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Kurt Hollocher

Guidebook to Selected Mineral Localities in the Northeast and their Geological Context

edited by
John B. Brady and John T. Cheney

Teaching Mineralogy Workshop

NATIONAL · SCIENCE · FOUNDATION

Smith College - June 1996

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Fort Edward
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Corinth
Benson
Mount McGregor
Argyle
West Hebron
Palmer
Arietta
Upper Benson
Northville


QUABBIN RESERVOIR
NORTHAMPTON
HOLYOKE
CHICOPEE
SPRINGFIELD
West Springfield
Amherst
Easthampton
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Ludlow
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Thompsonville
Wilbraham
Monson
Brookfield
West Brookfield
North Brookfield
Hardwick
Gilbertville
PRESCOTT PENINSULA
HOLYOKE RANGE
WESTOVER FIELD
(Military Reservation)
HAMPSHIRE CO
HAMPDEN CO
WORCESTER CO
MASSACHUSETTS
CONNECTICUT
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WINDHAM CO
Baffle Dam
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Quabbin Dike
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HARTFORD 19 MI.
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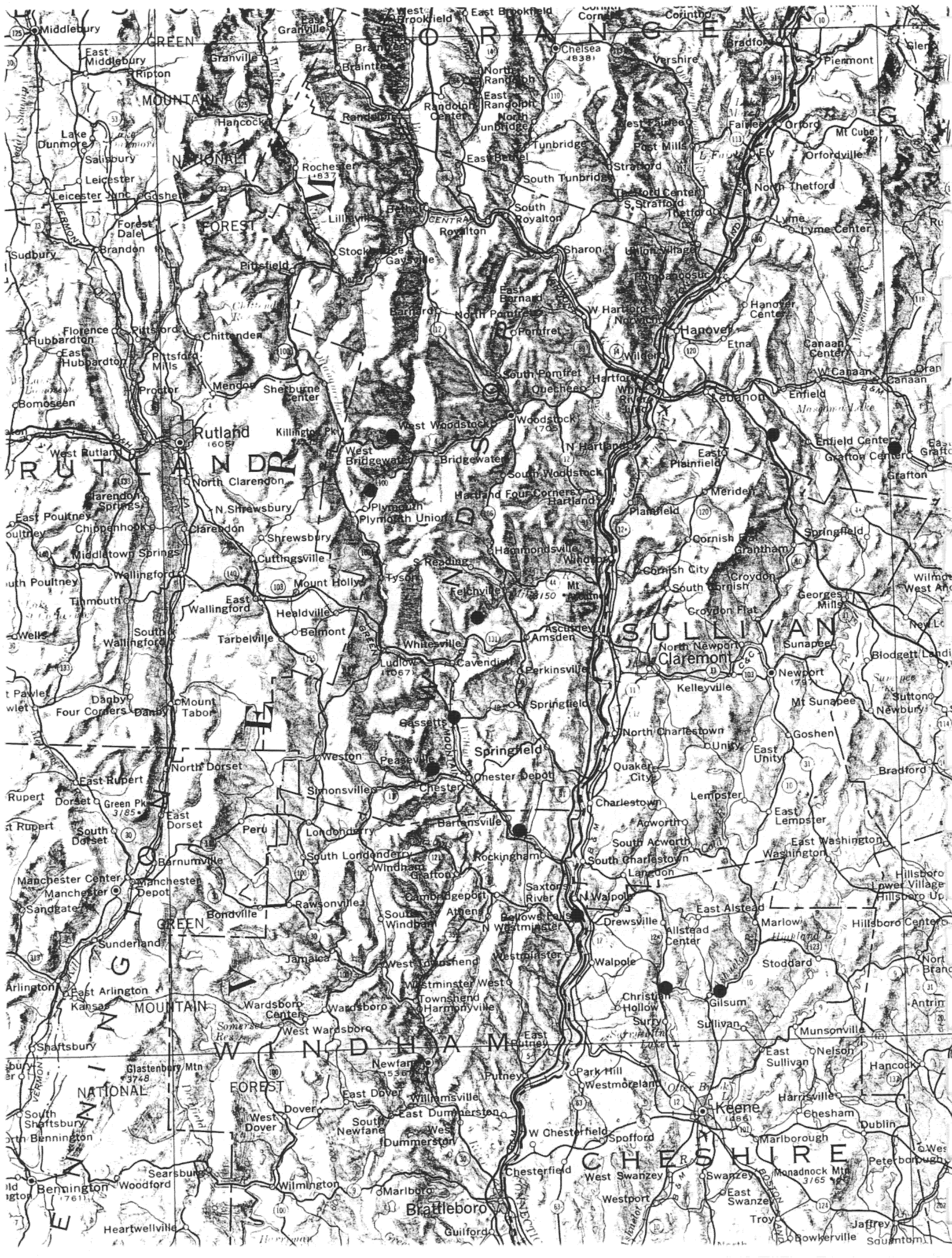

Middlebury
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NATIONAL
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Hancock
Granville
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Salisbury
Leicester
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Goshen
Rochester
Forest Dale
Brandon
Sudbury
Pittsfield
Florence
Pittsford
Hubbardton
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Middletown Springs
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White River Junc
Lebanon
W Canaan
Canaan
Enfield
Mascoma Lake
Killington Pk
West Woodstock
Woodstock
N Hartland
Enfield Center
Grafton Center
Grafton
West Bridgewater
Bridgewater
East Plainfield
South Woodstock
Meriden
Hartland Four Corners
Hartland
Plainfield
Plymouth
Plymouth Union
Cornish Flat
Springfield
Grantham
Hammondsville
Windsor
S. Reading
Cornish City
Croydon
Tyson
Felchville
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Marlborough
CHESHIRE
Chesterfield
West Swanzey
Swanzey
Monadnock Mtn
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Peterborough
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Marlboro
Westport
East Swanzey
Brattleboro
Troy
Jaffrey
Guilford
Bowkerville

GEOCHRONOLOGY AND PETROGENESIS OF ADIRONDACK IGNEOUS AND METAMORPHIC ROCKS

James M. McLelland, Department of Geology, Colgate University, Hamilton, NY 13346

INTRODUCTION AND GEOCHRONOLOGY

The Adirondacks form a southwestern extension of the Grenville Province (figure 1) and are physiographically divided into the Adirondack Highlands (granulite facies) and Lowlands (amphibolite facies) by a broad zone of high strain referred to as the Carthage-Colton Mylonite Zone (figs. 2,3) which is continuous with the Chibougamau-Gatineau line, or Labelle shear zone, in Canada (AB on figure 1). Together these two zones separate the Grenville Province into two major blocks with the Central Granulite Terrane (CGT) lying east of AB and the Central Metasedimentary Belt (CMB) and Central Gneiss Belt (CGB) lying to the west. Within the southwestern portion of the Grenville Province further subdivisions exist and are shown in figure 3.

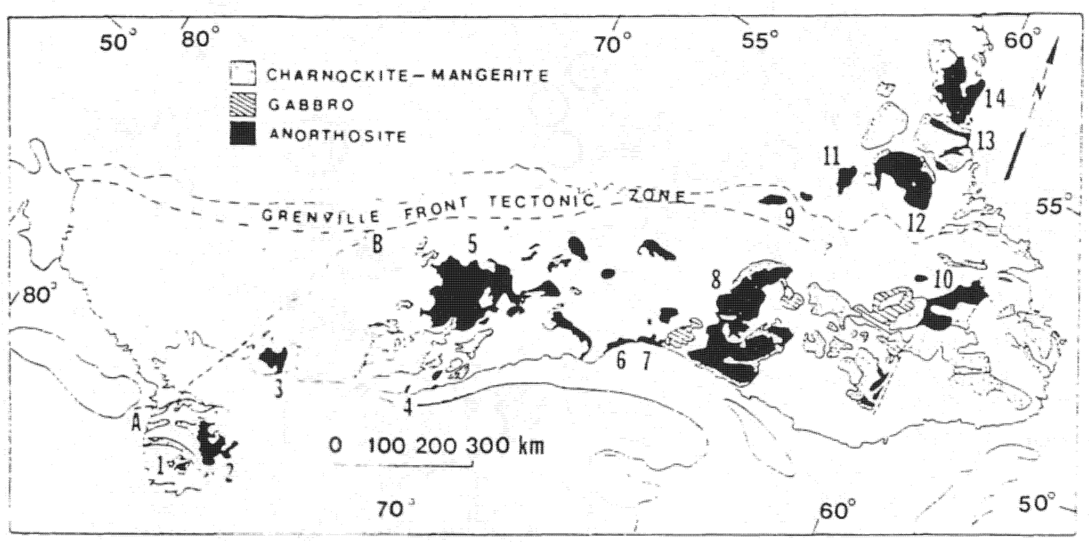

Figure 1. Generalized map of anorthositic massifs within the Grenville Province and adjacent Labrador. The dashed line, AB, separates terranes with anorthosite massifs on the east from ones lacking them on the west and corresponds to the Carthage-Colton-Gatineau-Chibougamau Line. 1-Snowy Mt. and Oregon domes (ca. 1130 Ma); 2-Marcy massif (ca. 1135 Ma); 3-Morin anorthosite and Lac Croche complex (1160±7 Ma); 4-St. Urbain anorthosite (ca. 1070 Ma); 5-Lac St. Jean complex (1148±4 Ma); 6-Sept Isles (1646±2 Ma); 7-8-Havre St. Pierre complex (1126±7 Ma) including the Pentecote (1365±7 Ma) anorthosite; 9-Shabagamo intrusives; 10-Mealy Mts. anorthosite (1646±2 Ma); 11-12-Harp Lake anorthosite (ca. 1450 Ma); 13-Flowers River complex (ca. 1260 Ma); 14-Nain complex (1295 Ma) including Kiglapait intrusive (1305±5 Ma). From McLelland (1989)

As demonstrated by recent U-Pb zircon and Sm-Nd geochronology summarized (table 1) by McLelland and Chiarenzelli (1990) and McLelland et al. (1993), the Adirondack-CMB sector of the Grenville Province contains large volumes of metaigneous rocks that represent recent (i.e., ca. 1400-1200 Ma) additions of juvenile continental crust. These results (figure 4) indicate that the Adirondack-CMB region experienced wide-spread calcalkaline magmatism ca. 1300-1220 Ma. Associated high grade metamorphism has been fixed at 1226±10 Ma by Aleinikoff (pers. comm.) who dated dust air abraded from metamorphic rims on 1300 Ma zircons. Identical rocks, with identical ages, have been described from the Green Mts. of Vermont by Ratcliffe et al. (1991), in northern Ireland by Menuge and Daly (1991), in the Llano uplift of Texas (Walker, 1993) and in the Texas-Mexico belt of Grenville rocks by Patchett and Ruiz (1990). It appears, therefore, that a major collisional-magmatic belt was operative along the present southern flank of the Grenville Province during the interval 1300-1220 Ma and may have been related to the assembly of a supercontinent. More locally, this magmatism and associated metamorphism, represents the Elzevirian Orogeny of the Grenville Orogenic Cycle. as defined by Moore and Thompson (1980). Within the Adirondacks, Elzevirian rocks are represented by 1300-1220 Ma tonalites and alaskites whose distribution is shown in figure 5. The apparent absence of this suite from the central Highlands is believed to be the combined result of later magmatic intrusion and recent doming along a NNE axis. Within the Frontenac-Adirondack region, the Elzevirian Orogeny was followed by 40-50 Ma of quiescence terminated at 1170-1130 Ma by voluminous anorogenic (figure 4) magmatism referred to as the

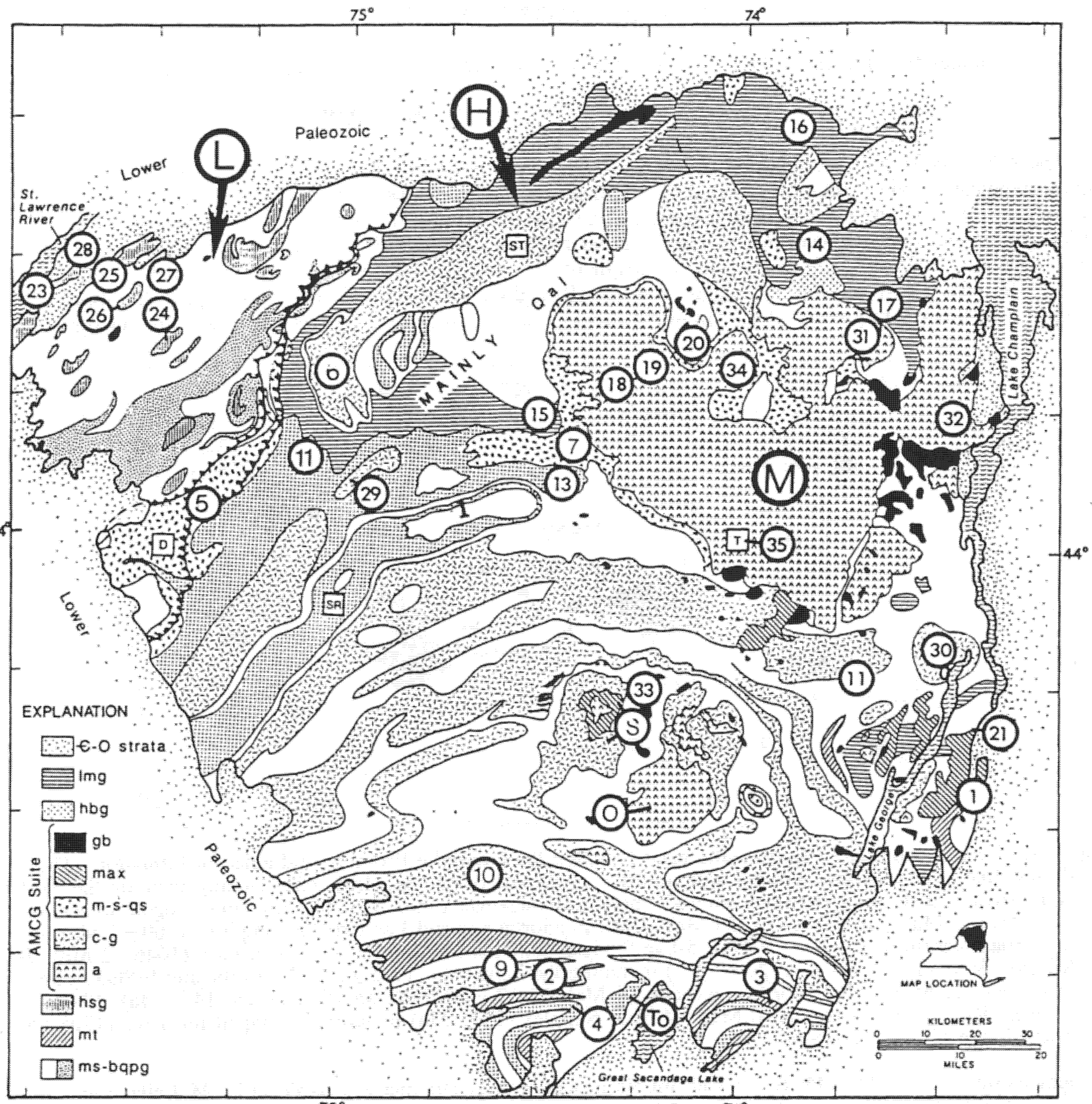

Figure 2. Generalized geologic map of the Adirondack Highlands (H) and Lowlands (L). The Carthage-Colton Mylonite Zone (CCMZ) is shown with sawteeth indicating directions of dip. Numbers refer to samples listed in Tables 1 and 2. Map symbols: lmg=Lyon Mt. Gneiss, hbg=hornblende-biotite granitic gneiss, gb-olivine metagabbro, max-mangerite with andesine xenocrysts, a=metanorthosite, m-s-qs=mangeritic-syenitic-quartz-syenitic gneiss, ms=metasediments, bqpg=biotite=quartz-plagioclase gneiss, hsg=Hyde School Gneiss, mt=metatonalitic gneiss. Locality symbols: A=Arab Mt. anticline, C=Carthage anorthosite, D=Diana complex, O=Oregon dome, S=Snowy Mt. dome, ST=Stark complex, SR=Stillwater Reservoir, T=Tahawus, To=Tomantown pluton. From McLelland and Chiarenzelli (1990) and Daly and McLelland (1991).

