PETROLOGY AND STABLE ISOTOPE SYSTEMATICS OF MIGMATITES IN PINKHAM NOTCH, NEW HAMPSHIRE

by

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INTRODUCTION

Migmatite is a textural term used to describe very heterogeneous metamorphic rocks consisting of intermingled light-colored material (leucosomes), dark-colored material (melanosomes) and intermediate material (mesosomes). The obvious question is how do these rocks form, and under what conditions? Since Tuttle & Bowen's (1958) experiments in the Granite System, the conventional wisdom has been that migmatites form as a result of anatectic partial melting processes (e.g., Winkler, 1979). However, leucosomes from many migmatites—including those in New Hampshire—do not have granitic minimum melt composition, containing quartz and albitic plagioclase but no K-feldspar (although muscovite is usually present; Dougan, 1979; Eusden, 1988; Allen, 1992). It is now generally recognized that other processes besides anatectic partial melting can play a role in forming migmatites, including sub-solidus differentiation, metasomatism, and injection of melt derived elsewhere (Ashworth, 1985; Olsen, 1985).

Dougan (1979, 1981, 1983) conducted detailed mass balance, compositional, and textural studies of leucosomes, melanosomes and mesosomes in migmatites samples from several localities in New Hampshire, including Pinkham Notch. He determined that these migmatites formed by "closed-system" partial melting, possibly driven by the infiltration of fluids. In geologic mapping of the Pinkham Notch area, Hatch & Wall (1986) identified a "migmatite front," separating un-migmatized schists from migmatitic gneisses both of the Silurian Rangeley Formation (Fig. 1). Such a front certainly suggests infiltration of fluids, and provides an opportunity to directly compare the migmatite with its protolith. In conjunction with geologic mapping in the area (trip B2, this volume; Allen, 1992), I have undertaken detailed stable isotope, petrologic and geochemical studies of these schists and migmatites to determine the migmatization reaction and the role of fluids. Complete analytical details and results are given in Allen (1992).

COMPARISONS ACROSS THE MIGMATITES FRONT

The mineral assemblage in the schists (#010, Stop 1b) is quartz + sillimanite + muscovite + biotite + quartz + garnet + plagioclase + ilmenite. Accessory minerals include pyrrhotite and tourmaline, and minor graphite, apatite, monazite and zircon. In these schists, sillimanite + muscovite + quartz pseudomorph andalusite, suggesting a polymetamorphic history for this region (Wall & Guidotti, 1986; Hatch & Wall, 1986). There is no potassium feldspar present in these schists, nor is there any evidence for potassium feldspar ever having been a component of the high-grade metamorphic assemblage.

The migmatites (#011, Stop 1c) contain exactly the same mineral assemblage as the schists, and belong to the same metamorphic zone, except that if the migmatization is due to partial melting then a melt phase was present in addition to the minerals listed above. This melt phase might now be represented by the leucosomes, which are composed of quartz, plagioclase and muscovite, with only minor amounts of the other minerals. Again, there is no evidence of potassium feldspar ever having been present in these rocks, not even in the leucosomes. Tourmaline, apatite, monazite, and zircon are important accessory minerals.

Figure 1: Map of Pinkham Notch

ALLEN

Figures 2 and 3: Compositional Diagrams

Comparison of Fig. 2a to 2b and Fig. 3a to 3b shows qualitatively a great deal of similarity between the compositions of the bulk migmatites and the schists, which suggests that the schists indeed represent the migmatite's protolith, and that the migmatization occurred without significant gain or loss of non-volatile components (e.g. silicate melt)—in a "closed system" (isotopic evidence presented below shows that the migmatization was an open system for hydrous fluids). However, the degree of heterogeneity in these rocks, both of the schists as well as the migmatites, makes it difficult to make quantitative comparisons given a limited number of samples. Other evidence for a non-volatile closed system comes from a modal analysis of the migmatite outcrops (Allen, 1994), which showed that the leucosomes and melanosomes occurred in equivalent proportions, each making up about 20 to 30 % of the rock; and from comparison of Figs. 2c & 2d to 2a & 2b and Figs. 3c & 3d to 3a & 3b, which suggests that the leucosome and melanosome compositions are roughly complementary, consistent with textural observations at the outcrops. Dougan's (1979) mass balance studies comparing leucosome and melanosome compositions to mesosomes (Olsen, 1983; 1985) quantitatively demonstrated "closed system" migmatization in his Pinkham Notch samples.

METAMORPHIC CONDITIONS

Figure 1 shows metamorphic temperatures calculated using the garnet-biotite Fe–Mg exchange geothermometer of Ferry & Spear (1978), as calibrated by Indares & Martignole (1985, model B). Metamorphic pressures were calculated at between 3.5 and 4.0 kilobars across the region, using the garnet-plagioclase-Al_2SiO_5-quartz calibration of Newton & Haselton (1981). (Full details of the temperature and pressure calculations are given in Allen, 1992.) The uncertainties of these calculations is on the order of $\pm 50°C$ and ± 2 Kbar, but these uncertainties do not strongly affect the precision of the method (Hodges & McKenna, 1987)—thus relatively small differences between samples may be real. . These results indicate a fairly steep temperature gradient across the migmatite front, consistent with the isograds mapped by Wall & Guidotti (1986) and Hatch & Wall (1986) (Fig. 1), and similar to the metamorphic "hot spots" studied by Chamberlain & Lyons (1983) and Chamberlain & Rumble (1988).

It is not clear, however, that the temperatures obtained in the migmatites are sufficiently high to cause partial melting. Water-saturated granitic melting requires temperatures of about 650°C at 4 Kbar (Johannes, 1984, 1985; Tuttle & Bowen, 1958; Luth *et al.*,1964), and water-saturated "trondhjemitoid" melting in the sodic-plagioclase + quartz system does not begin until almost 700°C at 4 Kbar (Johannes, 1978, 1985). However, in addition to water, the presence of P, F or B, particularly in combination with water, has been shown to significantly reduce the solidus temperature in these systems (Wyllie & Tuttle, 1964; Manning & Pichavant, 1983; Manning, 1981; Pichavant, 1981), even to temperatures below 550°C. The addition of these components also moves the minimum melt from "granite" towards "trondhjemite" compositions (Manning, 1981). Some evidence suggests that P and F may have been important in these rocks (Fig. 2 and Fig. 5), such that melting may have occurred at the recorded temperatures. It is also possible that the peak temperatures of these rocks were higher than recorded by the garnet-biotite thermometer.

