (MSWD = 2.3) with an initial $\epsilon_{Nd}(t)$ of +3.6 $\pm$ 2.8 (model 3 fit (Ludwig, 1991) recalculated from original data). The large errors in age and initial ϵ_{Nd} were partly attributed to a small range in Sm/Nd ratio. Some improvement in the errors can be achieved by omitting one aberrant sample (530), resulting in a regression age (MSWD = 1.2) of 2.884 $\pm$ 0.100 Ga with an initial $\epsilon_{Nd}(t)$ of +2.3 $\pm$ 1.8 (model 1 fit). The errors associated with both of these regressions overlap those of the hornblende metagabbro suite, and when the two suites are combined, a very precise regression age of 2.795 $\pm$ 0.028 Ga is obtained with an initial $\epsilon_{Nd}(t)$ of +0.66 $\pm$ 0.38 (MSWD = 1.09; model 1 fit, again omitting sample 530). In Fig. 6c we show the scatter about the best fit HMS line of all the Gruinard Bay data from the HMS and trondhjemite suite (this study; Whitehouse, 1989a; Burton et al., 1994). Clearly, this small amount of scatter is consistent with our hypothesis that the Gruinard Bay trond-

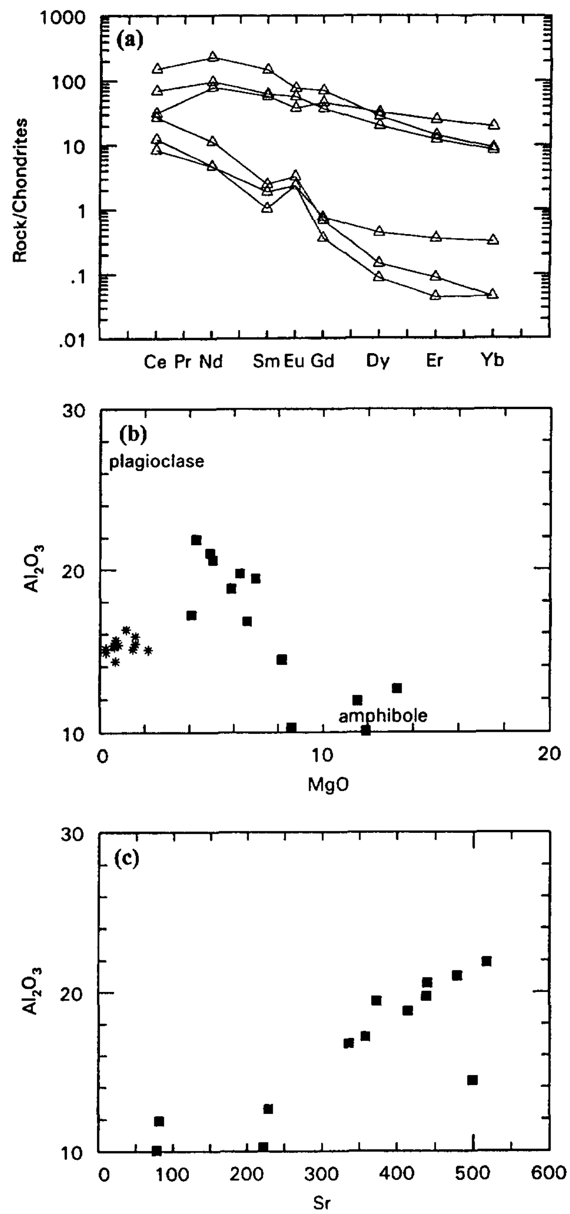

FIG. 5. Cumulate origin for HMS: (a) Calculated REE patterns of amphibole and plagioclase (triangles) in equilibrium with local TTG magmas (Rollinson and Fowler, 1987) for comparison with Fig. 4a; (b) Al_2O_3 vs. MgO showing that HMS samples plot within the triangle defined by igneous plagioclase, amphibole, and proposed trondhjemite parent; (c) Al_2O_3 vs. Sr showing strong correlation indicative of Sr partioning into proposed plagioclase cumulates.

Table 2. Sm-Nd analytical data

Sample	Sm (ppm)	Nd (ppm)	$^{143}Nd/^{144}Nd$	± 2σ	$^{147}Sm/^{144}Nd$	$\varepsilon_{Nd}(0)$	t_{CHUR} (Ga)	t_{DM} (Ga)
Main amphibolite suite (MAS)								
MBF93GB LG/1	2.421	10.30	0.511691	10	0.1421	-18.5	2.63	2.90
MBF93GB LG/2	2.109	6.927	0.512519	16	0.1841	-2.31	*n/a*	*n/a*
MBF93GB G/2	1.879	5.873	0.512710	14	0.1935	1.40	*n/a*	*n/a*
77779	4.733	21.29	0.511549	10	0.1344	-21.2	2.65	2.89
77789	1.547	4.027	0.513476	8	0.2323	16.4	*n/a*	*n/a*
77794	2.280	13.16	0.510996	13	0.1047	-32.0	2.71	2.87
77863	4.336	17.77	0.511864	12	0.1475	-15.1	2.39	2.73
Hornblendite - metagabbro suite (HMS)								
MBF93GB G/1	7.997	35.09	0.511600	9	0.1378	-20.3	2.68	2.92
MBF93GB UM/1	17.484	71.44	0.511776	10	0.1479	-16.8	2.68	2.97
MBF93GB UM/2	6.592	29.64	0.511517	9	0.1344	-21.9	2.73	2.95
MBF93GB UM/3	12.57	53.13	0.511686	10	0.1430	-18.6	2.69	2.95
MJW93GB-5B	2.545	13.61	0.511135	11	0.1130	-29.3	2.73	2.90
MJW93GB-5C	0.9908	5.333	0.511118	22	0.1123	-29.6	2.73	2.90
MJW93GB-5E	1.496	8.774	0.510927	12	0.1030	-33.4	2.77	2.92
MJW93GB-5F	0.7099	4.202	0.510924	10	0.1021	-33.4	2.75	2.90
MJW93GB-5G	1.498	8.334	0.511027	14	0.1086	-31.4	2.77	2.94
MJW93GB-5H	0.7235	3.413	0.511404	20	0.1281	-24.1	2.73	2.94

Depleted mantle model ages (t_{DM}) calculated using the model of DePaolo (1981); n/a implies that calculated values are unrealistic because of near-chondritic Sm/Nd ratios.

hjemite and hornblendite-metagabbro suites originated from a common parental magma at ca. 2.8 Ga.

6. DISCUSSION

The whole-rock Sm-Nd geochronological data which we have obtained from two sample suites which are distinct on the basis of field relationships, petrography, and geochemistry conflict with the ca. 3.3 Ga mineral isochrons presented by Burton et al. (1994) for samples which clearly belong to our ca. 2.85 Ga HMS. Resolution of this conflict rests with (1) objective assessment of the proposed ages based on the isotopic data themselves; (2) demonstration that only one of the alternatives can be correlated with the relative chronology derived from field relationships and petrography at Gruinard Bay; (3) consideration of resulting models for Lewisian evolution; and (4) identification of a mechanism for the development of spurious isochrons.