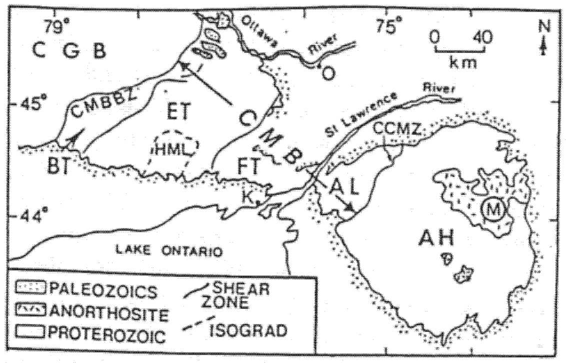

Figure 3. Southwestern Grenville Province. CMB=Central Metasedimentary Belt, CGB=Central Gneiss Belt, BT=Bancroft Terrane, ET=Elzevir Terrane, FT=Frontenac Terrane, AL=Adirondack Lowlands, HL=Adirondack Highlands, HML=Hastings metamorphic low, K=Kingston, O=Ottawa, CCMZ=Carthage-Colton Mylonite Zone, M=Marcy massif.

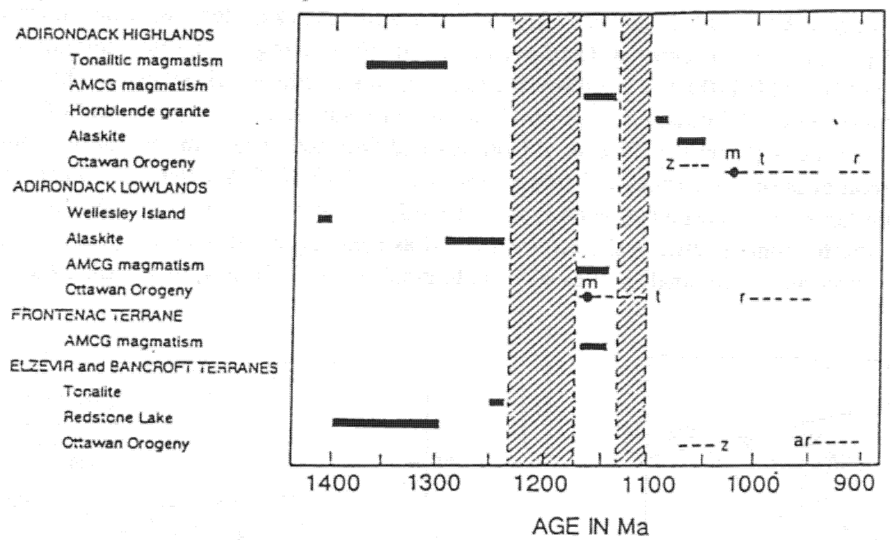

Figure 4. Chronology of major geological events in the southwestern Grenville Province. z=zircon, t=titanite, m=monazite, r=rutile, ar=Ar/Ar. Diagonal ruling=quiescence. From McLelland and Chiarenzelli (1991).

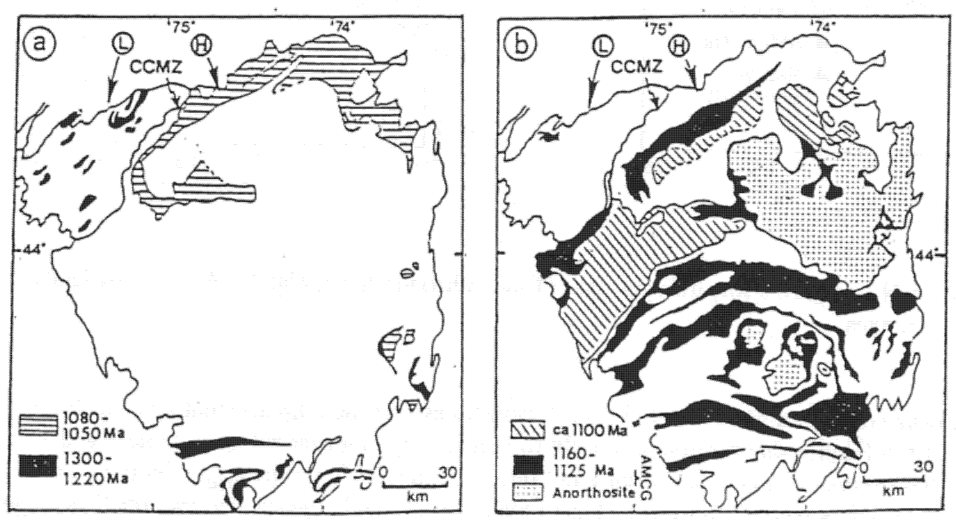

Figure 5. Chronological designation of Adirondack units. L=Adirondack Lowlands, H=Adirondack Highlands, CCMZ=Carthage Colton Mylonite Zone. From Chiarenzelli and McLelland (1991).

anorthosite-mangerite-charnockite-granite (AMCG) suite. The older ages are characteristic of AMCG magmatism in the Frontenac Terrane (including the Lowlands) while the Highlands commonly exhibit ages of 1150-1130 Ma (figure 5). The large Marcy anorthosite massif (figure 2) and its associated granitoid envelope were emplaced at ca. 1135 Ma (McLelland and Chiarenzelli, 1990). These ages are similar to those determined (Emslie and Hunt, 1990) for the Morin, Lac St. Jean, and several other large massifs farther northeast in the Grenville Province (figure 1). Rocks of similar age and chemistry (i.e., Storm King Granite) have been described within the Hudson Highlands (Grauch and Aleinikoff, 1985). The extremely large dimensions of the AMCG magmatic terrane emphasize its global-scale nature corresponding, perhaps, to supercontinent rifting with the rifting axis located farther to the east. Valley et al. (1990) and McLelland et. al. (1991) have provided evidence that contact, and perhaps also regional, metamorphism accompanied emplacement of hot (~1100°C), hypersolvus AMCG magmas. Wollastonite and monticellite occurrences related to thermal pulses from AMCG intrusions occur in proximity to AMCG intrusions (Valley et al., 1990). In the Lowlands and the Canadian sector of the Frontenac Terrane, monazite (table 1., no. 28), titanite, and garnet ages (Mezger et al., 1992) all indicate high temperatures (~600-800°C) at ca. 1150 Ma. Rutile ages and Rb/Sr whole rock isochron ages document temperatures not exceeding ~400 °C at ca. 1050-1000 Ma.

Following approximately 30 Ma of quiescence (figure 4), the Adirondacks, along with the entire Grenville Province, experienced the onset of the Ottawan Orogeny of the Grenville Orogenic Cycle. Initially the Ottawan Orogeny was represented by 1090-1100 Ma hornblende granites in the northwest Highlands. These rather sparse granites were followed by deformation, high grade metamorphism, and the emplacement of trondhjemitic to alaskitic magnetite-rich rocks (Lyon Mt. Gneiss of Whitney and Olmsted, 1988) in the northern and eastern Adirondacks. The zircon ages of these rocks fall into an interval of 1050-1080 Ma (table 1) which corresponds to the peak of granulite facies metamorphism when crust, currently at the surface, was at ~25 km. Accordingly, the alaskitic to trondhjemitic rocks are interpreted as synorogenic to late-orogenic intrusives. They were followed by the emplacement of small bodies of fayalite granite (ca. 1050 Ma) at Wanakena and Ausable Forks (figure 2).

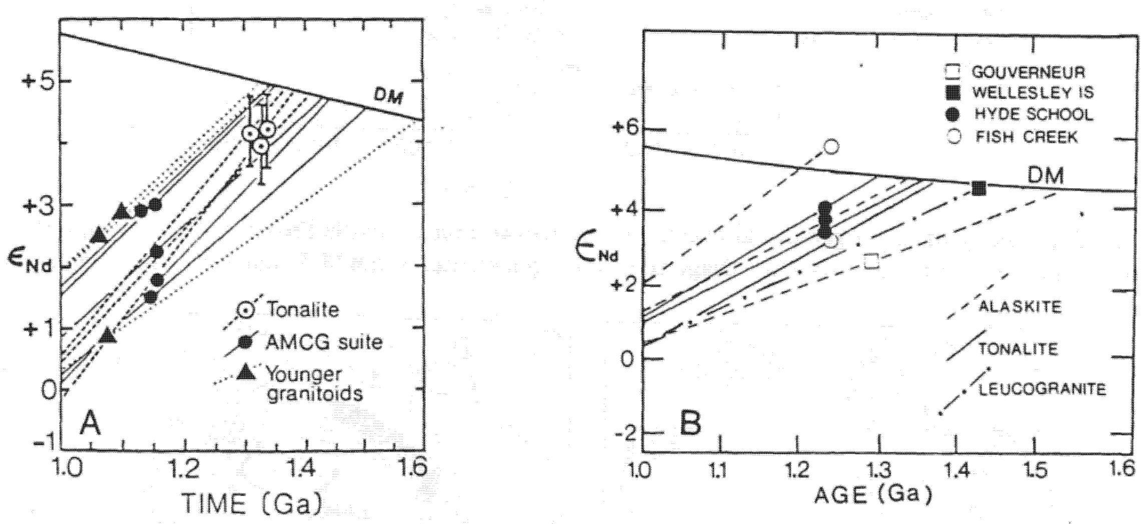

Figure 6. Epsilon-Nd diagrams for orthogneisses of the Adirondack Highlands (A) and Lowlands (B). Symbols fixed by zircon ages.

Sm-Nd analysis (Daly and McLelland, 1991) demonstrates that the emplacement ages of the ca. 1300 Ma tonalitic rocks of the Highlands correspond closely to their neodymium model ages (table 1 and figure 6) indicating that these represent juvenile crustal additions. As seen in figure 6, ϵ_{Nd} evolution curves for AMCG and younger granite suites pass within error of the tonalitic rocks and suggest that the tonalites, together with coeval granitoids, served as source rocks for succeeding magmatic pulses. Remarkably, none of these igneous suites gives evidence for any pre-1600 Ma crust in the Adirondack region and the entire terrain appears to have come into existence in the Middle to Late Proterozoic. Significantly, Sm-Nd analysis for the ca. 1230-1300 Ma tonalitic to alaskitic Hyde School Gneiss at the Lowlands (table 1, figure 6) demonstrates that it has model neodymium ages and ϵ_{Nd} values similar to Highland tonalites (McLelland et al., 1993). These results are interpreted to reflect the contiguity of the Highlands and Lowlands at ca. 1300 Ma. Given this, the Carthage-Colton Mylonite Zone is interpreted as a west-dipping extensional normal fault that formed during the Ottawan Orogeny in response to crustal thickening by thrust stacking (McLelland et al., 1993). East dipping extensional faults of this sort and age have been described by van der Pluijm and Carlson (1989) in the Central Metasedimentary Belt. Motion of this sort along the Carthage-Colton Mylonite Zone would help to explain the juxtaposition of amphibolite and granulite facies assemblages across the zone. A downward displacement of 3-4 km would satisfactorily explain the somewhat lower grade of the Lowlands terrane.

PETROLOGIC CHARACTERISTICS OF THE PRINCIPAL ROCK TYPES IN THE ADIRONDACKS

The following discussion is divided into igneous and metasedimentary sections. Whole rock analyses for granitoids are given in table 2 while those for anorthositic and gabbroic rock appear in tables 3 and 4.

Igneous Rocks

Tonalites and related granitoids. Typical whole rock chemistries for these rocks are shown in figures 7-9. Figure 8 shows the normative anorthite (An)-albite (Ab)-orthoclase (Or) data for these rocks and compares them to similar rocks in the Lowlands. AFM plots are given in figure 8 and calc-alkali index versus silica plots in figure 9; both figures illustrate the strongly calcalkaline nature of the Highland tonalite to granitoid suite. The tonalitic rocks, which will be visited at Stop 1, outcrop in several belts within the southern and eastern Adirondacks. In the field they can be distinguished from, otherwise similar, charnockitic rocks by the white alteration of their weathered surfaces and the bluish grey on fresh surfaces. A distinctive characteristic is the almost ubiquitous presence of discontinuous mafic sheets. These have been interpreted as disrupted mafic dikes coeval with emplacement of the tonalites. Associated with the tonalitic rocks are granodioritic to granitic rocks containing variable concentrations of orthopyroxene.