ROLE OF FLUIDS

Figure 1 also shows the distribution of whole rock oxygen isotope values ($\delta^{18}O_{WR}$) in the region near the migmatite front. In addition, numerous samples were analyzed from the schist (#010, Stop 1b) and migmatite (#011, Stop 1c) outcrops immediately adjacent to the migmatite front, and the average and range of these data are reported on Fig. 1. No clear overall trends are apparent in the regional data, because of the variety of rock types sampled, but there is a measurable lowering of the $\delta^{18}O_{WR}$ from the schist to the migmatite outcrops immediately adjacent to the migmatite front.

If the migmatization process were completely closed, then the "bulk" migmatites and the original schists should have identical whole-rock oxygen isotope values, regardless of the mineral changes that may have occurred. Without introducing lighter oxygen by injecting melt or removing heavier oxygen by extracting melt, the lowering of $\delta^{18}O_{WR}$ observed in the migmatites could be achieved by de-volatilization of fluids with heavier oxygen or by isotopic exchange with an infiltrating fluid of lower $\delta^{18}O$. Among the migmatites, the lowest $\delta^{18}O_{WR}$ values are from samples spatially associated with pegmatites, while the highest $\delta^{18}O_{WR}$ values are from the "least migmatized" samples, suggesting a relationship between the migmatization process and infiltrating fluids.

Whole-rock isotope measurements tell only part of the story. The quartz $\delta^{18}O$ values from three samples of the schist outcrop (#010, Stop 1b) are all nearly identical, as are the biotite $\delta^{18}O$ values (Fig. 4). These three samples have very different $\delta^{18}O_{WR}$, from 13.2‰ to 14.1‰, directly attributable to different modal mineralogies. That the minerals from all three of these samples have identical $\delta^{18}O$ values suggests these samples were in equilibrium with a common fluid, at least on the outcrop scale. Using the fractionation factors of Bottinga & Javoy (1973, 1975), this fluid is determined to have $\delta^{18}O_{water}$ of about 13.9‰, and the equilibrium temperature is determined to be about 546°C, in reasonable agreement with that determined by garnet-biotite thermometry (520°C, Fig. 1). The homogeneity of mineral values within schist outcrop #010 is suggestive of the kind of isotopic homogenization that is associated with large scale fluid infiltration (Valley, 1986), however in comparison with mineral values from adjacent migmatite outcrops, there is no large scale homogenization. The small variation in these schist samples is in the direction expected for closed-system behavior (Fig. 4, Gregory & Criss, 1986).
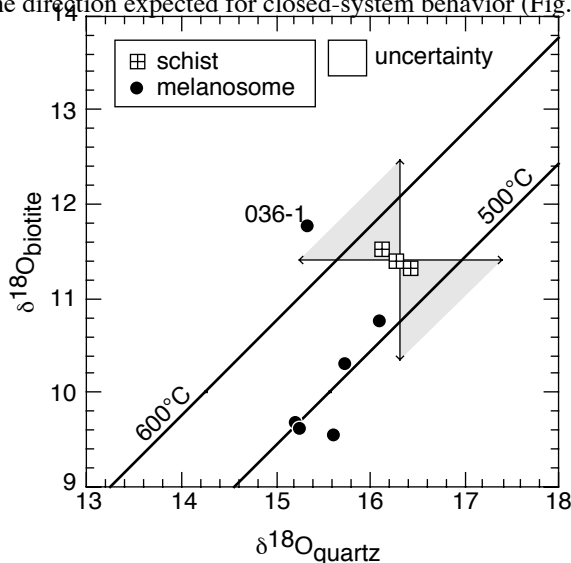

Figure 4: Biotite and quartz oxygen isotope values for schist and migmatite melanosome samples. Isotherms determined from the fractionation factors of Bottinga & Javoy (1973, 1979). Shaded triangles show region of closed-system behavior for schist samples (Gregory & Criss, 1986). Sample #036-1 is considered anomalous since biotites in this rock contain a large number of included minerals (monazite, etc.), which may contaminate its $\delta^{18}O_{Bt}$ value. The uncertainty associated with these analyses is ±0.2‰ and is shown by the labelled box.

It should be noted that in the migmatites, biotite is present only in trace quantities in the leucosomes. Thus only biotite-quartz pairs from melanosomes are shown in Fig. 4. With the exception of sample #036–1, the biotite-quartz values from the migmatites define an isothermal array, at a temperature of about 500°C. If the quartz and biotite $\delta^{18}O$ values of the schist samples are taken as a starting point for the material that formed the migmatites then, as shown in Fig. 4, the values for the migmatites lie outside the area expected for closed system behavior (Gregory & Criss, 1986), further confirming the role of infiltrating fluids in driving migmatization. Both biotite and quartz from melanosomes have $\delta^{18}O$ values that are lighter than the original schist, consistent with exchange with an isotopically light fluid. From the migmatite samples with the lowest $\delta^{18}O$ values, furthest down and to the left in the array (those presumably most exchanged with the fluid), I estimate $\delta^{18}O_{water}$ for the fluid to be 12‰ or less, using the fractionation factors of Bottinga & Javoy (1973, 1975), at the temperature determined by the array (500°C). This value is consistent with a $\delta^{18}O_{water}$ value estimated to be in equilibrium with a pegmatite (#011–2) cutting

the outcrop.

The temperature of this isothermal array is not consistent with the temperature obtained by garnet-biotite thermometry in the migmatites, and in fact is lower than the temperature obtained in the schists by either garnet-biotite or oxygen isotope thermometry. It is not clear why this is, but one possibility may be that continued isotopic exchange of minerals in migmatites occurred during cooling after migmatization. Due to the fluid infiltration-driven migmatization, these rocks may have had a higher water content than the schists, perhaps facilitating such exchange during cooling.

In Figure 5 we can clearly see that migmatite biotites are lighter than those in the schists (with the exception of #036–1), and that the migmatite biotites are somewhat enriched in F relative to the schists. Measurements by Hamza & Epstein (1980) have suggested that there is a large oxygen isotope fractionation between oxygen in the hydroxl site and oxygen in the rest of the biotite, and that as F substitution displaces OH in biotite, the overall $\delta^{18}O_{Bt}$ should become higher. The trend observed in our migmatite samples is in the opposite direction, which clearly indicates that the $\delta^{18}O_{Bt}$ is not a function of F–OH composition, but may be influenced by infiltrating fluids. As the biotites with the lightest $\delta^{18}O_{Bt}$ values are the most enriched in F, it is possible that the infiltrating fluid is also the source of F. Hydrogen isotope measurements from these biotites show no trends, as all samples have δD_{Bt} values of about -75‰ (Allen, 1992), within the "normal" range for biotites from metamorphic rocks (Taylor & Sheppard, 1986).
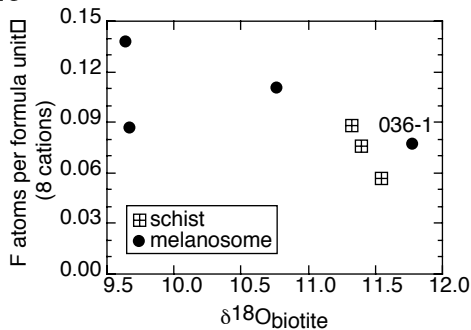

Figure 5: Fluorine content determined by microprobe analyses of biotites versus biotite oxygen isotope value for schist and melanosome samples. Sample #036-1 is considered anomalous since biotites in this rock contain a large number of included minerals (monazite, etc.), which may contaminate its $\delta^{18}O_{Bt}$ value.