6.1. Assessment of Isochrons Using Isotopic Data

Spurious isochrons (both whole-rock and mineral) should be expected when either, or both, of the critical isochron prerequisites of common initial ratio and subsequent closed-system evolution are not achieved. Unfortunately, it is often difficult to confirm that the prerequisites have been satisfied on the basis of the isotopic data alone, and even if they have been, unambiguous assignment of the ages to magmatic or metamorphic events may be difficult. The following section gives examples of isochrons which have later been demonstrated to contravene the prerequisites, or are open to alternative interpretation as magmatic or metamorphic events.

6.1.1. Whole-rock isochrons

The common initial ratio condition can be demonstrated as invalid for a number of whole-rock examples. In some cases this is due to inhomogeneous initial ratios even within a single magmatic body. Such inhomogeneity has been demonstrated in young plutonic rocks such as the Peninsula Ranges and Sierra Nevada batholiths (DePaolo, 1981b), and is generally attributed to the effect of assimilation of older, and, therefore, potentially different isotopic composition, continental crust. In principle, such inhomogeneities should be less significant in the early evolution of Archaean terrains because of the relative paucity of significantly older crust, although the possibility of such contamination cannot always be discounted and must be considered on an individual terrain evolution basis. An example of probable contamination affecting initial ratios in late-Archaean rocks has been discussed for basalts from the Kambalda greenstone belt (Chauvel et al., 1985). In this case, a Sm-Nd whole-rock isochron age of ca. 3.2 Ga (Claoué-Long et al., 1984) has convincingly been shown to result from contamination of juvenile mafic magmas with older continental crust (or enriched mantle) at ca. 2.73 Ga. It is interesting to note, in the context of the present study, that separated minerals from the Ora Banda sill yield an isochron (Chauvel et al., 1985) within error of the intrusion age, but *younger* than the spurious whole-rock age.

Initial ratio heterogeneities may also arise by inappropriate combination on an isochron of samples that are clearly not co-magmatic (or in the case of a metamorphic isochron, have never achieved equilibrium). In the first Sm-Nd study from the Lewisian, Hamilton et al. (1979) combined a suite of TTG gneisses from both the central and northern region (granulite and amphibolite facies, respectively) together with samples from central region mafic enclaves to obtain a regression age of 2.92 ± 0.05 Ga, which was interpreted as a magmatic protolith age for the entire Lewisian Complex, approximately 200 Ma earlier than contemporary estimates for the timing of high-grade metamorphism. However, it is unlikely that the mafic and felsic components are cogenetic, and subsequent studies have revealed quite different ages and petrogeneses for the felsic components included in the Hamilton et al. (1979) regression. For the Scourie tonalites, Whitehouse (1988) reported a ca. 2.6 Ga Sm-Nd whole-rock regression age with negative $\epsilon_{Nd}(t)$ which was interpreted as metamorphic disturbance and resetting of the Sm-Nd system. Burton et al. (1994) report a similar age and initial $\epsilon_{Nd}(t)$ by combining these data with tonalite analyses from Hamilton et

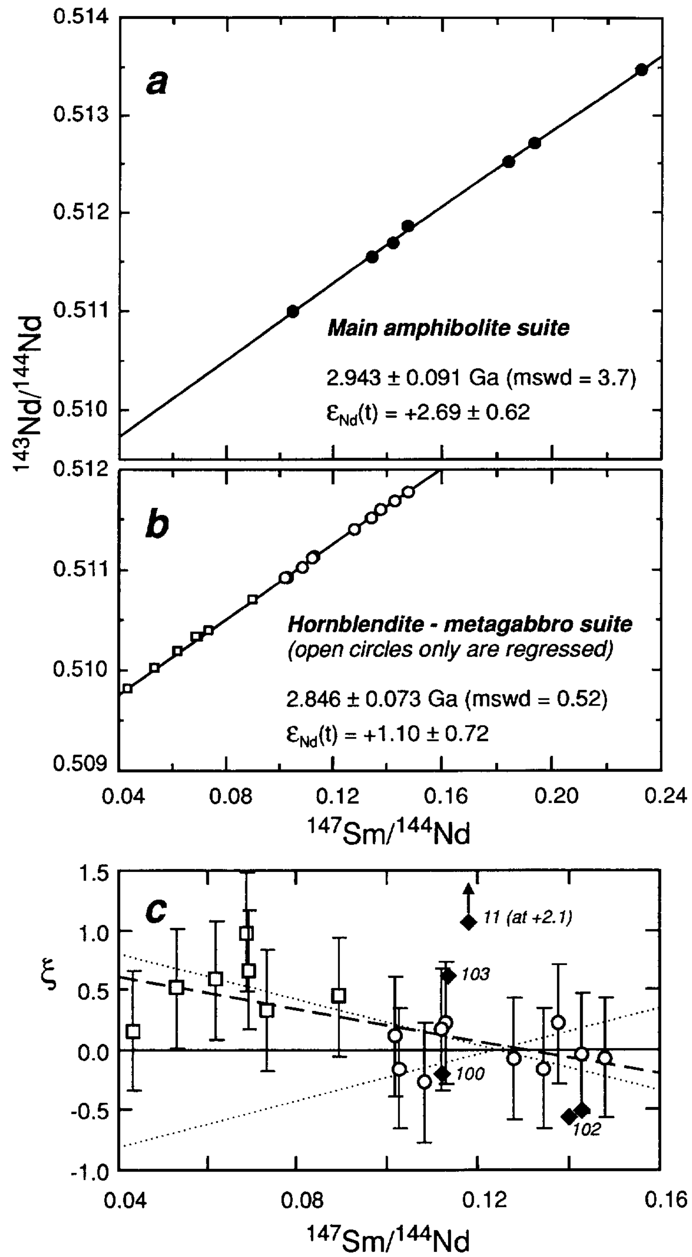

FIG. 6. $^{143}Nd/^{144}Nd$ vs. $^{147}Sm/^{144}Nd$ regression diagrams for (a) the MAS, and (b) the HMS. In both diagrams, regressed data points are shown as circles. Squares in (b) represent analyses of Gruinard Bay trondhjemites (Whitehouse, 1989a). (c) Deviation of analyses in ϵ_{Nd} units (ξ) from the best fit HMS isochron. Horizontal line at $\xi = 0$, and dotted lines ($\pm 2\sigma$); filled, labelled diamonds represent whole-rock data of Burton et al. (1994) from which error bars have been omitted for clarity; other symbols as in (b). Dashed line represents the combined isochron through the HMS and trondhjemite data presented in this study (open symbols only).

al. (1979) and Cohen et al. (1991). However, Whitehouse (1989a, 1990) demonstrated a significant difference in the neodymium isotopic evolution between the northern and central regions and suggested that these might indicate different protolith ages, an observation confirmed by recent U-Pb zircon studies which also indicate a ca. 2.5 Ga high-grade metamorphism in the central region that is not seen in the northern region (Corfu et al., 1994; Kinny and Friend, 1994).