AMCG Suite. Within the Adirondack Highlands AMCG rocks are widely developed and abundantly represented in the Marcy massif as well as the Oregon and Snowy Mt. Domes. The chemistry of granitoid (mangeritic to charnockitic) facies of these rocks is given in Table 2 and figures 9 and 10, both for the older as well as the younger anorogenic plutonic rocks. As shown in figure 9, the AMCG rocks have calcalkali-silica trends that are distinctly different than those shown by the tonalitic suites. McLelland and Whitney (1991) have shown that the AMCG rocks exhibit anorogenic geochemical characteristics and constitute bimodal magmatic complexes in which anorthositic to gabbroic cores are coeval with, but not related via fractional crystallization to, the mangeritic-charnockitic envelopes of the AMCG massifs (i.e., Marcy massif, figure 2). Bimodality is best demonstrated by the divergent differentiation trends (figure 11) of the granitoid members on the one hand and the anorthositic-gabbroic rocks on the other (Buddington, 1972). Eiler and Valley (pers. comm.) report that $\delta^{18}O$ values for AMCG granitoids are magmatic in origin and demonstrate that these rocks are related by fractional crystallization and were metamorphosed under vapor-absent conditions. Anorthositic members of the AMCG suite have distinctly different $\delta^{18}O$ values. The extreme low-SiO_2, high-iron end members of the anorthosite-gabbro family will be seen at Stop 10 and are believed to represent late liquids developed by plagioclase fractionation under conditions of low oxygen fugacites (i.e., dry, Fenner-type trends). Associated with these are large magnetite-ilmenite deposits which will be visited at Sanford Lake.

Younger Hornblende Granitic Rocks. The distribution of these rocks is shown in figure 5a, and their ages are given in table 1. An example of these rocks will be visited at Stop 14. In the field these rocks consist of medium grained, pink, streaky granitic rocks containing hornblende and minor biotite. They are difficult to distinguish from the granitic facies of the AMCG suite. As pointed out by Chiarenzelli and McLelland (1991), their restriction to the northwestern Highlands is intriguing but not yet understood.

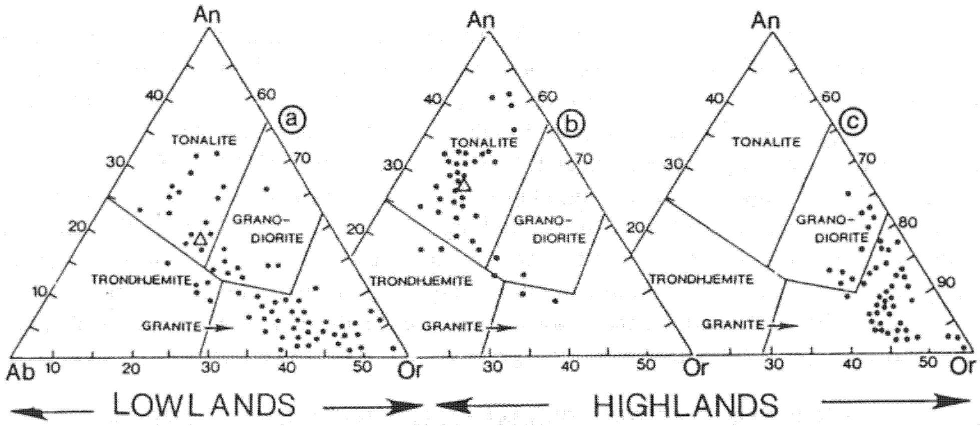

Figure 7. Plots of normative albite (Ab)-anorthite (An)-orthoclase (Or) for (a) Hyde School Gneiss, (b) Highlands tonalites, and (c) Tomantown pluton. Open triangles give average values for tonalitic samples. Definition of fields due to Barker (1979).

A: Anorogenic Complexes
- a) Klokken Complex
- b) Anorogenic granitoids, Labrador
- c) hypersolvus anorogenic granitoids Colorado, Nigeria, Scandinavia
- d) Puklen Complex

B: AMCG Suite
- Filled circles - granitoids
- Filled triangles - jotunites

C: Tonalites
- Filled circles - tonalite
- Stars - gabbros

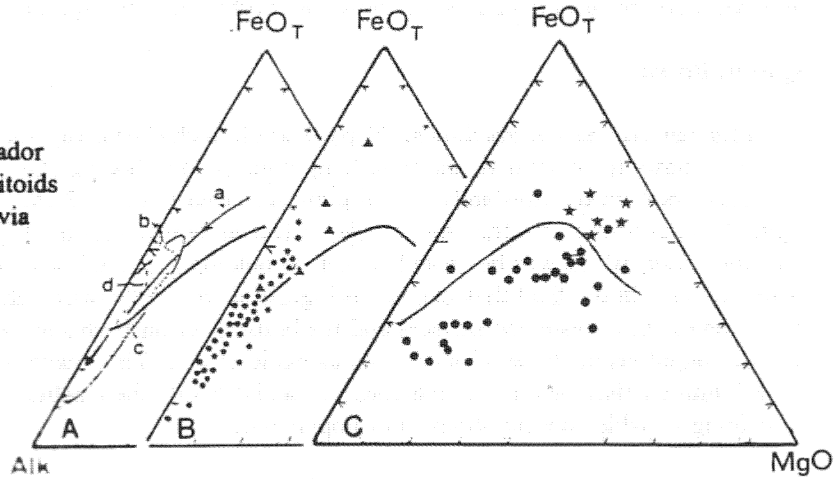

Figure 8. AFM variation diagrams for A) anorogenic complexes, B) AMCG suite, C) Highlands tonalites and associated gabbro (see McLelland 1991 for sources).

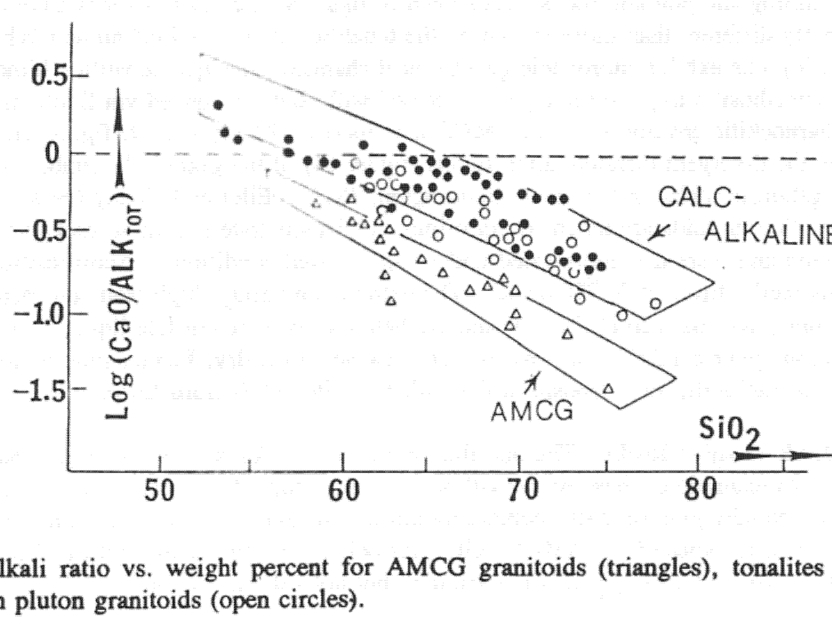

Figure 9. Calcalkali ratio vs. weight percent for AMCG granitoids (triangles), tonalites (closed circles) and Tomantown pluton granitoids (open circles).

TABLE 2

	EARLY CALCALKALINE ROCKS			OLDER ANOROGENIC PLUTONIC ROCKS					
	AM-87-13	TOE	CL-6	AM-86-17	AM-86-1	AC-85-1	AM-86-9	AM-86-15	AC-85-2
SiO_2	65.00	65.68	74.63	68.90	71.88	73.72	69.14	67.47	75.17
TiO_2	.75	1.16	.37	.59	.36	.04	.89	.72	.20
Al_2O_3	15.10	14.97	14.22	14.50	14.82	13.54	13.78	15.12	12.63
FeO	nd	nd	nd	2.16	1.27	.87	2.83	3.34	1.11
Fe_2O_3	6.03	6.79	1.53	1.1	.96	.11	1.82	1.59	1.07
MnO	.10	.08	.04	.02	.03	.01	.04	.10	.02
MgO	.46	1.15	.55	.84	.43	.20	.45	.51	.19
CaO	2.71	2.56	1.66	2.3	1.87	.85	2.26	2.57	.88
Na_2O	4.10	2.80	3.56	3.06	3.93	5.71	3.07	3.41	2.99
K_2O	5.13	4.52	4.26	4.18	3.99	4.42	4.91	5.18	5.49
P_2O_5	.10	.51	.10	.24	.09	.01	.23	.19	.04
LOI	.39	.74	...	.40	.27	.17	.19	.39	.17
Σ	99.87	100.96	99.42	99.63	99.65	99.61	100.59	99.96	99.96
Ba(ppm)	1230	710	510	680	660	1100	736	810	442
Rb (ppm)	100	97	170	160	200	406	81	128	230
Sr (ppm)	260	230	260	200	130	26	211	215	99
Y (ppm)	70	70	37	40	110	321	60	71	77
Nb (ppm)	20	19	17	30	30	15	19	21	15
Zr (ppm)	790	345	160	270	670	118	538	546	284

	YOUNGER ANOROGENIC PLUTONIC ROCKS							
	AC-85-6	AC-85-10	AM-86-7	WPG	SLC	AM-87-9	AM-86-8	AM-87-10
SiO_2	62.15	54.94	58.50	69.20	6.64	61.05	6.94	62.14
TiO_2	.88	1.55	.68	.51	1.14	.78	1.39	.36
Al_2O_3	16.40	14.87	2.32	13.90	15.27	15.98	15.91	12.35
FeO	3.96	1.25	2.81	3.1	9.28	4.60	6.51	1.32
Fe_2O_3	1.49	2.80	.43	1.34	1.77	2.10	1.02	1.7
MnO	.09	.24	.01	.05	.19	.05	.14	.01
MgO	1.06	.96	1.47	.52	.74	1.64	1.70	.83
CaO	3.27	5.52	6.16	2.03	3.97	3.63	4.53	3.65
Na_2O	4.81	3.45	5.02	3.02	3.34	3.41	3.55	6.05
K_2O	5.13	3.83	3.35	5.48	3.70	4.76	3.86	1.26
P_2O_5	.30	.65	.32	.11	.31	.42	.46	.09
LOI	.41	.37	.43	.39	.01	.91	.37	.67
Σ	99.95	99.43	99.50	99.65	100.46	99.63	100.38	99.24
Ba(ppm)	850	625	nd	1279	823	nd	1100	nd
Rb (ppm)	106	47	nd	124	87	nd	83	29
Sr (ppm)	335	367	nd	184	215	nd	410	180
Y (ppm)	60	55	nd	48	121	nd	110	63
Nb (ppm)	21	14	nd	14	38	nd	25	79
Zr (ppm)	464	431	nd	382	647	nd	1200	309

	YOUNGER GRANITIC ROCKS						LATE LEUCOGRANITIC ROCKS				
	AM-86-3	AM-86-6	NO-Fo1	AM-86-13	AM-87-6	GHA	AM-86-11	AM-87-7	AM-86-4	AM-86-10	AM-86-14
SiO_2	68.62	68.05	71.75	76.30	69.00	73.2	69.98	67.80	7.01	69.05	72.39
TiO_2	.48	.55	.41	.18	1.43	.35	.46	.42	.69	.57	.38
Al_2O_3	14.37	14.67	13.49	11.64	12.12	13.1	12.37	15.76	12.43	13.06	12.63
FeO	3.01	3.51	2.39	1.13	4.94	.84	5.13	1.2	4.23	3.50	1.6
Fe_2O_3	.93	1.18	1.12	.61	1.16	2.1	1.11	2.8	1.5	1.42	4.13
MnO	.06	.07	.05	.01	.02	.02	.14	.03	.03	.01	.03
MgO	.49	.45	.11	.01	.67	.33	.08	.56	.01	.17	.29
CaO	1.99	1.81	1.43	.45	.75	1.55	1.25	2.35	.94	.35	1.07
Na_2O	3.71	3.81	2.99	3.32	2.79	4.16	3.99	3.71	1.92	1.08	6.63
K_2O	5.67	5.61	5.79	5.22	6.50	4.01	4.91	4.91	8.34	9.64	.52
P_2O_5	.13	.12	.07	.03	.17	.07	.04	.13	.17	.12	.70
LOI	.40	.58	.5	.30	.25	.39	.21	.31	.23	.41	.10
Σ	99.86	100.41	100.00	99.20	99.80	100.3	99.67	99.98	100.50	99.38	100.47
Ba(ppm)	861	715	692	nd	1014	1249	160	nd	840	290	98
Rb (ppm)	161	148	182	188	214	178	190	nd	315	330	16
Sr (ppm)	209	240	115	132	303	211	20	nd	73	60	42
Y (ppm)	65	62	72	157	35	86	120	nd	66	75	117
Nb (ppm)	20	20	25	29	17	19	30	nd	18	23	27
Zr (ppm)	394	542	595	392	507	338	1230	nd	414	600	786

Alaskitic and Leucogranitic Rocks. The distribution of these distinctive rocks is shown in figure 5a, and their geochronology is summarized in table 1. An example of these rocks will be visited at Stop 12. They consist principally of pink quartz-mesoperthite gneiss commonly with magnetite as the only dark phase. A less voluminous, but important, trondhjemitic facies is also common and is commonly associated with low-Ti magnetite deposits in the unit. Granitic facies also occur within this group which, together, constitutes the Lyon Mt. Gneiss (Whitney and Olmsted, 1989). U-Pb zircon ages of 1080-1050 Ma for the alaskites are interpreted as dating emplacement, and, since this time interval corresponds to granulite facies metamorphism at ~25 km, the Lyon Mt. Gneiss is interpreted as intrusive (Chiarenzelli and McLelland, 1991). This is in contrast to Whitney and Olmsted (1989) who have interpreted the Lyon Mt. Gneiss as a metamorphosed series of altered acidic volcanics. This issue is discussed in detail in the text for Stop 12.