Figure 6 shows $\delta^{18}O_{Ms}$ versus $\delta^{18}O_{Qtz}$ for schist, leucosome and melanosome samples. Muscovite is abundant in both melanosomes and leucosomes, so muscovite-quartz points from both leucosomes and melanosomes. Leucosome and melanosome pairs are connected with a line segment, resulting in the black and white "dumbbells." In contrast to the biotite-quartz system (Fig. 4), the muscovite-quartz points produce a disequilibrium array. This has also been observed in muscovite-quartz pairs from migmatites studied by Mazurek (1992). Certainly some of the muscovite in these samples is retrograde, as indicated by the mesoscopic muscovite spangle texture prevalent in these rocks. It is not clear what factors are responsible for the wide distribution of $\delta^{18}O_{Ms}$ values, although there may be a trend of increasing $\delta^{18}O_{Ms}$ away from the Wildcat Granite (trip B2, this volume; Allen, 1992) on a regional scale, and away from pegmatites on an outcrop scale (Allen, 1992)
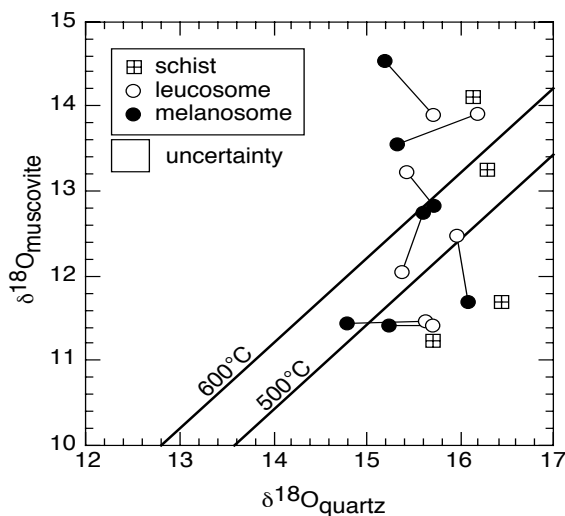
Figure 6: Muscovite and quartz oxygen isotope values for schist and migmatite leucosome and melanosome samples. Isotherms determined from the fractionation factors of Bottinga & Javoy (1973, 1979). Lines connect leucosome (open circle) and melanosome (filled circle) pairs from the same migmatite sample. The uncertainty associated with these analyses is ±0.2‰ and is shown by the labelled box.

MIGMATIZATION REACTIONS

J. B. Thompson, Jr., (1982) developed methods for derivation of a sufficient set of linearly independent net transfer and exchange reactions that describe all possible reactions in a given mineral assemblage. Ferry (1984) adapted this method to calculate the overall reaction at an isograd. In this study I have applied these methods to determine the reactions that occurred across the migmatite front, including possible melting reactions. The additive phase components, exchange phase components, and system components for the assemblage studied here (quartz + biotite + muscovite + garnet + plagioclase + sillimanite + ilmenite) are listed in Table 1. Only net transfer reactions are important in controlling the texture and mineral modes in metamorphic rocks. Exchange reactions only change the compositions of minerals, for example the Fe–Mg ratio in garnet and biotite, and do not affect the modal abundance of the minerals. The first set of net transfer reactions (1–4, Table 1) are determined for a completely closed system with the above mineral assemblage (where potassium and hydroxl are coupled). Opening the system to Na or K exchange with a hydrous fluid introduces an additional net transfer reaction (5, Table 1). The enrichment of fluorine in migmatite biotites (Fig. 5) is accommodated by the introduction of HF as a system component and a phase component, and $F(OH)_{-1}$ as an exchange component, which yields another independent net transfer reaction (6, Table 1).

If the migmatization occurred without partial melting, then these six reactions completely describe the possible reaction space of the migmatization. If, on the other hand, partial melting is involved in the migmatization process, then we have to introduce yet another phase component to our assemblage, the melt phase. Silicate liquids probably have a wide range of possible compositions, with no enforced stoichiometry that would enable writing carefully balanced reactions. For the partial melting scenario, it is assumed that the leucosomes represent the melt phase (and that no melt was extracted from or injected into the rocks). Estimated melt phase "stoichiometry" was determined from the chemical compositions of the leucosomes (Table 1), on the basis of 1000 anhydrous oxygen atoms per formula unit. With this we can determine a linearly-independent net-transfer "melting" reaction for each of the leucosome samples (7, Table 1).

If we allow the water content of the melt phase to vary independently of the melt phase stoichiometry, then no specific amount of water is needed to balance reactions involving melt, and these reactions may equally represent either water-saturated melting or dehydration melting (A. B. Thompson, 1982; Vielzeuf &

CLOSED SYSTEM ASSEMBLAGE

Mineral	Additive Component	Exchange Components
Quartz (Qtz)	SiO_2	—
Sillimanite (Sil)	Al_2SiO_5	—
Muscovite (Ms)	$KAl_3Si_3O_{10}(OH)_2$	NaK_{-1} (nk), $Al_2Fe_{-1}Si_{-1}$ (tk), $FeTiAl_{-2}$ (tx), $MgFe_{-1}$
Biotite (Bt)	$KFe_3AlSi_3O_{10}(OH)_2$	NaK_{-1} (nk), $Al_2Fe_{-1}Si_{-1}$ (tk), $FeTiAl_{-2}$ (tx), $MgFe_{-1}$, $MnFe_{-1}$
Garnet (Grt)	$Fe_3Al_2Si_3O_{12}$	$CaFe_{-1}$ (cf), $MgFe_{-1}$, $MnFe_{-1}$
Plagioclase (Ab)	$NaAlSi_3O_8$	$CaAlNa_{-1}Si_{-1}$ (pl)
Ilmenite (Ilm)	$FeTiO_3$	$MgFe_{-1}$, $MnFe_{-1}$

System Components: SiO_2, TiO_2, Al_2O_3, CaO, FeO, MgO, MnO, Na_2O, $K_2H_4O_3$

Linearly Independent Net Transfer Reactions: ξ

(1)	Ilm + Qtz = Sil + tx	(Ilm)
(2)	Bt + 5 Qtz +3 tk = 2 Sil + Ms	(Bt)
(3)	Ab + Qtz + tk + pl = 2 Sil +cf	(Ab)
(4)	Grt + 4 Qtz + 3 tk = 4 Sil	(Grt)