Dating has proved particularly problematic for the layered (Sills et al., 1982) mafic/ultramafic enclaves, and illustrates the general problems inherent in assigning magmatic or

Table 3. Summary of Sm-Nd geochronology for Lewisian mafic/ultramafic bodies

Mafic/ultramafic body	Method[1]	mswd	Age (Ga)[2]	$\epsilon_{Nd}(t)$	Data source[3]
Achiltibuie	w-r (4)	0.73	2.849 ± 0.091	+2.0 ± 0.5	Whitehouse (1989a)
Drumbeg	w-r (4)	0.16	2.923 ± 0.054	+1.7 ± 0.3	Whitehouse (1989a)
Camas nam Buth	w-r (4)	5.0	2.675 ± 0.250	+0.9 ± 1.6	Whitehouse (1989a)
Geodh Eanruig	w-r (5)	9.0	2.600 ± 0.280	+2.1 ± 1.1	Cohen *et al.* (1991)
First Inlet	w-r (6)	12.9	2.836 ± 0.110	+2.7 ± 0.7	Cohen *et al.* (1991)
Combined Geodh Eanruig & First Inlet	w-r (10)	8.6	2.707 ± 0.052	+1.8 ± 0.3	Cohen *et al.* (1991)
Combined Achiltibuie, Drumbeg & First Inlet	w-r (14)	13.3	2.856 ± 0.082	+2.2 ± 0.5	Whitehouse (1989a) & Cohen *et al.* (1991)
Gruinard "amphibolite"	min. (5)	0.27	3.298 ± 0.073	+3.9 ± 0.6	Burton *et al.* (1994)
Gruinard MAS	w-r (7)	3.7	2.943 ± 0.091	+2.7 ± 0.7	this study
Gruinard HMS	w-r (10)	0.52	2.846 ± 0.073	+1.2 ± 0.7	this study

1) Whole-rock and mineral isochrons are designated w-r and min. respectively. Number of data points regressed is given in parentheses.
2) Age regressions have been recalculated using ISOPLOT (Ludwig, 1991); analytical errors reported in the original data sources have been applied.
3) Recalculated age from Camas nam Buth uses data for one sample from Whitehouse (1987); data reported by Whitehouse (1989a) contains a typographical error.

metamorphic events to whole-rock isochrons. Table 3 summarises the existing Sm-Nd geochronology, including data from Gruinard Bay for comparison. Whitehouse (1989a) published whole-rock Sm-Nd isochron ages (recalculated here) of 2.923 ± 0.054 Ga for the Drumbeg body and 2.849 ± 0.091 Ga for the Achiltibuie body. Cohen et al. (1991) presented an age of 2.71 ± 0.05 Ga for samples from two separate bodies near Scourie (First Inlet and Geodh Eanruig) and, together with Burton et al. (1995a), questioned the interpretation of the ca. 2.9 Ga whole-rock ages from the Achiltibuie and Drumbeg bodies as magmatic protolith ages on the basis that correlation between 1/[Nd] and $\epsilon_{Nd}(2.7)$ (see Fig. 7a) would be consistent with incomplete mixing between older tonalite and younger (2.7 Ga) mafic magma. However, we show below that the application of 1/[Nd] vs. $\epsilon_{Nd}(t)$ plots for determining whether an age regression might be the result of mixing is highly inappropriate in layered mafic/ultramafic complexes unless there is unambiguous evidence for a younger magmatic age than that obtained. For Kambalda, where such evidence exists (Chauvel et al., 1985; Claoué-Long et al., 1984), a correlation is seen at 2.7 Ga. However, if a range of compositions is generated from a homogeneous initial magma, the more evolved endmembers will normally be more light-REE-enriched (i.e., lower 1/[Nd] and lower Sm/Nd) than the less evolved endmembers and will, therefore, evolve to less radiogenic neodymium isotopic compositions (i.e., lower ϵ_{Nd} values), and produce a similar correlation at any later date. Two examples are given here. A correlation between low 1/[Nd] and low Sm/Nd can be seen in data from the ca. 2.7 Ga Stillwater Complex, Montana reported by DePaolo and Wasserburg (1979), where there is convincing evidence both for intrusion age and local initial isotopic homogeneity, and a correlation is produced on a plot of $\epsilon_{Nd}(t)$ vs. 1/[Nd] when an arbitrarily young age of 2.4 Ga is chosen (Fig. 7b). Similarly, data from the ca. 1.67 Ga Bridges intrusion in Labrador (Ashwal et al., 1992) shows this relationship on the scale of a single modally graded layer in which the effects of variable contamination are likely to be negligible. Therefore, we conclude that calculating $\epsilon_{Nd}(t)$ values at some arbitrary later time will *inevitably* result in a correlation between 1/[Nd] and $\epsilon_{Nd}(t)$. The technique thus cannot be used to demonstrate unambiguously that mixing has occurred to generate a spurious whole-rock isochron, particularly in the absence of convincing geochronological data for an appropriate younger magmatic event.

The preference of Cohen et al. (1991) for a 2.7 Ga magmatic age for the Lewisian mafic-ultramafic rocks is based largely upon their own composite age of 2.71 ± 0.05 Ga obtained from separate bodies at First Inlet and Geodh Eanruig. Whitehouse (1989a) also reported a poorly constrained age of 2.68 ± 0.25 Ga from a third body in the Scourie area at Camas nam Buth, but attributed the younger age and slightly lower initial ϵ_{Nd} of this body to disturbance of the Sm-Nd isotope systematics during high-grade metamorphism, a phenomenon also observed in the host tonalitic gneisses (Whitehouse, 1988). Recent ion-probe U-Pb geochronology of zircons from a granulite-facies tonalitic sheet which crosscuts the mafic/ultramafic body at First Inlet (ca. 1 km NE of the Camas nam Buth body dated by Whitehouse (1989a) and included together with data from Geodh Eanruig in the age regression of Cohen et al., 1991), indicates a protolith age of ca. 2.95 Ga (Friend and Kinny, 1995). It is interesting to note (Table 3) that consideration of the Cohen et al. (1991) First Inlet data alone yields a regression age of 2.84 ± 0.11 Ga which is within error of the U-Pb age for the crosscutting tonalite, and also of the ages of the Achiltibuie and Drumbeg mafic/ultramafic bodies. Further, all three bodies can be combined to give a regression age of 2.856 ± 0.082 Ga (Table 3).