Olivine Metagabbro. Numerous bodies of tholeiitic metagabbros are scattered throughout the Adirondacks and are especially abundant near the anorthosite massifs with which they are coeval. Most of these bodies are coronites whose evolution has been discussed by McLelland and Whitney (1980) and which will be visited at Stops 2 and 17.

Metasedimentary Rocks. Within the southern and eastern Adirondacks the metasedimentary sequence is dominated by quartzites and metapelites with marbles being virtually absent. The major quartzite of the southern region is exceptionally pure and comprises an ~1000 m-thick unit. Of even greater extent, as well as thickness, are garnet-biotite-quartz-oligoclase ± sillimanite gneisses (referred to as kinzigite) together with sheets, pods, and stringers of white, garnetiferous anatectite (Stop 2). The common presence of hercynitic spinel supports an anatectic origin for the leucosomes. Kinzigitic rocks grade into khondalites as sillimanite and garnet increase at the expense of biotite.

In contrast to the southern and eastern Adirondacks, the central and northern Adirondacks contain only sparse kinzigite, and metasediments are principally represented by synclinal keels of marble and calcsilicate with minor quartzite (Stop 15). It is possible that the change from carbonate to pelitic metasediments corresponds to an original shelf to deep water transition, now largely removed by intrusion, doming, and erosion.

STRUCTURAL GEOLOGY

The Adirondacks region is one of intense ductile strain, essentially all of which must postdate the ca. 1150 Ma AMCG suite which is involved in each of the major phases of deformation, i.e., the regional strain is associated with the Ottawan Orogeny. Earlier Elzevirian fabric may be present and rotated into parallelism. Here we describe the structure of the southern Adirondacks which is best known and representative of the entire region. A complete set of references is given in McLelland and Isachsen (1986).

As shown in figures 2 and 11, the southern Adirondacks are underlain by very large folds. Four major phases of folding can be identified and their intersections produce the characteristic fold interference outcrop patterns of the region (figure 11). The earliest recognizable map-scale folds (F_1) are exceptionally large isoclinal recumbent structures characterized by the Canada Lake, Little Moose Mt., and Wakely Mt. isoclines, whose axes trend E-W and plunge 10-15° about the horizontal. The Little Moose Mt. isocline is synformal and the other two are antiformal, and suspected to be anticlinal, but the lack of any facing criteria precludes any age assignments. All of these structures fold an earlier tectonic foliation consisting of flattened mineral grains of unknown age and origin. An axial planar cleavage is well developed in the Canada Lake isocline, particularly in the metapelitic rocks.

F_2-folds of exceptionally large dimensions trend E-W across the region and have upright axial planes (figure 11). They are coaxial with the F_1 folds suggesting that the earlier fold axes have been rotated into parallelism with F_2 and that the current configurations of both fold sets may be the result of a common set of forces. An intense ribbon lineation defined by quartz and feldspar rods parallels the F_2-axes along the Piseco anticline, Gloversville syncline, and Glens Falls syncline and documents the high temperatures, ductile deformation and mylonitization that accompanied the formation of these folds. Large NNE trending upright folds (F_3) define the

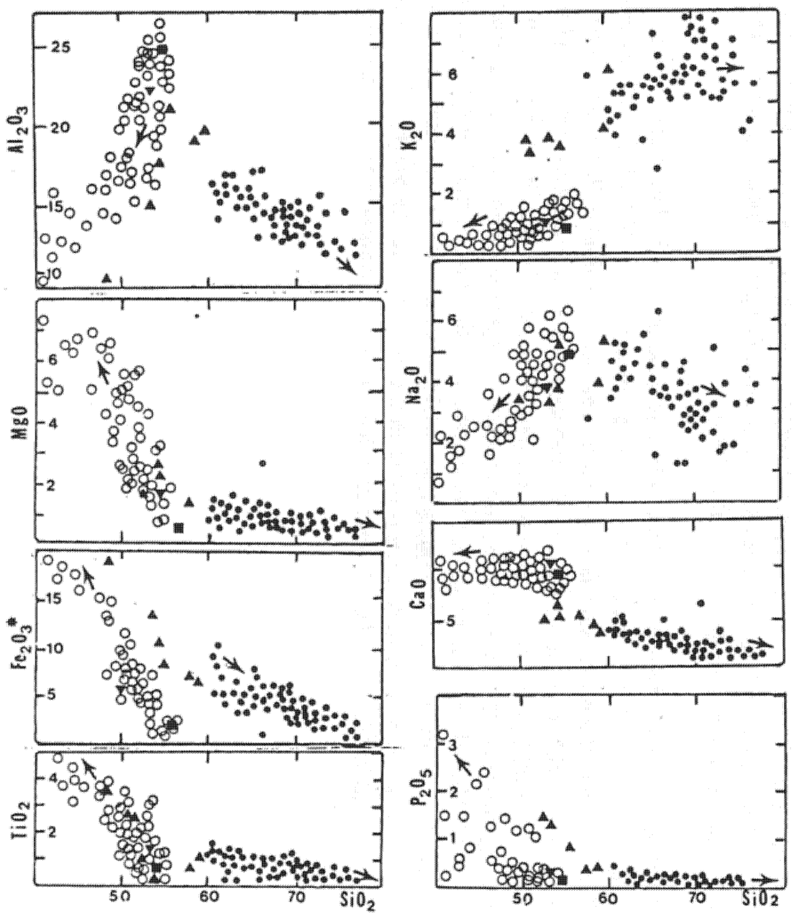

Figure 10. Harker variation diagrams or AMCG-rocks of the Marcy massif. Open circles=anorthositic suite, filled circles=granitoid suite, upright triangles=mixed rocks, inverted triangles=Whiteface facies, square=Marcy facies. Arrows indicate differentiation trends.

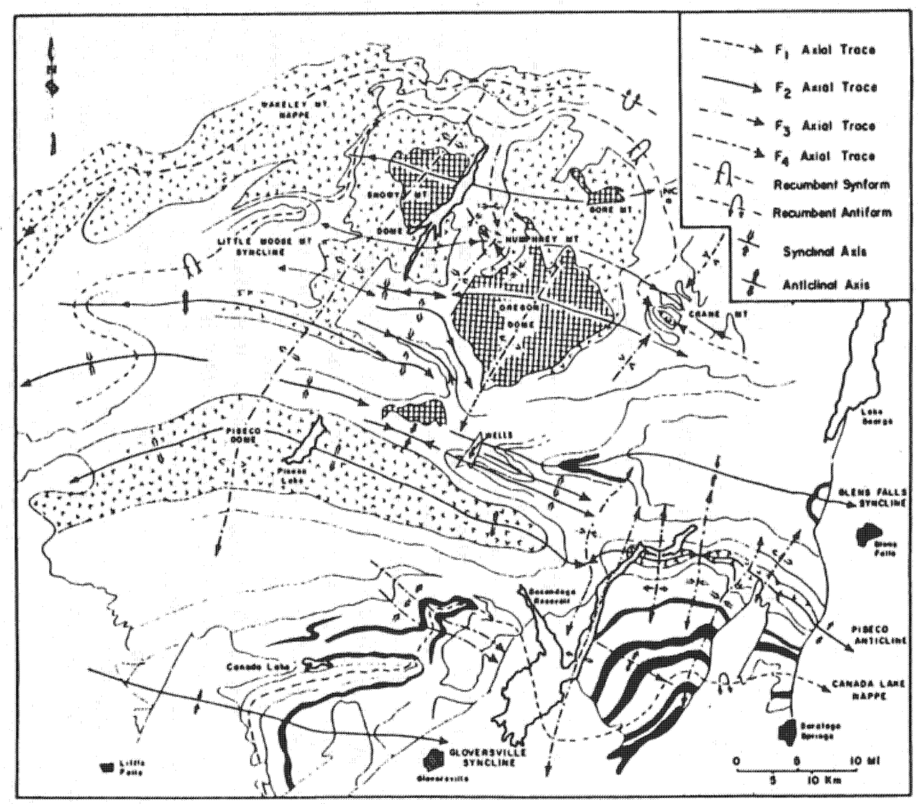

Figure 11. Fold axes within the southern and central Adirondacks. Designation of folds as synclines and anticlines is provisional, since facing directions are not yet known.

Snowy Mt. and Oregon domes (figure 11). Where the F_3 folds intersect F_2 axes structural domes (i.e., Piseco dome) and intervening saddles result. A late NW-trending fold set results in a few F_4 folds between Canada Lake and Sacandaga Reservoir (figure 11).

Kinematic indicators (mostly feldspar tails) in the area suggest that the dominant displacement in the region involved motion in which the east side moved up and to the west (McLelland, 1984). In most instances this implies thrusting motion, however, displacement in the opposite sense has also been documented. This suggests that relative displacement may have taken place in both senses during formation of the indicators.

METAMORPHISM

Figure 12 shows the well known pattern of paleoisotherms reported by Bohlen et al. (1985). paleotemperatures have been established largely on the basis of two-feldspar geothermometry but (Fe, Ti)-oxide methods have also been used and, locally, temperature-restrictive mineral assemblages have been employed (Valley et al., 1990). The bull's eye pattern of paleoisotherms, centered on the Marcy massif, is believed to result from late doming of the massif. Paleopressures show a similar bull's eye configuration with pressures of 7-8 kbar decreasing outward to 6-7 kbar away from the massif and reaching 5-6 kbar in the Lowlands (Bohlen et al., 1985). The P,T pattern of figure 12 is interpreted as reflecting peak metamorphic conditions, although microtextures suggest that some retrogression exists. Generally, the P,T conditions of the Adirondack are those of granulite facies metamorphism, and most commonly correspond to the hornblende-clinopyroxene-almandine subfacies of the high-pressure range of the granulite facies. These conditions must have been imposed during the Ottawan Orogeny since they affect rocks as young as 1050 Ma. The identification of ca. 1050-1060 Ma metamorphic zircons by McLelland and Chiarenzelli (1990) fixes the time of peak metamorphic conditions and corresponds well with garnet and titanite U-Pb ages of ca. 1050-1000 Ma in the Highlands (Mezger et al., 1992). Rb-Sr whole rock isochron ages of ca. 1100-1000 Ma also reflect Ottawan temperatures and fluids. Despite the high-grade, regional character of the Ottawan Orogeny, the preservation of foliated garnet-sillimanite xenoliths in an 1147 ± 4 Ma metagabbro (McLelland et al., 1987), and the report of several 1150 Ma U-Pb garnet ages (Mezger et al., 1992), reveals that earlier assemblages from the Elzevirian and AMCG metamorphic pulses managed to survive locally. The dehydrating effects of these high temperature events, as well as the anhydrous nature of the AMCG rocks themselves, are thought to be responsible for creating a water-poor terrane throughout the Adirondack Highlands prior to the Ottawan Orogeny which appears to have proceeded under generally vapor-absent conditions (Valley et al., 1990).

The present depth to the Moho beneath the Adirondack Highlands is $\sim$35 km (Katz, 1955; Hughes and Luetgert, 1992). Since metamorphic pressures of 7-8 kbar correspond to $\sim$20-25 km depth of burial, it follows that during Ottawan metamorphism the Adirondack region consisted of a double thickness of continental crust, and this portion of the Grenville orogen may have corresponded to a Himalayan-type collisional margin at 1050-1080 Ma. Bohlen et al. (1985) proposed a counterclockwise P-T-t path for the Ottawan Orogeny, including an almost isobaric cooling history. If this is correct, the necessary magmatic component of heat may have been derived from 1130-1150 Ma gabbroic magmas that were ponded at the base of the lithosphere during the AMCG magmatism. Upward transfer of this heat by <u>conduction</u> would require $\sim$80 Ma to reach the surface (Emslie and Hunt, 1990) and would, therefore, have been present in the crust during the height of the Ottawan Orogeny. Granitic rocks of the ca. 1050 Ma Lyon Mt. Gneiss may have helped to transport this thermal energy.

□ ILMENITE-SILLIMANITE-QUARTZ-GARNET-RUTILE
■ FERROSILITE-FAYALITE-QUARTZ
◪ SPHALERITE-PYRRHOTITE-PYRITE
○ FAYALITE-ANORTHITE-GARNET
● FERROSILITE-ANORTHITE-GARNET-QUARTZ
◑ ANORTHITE-GARNET-SILLIMANITE-QUARTZ
▲ KYANITE-SILLIMANITE
△ AKERMANITE

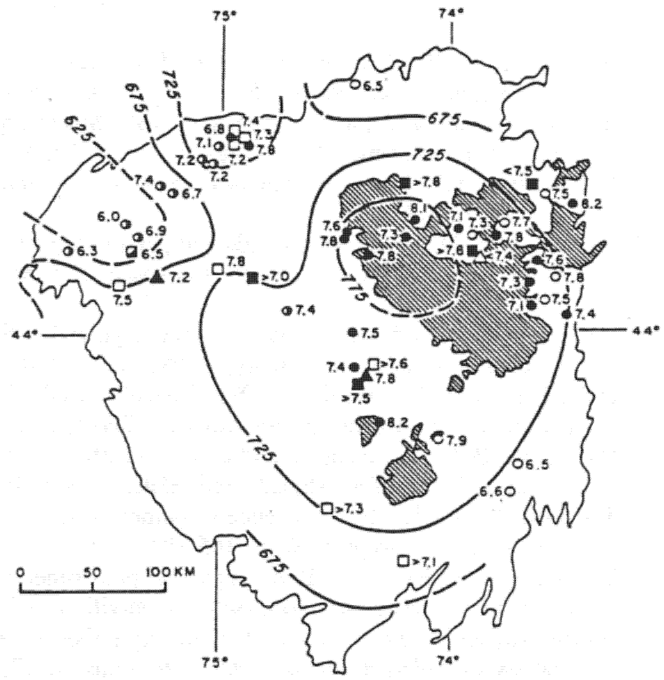

Figure 12. Peak metamorphic P,T conditions for the Adirondack MB. Isotherms are continued from feldspar and Fe-Ti thermometry and Pressures are estimated from the geobarometers given above and indicated on the map. (From Bohlen et al., 1985)