ALLOW NaK_{-1} and $F(OH)_{-1}$ EXCHANGE WITH ENVIRONMENT

Additional Phase Components:

Muscovite	$F(OH)_{-1}$
Biotite	$F(OH)_{-1}$ (Apatite also provides F-OH exchange capacity)
Water	H_2O, NaK_{-1} (nk), HF

System Components: SiO_2, TiO_2, Al_2O_3, CaO, FeO, MgO, MnO, Na_2O, K_2O, H_2O, HF

Additional Independent Net Transfer Reactions: ξ

(5)	nk + Ms + 2 Qtz + tk + pl = 3 Sil + cf + H_2O	(nk)
(6)	HF = $F(OH)_{-1}$ + H_2O	$(F(OH)_{-1})$

PARTIAL MELTING

Additional Phase Components:

MELT (011–1L) $Si_{383}Al_{116}Na_{15}K_{31}Ca_4(Fe+Mg+Mn)_{29}Ti_2O_{1000}(H2O)_X$
MELT (011–3L) $Si_{406}Al_{99}Na_{35}K_{14}Ca_7(Fe+Mg+Mn)_6Ti_1O_{1000}(H2O)_X$
MELT (036–2L) $Si_{397}Al_{104}Na_{51}K_{11}Ca_9(Fe+Mg+Mn)_8Ti_1O_{1000}(H2O)_X$
<div align="center">X might be between 20 and 30</div>

Exchange Components: NaK_{-1} (nk), $CaFe_{-1}$ (cf), $F(OH)_{-1}$, $MgFe_{-1}$, $MnFe_{-1}$, $CaAlNa_{-1}Si_{-1}$ (pl), $Al_2Fe_{-1}Si_{-1}$ (tk), $FeTiAl_{-2}$ (tx)

Additional Independent Net Transfer Reactions: ξ measured by melt

(7a) 011–1L: 228 Qtz + 9 Sil + 9 Bt + 22 Ms + 19 Ab + 4 pl + 2 Ilm + X H_2O = MELT (011–1L)
(7b) 011–3L: 239 Qtz + 4 Sil + 1 Bt + 13 Ms + 42 Ab + 7 pl + 1 Ilm + X H_2O = MELT (011–3L) + 2 tk
(7c) 036–2L: 278 Qtz + 4 Sil + 2 Bt + 9 Ms + 60 Ab + 9 pl + 1 Ilm + X H_2O = MELT (036–2L) + 1 tk
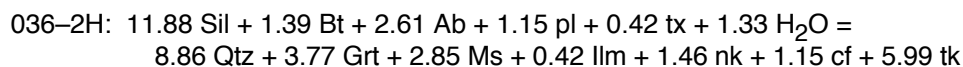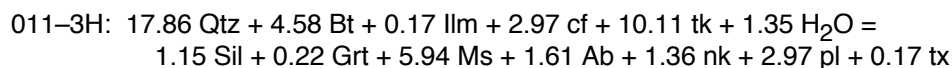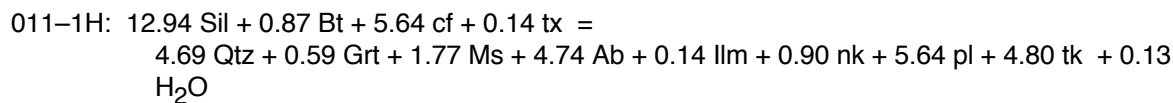<div align="center">X might be between 20 and 30</div>

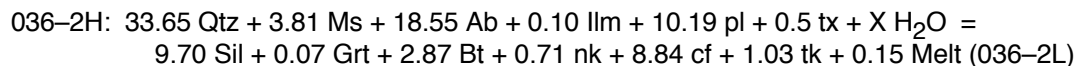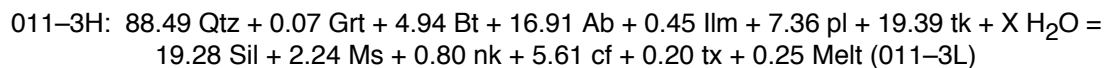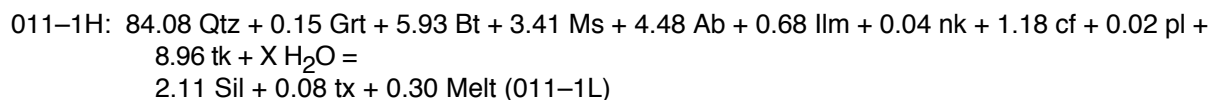Table 1: Mineral assemblage, phase and system components, and linearly independently variable reactions.

ALLEN

EXTENTS of REACTION, if no melting

Reaction	ξ	011–1	011–3	036–2
1	llm	0.14	-0.17	0.42
2	Bt	-0.87	-4.58	-1.39
3	Ab	4.74	1.61	-2.61
4	Grt	0.59	0.22	3.77
5	nk	0.90	1.36	1.46
6	$F(OH)_{-1}$	1.03	0.01	0.13

NET Reactions, assuming no melt was present:

011–1H: 12.94 Sil + 0.87 Bt + 5.64 cf + 0.14 tx =
4.69 Qtz + 0.59 Grt + 1.77 Ms + 4.74 Ab + 0.14 llm + 0.90 nk + 5.64 pl + 4.80 tk + 0.13 H_2O

011–3H: 17.86 Qtz + 4.58 Bt + 0.17 llm + 2.97 cf + 10.11 tk + 1.35 H_2O =
1.15 Sil + 0.22 Grt + 5.94 Ms + 1.61 Ab + 1.36 nk + 2.97 pl + 0.17 tx

036–2H: 11.88 Sil + 1.39 Bt + 2.61 Ab + 1.15 pl + 0.42 tx + 1.33 H_2O =
8.86 Qtz + 3.77 Grt + 2.85 Ms + 0.42 llm + 1.46 nk + 1.15 cf + 5.99 tk

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EXTENTS of REACTION, with melting

Reaction	ξ	011–1	011–3	036–2
1	llm	-0.08	-0.20	0.05
2	Bt	-3.23	-4.69	3.17
3	Ab	1.22	-6.41	-9.55
4	Grt	-0.15	-0.07	0.07
5	nk	-0.36	0.80	0.71
6	$F(OH)_{-1}$	0.70	0.00	0.53
7	MELT	0.30	0.25	0.15

NET Reactions, assuming melt present:

011–1H: 84.08 Qtz + 0.15 Grt + 5.93 Bt + 3.41 Ms + 4.48 Ab + 0.68 llm + 0.04 nk + 1.18 cf + 0.02 pl + 8.96 tk + X H_2O =
2.11 Sil + 0.08 tx + 0.30 Melt (011–1L)

011–3H: 88.49 Qtz + 0.07 Grt + 4.94 Bt + 16.91 Ab + 0.45 llm + 7.36 pl + 19.39 tk + X H_2O =
19.28 Sil + 2.24 Ms + 0.80 nk + 5.61 cf + 0.20 tx + 0.25 Melt (011–3L)

036–2H: 33.65 Qtz + 3.81 Ms + 18.55 Ab + 0.10 llm + 10.19 pl + 0.5 tx + X H_2O =
9.70 Sil + 0.07 Grt + 2.87 Bt + 0.71 nk + 8.84 cf + 1.03 tk + 0.15 Melt (036–2L)

X might be 5 or 6

Table 2: Extents of reactions specified in Table 1, and overall net-transfer reactions, without and with melting.