6.1.2. *Mineral isochrons*

The generation of mineral isochrons from equilibrium mineral assemblages is a powerful tool in geochronology and there are numerous literature examples using both igneous and high-grade metamorphic assemblages. For example, the recent studies by Mezger et al. (1992) and Burton et al. (1995b) have utilised the different diffusion behaviour of Sm-Nd and U-Pb to generate precise chronologies of metamorphic evolution. However, these studies were performed on assemblages for which there is demonstrable textural evidence that equilibrium has been achieved. It is questionable

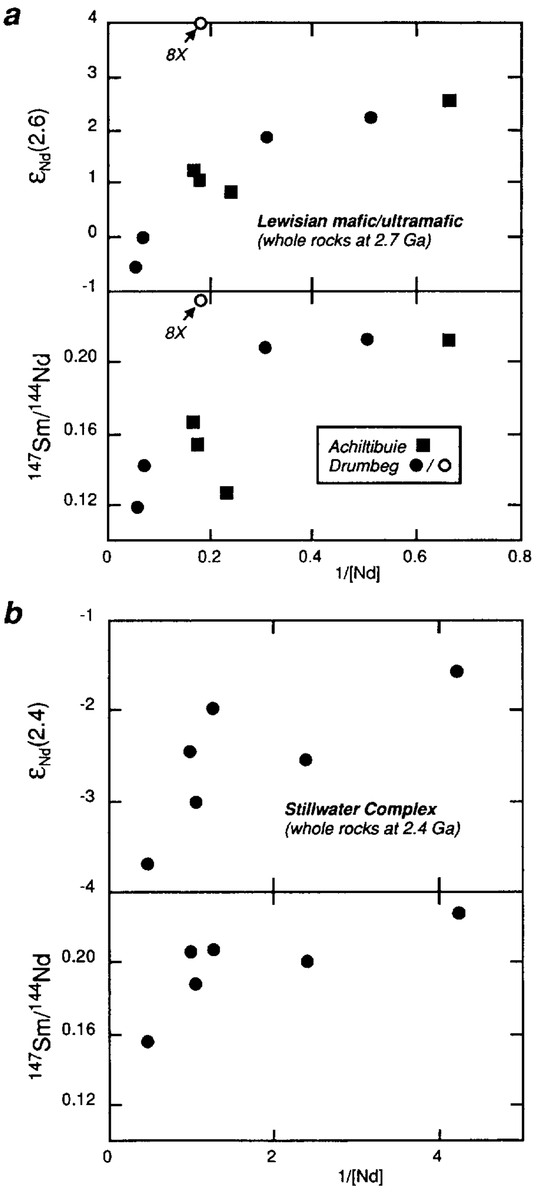

FIG. 7. Plots of (t) and $^{147}Sm/^{144}Nd$ against 1/[Nd] for whole rock samples from (a) Lewisian mafic/ultramafic bodies at Achiltibuie and Drumbeg (Whitehouse, 1989a; data for Drumbeg sample 8X (open circle) from Hamilton et al., 1979) calculated at 2.7 Ga and (b) Stillwater Complex (DePaolo and Wasserburg, 1979) calculated at 2.4 Ga.

whether reliable chronological data can be obtained from manifestly disequilibrium and/or retrogressive mineral assemblages because both the common initial ratio and closed-system requirements for an isochron are likely to be invalid. This will be particularly evident where partial mineral recrystallisation occurs significantly after whole-rock closure when the isotopic composition contrasts between different phases will be most pronounced. On the other hand, the ambiguities of interpretation as magmatic or metamorphic ages should be less troublesome.

6.2. Reconciliation with Established Relative Chronologies and Local Geology

We have shown above that assessment of the validity of isochrons on the basis of the isotopic data is difficult, that spurious isochrons may be produced in a number of ways, and that even valid isochrons are open to varied interpretation. In this section, therefore, we seek to assess the data in the light of existing chronology and local geology. Figure 8 presents three schematic stratigraphic sequences for development of lithologies and metamorphic events at Gruinard Bay. The columns represent that deduced from Burton et al. (1994), that proposed from this study, and the accepted chronology of the central region (Park et al., 1994). Using the MAS and associated tonalites as a base, the sequence of events produced by this study from field relations, petrography, geochemistry, and whole-rock geochronology clearly correlates with the accepted chronology (Fig. 8). In particular, the age obtained for the MAS, together with its field relationships and geochemistry suggest that correlation with the early mafic rocks in the Scourie area is plausible. If correct, the older tonalites at Gruinard Bay (Rollinson and Fowler, 1987) might represent the ca. 3.0 Ga tonalites from Scourie. The presently undated granulite facies trondhjemites at Badcall and Scourie, which crosscut the main gneissic banding, could be equated with our HMS-trondhjemite suite. Within this magmatic framework the metamorphic events and subsequent retrogression(s) closely correlate. In contrast, the sequence developed by Burton et al. (1994) bears little resemblance to that seen elsewhere in the central region (Fig. 8). This suggests either that there is a problem with the mineral isochron geochronology, or that Gruinard Bay indeed represents a unique part of the Lewisian Complex and consequently that the whole-rock isochrons are spurious.

In detail, the mineral regression ages conflict with observed field relationships and petrography. Burton et al. (1994) interpreted their ca. 3.3 Ga ages for the Gruinard Bay amphibolites as dating mineral equilibration shortly after differentiation from a depleted mantle source, and considered them "unaffected by the later granulite-facies event" which was correlated with the regional Badcallian high-grade event. This opinion challenges the general consensus (e.g., Park et al., 1994), based on petrographical and geochemical evidence (e.g., Fowler 1986; Rollinson and Fowler, 1987), that all the Lewisian gneisses at Gruinard Bay are retrogressed from hornblende-granulite facies. The geochemical evidence is particularly important as it shows moderate depletion in LILE, interpreted by Fowler (1986), to indicate that the Gruinard Bay area was transitional between highly depleted Scourie granulites and undepleted amphibolite facies counterparts, not locally exposed. Therefore, their assertion requires an unrecognised prograde amphibolite to granulite facies boundary between the sample site and the orthopyroxene-bearing granulites some 1.5 km to the north (Field, 1978; Fowler, 1986) and the retrogressed granulites 3 km to the southeast (Crane, 1978).