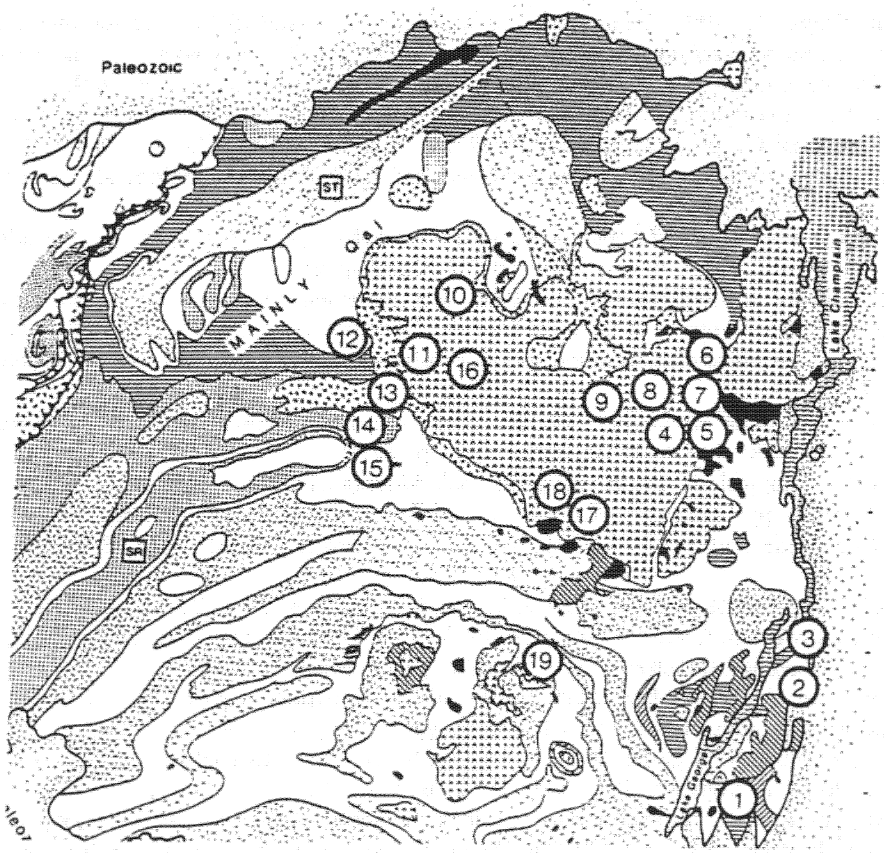

Figure 13. Locations of stops in the Adirondack Highlands. Symbols as in figure 2.

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STOP A-1: THE GLEN

Description modified from Isachsen (1965)

The calcite marble and calc-silicate beds here are highly-deformed, having been isoclinally folded and refolded. The relatively competent amphibolite layers have been stretched, attenuated, and dismembered well beyond the boudinage stage so that they occur as disconnected lenses, hooks, and clots. Such floating fragments have sometimes been picturesquely tagged "tectonic "fish." Minerals that can be readily found and identified in the marble include calcite, graphite, diopside, forsterite, and titanite. Calc-silicate reaction zones can be observed between the marble and many of the mafic units. Local pockets of very course calcite occur with obvious twin lamellae.

STOP A-2: GORE MOUNTAIN

Description modified from Kelly and Peterson (1993).

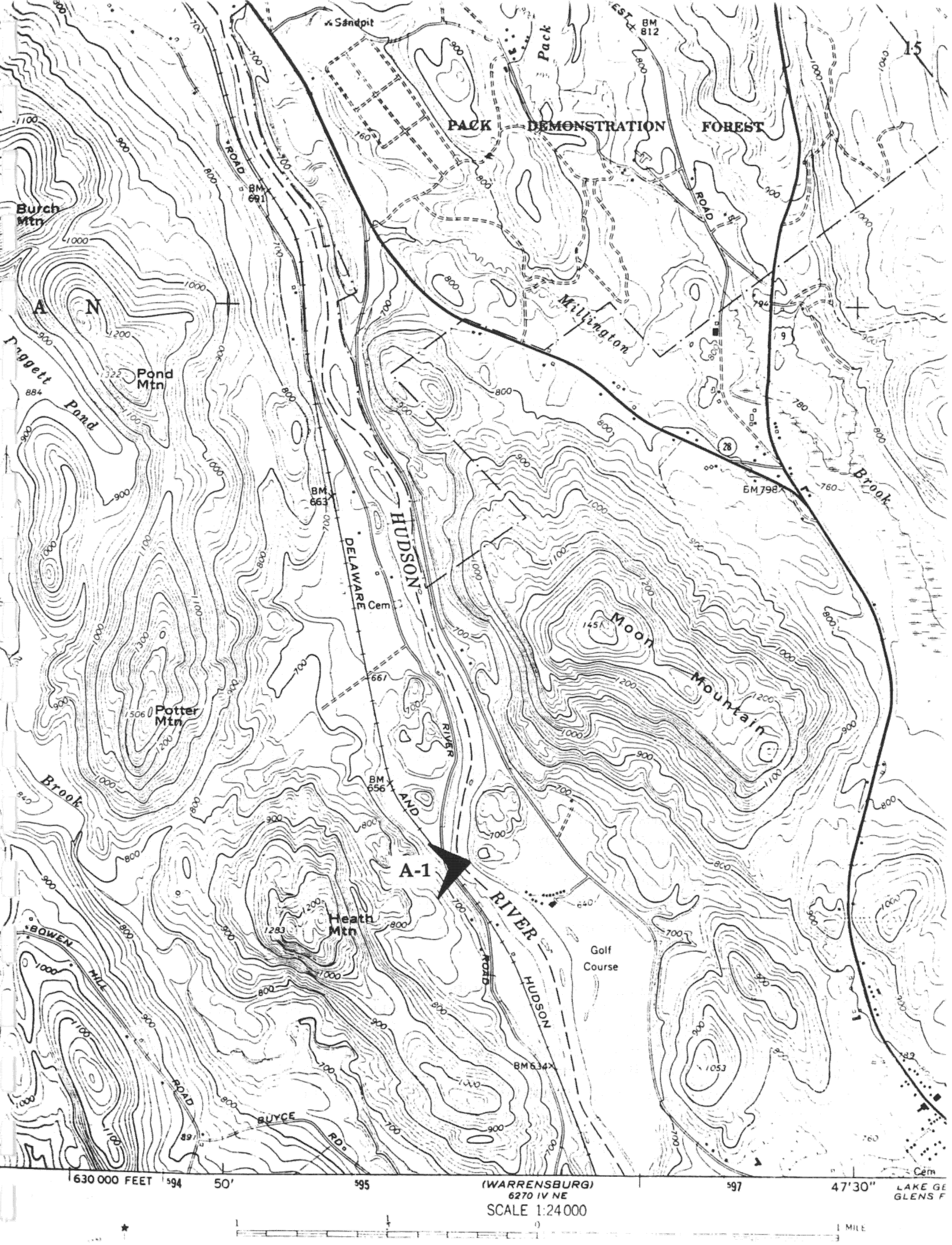
The Barton Mines Corporation open pit mine is located at an elevation of about 2600 feet on the north side of Gore Mountain. For 105 years, this was the site of the world's oldest continuously operating garnet mine and the country's second oldest continuous operating mine under one management. The community at the mine site is the highest self-sufficient community in New York State. It is 10 miles from North Creek and 5 miles from State Route 28 over a Company-built road that rises 300 feet per mile. This road, like others in the vicinity, is surfaced with coarse mine tailings. About eleven families can live on the property. The community has its own water, power, and fire protection systems. On the property are the original mine buildings and the Highwinds Inn, built by Mr. C.R. Barton in 1933 as a family residence. The Inn is now privately leased from the corporation and operates 10 months per year. It offers a four-bedroom lodge, a four star dining room, cross-country skiing and fantastic views of the Siamese Wilderness Area.

The garnet is used in coated abrasives, glass grinding, metal and glass polishing, and even to remove the red hulls from peanuts. Paint manufacturers add garnet to create non-skid surfaces and television makers use it to prepare the glass on color picture tubes. Barton sells between 10,000 and 12,000 tons of garnet abrasive annually. About 40% of the company's shipments are to foreign countries. All current U.S. production of technical grade garnet is limited to the Barton mines from where it is shipped world wide for use in coated abrasives and powder applications (Austin, 1993a, b).

Garnet has been designated as the official New York State gemstone. Barton produces no gem material but collectors are still able to find gem rough. Stones cut from Gore Mountain rough material generally fall into a one to five carat range. Garnets from this locality are a dark red color. Special cutting schemes have been devised for this material in order to allow sufficient light into the stone.

History

The first 80 years of the history of the Barton garnet mines has been compiled by Moran (1956) and is paraphrased below. Mr. Henry Hudson Barton came to Boston from England in 1846 and worked as an apprentice to a Boston jeweller. While working there in the 1850's, Barton learned of a large supply of garnet located in the Adirondack Mountains. Subsequently, he moved to Philadelphia and married the daughter of a sandpaper manufacturer. Combining his knowledge


Sandpit
BM 812
Pack
PACK DEMONSTRATION FOREST
ROAD
BM 691
Burch Mtn
Millington
Pond Mtn
1322
884
Daggett Pond
28
9
BM 798
Brook
BM 663
HUDSON
DELAWARE
Cem
Moon Mountain
1451
661
RIVER
1506
Potter Mtn
Brook
BM 656
AND
A-1
RIVER
Heath Mtn
1283
BOWEN HILL ROAD
Golf Course
HUDSON
BM 634
1053
BUYCE RD
891
Cem
630 000 FEET
594
50′
595
(WARRENSBURG)
6270 IV NE
597
47′30″
SCALE 1:24 000
1 MILE

of gem minerals and abrasives, he concluded that garnet would produce better quality sandpaper than that currently available. He was able to locate the source of the Adirondack garnet stones displayed at the Boston jewelry store years before. Barton procured samples of this garnet which he pulverized and graded. He then produced his first garnet- coated abrasive by hand. The sandpaper was tested in several woodworking shops near Philadelphia. It proved to be a superior product and Barton soon sold all he could produce.

H.H. Barton began mining at Gore Mountain in 1878 and in 1887, bought the entire mountain from New York State. Early mining operations were entirely manual. The garnet was hand cobbed i.e. separated from the waste rock by small picking hammers and chisels. Due to the obstacles in moving the ore, the garnet was mined during the summer and stored on the mountain until winter. It was then taken by sleds down to the railroad siding at North Creek whence it was shipped to the Barton Sandpaper plant in Philadelphia for processing. The "modern" plant at Gore Mountain was constructed in 1924. Crushing, milling, and coarse grading was done at the mine site. In 1983, the Gore Mountain operation was closed down and mining was relocated to the Ruby Mountain site, approximately four miles northeast, where it continues at present.

Mining and Milling

The mine at Gore Mountain is approximately one mile in length in an ENE-WSW direction. The ore body varies from 50 to 400 feet and is roughly vertical. Mining was conducted in benches of 30 feet using standard drilling and blasting techniques. Oversized material was reduced with a two and one-half ton drop ball. The ore was processed through jaw and gyratory crushers to liberate the garnet and then concentrated in the mill on Gore Mountain. Garnet concentrate was further processed in a separate mill in North River at the base of the mountain. Separation of garnet is accomplished by a combination of concentrating methods including heavy media, magnetic, flotation, screening, tabling, and air and water separation. Processes are interconnected and continuous or semi-continuous until a concentrate of 98% minimum garnet for all grades is achieved (Hight, 1983). Finished product ranged from 1/4 inch to 1/4 micron in size.

Characteristics of Gore Mountain Garnet

The garnet mined at Gore Mountain is a very high quality abrasive. The garnets display a well developed tectonic parting that, in hand specimen, looks like a very good cleavage. This parting is present at the micron scale. Consequently, the garnets fracture with chisel-like edges yielding superior cutting qualities. The garnet crystals are commonly 12 inches in diameter and rarely up to thirty-six inches with an average diameter of 4 inches (Hight, 1983). The composition of the garnet is approximately 37-43% pyrope, 40-49% almandine, 13-16% grossular and 1% spessartine (Levin, 1950; Harben and Bates, 1990). Chemical zoning in garnet, where present, is very weak and variable (Luther, 1976). Typical chemical analyses of the garnet and associated tschermakitic hornblende (Leake, 1978), andesine and hypersthene are given in Table A-1. The garnet has a hardness between eight and nine and an average density of 3.95 gm/cm^{-3}.