Holloway, 1988; LeBreton & Thompson, 1988). However, it is probable that a significant amount of water is involved as a reactant in the melting reactions (7), or at least that the melt phase is capable of dissolving a relatively large amount of excess water.

These reactions (7) may not be the melting reactions, just as none of the other six linearly independent net transfer reactions are necessarily the reactions occurring within the rock. Rather, all possible reactions within this assemblage, including melting reactions, can be described by some linear combination of these seven independent reactions. The overall reaction responsible for the migmatization of these rocks can be determined by measuring the extent of progress for each of the linearly independent net-transfer reactions. Since each of these reactions involves changes in the modal abundance of minerals, in a system closed to gain or loss of non-volatile components, the progress of each reaction (ξ_r, Table 1) can be determined by comparing modes of minerals before (schists) and after (migmatites) migmatization.

Due to the heterogeneity in these rocks, particularly within the migmatites, "bulk" mineral modes were calculated from whole rock chemical compositions and the compositions of the constituent mineral phases, following the techniques of Ferry (1984). The details of these calculations are provided in Allen (1992). The average of the schist compositions was used as the best approximation of the protolith material. The calculations were done on an oxygen basis, rather than the volume basis used by Ferry (1984), because the molar volume of the "melt" phase is unknown and the stoichiometry of the melt phase was defined on an oxygen basis (Table 1). An oxygen basis is considered to be approximately equivalent to a volume basis (Chamberlain, 1986). The modes of minerals present in the "bulk" migmatites, and thence the overall migmatization reactions, were calculated both with and without a melt phase. In the melting scenario, the proportion of leucosome determined by a modal analysis of the migmatite outcrops (Allen, 1994) was taken to be the mode of the melt phase. The amount of components Ab, nk and $F(OH)_{-1}$, which measure the ξ_r for reactions (3), (5) and (6), were determined from the calculated modes of the minerals containing these components together with the measured composition of the minerals (Allen, 1992). The ξ_r for the six (or seven, if melt is present) linearly independent net transfer reactions possible in the assemblage (Table 1), and the overall net-transfer reactions determined for the migmatization of these samples (with and without melting) are given in Table 2.

Regardless of whether melting is involved or not, water is generally consumed in the migmatization reactions presented in Table 2. This suggests that infiltration of water from some external source is required to drive the migmatization process. In the case of melting, it is possible that much larger amounts of water may be consumed than if there is no melt. The loss on ignition during XRF analysis, expected to be mostly water, averaged 2.4% for the migmatite whole rock samples, while the loss on ignition from the schist samples averaged only 1.9%. This difference suggests an increase in the water content of the migmatites relative to the schists, consistent with consumption of water by the migmatization reactions.

LEUCOSOMES AS MELTS?

Leucosome compositions shown in Figs. 2c and 3c are similar in character to the compositions of the Wildcat Granite (trip B2, this volume; Allen, 1992) samples (Figs. 2e and 3e), but—unlike the "G" phase of the Wildcat Granite (sample #105-1)—they do not appear to be of granitic minimum melt composition (Fig. 7). The "G" phase of the Wildcat Granite is the only rock in the region that actually contains potassium feldspar. The other samples contain quartz, albitic plagioclase and muscovite as the major components. Thus these leucosomes may fit Ashworth's (1976, 1985) definition of "trondhjemitoid", as distinguished from K–feldspar bearing "granitoid", leucosomes. Trondhjemitoid melts are possible in water-saturated, micaceous but K–feldspar absent rocks, particularly with a more sodic plagioclase (Ashworth, 1985; Johannes, 1985, 1983, 1978), as is present in these

rocks. Also, the addition of F and increases in P_{H2O} both move the minimum melt in the granite system (Qtz–Ab–Or) towards a more trondhjemitic composition (Qtz–Ab) (Manning, 1981; Luth *et al.,* 1964).
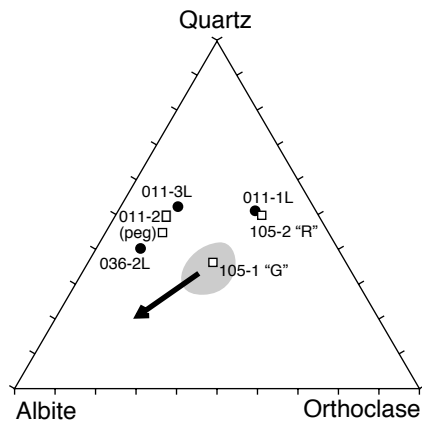

Figure 7: Normative Qtz-Ab-Or mineralogy of migmatite leucosomes (filled circles) and granite samples (open squares). Gray field shows area of commonly observed granite compositions (LeMaitre c.f. Best, 1982, p. 114), which corresponds with the granitic "minimum melt" composition (Tuttle & Bowen, 1958). Arrow shows direction that "minimum melt" composition moves with both increasing water pressure (Luth et al., 1964; Johannes, 1978) and increasing fluorine content (Manning, 1981). Samples 011–1L and 105–2 "R" had total normative Qtz + Ab + Or < 80%, indicating that their compositions are not well represented by this diagram.