The field relationships and petrography described above

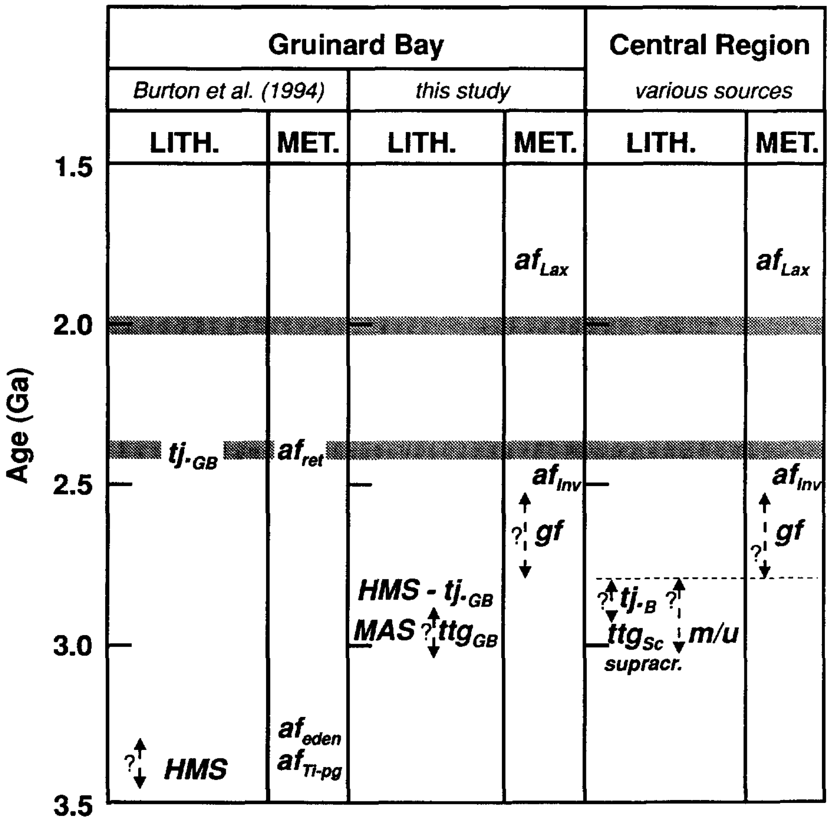

FIG. 8. Schematic stratigraphic column showing correlation between dated lithologies (LITH.) and metamorphic events (MET.) for the Central Region of the Lewisian and Gruinard Bay according to this study and Burton et al. (1994). The two shaded lines represent presently known intrusive ages for Scourie dykes (Heaman and Tarney, 1989; Waters et al., 1990) in the Central Region; extension of these lines to Gruinard Bay simply provides a time marker and does not imply the age of any particular dyke at this locality or the presence of dykes of both (or either) age(s). Lithology abbreviations: trondhjemite—tj (subdivided into localities Badcall (B) and Gruinard Bay (GB)); early tonalite-trondhjemite-granodiorite magmas—ttg (subdivided into localities Scourie (Sc) and Gruinard Bay (GB)); mafic/ultramafic enclaves—m/u; pre-TTG supracrustals—supracr.; abbreviations HMS, and MAS follow the conventions adopted throughout this paper. Metamorphism abbreviations: granulite facies events—gf; amphibolite facies events—af (subdivided into Inverian (Inv), Laxfordian (Lax), Ti-pargasite growth (Ti-pg) and edenite (eden) pseudomorphing in the Gruinard Bay amphibolites, and retrogressive metamorphism (ret) in the trondhjemites). Uncertainties in the age for either lithologies or metamorphic events are indicated by vertical dashed lines, except for m/u where the dashed line represents the possibility that there might be more than one intrusive episode.

clearly suggest that the area has suffered a complicated metamorphic history and exposure to retrogressive fluids over a long period of time. The metamorphic minerals included on the amphibolite regression lines by Burton et al. (1994) are at least second generation, and these authors' description of "high-Ti pargasitic hornblende pseudomorphed by edenite with exsolved ilmenite" suggests retrogression from granulite facies. Burton et al. (1994) do not specify which metamorphic event produced this high-Ti pargasite, but it follows from their arguments and interpretation that this event must be pre-3.3 Ga ($af_{Ti\text{-}pg}$ in Fig. 8). Additionally, Burton et al. (1994) describe a second origin for amphibole as hornblende pseudomorphs (edenite + quartz) after clinopyroxene (af_{eden} in Fig. 8), but no indication is given as to which amphibole was analysed. Analysis of mixed amphiboles which are not in equilibrium has important interpretative implications, particularly when derived ages are not recorded elsewhere in the Lewisian Complex.

Concordant ca. 2.4 Ga Sm-Nd and Pb-Pb mineral regressions from the host trondhjemites (Burton et al., 1994) were interpreted to approximate magmatic protolith ages on the basis that the alternative of metamorphic recrystallisation would require an event that might also be expected to disturb the amphibolites. The trondhjemites preserve evidence for a hornblende-granulite facies metamorphic event (Rollinson and Fowler, 1987) which, if the mineral ages are correctly interpreted, would have to post-date 2.4 Ga. Indeed, Burton et al. (1994) describe their trondhjemite mineral assemblages as retrogressive (af_{ret} in Fig. 8), although the timing of preceeding high-grade event is not discussed. Previously reported ca. 2.4–2.5 Ga mineral ages from the Lewisian (Humphries and Cliff, 1982; Cohen et al., 1988) have been interpreted to represent late cooling of the complex through the Sm-Nd closure temperature following ca. 2.7 Ga metamorphism. Close concordance between Rb-Sr, Pb-Pb, and Sm-Nd mineral regression ages in the Scourie area (Cohen et al., 1988) was further taken to indicate that final cooling was relatively rapid. However, the recent U-Pb ages of ca. 2.5 Ga (Corfu et al., 1994; Friend and Kinny, 1995) have revealed a probable metamorphic event at this time. Thus we consider it likely that the trondhjemite mineral ages record metamorphic recrystallisation during the ca. 2.5 Ga event, and it is possible that disturbance of the amphibolite mineral systematics would also have occurred.

Previous ca. 2.8–3.0 Ga Sm-Nd and Pb-Pb whole-rock regression ages for the trondhjemites obtained by Whitehouse (1989a,b) are considered by Burton et al. (1994, 1995a) to reflect the age of their source region, an interpretation that requires preservation of both isotopic and parent/daughter element ratios during magma genesis and emplacement. A similar scenario has been described for whole-rock Sm-Nd data from the Scourie dykes which have been suggested to record the ca. 3.0 Ga age of their lithospheric source (Waters et al., 1990). However, Fowler et al. (1995) have pointed out that the expected whole-rock age should, therefore, be ca. 3.3 Ga, not 2.8–3.0 Ga.

Finally, the abundant crosscutting Scourie dykes at Gruinard Bay, some of which cut the sampled locality of Burton et al. (1994), have been converted to amphibolites clearly demonstrating that there has been a post-dyke emplacement amphibolite facies metamorphism (Fig. 8). It is probable that at least some (if not all) of the amphibole in the older amphibolites must have re-equilibrated during this event, usually regarded as Proterozoic. On a regional basis, even younger events are recorded in titanite and rutile (Corfu et al., 1994).

6.3. Models for Lewisian Evolution

6.3.1. Evolution of the Lewisian based upon Gruinard Bay mineral ages

Burton et al. (1994) extend their observations at Gruinard Bay to present a general re-interpretation of Lewisian isotopic evolution and geochronology (their Fig. 2). They claim that their model is "consistent with an origin for the (Scourie) tonalites, and ultimately the trondhjemites, by partial melting of the older amphibolites, or similar material" that is similar in age, $f_{Sm/Nd}$ and μ_2 with the dated amphibolite samples. A magmatic protolith age for tonalites from Scourie of ca. 2.93 Ga has been suggested on the basis of Sm-Nd model ages (Whitehouse, 1989a), and recently confirmed by a U-Pb SHRIMP age of ca. 2.95 Ga (Friend and Kinny, 1995). We agree with previous interpretations (Whitehouse, 1988, 1989a,b; Burton et al., 1994) that the ca. 2.7 Ga whole-rock ages of these gneisses result from metamorphic reworking. Therefore, the significance of incorporating such metamorphic ages in a plot that is used to support a partial melting origin from older amphibolitic crust for many of the Lewisian TTG lithologies is not readily apparent. Furthermore, the model suggested by Burton et al. (1994) cannot be applied to the Lewisian Complex TTG suites as a whole. The exception of the northern region, which clearly does not lie on the postulated evolution trend (Fig. 2a of Burton et al., 1994), has been discussed elsewhere (Fowler et al., 1995; Burton et al., 1995a), as has the case for gneisses from Tiree, Inner Hebrides (Whitehouse and Robertson, 1995).