Geology

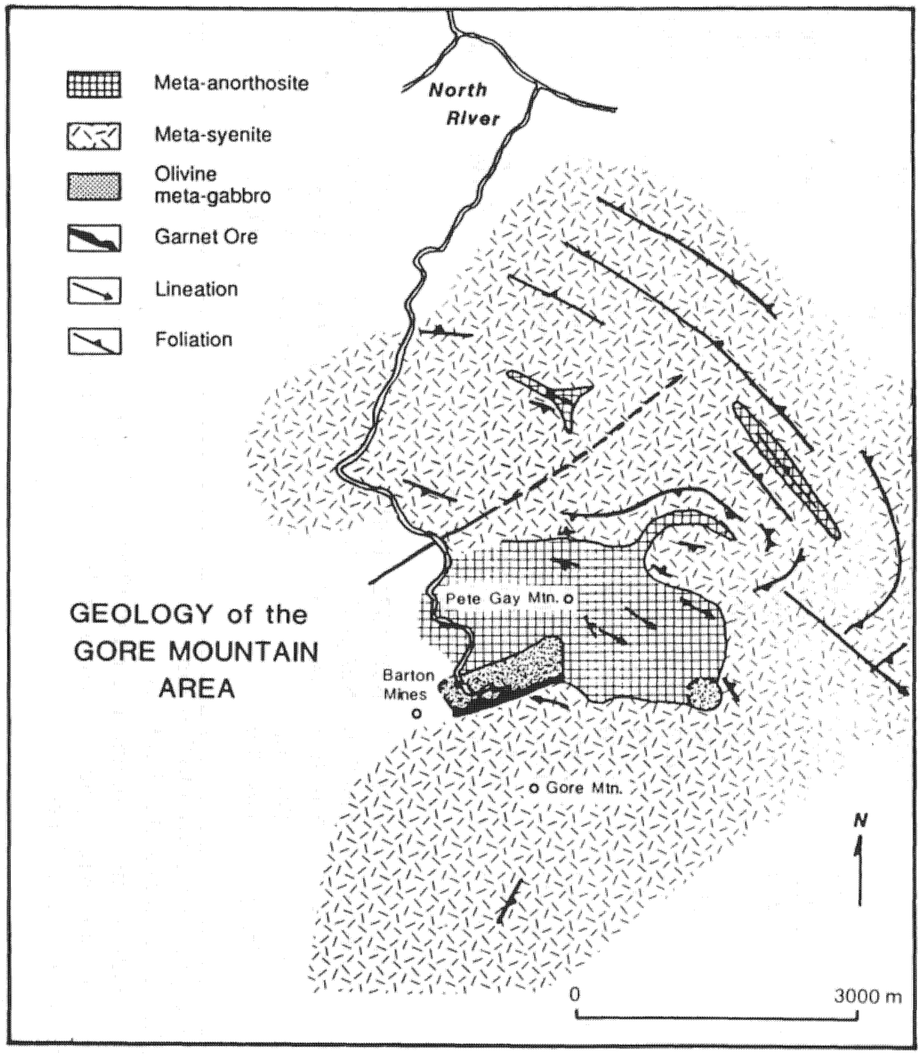
The garnet mine is entirely hosted by a hornblende-rich garnet amphibolite unit sandwiched between a small olivine meta-gabbro body which contacts meta-anorthosite to the north and is in fault contact with meta-syenite to the south (Fig. A-1,A-2). Preserved in the olivine meta-gabbro are igneous textures and faint igneous layering, and a xenolith of anorthosite has been reported in the gabbro (Luther, 1976). Prior to metamorphism, the rock was composed of plagioclase, olivine, clinopyroxene, and ilmenite. During metamorphism, coronas of orthopyroxene, clinopyroxene, and garnet formed between the olivine and the plagioclase and coronas of biotite,

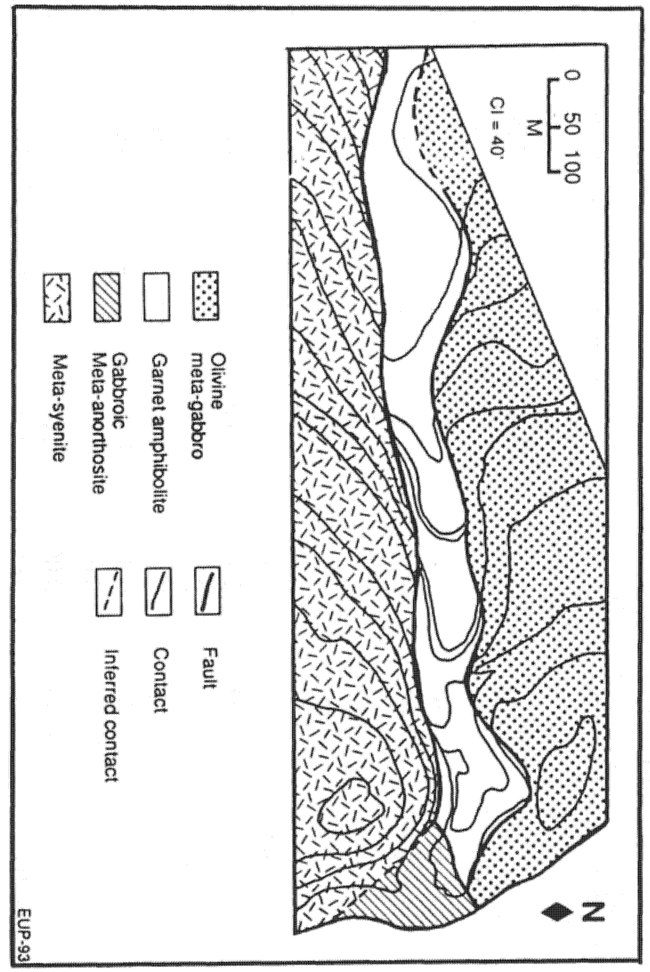
Table A-1. Microprobe Analyses of Gore Mt. Garnet, Hornblende, Orthopyroxene and Feldspar

	GARNET #29	GARNET #41	HBL #30	HBL #31	OPX #37	PLAG #33	PLAG #34
Oxide Weight Percent							
SiO_2	39.43	39.58	44.13	44.26	25.41	27.16	27.29
Al_2O_3	21.40	21.20	12.36	12.53	1.22	13.56	13.67
TiO_2	0.05	0.10	1.28	1.45	0.04	0.00	0.00
FeO^*	22.80	24.45	3.49	3.03	14.82	0.05	0.05
$Fe_2O_3{}^*$	1.44	0.72	8.12	8.93		0.00	0.00
MgO	10.65	9.60	14.51	14.50	14.87	0.00	0.00
MnO	0.48	0.74	0.06	0.08	0.11	0.00	0.00
CaO	3.85	3.97	10.56	18.48	0.32	6.38	6.47
Na_2O	0.00	0.00	2.48	2.43	0.02	4.65	4.70
K_2O	0.00	0.00	0.57	0.55	0.00	0.17	0.18
Cl	0.00	0.00	0.02	0.00	0.00	0.00	0.00
F	0.00	0.00	0.03	0.00	0.00	0.00	0.00
H_2O $(OH)^*$			2.06	2.10			
O = Cl	0.00	0.00	0.00	0.00	0.00	0.00	0.00
O = F	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	100.09	100.36	99.66	100.33	101.08	99.20	99.87
Atoms							
Si	2.997	3.020	6.348	6.320	0.984	2.633	2.627
Al^{iv}			1.652	1.680	0.016	1.368	1.370
Al^{vi}	1.918	1.907	0.444	0.429	0.013		
Ti	0.003	0.006	0.139	0.156	0.002	0.000	0.000
Fe^{2+}	1.449	1.561	0.420	0.362	0.289	0.002	0.003
Fe^{3+}	0.082	0.041	0.878	0.959	0.000	0.000	0.000
Mg	1.206	1.092	3.111	3.085	0.666	0.000	0.000
Mn	0.031	0.048	0.008	0.009	0.002	0.000	0.000
Ca	0.313	0.325	1.628	1.604	0.009	0.433	0.436
Na	0.000	0.000	0.692	0.672	0.001	0.551	0.553
K	0.000	0.000	0.104	0.101	0.000	0.012	0.012
Cl	0.000	0.000	0.005	0.000	0.000	0.000	0.000
F	0.000	0.000	0.014	0.000	0.000	0.000	0.000
O	12.000	12.000	22.000	22.000	3.009	8.036	8.029
OH			1.981	2.000			
	And 4.1	And 2.1	NaB 0.372	NaB 0.396	En 69.1	A 0.996	A 1.001
	Alm 48.3	Alm 52.0	NaA 0.320	NaA 0.275	Fs 30.0	T 4.002	T 3.997
	Pyp 40.2	Pyp 36.4	KA 0.104	KA 0.101	Di 0.6	Ab 55.3	Ab 55.2
	Sps 1.0	Sps 1.6			Hd 0.3	An 43.5	An 43.6
	Grs 6.3	Grs 7.9				Or 1.2	Or 1.2
	10.4	10.0					

EUP-93

OPX (hypersthene) normalized to 2 cations
PLAG (andesine) normalized to 5 cations
HBL (tschermakitic hornblende) normalized to 13 M1+M2+M3+T cations
GARNET (almandine-pyrope) normalized to 8 cations and 12 anions
* Calculated by charge balance





Figures A-1, A-2. Geologic maps of the Barton Garnet Mine. A-1 (left) is modified from Bartholomé (1956). A-2 (right) is modified from Goldblum and Hill (1992).

Table A-2. Whole-rock Chemical Analyses

Oxide	Olivine Meta-gabbro	Garnet Ore
SiO_2	47.14	45.68
TiO_2	0.81	0.78
Al_2O_3	16.98	17.32
Fe_2O_3	0.69	1.30
FeO	11.13	9.67
MnO	0.16	0.15
MgO	11.04	10.97
CaO	8.05	8.58
Na_2O	2.54	2.85
K_2O	0.56	0.59
P_2O_5	0.10	0.10
H_2O	0.44	1.16
Total	99.64	99.15

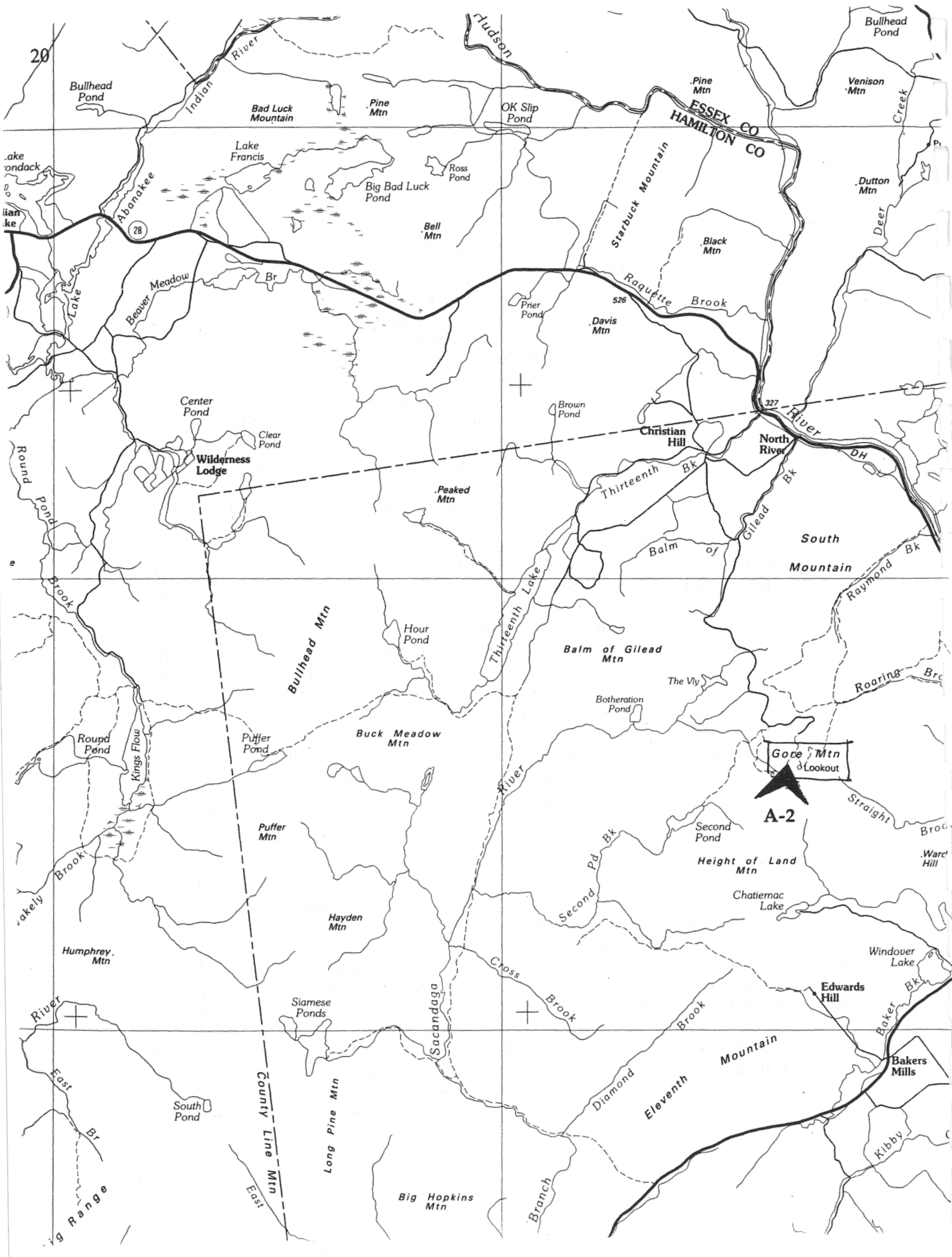
hornblende and ilmenite formed between plagioclase and ilmenite (Whitney and McLelland, 1973, 1983). The contact between the olivine meta-gabbro and the garnet amphibolite is gradational through a transition zone 2-3 meters wide. Garnet size increases dramatically across the transition zone from <1 mm in the olivine meta-gabbro, to 3 mm in the transition zone to 50 to 350 mm in the amphibolite (Goldblum and Hill, 1992). Crossing the transition zone the increase in garnet size is paralleled by a ten-fold increase in the size of hornblende and biotite, olivine disappears, modal clinopyroxene decreases as it is replaced by hornblende and spinel-included plagioclase grades to inclusion-free plagioclase. The modal percent garnet varies from 5 to 20 percent but averages 13 percent in both the olivine meta-gabbro and the amphibolite (Luther, 1976; Hight, 1983; Goldblum, 1988). At the west end of the mine, a garnet hornblendite with little or no feldspar is locally present. This rock probably represents original ultramafic layers in the gabbro (Whitney *et al.*, 1989). In the more mafic portions of the ore body, the garnets are rimmed by hornblende up to several inches thick. Elsewhere, the ore is less mafic, and the rims contain plagioclase and orthopyroxene. The amphibolite is thought to represent a retrograded zone of granulite facies rocks (approximately 750 C and 7.5 kilobars).

A strong, consistent lineation and a weak planar fabric coincides with the zone of large garnets and is an important characteristic of the ore zone (Goldblum and Hill, 1992). The lineation is defined by parallel alignment of prismatic hornblende grains, elongate segregations of felsic and mafic minerals, plagioclase pressure shadows and occasional elongate garnet. The foliation is defined by a slight flattening of the felsic and mafic aggregates.

Chemical analyses of the olivine meta-gabbro and garnet amphibolite show that the garnet ore was derived by retrograde isochemical metamorphism, except for an increase in H_2O and fO_2, of the olivine meta-gabbro (Table A-2; Luther, 1976).