Data presented in Figs. 3c & 3d and Fig. 8 suggest that the Rare Earth Elements (REE) and U, Th and Zr are preferentially retained in the melanosomes, to become more concentrated than in the parent schist, and do not enter the leucosomes. The distribution of the REE and U, Th and Zr in high-grade pelitic rocks and in granitic systems is strongly controlled by the behavior of accessory mineral phases such as monazite and zircon (Watson & Harrison, 1984). Melanosome biotites observed in thin section tend to have a high abundance of pleochroic radiation damage halos due to small inclusions of monazite and/or zircon. These halos are much less abundant in biotites from the schists. Studies showing that these minerals may have very low solubility in granitic melts (Rapp & Watson, 1986; Rapp *et al.,* 1987; Harrison & Watson, 1983) further suggest that the leucosomes may indeed be melts. Similar REE and U, Th, and Zr distributions have been
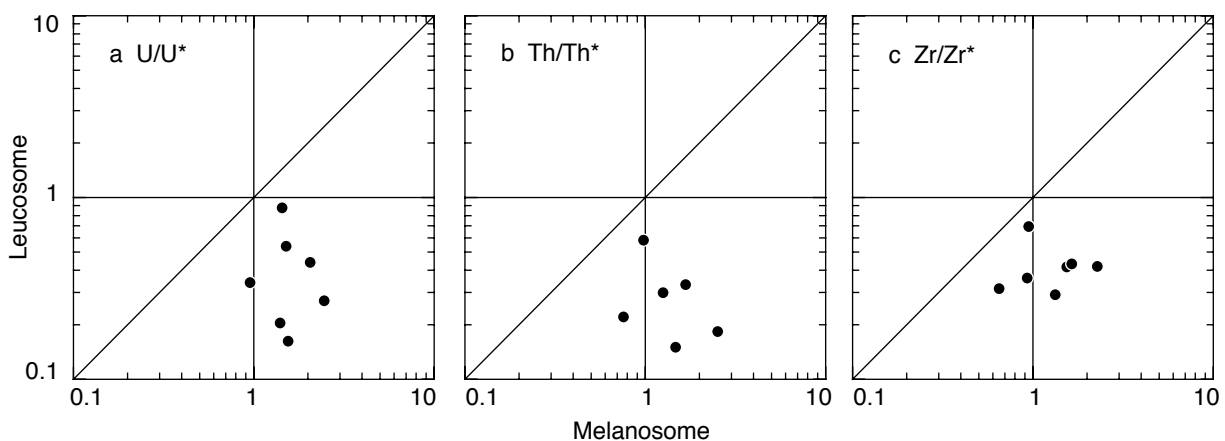

Figure 8: Uranium (a), Thorium (b), and Zirconium (c) concentrations in migmatite leucosomes and melanosomes, normalized by the concentration in the average schist (U*, Th*, Zr*; Allen, 1992). The diagonal lines represent the loci of points for which leucosome and melanosome concentrations are equal. The horizontal and vertical lines at unity represent the loci of points where the leucosome or melanosome concentrations are the same as in the average schist.

observed in migmatites studied by Weber *et al*. (1985) and by Barbey & others (1989, 1990). Petrochemical modelling by these authors suggests that their observations are are best explained by complex melt-fluid interactions

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(Weber *et al.,* 1985) or disequilibrium partial melting (Barbey & others, 1989, 1990).

ISOTOPIC FRACTIONATION

Fig. 6 shows that for several of the migmatite samples studied, melanosome $\delta^{18}O_{Qtz}$ is lower than leucosome $\delta^{18}O_{Qtz}$ by up to 1‰, a significant and measurable difference (Allen, 1992). Similar results were obtained by Mazurek (1992) studying migmatite samples from drill cores of crystalline basement in Switzerland. This difference is better depicted in Fig. 9, constructed in a similar manner to Fig. 8, comparing leucosome to melanosome.
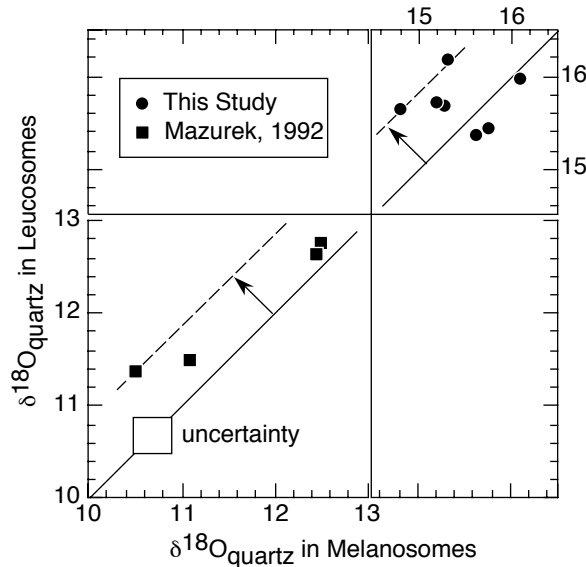

Figure 9: Oxygen isotope compositions of quartz from migmatite leucosome and melanosome pairs from this study and from Mazurek (1992). The diagonal line across the center of the diagram represents the loci of points for which quartz in the leucosome has the same isotopic composition as quartz in the melanosome. The dashed line parallel to the central diagonal indicates possible fractionation of about 1 ‰ during partial melting. The uncertainty associated with these analyses is ±0.2‰ and is shown by the labelled box. Vertical and horizontal scales were broken to keep the diagram at the same scale as Figs. 4 and 6.

Possible mechanisms for producing such a shift include: (1) re-equilibration of the melanosome mineral assemblage during cooling; (2) selective post-migmatization alteration, in which late fluids are able to infiltrate only melanosome or leucosome portions; (3) oxygen isotope fractionation between melt and solids; and (4) disequilibrium melting, in which the melt incorporates infiltrated fluids, thereby changing its isotopic composition.

Closed-system isotopic re-equilibration of the mineral assemblages in leucosome and melanosome during post-migmatization cooling could have produced differences in the isotopic composition of minerals in the leucosome and melanosome, because of the very different modal mineralogies. However, given the mineral assemblages in these rocks, one would expect melanosome quartz to take on higher $\delta^{18}O_{Qtz}$ values compared to leucosome $\delta^{18}O_{Qtz}$, rather than lower (Allen, 1992, 1996).

Continued isotopic exchange with an external fluid after migmatization could have produced the observed differences, if the exchange occurred only in the melanosomes or the leucosomes. While it is possible that some difference in permeability may have existed, the presence of retrograde muscovite in both leucosome and melanosome argues against this. Fig. 6 shows that the $\delta^{18}O$ of muscovites from leucosome and melanosome are essentially identical given the uncertainties.

Incorporation of infiltrating fluids into the melt phase could have produce the differences, if the fluid had an isotopic composition that was relatively heavy compared to the rocks undergoing melting, as Mazurek (1992) argued. In Pinkham Notch, however, the infiltrating fluids driving the migmatization were isotopically light relative to the rocks. Incorporation of this fluid into molten leucosomes should have lowered the $\delta^{18}O$ values of the

leucosomes relative to the melanosomes, rather than raised them.

Of the four possibilities for producing the observed shift in leucosome $\delta^{18}O_{Qtz}$, we are left with oxygen isotope fractionation between melt and solids during partial melting and crystallization. Taylor & Sheppard (1986) have considered the possibility of such fractionation and concluded that any fractionation between rocks and melts of identical composition is negligible. Garlick (1966) has also demonstrated that oxygen isotope fractionation is independent of physical state—molten or solid. Mazurek (1992) ruled out fractionation between minerals and melt as an explanation of the observed difference in $\delta^{18}O_{Qtz}$ from leucosome and melanosomes on the basis of these studies. However, these studies neglect the fact that in partial melting of pelitic rocks, the initial melt produced can be of very different composition than the residual solids.