6.3.2. Evolution of the Lewisian based upon whole-rock Sm-Nd isochrons

Figure 9 plots the age and initial ϵ_{Nd} information from our two amphibolite whole-rock regressions in a neodymium isotope evolution diagram (ϵ_{Nd} vs. t), using the error polygon method (Fletcher and Rosman, 1982; Sanz and Wasserburg, 1969) to represent the covariation of errors in age and initial ϵ_{Nd}. Also plotted on this diagram are the error polygons derived from isochrons obtained on two other mafic/ultramafic bodies within the Lewisian complex central region (Achiltibuie and Drumbeg, ca. 30 km north of Gruinard Bay) which have been suggested as possible TTG suite mafic precursors (Whitehouse, 1989a). The two solid evolution lines are average trajectories (weighted based on Nd ppm) for the HMS (A; $\epsilon_{Nd}(0) = -21.0, f_{Sm/Nd} = -0.31$) and the MAS (B; $\epsilon_{Nd}(0) = -15.6, f_{Sm/Nd} = -0.24$); the two dotted line trajectories represent the whole-rock amphibolite analyses presented by Burton et al. (1994) and the dashed line represents their preferred amphibolite evolution trend ($\epsilon_{Nd}(3.3) = 4, f_{Sm/Nd} = -0.33$). The first important point illustrated by this diagram is that it is possible to derive the ca. 2.8 Ga HMS-trondhjemite magma from the MAS only by using the most light-REE-enriched members of our sample suite, in accordance with the modelling of Rollinson and Fowler (1987). For example, the large arrow starting at the centre of the MAS error polygon shows evolution of the most light REE-enriched MAS composition which has $f_{Sm/Nd} = -0.47$ compared with the average MAS $f_{Sm/Nd}$ value of -0.24. However, the Gruinard Bay Lewisian is agmatitic, not migmatitic: there is no field evidence for in situ partial melting of the MAS, and no geochemical evidence that this suite represents a partial melting residue. Instead, we prefer an interpretation in which the trondhjemitic magma (and its cumulates as represented by the HMS) was derived at ca. 2.8 Ga from a slightly evolved (ϵ_{Nd} of ca. +1) basaltic source similar geochemically to the MAS at Gruinard Bay, but perhaps with slightly different protolith age and initial ϵ_{Nd} characteristics. For example, in Fig. 9 the Drumbeg mafic/ultramafic body has ideal precursor characteristics (although we are not suggesting that this body was itself partially melted at any time). In our model, the MAS simply represents pre-existing basaltic crust which was disrupted and entrained into a younger TTG magma. The second important point is that of the Burton et al. (1994) actual and inferred amphibolite evolution trends, only GR-102 appears to bear any relation to the age and initial ϵ_{Nd} data derived from our sample suites.

6.3.3. Late-Archaean crustal evolution

Strongly light rare earth element-enriched TTG suite magmas, which form the dominant granitoid component of Archaean cratons, cannot be derived directly from the mantle,

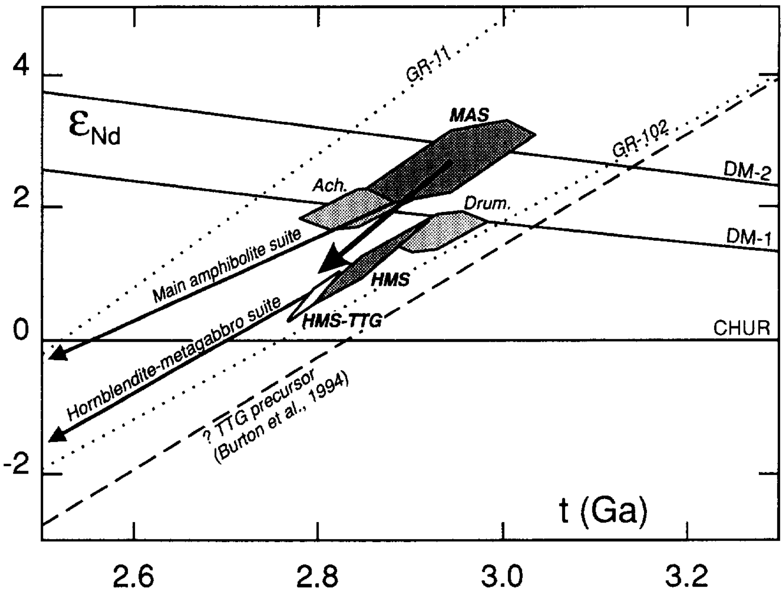

FIG. 9. Neodymium isotope evolution diagram (ϵ_{Nd} vs. t) presenting initial ϵ_{Nd} and age data from regressions as error polygons. Dark shaded polygons represent the HMS and the MAS. Light shaded polygons show data from the Achiltibuie (Ach.) and Drumbeg (Drum.) mafic/ultramafic bodies (Whitehouse, 1989a; errors recalculated from original data). White unlabelled polygon represents the combined HMS-TTG data. Terrane averaged evolution lines for MAS and HMS are shown as solid lines, and the thick arrow extending from the centre of the MAS polygon represents the evolution line for the most light REE-enriched member of this suite ($f_{Sm/Nd} = -0.47$). The dashed evolution line represents the TTG precursor proposed by Burton et al. (1994) and the two dotted evolution lines labelled GR11 and GR102 represent their whole-rock analyses. CHUR is the evolution line for the chondritic reservoir and two depleted mantle models are shown—DM-1 (DePaolo, 1981) and DM-2 (DePaolo et al., 1991).

but require a precursor mafic (basaltic) crust in which heavy REE-enriched mineral phases such as garnet and amphibole are residual during partial melting (Arth and Barker, 1976; Barker and Arth, 1976; Jahn et al., 1981; Rudnick and Taylor, 1986; Martin, 1987). Given the volume of TTG magmatism observed in the Archaean, an analogue of modern oceanic crust might represent the source region, albeit under very different conditions imposed by higher heat flow (de Wit et al., 1992). Unambiguous identification of such mafic crust is complicated by the high strain typically observed in Archaean cratons, and often relies upon identifying components with appropriate geochemistry and/or geochronology. By linking the Gruinard Bay 3.3 Ga amphibolites and 2.4 Ga trondhjemites as source and derived magma respectively, as Burton et al. (1994) have done implies that Lewisian TTG generation occurred up to 900 Ma, after mafic precursor generation, considerably longer than the present survival time of oceanic crust (250 Ma maximum). Such longevity goes against current evolutionary models for Archaean terrains when heat flow is thought to have been higher and turnover of oceanic crust more rapid (Martin, 1986; deWit et al., 1992).