Origin of Garnet

The formation of the garnets is not completely understood. Whereas the Gore Mountain deposit is the largest known, it is not unique in the Adirondacks. Elsewhere, there are occurrences of garnet amphibolite that are texturally and mineralogically similar. These are usually located on the margins of gabbroic rock bodies. Although the garnets at Gore Mountain are atypical in size, the modal amount of garnet is not unusually high for Adirondack garnet amphibolites. The ore at the currently operating Barton Corporation mine at Ruby Mountain, for example, is of the same tenor but the garnets rarely are larger than one or two inches. Petrologic studies (Buddington, 1939, 1952; Bartholomé, 1956, 1960; Luther, 1976; Sharga, 1986; Goldblum, 1988; Goldblum and Hill, 1992) have agreed that the growth of the large garnets is related to a localized influx of water at the margin of a competent meta-gabbro body during amphibolite facies metamorphism. Southward across the transition zone increased ductile deformation resulted in grain size reduction of plagioclase and clinopyroxene. The presence of deformation lamellae, undulose extinction,


20
Bullhead Pond
Indian River
Hudson
Bad Luck Mountain
Pine Mtn
OK Slip Pond
Pine Mtn
ESSEX CO
HAMILTON CO
Venison Mtn
Creek
Lake Francis
Ross Pond
Big Bad Luck Pond
Starbuck Mountain
Dutton Mtn
Abanakee
28
Bell Mtn
Black Mtn
Deer
Lake
Beaver Meadow
Br
526
Raquette Brook
Prier Pond
Davis Mtn
327
River
Center Pond
Brown Pond
Clear Pond
Wilderness Lodge
Christian Hill
North River
DH
Round Pond Brook
Peaked Mtn
Thirteenth Bk
Bk
Gilead
South Mountain
Balm of
Raymond Bk
Thirteenth Lake
Bullhead Mtn
Hour Pond
Balm of Gilead Mtn
The Vly
Roaring Brook
Botheration Pond
Round Pond
Kings Flow
Puffer Pond
Buck Meadow Mtn
Gore Mtn
Lookout
River
Straight Brook
A-2
Puffer Mtn
Brook
Second Pond
Second Pd Bk
Height of Land Mtn
Ward Hill
Chatiemac Lake
Hayden Mtn
Humphrey Mtn
Windover Lake
Cross Brook
Edwards Hill
Baker Bk
River
Siamese Ponds
Brook
Sacandaga
Mountain
Bakers Mills
East Br
County Line Mtn
South Pond
Diamond
Eleventh
Long Pine Mtn
Kibby
Range
East
Branch
Big Hopkins Mtn

deformation twins, bent twins, and subgrain boundaries in plagioclase are evidence for plastic deformation and high strain, and abundant hornblende is a testament to the large amount of fluid that has permeated the rocks. Recognition that the orebody, retrograde metamorphism and L-S deformation fabric all coincide with the southern margin of the olivine meta-gabbro led Goldblum and Hill (1992) to hypothesize that the high fluid flow required for growth of large garnets was the result of high-temperature shear zone that crossed the contact of lithologies with contrasting rheologies and propose that ductility contrasts at this lithologic contact was responsible for localizing garnet growth, retrograde metamorphism and fabric development. Grain size reduction by cataclasis was replaced by recrystallization as the hydrated ore body replaced the olivine meta-gabbro during ductile deformation.

The Gore Mountain garnets are chemically homogeneous indicating that (a) the garnets grew under conditions in which all chemical components were continuously available and (b) that temperature and pressure conditions must have been uniform during the period of garnet formation. If the garnet amphibolite zone within the the original gabbro represents a zone wherein fH_2O was higher than elsewhere during the granulite facies metamorphism, this may have facilitated diffusion and favored growth of very large garnets and thick hornblende rims at the expense of plagioclase and pyroxene. The presence of volatile components, particularly H_2O, promotes the growth of large crystals. The physical and chemical conditions necessary for the nucleation of a mineral may be different from the conditions necessary for the growth of that mineral. It has been speculated (Luther, 1976) that the physico-chemical environment was poor for the nucleation of garnet but that the environment was conducive to the growth of garnet. Therefore, the garnet crystals that did nucleate grew to large size. Growth might have been abetted by aqueous transport of components due to elevated fH_2O in this portion of the ore body.

STOP A-3: MAGNETITE-ILIMENITE DEPOSIT AT SANFORD LAKE (TAHAWAS)

Description modified from McLelland, Kelly, and Peterson (1993).

Introduction

According to Stephenson (1945) the Sanford Lake magnetite-ilmenite ores were discovered in 1826 when a party, entering from Indian Pass, encountered the now mined-out "Iron Dam" of ore which extended across the Hudson River at the present site of the Tahawus Club. Mining began in the 1830's and by the 1840's was supplying ore for the first cast-steel plant in America (Adirondack Iron and Steel Company, Jersey City, N.J.). In 1851 steel from this plant was awarded a gold medal at the World's Fair in London. Production halted in 1858; was reorganized as the MacIntyre Iron Co. in 1894; and resumed production in 1906. Despite extensive planning, little ore was produced or shipped. In 1908 a French metallurgist, A. Rossi, employed by the MacIntyre Iron Co., discovered the suitability of titanium as a white paint pigment. Continued transportation difficulties plagued mining operations until 1941 when N.L. Industries, Titanium Division, acquired the Sanford Hill-South Extension ore body. By 1942 ilmenite concentrates were being shipped. The mine was extensively developed during, and after, World War II where it was exploited for titanium, and a railroad was built to North Creek. Since 1980, mining activity has slowed, and at present a skeleton crew works the deposits for a variety of purposes. The Sanford Hill-South Extension pit is flooded. Thirty-three and one-half million tons of ore was shipped to the crusher between 1942 and 1968 (Gross, 1968).

Geology

Gross (1968) has described the geology of the Sanford Lake district and the following is a brief summary.

Rock types

The rocks of the Sanford Lake district are generally regarded as members of a genetically related Adirondack anorthosite series (Marcy-type and Whiteface-type). Locally anorthosite grades into gabbro by an increase in mafic minerals. Buddington (1939) divided the anorthosite-to-gabbro sequence into four rock types on the basis of the mafic mineral content: anorthosite (0-10%), gabbroic anorthosite (10-22.5%), anorthositic gabbro (22.5-35%) and gabbro (>35%). However, in local mapping the term anorthositic gabbro was not used.

Marcy-type anorthosite predominates in the mine area. It is a coarse-grained, somewhat porphyritic, dark, blue-gray to greenish rock composed mainly of labradorite feldspar ($An_{37\text{-}65}$, Stephenson, 1945; $An_{50\text{-}53}$, Avenius, 1948). The dark color is attributed to an abundance of very fine inclusions of mafic minerals within the crystals of labradorite. Pyroxene (clinopyroxene and orthopyroxene) and hornblende are the major accessory minerals and biotite, grossularite, magnetite, ilmenite and sulfides are present in minor amounts in the anorthosite. Marcy-type anorthosite shows very weak foliation.

Whiteface-type anorthosite, which is generally considered a border phase, occurs in variable amounts. It is of a medium gray color, and is considerably finer-grained than Marcy-type anorthosite. In addition, it is more equigranular and phenocrysts are less abundant. Andesine plagioclase is the major phase and pyroxene and amphibole the most abundant accessory phases. Trace amounts of garnet and biotite are present.

Anorthosite grades into gabbroic anorthosite with an increase in fine- to medium-grained intergranular plagioclase ($An_{40\text{-}50}$) and mafic minerals. Up to 50 percent of plagioclase phenocrysts show a preferred orientation paralleling that of the orebodies. The phenocrysts are dark green in color, contrasting with the light greenish-gray color of the finer-grained plagioclase matrix. Plagioclase tend to be free of inclusions compared to plagioclase in anorthosite.

Gabbro occurs as a fine- to medium-grained rock having a uniform texture and a distinct foliation. The color ranges from gray to brownish-gray to black, depending on the amount and composition of the ferromagnesian and ore minerals present. Gabbro consists of plagioclase ($An_{30\text{-}40}$) and 35 to 60 modal percent mafic minerals. Fine-grained garnet and hornblende increase as the fraction of ore minerals decrease. Pyroxene commonly rims magnetite and ilmenite and is in turn rimmed by fine-grained garnet or hornblende or both. The gabbroic units parallel the ore lenses.

Faulting in the district conforms to fault patterns present throughout the region as a whole. Prominent faults strike NE-SW with steep dips to the NW. A less prominent fault set is developed at almost right angles to the prominent one. Three major joint sets are exposed in the pits; two sets have vertical dips and the third set is horizontal. Joint sets generally parallel regional structures.

Orebodies

Although there are four important mineral bodies in the district -- Sanford Hill-South Extension (Sanford Hill), Cheny Pond, Mount Adams (Iron Mountain or Ore Mountain) and Upper Works (Calamity-Mill Pond) -- all ore production has come from Sanford Hill-South Extension (Fig. A-3) except for a few thousand tons mined from Upper Works prior to 1900. For practical purposes mine mapping was done strictly on the basis of sample assays (percent TiO_2) for which the following classification evolved: anorthosite (0-5.4%), gabbro (5.5-9.4%), low grade protore (9.5-13.4%), medium grade ore (13.5-17.5%) and high grade ore (>17.5%). The ore in the Sanford Lake district occurs in two major modes: 1) as lean or disseminated ore in gabbro ("gabbroic ore"), and 2) as massive, rich ore generally in anorthosite but locally within gabbro ("anorthositic ore"). The lean ore within gabbro is gradational into the host rock with which it is commonly layered (Ashwal, 1978, p. 106), but in anorthosite the ore exhibits sharp contacts relative to the host rock. Gabbroic ores are fine-grained and have a well defined foliation similar to that in gabbro. Ore-bearing gabbro may sharply crosscut anorthosite. The anorthositic ores are coarse-grained, show no flow structures, have irregular dimensions and exhibit sharp contacts with host anorthosite and with disseminated ore in anorthosite.

Both types of ores occur in the Sanford Hill-South Extension zone. Anorthositic ore forms a footwall orebody and gabbroic type forms a hanging-wall orebody (Fig A-4.). These orebodies are separated by various widths of anorthosite and/or gabbro rock. The Mt. Adams orebody averages 21.3 m in width over a length of 530 m and the portion that hosts ore grade mineralization is nearly all of the anorthositic type. Both types of ore are present in the Upper Works orebody. At Cheney Pond only gabbroic ore is present, and it has sharp contacts between anorthosite and gabbro at both the upper and lower limits.

Ore Mineralogy

The ore in the Sanford Lake district consists of titaniferous magnetite and hemo-ilmenite in subequal amounts with ilmenite generally being slightly more abundant. Lamellae of ilmenite in magnetite originated via subsolidus oxidation-exsolution (Haggerty, 1976). Green spinel (spinel-hercynite solid-solution) commonly forms as an exsolution product in magnetite. Iron sulfides occur as accessory phases. Both titanomagnetite and hemo-ilmenite form abundant small, rod-like inclusions in associated plagioclase sometimes rendering them black and opaque. Gangue minerals include feldspar (10-20%), garnet (grossular-almandine-andradite solid solution; 3-8%), pyroxene (ortho- and clinopyroxene; 4-7%), hornblende (1-3%), sulfides (≈1.5%), iron-rich biotite (≈1%) (Hayburn, 1960). The sulfides include chalcopyrite, sphalerite, molybdenite, pyrrhotite and pyrite. Other phases present in minor amounts include apatite, prehnite, barite, orthoclase, leucoxene, scapolite, epidote and quartz (Gross, 1968).

Table A-3. Compositions of ilmeno-magnetite in rocks of anorthosite series.

	Ilmeno-magnetite Wt % TiO_2	Wt. % MnO	Ilmenite Mole % Fe_2TiO_4	Mole % Fe_2O_3	Wt. % MnO	Estimated Values T^oC ±50	log fO_2 ±1.0
Sr-153	11.2	0.30	31.9	9.4	0.57	810	-13.6
Sr-289	8.7	0.20	24.0	8.1	0.45	730	-13.8
Sr-181	4.1	0.02	11.0	12.9	0.20	635	-17.1

Sr-153 Feldspathic oxide mineral-rich pyroxenite with sparse garnet. One-half mile west of Derrick, St Regis quadrangle.
Sr-289 Feldspathic oxide mineral-rich pyroxinite with considerable garnet. South center of St. Regis quad.
Sr-181 Garnetiferous gabbroic anorthosite gneiss. One-half mile south of Wabeek. Saranac quadrangle.

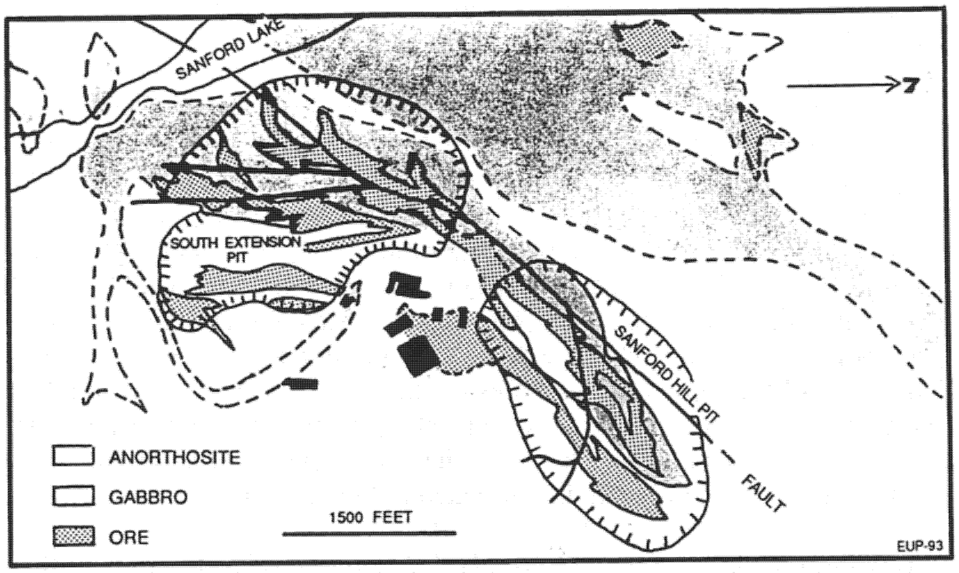


Figure A-3. Geologic map of the Sanford Hill-South Extension orebody (modified from Gross, 1968).