Using the techniques of Chamberlain *et al.* (1990), which provide a means of calculating the effect of net transfer reactions on the isotopic composition of minerals, I have modelled the isotopic evolution of quartz in leucosomes and melanosomes in these rocks using the migmatization reactions presented in Table 2 (Allen, 1992, 1996). I believe that the observed shift in the $\delta^{18}O_{Qtz}$ of migmatite leucosome relative to melanosome may be best explained by oxygen isotope fractionation during partial melting. Such fractionation has here-to-fore not been recognized. The fact that not all migmatite leucosome-melanosome quartz pairs have a difference in $\delta^{18}O$ does not invalidate the model, but suggests that either more than one migmatization process may be operating, or that the fractionation due to melting may vary depending on details of the melt composition.

DISCUSSION AND CONCLUSIONS

The schists at Stop 1b represent the parent material of the migmatites at Stop 1c, and the migmatization process appears to have been largely closed to gain or loss of non-volatile components. There is a steep metamorphic and thermal gradient across the migmatite front, and the migmatization reactions involved infiltration of hydrous fluids, resulting in a lowering of $\delta^{18}O_{WR}$ in the migmatites and changes in $\delta^{18}O_{mineral}$ systematics from closed to open system behavior. Trace element and oxygen isotope partitioning suggest an anatectic partial melting migmatization process, producing trondhjemitoid melts, driven by the infiltrating fluids. The presumed oxygen isotope composition of the fluids, its relationship to the isotopic composition of pegmatites and granites in the area, and the role of F and P as fluxes for melting, suggests that these fluids may have been of magmatic origin.

There are abundant pegmatites and scattered small bodies of granite throughout the migmatite zone. It is not clear, however, that these small bodies could have supplied the volume of fluids necessary to cause the isotopic and thermal changes observed. Large volumes of fluid must have exchanged oxygen isotopes with the rock in order to produce the observed lowering of $\delta^{18}O_{WR}$ (Allen, 1992, 1996).

Since they are not exposed at the surface, the plutons from which these large volumes of fluids might have originated must either be buried below the current erosion surface, or they passed through this zone and were emplaced in rocks above those now exposed and have been eroded away (Fig. 5 of trip B2, this volume). While there is no direct evidence to rule out the first of these two possibilities, evidence from geologic and structural mapping (trip B2, this volume; Allen, 1992) may support upward pluton migration through the area. This model may also explain the variations in metamorphic P–T paths observed along strike in the Central Maine Terrane. In Maine, P–T paths show isobaric heating (Lux *et al.,* 1986; DeYoreo *et al.,* 1989), while in central New Hampshire, P–T paths show increasing pressure during heating, and in Massachusetts, P–T paths are "counterclockwise" loops showing continued increasing pressure during cooling (Tracy & Robinson, 1980; Robinson *et al.,* 1989). These P–T paths can be explained by the transfer of heat and mass by pluton migration upward through the crust during metamorphism. Granite emplacement in the upper levels of the crust could produce the heating observed in Maine

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and the tectonic loading observed in New Hampshire and Massachusetts.

ROAD LOG

This trip will re-visit many of the stops of the 1986 NEIGC trip (B3) lead by Ellen Wall and the late Norman Hatch. Many of these stops are pertinent to the stratigraphy and structure in the region as discussed in trip B2, this volume; participants are urged to read the article for that trip. All of these stops can be located on Washburn's 1:20,000 map of "Mount Washington and the Heart of the Presidential Range" (1988).

Mileage

0.0 Great Gulf Wilderness Parking Area on the west side of NH 16, about 6 miles south of its intersection with US 2 in Gorham, or 3 miles north of Wildcat Ski Area.. Walk down the trail and across footbridge over the Peabody River; we will return to these outcrops later (Stop 1C). About 150 feet beyond the footbridge, the old Great Gulf Link trail enters sharply from the left; follow this abandoned trail north past some rock ledges on the west bank of the Peabody River (Stop 1B) until it joins the re-located Great Gulf Link trail, and then about 300 feet further to rock ledges on the west bank of the Peabody River.

STOP 1A (OPTIONAL) SCHISTS and QUARTZITES of the PERRY MOUNTAIN FORMATION (Stop 1A/2A of Hatch and Wall, 1986) (15 minutes): Interbedded quartzites and mica schists, with well preserved bedding, locally graded, with some boudined calc-silicate beds forming "pods."

STOP 1B SCHISTS of the RANGELEY FORMATION (Stop 2C of Hatch and Wall, 1986, Outcrop #010 of Allen, 1992) (30 minutes): Coarse-grained schists with some interbedded quartzites and granofels. Calc-silicate "pods" occur in groups along bedding planes. Sillimanite nodules (pseudomorphs after andalusite) and un-oriented spangles of retrograde muscovite give the bedding/foliation surfaces a knobby texture. While there are some bed-to-bed variations, the major element, trace element, and stable isotope compositions are all consistent with a metamorphosed shale. There is no evidence that these rocks reached the sillimanite-potassium feldspar (muscovite absent) grade.

STOP 1C MIGMATITES of the RANGELEY FORMATION (Stop 2D of Hatch and Wall, 1986; Outcrop #011 of Allen, 1992) (45 minutes): The migmatite leucosomes are usually elongate blebs or stringers, rather than continuous layers and the intensity of migmatization is highly variable. Calc-silicate "pods" are abundant, and usually lie parallel to the migmatite layering, although sometimes they are at an angle to it. These rocks are obviously quite heterogeneous, but in bulk they are geochemically similar to the protolith schists at Stop 1B (Figs.

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2 & 3). This suggests that the migmatization of these rocks is the result of metamorphism in an iso-chemical or "closed" system. The migmatites have a lower oxygen isotope value than the schists (Fig. 1) and exhibit open-system isotopic behavior (Fig. 4), which suggests infiltrating fluids may have been important in driving the migmatization process. Isotopic fractionation between the melanosome and leucosome components of the migmatites may be the result of partial melting (Fig. 9). These migmatites have an identical metamorphic mineral assemblage to that at Stop 1B (sillimanite-muscovite), but garnet-biotite geothermometry indicates temperatures were higher. Two generations of pegmatite cut across the outcrop.

0.0 Return to vehicles and drive south on NH 16, towards Wildcat Ski Area and Jackson, NH.
0.3 Turn left into 19 Mile Brook trailhead parking area. Walk across the highway bridge over 19 Mile Brook, cross route 16 (watch for traffic!), and head down a gated access trail along the brook.