6.4. Development of Spurious Mineral Isochrons during Retrogression

We have argued above that the mineral regression ages of Burton et al. (1994) cannot be reconciled with the observed geology at Gruinard Bay, and that the Lewisian evolutionary model developed from these ages cannot be applied to the Lewisian as a whole. In this section we consider whether the data and interpretations presented by Burton et al. (1994) might be the result of later isotopic disturbance on a mineral scale.

Despite the obvious high precision of the analyses and derived isochrons presented by Burton et al. (1994), features of their geochronological interpretation cast doubt upon the significance of the ca. 3.3 Ga regressions as ages for the amphibolites. First, we find it unusual that both samples yield Pb-Pb regression lines but only one yields a meaningful Sm-Nd isochron, the other sample having clearly been disturbed. Isotopic disturbance at the mineral scale which so thoroughly affects Sm-Nd systematics is unlikely to leave U-Pb systematics so completely undisturbed in the same minerals (although we note that our recalculation of GR-11 Pb-Pb data yields an age of 3332 ± 240 Ma which is of considerably lower precision than that claimed by Burton et al., 1994). Second, the significant difference of model μ_1 values between the two Pb-Pb mineral isochrons (8.43 ± 0.21, recalculated as 8.4 ± 0.5, and 8.02 ± 0.08) requires lengthy pre-3.3 Ga development of the two amphibolites in very different reservoirs (for example, a μ value as high as 15 would be required over a ca. 200 Ma period to develop the $\mu = 8.4$ source from the $\mu = 8$ source). Furthermore, the μ value of 8.43 for GR-102 is unusually high for the middle- to late-Archaean North Atlantic Craton which ranges from 7.5–8 (Moorbath and Taylor, 1981). In the Lewisian specifically, Whitehouse (1990) has argued for a source μ value as low as ca. 7.5 based upon unradiogenic lead isotopic compositions from the southern Outer Hebrides gneiss complex. Fractionation in the U-Pb system could have occurred during igneous emplacement, but this would have to be considerably earlier than 3.3 Ga, an alternative not considered by Burton et al. (1994).

The concordance between mineral isochron ages obtained from two independent isotopic systems applied to one of the Burton et al. (1994) samples is, at first sight, convincing evidence for a real event at ca. 3.3 Ga. However, we believe that given the isotopic arguments above together with the petrographic details discussed earlier, the younger whole-rock ages presented here are more likely to reflect true magmatic protolith ages for the Gruinard Bay amphibolites, and the ages presented by Burton et al. (1994) most probably reflect post 2.8 Ga disturbance of the isotope systematics on the mineral scale, coincidentally yielding concordant ages. Figure 10 presents schematic ϵ_{Nd} vs. t and $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagrams which demonstrate how such spurious isochrons may develop.

In the ϵ_{Nd} vs. t diagram (Fig. 10a) we consider the case of a closed whole-rock system generated at protolith age t_{proto} and initial ϵ_{Nd} at point A. Within the whole-rock, individual minerals will be evolving with both more evolved Sm/Nd ratios (e.g., plagioclase) and more depleted Sm/Nd ratios (e.g., hornblende) than the overall whole-rock; these are represented by the dashed lines from point A. At some later time t_{dist}, the system experiences internal disturbance of Sm/Nd ratios without significant neodymium isotopic homogenisation (although some degree of homogenisation would not affect the result here) and the mineral phases then develop along different Sm/Nd evolution lines to their present-day composition. If the whole-rock system has remained closed then the intersection of the mineral phases with the whole-rock evolution line at point B will represent the apparent isochron age, t_{app}, and its initial ϵ_{Nd}. Mørk and Mearns (1986) have presented a case for incomplete homogenisation of neodymium isotope systematics during eclogitisation of gabbros which retain their original igneous neodymium isotopic composition. This model can be quantified using the data for GR-102 presented by Burton et al. (1994). A diagram of $^{147}Sm/^{144}Nd$ vs. t_{dist} (Fig. 10b) shows the $^{147}Sm/^{144}Nd$ ratio that would be required in plagioclase and hornblende assuming a t_{proto} of 2.8 Ga, Sm-Nd fractionation at t_{dist} and absence of whole-rock disturbance. Fractionation in plagioclase might be achieved by the growth of abundant epidote during retrogression. This model assumes only Sm-Nd fractionation and probable partial neodymium isotopic homogenisation at t_{dist} would make these values limiting cases, i.e., the values for $^{147}Sm/^{144}Nd$ in plagioclase between t_{proto} and t_{dist} would be lower than indicated in the inset diagram, and those for hornblende higher. The $^{147}Sm/^{144}Nd$ ratios modelled for GR-102 plagioclase are very low. For comparison, an extreme value of ca. 0.04 has been obtained from plagioclase in a Nain anorthosite (L. D. Ashwal et al., unpubl. data), and DePaolo (1985) reports a value of ca. 0.07 from the Kiglapait anorthosite intrusion. Clearly realistic plagioclase $^{147}Sm/^{144}Nd$ ratios constrain t_{dist} to a probable Laxfordian event.

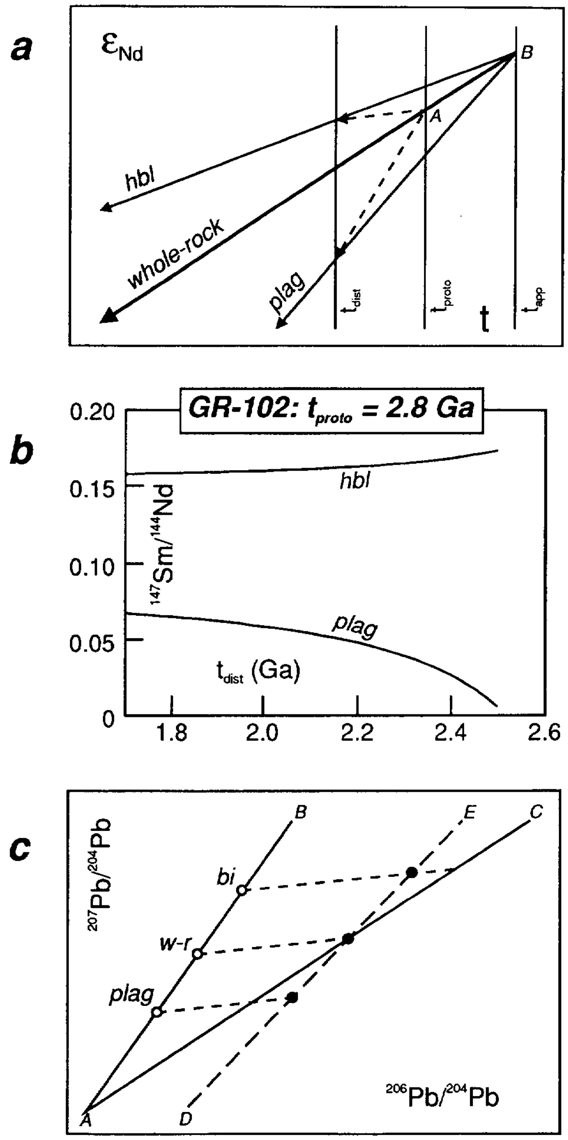

FIG. 10. Schematic diagrams showing development of spurious ages in the U-Pb and Sm-Nd systems. (a) ϵ_{Nd} vs. t; (b) $^{147}Sm/^{144}Nd$ vs. t_{dist} quantifying the model for the GR-102 data presented by Burton et al. (1994), showing the $^{147}Sm/^{144}Nd$ ratio that would be required in plagioclase and hornblende assuming a protolith age of 2.8 Ga, Sm-Nd fractionation at t_{dist} and absence of whole-rock disturbance; (c) $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$. For detailed discussion, see text.