STOP 2 RANGELEY MIGMATITES (Stop 7 of trip B2, this volume) (20 minutes): Extensive washed and pot-holed outcrops of gray to rusty orange migmatitic gneiss. The rocks are generally gray adjacent to granite/pegmatite intrusions, rusty elsewhere. Locally the gneiss appears to "grade" into granite. There are "diffuse" pegmatite bodies, as well as pegmatites with distinct sharp contacts. Pods with moats appear to occur in horizons, generally parallel to the layering. A nice polished surface on the floor of a pool in 19 Mile Brook, approximately 100 feet upstream from its confluence with the Peabody River, shows leucosome/melanosome layering in the migmatite gneiss and the internal structure of pods and their relationship to one another (a family of pods within a larger pod!).

1.3 pass entrance to Mount Washington Auto Road on the right
1.6 paved parking area on the right. The road cuts opposite are the destination of a traverse made in Stop 3B.
2.0 Pull off into a paved parking area on the right hand side of the road. Park near the upper end. Make your way down to Emerald Pool and then right over ledges and through woods to a small beach just upstream of the pool.

STOP 3A EMERALD POOL (Stop 8 of trip B2, this volume; Stop 3A of Hatch & Wall, 1986) (30 minutes): Here we can examine almost the entire section in one spot, including orange colored Rangeley migmatite gneiss, rusty and flaggy Smalls Falls quartzites and schists, well banded light- and dark-green calc-silicate granofels of the Madrid, and aluminous gray Littleton gneiss.

Return to the parking area and carefully cross the highway to roadcuts on the east side.

STOP 3B LITTLETON to RANGELEY SECTION (Stop 3B of Hatch & Wall, 1986) (45 minutes): Walk north along this half-mile long series of outcrops on the east side of NH 16 and find the transition from the gray "sinewy" migmatites of the lower Littleton rocks back to the rusty of the Rangeley Formation, with distinctive pods and exotic cobbles. An outcrop of clean quartzite and schist just off the road in the woods between these roadcuts may be Perry Mountain. The rocks at Emerald Pool and along these roadcuts appear to form an isolated septum of Smalls Falls, Madrid and Littleton rocks within the migmatites of the Rangeley Formation (Fig. 1).

2.3 Pull off into a paved parking area on the right hand side of the road. Park near the lower (northern) end, where access to pavement outcrops in the Peabody River is obvious.

STOP 4 WILDCAT GRANITE (Stop 9 of trip B2, this volume; Stop 4 of Hatch and Wall, 1986; Outcrop #105 of Allen, 1992): For those familiar with the controversy, *Granitization* rears its ugly head! This is the type locality for the rock I have called Wildcat Granite (trip B2, this volume; Allen, 1992). There are clearly two different phases: "G" consists of medium grained, whitish-weathering two-mica granite; "R" is much coarser grained,

orangish-weathering granitoid, also bearing both muscovite and biotite, but with much more abundant biotite. "R" contains abundant calc-silicate pods, identical to those found in the metasediments, rimmed by strong reaction zones, and its textures and mineralogy suggest that it may be formed from completely melted and recrystallized Rangeley schists. Both the "G" and "R" phases are extensively intermingled in a complex fashion, with wispy biotite-rich schlieren observed throughout the outcrop. Geochemically, the "G" phase is characteristic of an "S" type granite (Figs 1, 2, 3 & 7); the "R" phase has a composition intermediate between that of the "G" phase and those of the schists and migmatites. Second-generation white, tourmaline-bearing pegmatites cross-cut both phases

3.1 Turn left into Wildcat Ski Area.

   STOP 5 (OPTIONAL) THOMPSON'S FALLS (45 minutes): Follow the Thompson Falls Trail 0.8 miles to its end at ledges of Wildcat Granite grading into Rangeley migmatites. The contact between the Wildcat Granite and the surrounding migmatitic metasedimentary rocks is gradational—not a sharp intrusive contact. Excellent views of Mt. Washington and its ravines.

4.0 Turn right into the AMC's Pinkham Notch Base Camp. Parking may be difficult to find, in which case continue to the auxiliary parking lot on the south side of the Cutler River. Several optional stops may be made from here.

STOP 6A (OPTIONAL) CRYSTAL CASCADE (30 minutes): Follow the Tuckerman Ravine Trail 0.3 miles to the Crystal Cascade of the Cutler River. The rocks at the Cascade are mesozoic volcanic vent agglomerates (Billings, *et al.,* 1979). Downstream are exposures of granite and pegmatite, and laminated rusty schists of the Smalls Falls Formation.

STOP 6B (OPTIONAL) OLD JACKSON ROAD to LOWE'S BALD SPOT (2 hours): Follow the Old Jackson Road (trail) 1.8 miles to the Mount Washington Auto Road, roughly following the contact between the Smalls Falls and Madrid or Littleton Formations.This belt of Smalls Falls and Madrid is distinct from that at Emerald Pool (Stop 3A) and they do not connect. Continue 0.2 miles on the Madison Gulf Trail to Lowe's Bald Spot, a nice exposure of Littleton schists.

STOP 6C (OPTIONAL) SQUARE LEDGE (1 hour): Cross to the east side of NH 16 carefully. Follow the Square Ledge Trail 0.5 miles to the top of Square Ledge, an outcropping of non-descript gray Rangeley migmatites with an excellent overlook of Pinkham Notch and view of Mt. Washington and its ravines.

4.5 Pull off on right shoulder of NH 16, at a small waterfall opposite a turnout on the northbound side and just before a turnout on the southbound side.

STOP 7 SMALLS FALLS, MADRID & LITTLETON FORMATIONS (15 minutes): The rocks at the base of the falls are laminated rusty schists of the Smalls Falls Formation. These are overlain by layered green calc-silicate granofels of the Madrid Formation near the top of the falls, and above that by well bedded aluminous schists of the Littleton Formation.

4.7 Turn right into Glenn Ellis Falls Scenic Area. Follow path under NH 16 and down to falls.

STOP 8 (OPTIONAL) GLENN ELLIS FALLS (30 minutes): Abundant pegmatite and pod-bearing Wildcat Granite. Extremely punky weathering and possibly brecciated.

5.2 (add 0.4 if you went in to Glenn Ellis Falls) Turn left into turnout on east side of NH 16 overlooking the Ellis River valley and Jackson, NH. Carefully cross the highway to roadcuts on west side.

STOP 9 OVERLOOK, more WILDCAT GRANITE (15 minutes): A coarse grained granular rock, with abundant calc silicate pods and dark rusty red stains. Small weathering pits on the surface may represent cordierite. I have not studied this rock in detail, but would assign it to the Wildcat Granite, representing melted and recrystallized Rangeley and possibly Smalls Falls Formations.

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