A similar scenario of parent-daughter ratio fractionation in the absence of significant isotopic homogenisation is considered in the $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagram (Fig. 10c). Here the system is generated at t_{proto} with an initial composition represented by point A. At t_{dist} minerals and whole-rock compositions will lie along the t_{proto} at t_{dist} isochron represented by line AB. If no disturbance of the system occurs, samples will then evolve to their present day compositions on line AC which is the t_{proto} isochron. If there is fractionation of U/Pb ratios at t_{dist}, however, the samples will evolve to some point along the dashed lines which represent t_{dist} isochrons. In the example shown, there has been retardation of U/Pb ratios in the more radiogenic composition (e.g., biotite), increase in the least radiogenic composition (e.g., plagioclase), and no change in the intermediate composition (e.g., whole-rock), so that they evolve to the compositions represented by the filled circles on line DE. Clearly, in this case line DE has a higher slope than AC and represents the older t_{app}; it will also have a higher μ_1 value. This model of evolution is similar to that proposed to explain transposed palaeoisochrons in whole-rock Pb-Pb data (Moorbath and Taylor, 1986; Whitehouse, 1990). Examination of the two Pb-Pb regression lines of Burton et al. (1994) shows that both (in particularly the more precise GR-102) are strongly controlled by plagioclase and biotite only, suggesting that the model presented above might be applicable.

7. CONCLUSIONS

On the basis of field relationships, petrography and major and trace-element (including REE) geochemistry we recognise two distinct suites of amphibolitic enclaves in the trondhjemites of the Gruinard Bay Lewisian Complex. A suite of essentially basaltic composition, the MAS, is considered to represent potential TTG suite mafic precursor material and yields a Sm-Nd whole-rock regression age of 2.943 ± 0.091 Ga with an initial ϵ_{Nd} value of +2.69 ± 0.62. A suite of hornblendites and metagabbros, the HMS, display geochemistry consistent with their origin as amphibole and plagioclase cumulates from the host trondhjemitic magmas, and yield a Sm-Nd isochron (MSWD = 0.52) age of 2.846 ± 0.073 Ga with an initial ϵ_{Nd} value of +1.10 ± 0.72. Combination of the HMS data with previously published data from the Gruinard Bay trondhjemites (Whitehouse, 1989a), on the basis of our established genetic link, yields a Sm-Nd isochron (MSWD = 1.09) age of 2.795 ± 0.028 Ga with an initial ϵ_{Nd} value of +0.66 ± 0.38. These data show that crustal evolution at Gruinard Bay began at ca. 3.0 Ga with production of mafic magmas from a depleted mantle source region ($\epsilon_{Nd}(t)$ ca. +2 to +3). Equivalents of the most light REE-enriched components of this early mafic suite subsequently provided the source for trondhjemitic magmas ca. 100–200 Ma later ($\epsilon_{Nd}(t)$ ca. +1), with plagioclase and amphibole fractionation producing cumulates, fragments of which are preserved along with those of the early basic magmas in the exposed agmatite trains.

There is a clear discrepancy between the 3.0–2.8 Ga whole rock Sm-Nd ages reported here and previously published ca. 3.3 Ga Sm-Nd and Pb-Pb mineral regression ages (Burton et al., 1994) for two samples from the HMS. Acceptance of these mineral regression ages, together with ca. 2.4 Ga Sm-Nd and Pb-Pb mineral age regressions for the host trondhjemites (Burton et al., 1994) has profound implications for the geological evolution of the Lewisian Complex, requiring radical revision of the magmatic and metamorphic time-scale established over the past few decades. In accord with conventional assumptions, we consider that the whole-rock Sm-Nd systematics are more robust than mineral systematics and the ages we present reflect true magmatic protolith ages for both of the amphibolite suites recognised at Gruinard Bay.

The complexity of the metamorphic evolution of this part of the Lewisian Complex (Park et al., 1994), including extensive retrogression to the present disequilibrium mineral assemblages in the amphibolites, makes these lithologies unsuitable for geochronology based upon mineral separates. We consider the amphibolite ages presented by Burton et al. (1994) to be a spurious product of retrogressive metamorphism, with incomplete isotopic homogenisation during parent/daughter element fractionation, probably Laxfordian (ca. 1.8–1.6 Ga). The trondhjemite ages presented by these authors are almost certainly misinterpreted as emplacement ages and most likely reflect cooling following a ca. 2.5 Ga metamorphism that has recently been recognised in the Lewisian (Corfu et al., 1994; Kinny and Friend, 1994; Friend and Kinny, 1995).

There are no reliable objective ways to assess the validity of any given whole-rock (or mineral) isochron using the regressed data alone, or in combination with element concentration data. Isochron data should be assessed in the light of all available field, petrographic, geological, geochemical, and other geochronological evidence. Whilst mineral regression ages potentially can yield detailed chronological information on metamorphic evolution, they should be used with extreme caution in high-grade terrains which have experienced a long and complex metamorphic evolution, particularly involving retrogressive metamorphism(s). At Gruinard Bay, detailed consideration of field, petrographic, and geochemical data has enabled us to select probable cogenetic groups for inclusion on whole-rock isochrons which, in principle, should be less susceptible to the effects of pervasive retrogression. This approach yields information which leads to a consistent, unambiguous sequence of events in complete accord with that established in other parts of the Lewisian Complex.

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APPENDIX

Kyanite-bearing metasedimentary rocks and pegmatites occur on the southwestern side of the Inveranvie River (map reference NG 9574 8871, Fig. 1b). The kyanite occurs as variably oriented, poikiloblastic blades overgrowing the quartzo-feldspathic groundmass and appears to have no crystallographic relationship with biotite (now largely chlorite) and quartz ribbons which were the main phases defining an early, pre-retrogression foliation. Two distinct generations of muscovite occur. First, the main phase has either a foliated or a random crystallographic orientation. Where foliated it is in close association with biotite and retrogressive chlorite. In zones around kyanite poikiloblasts this generation appears to be overgrowing kyanite which often has a ragged outline and shows the development of kink bands or undulose extinction. Second, and quite distinct, are fine-grained, randomly oriented muscovite aggregates which pseudomorph the kyanite. Biotite is frequently replaced, largely along its cleavage traces, by chlorite. The orientations of the retrogressive minerals suggest that this took place in an essentially static environment, though some weak deformation is indicated by the kink-bands and undulose extinction.