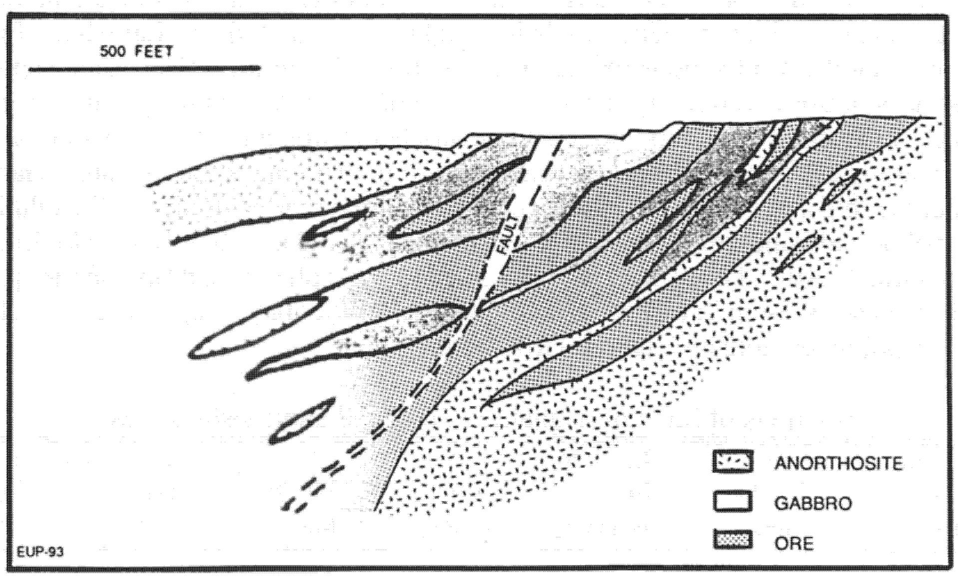


Figure A-4. Representative geologic cross section of the Sanford Hill pit (modified from Gross, 1968)

The average composition of titanomagnetite and hemo-ilmenite in the principal ore deposits is $Mt_{81}Usp_{18}$ and $Ilm_{94}Hm_6$, respectively (Kelly, 1979). Table A-3 shows that this composition yields a crystallization temperature of 650°C and an oxygen fugacity of 10^{-19} atm, both indicative of metamorphic conditions. Clinopyroxene compositions yield temperatures of 550° to 850°C when plotted on a temperature determinative curves of Ross and Huebner (1975).

Chemistry of Ores

With the exception of apatite-rich, and possibly nelsonitic rocks near Cheney Pond (Kolker, 1980), the concentrations of P_2O_5 in the ore deposits are strikingly low (Table A-4).

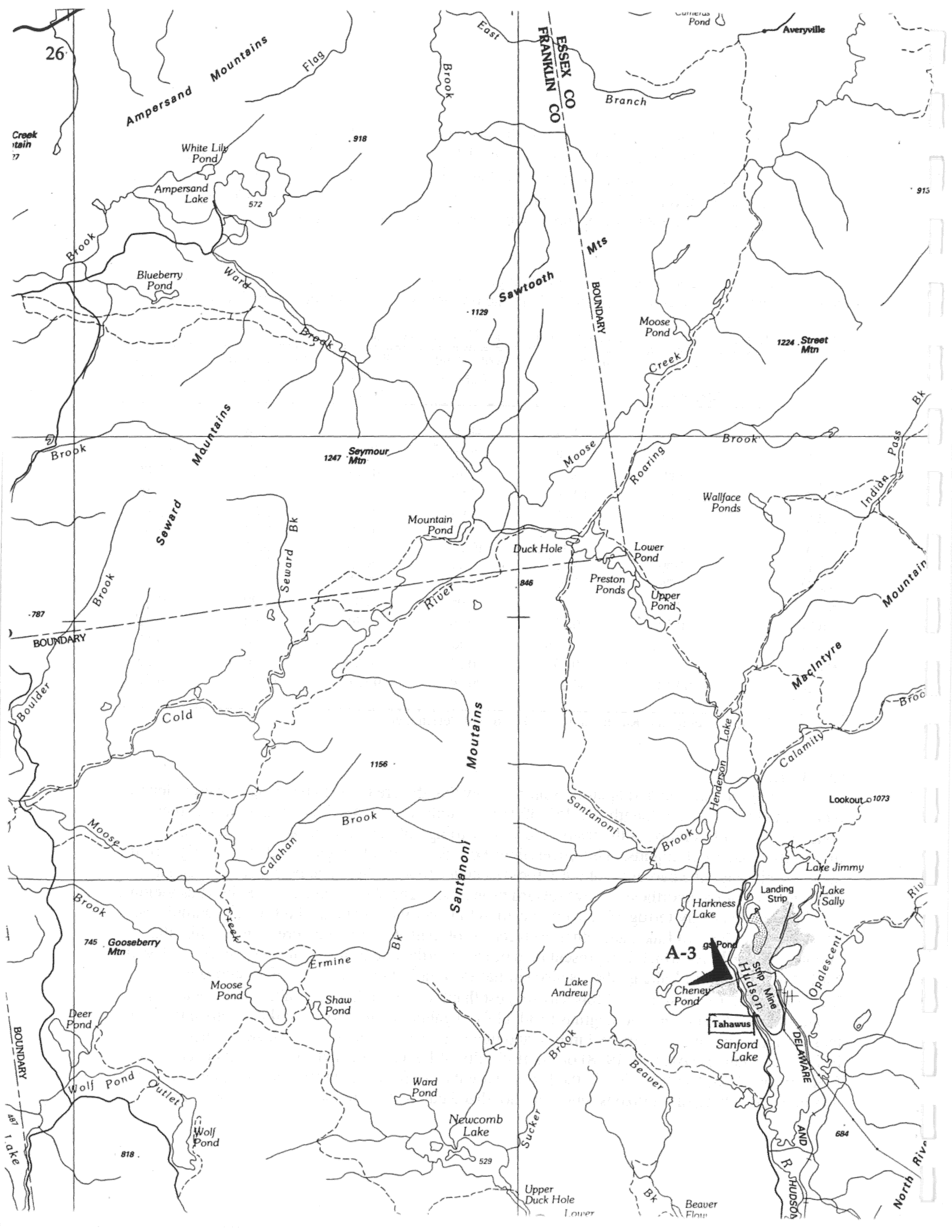
Table A-4. Chemical Analyses of Sanford Lake District Rocks

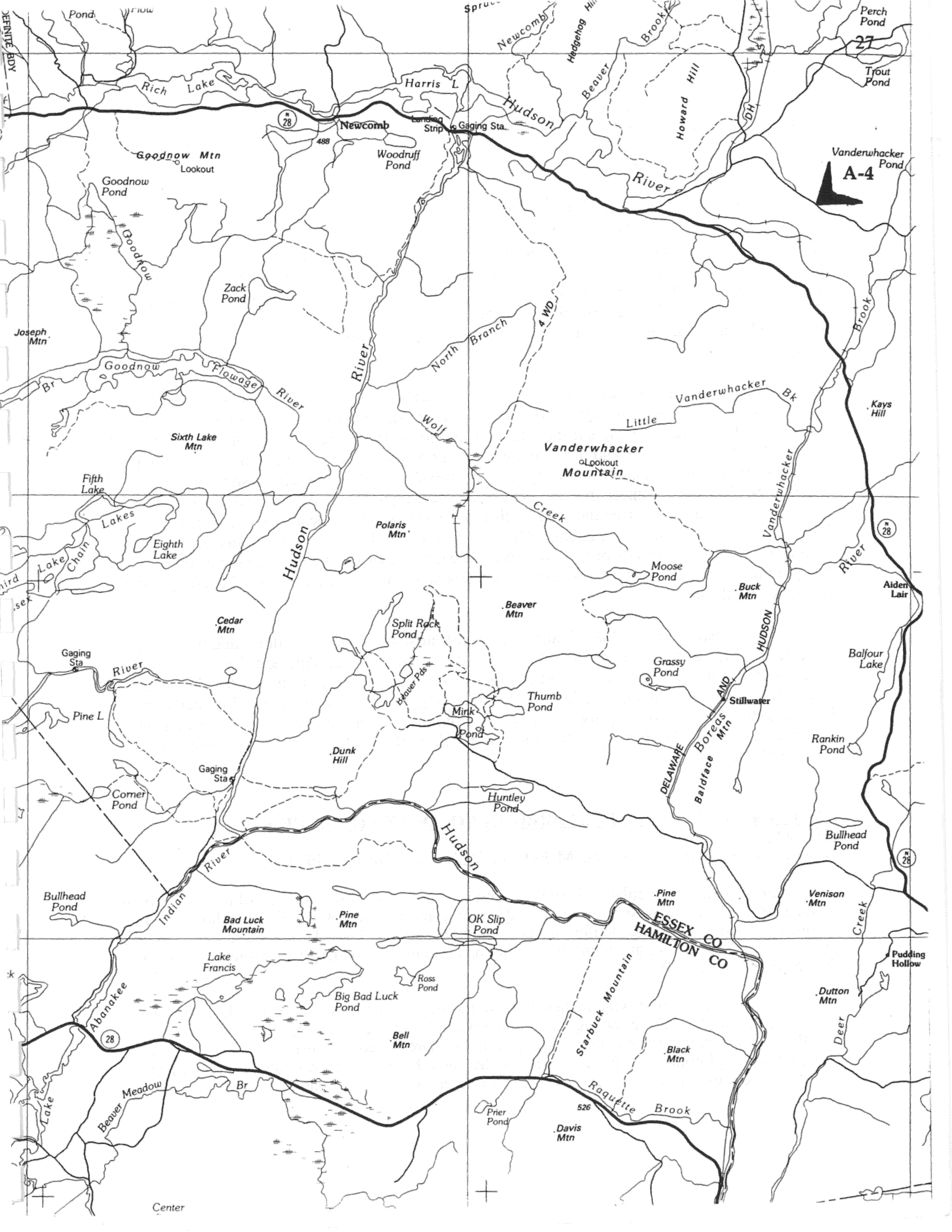
	Tahawus Olivine Metagabbro	Sanford Lake Gabbro[1]	Lincoln Pond Gabbro[2]	Westport Mafic Gabbro[2]	Woolen Mill Gabbro[2]	Sanford Lake Ore[1]
SiO_2	47.62	39.04	44.70	47.88	47.16	4.59
TiO_2	0.82	6.78	5.26	1.20	3.37	18.58
Al_2O_3	18.69	13.09	12.46	18.90	14.45	5.48
Fe_2O_3	11.40	19.09	4.63	1.39	1.61	nd
FeO	nd	nd	12.99	10.45	13.81	66.37
MnO	0.14	0.24	0.17	0.16	0.57	0.28
MgO	8.85	5.31	10.20	7.10	5.24	3.39
CaO	8.29	9.77	5.34	8.36	8.13	0.31
Na_2O	2.76	2.02	2.47	2.75	3.09	0.22
K_2O	0.41	0.66	0.95	0.81	1.20	0.09
P_2O5_5	0.12	0.23	0.28	0.20	nd	0.01
V_2O_5	nd	nd	nd	nd	nd	0.45
H_2O		0.03	0.64	0.61	0.60	0.10
Total	99.10	96.26	100.09	99.81	99.23	99.87

1) Kelly, 1979; 2) Kemp and Ruedemann, 1910; nd, not determined

Ore Genesis

Diamond drilling and open-pit mining show that the ore tends to be concentrated in lenses measuring 600-700 m in length and 150-300 m in width (Fig. A-4). It is not known whether this conformable configuration is the result of crystal settling, intrusion, or the accumulation of immiscible oxide-rich liquids. This uncertainty extends to the petrologic details of the origin of the deposits, although the evolution of these rocks is understood in the broad perspective. The late differentiates of the anorthosite move toward pronounced enrichment in Fe-Ti-oxides thus yielding liquids of increasingly ferrogabbroic composition together with associated ultramafic cumulates. The gabbro at Sanford Lake, and other occurrences of magnetite-ilmenite ore, is not unlike the Woolen Mill gabbro, which is representative of late anorthositic differentiates (Table A-4). Except for P_2O_5 the Sanford Lake gabbro is also similar to Buddington's (1939) mafic gabbro from McCauley Mt. Comparisons of this sort suggest that the ores at Sanford Lake are the result of progressive differentiation of magmas residual from gabbroic anorthosite and that, at some point, these magmas became so enriched in iron and titanium that they either crystallize magnetite-ilmenite cumulates (Ashwal, 1978) or immiscibility of Fe-Ti oxide and silicate melts occurs (cf. Stephenson 1945; Kelly 1979). In the former case the conformable layers represent cumulate beds and in cross-cutting ore horizons represent mobilized cumulates. In the latter case, both layered


26
Ampersand Mountains
Flag
Brook
East
ESSEX CO
FRANKLIN CO
Branch
Averyville
Pond
Creek
White Lily Pond
Ampersand Lake
572
918
913
Brook
Ward
Blueberry Pond
Brook
Sawtooth
Mts
BOUNDARY
1129
Moose Pond
Creek
1224 Street Mtn
Seward Mountains
Brook
1247 Seymour Mtn
Moose
Roaring
Brook
Indian Pass Bk
Wallface Ponds
Seward
Seward Bk
Brook
Mountain Pond
Duck Hole
Lower Pond
Preston Ponds
Upper Pond
846
River
787
BOUNDARY
Mountain
MacIntyre
Boulder
Cold
Moutains
Lake
Henderson
Calamity
Brook
1156
Lookout 1073
Santanoni
Moose
Brook
Calahan
Brook
Lake Jimmy
Santanoni
Landing Strip
Lake Sally
Harkness Lake
Brook
745 Gooseberry Mtn
Creek
Bk
Ermine
A-3
Pond
Hudson
Strip Mine
Opalescent
Lake Andrew
Cheney Pond
Moose Pond
Shaw Pond
Tahawus
Deer Pond
Sanford Lake
DELAWARE
Brook
Beaver
BOUNDARY
Wolf Pond Outlet
Ward Pond
Newcomb Lake
Sucker
Wolf Pond
684
AND
818
529
R HUDSON
North River
Lake
487
Upper Duck Hole
Bk
Beaver



A-4
Perch Pond
27
Trout Pond
Vanderwhacker Pond
Rich Lake
Harris L
Newcomb
Landing Strip
Gaging Sta
28
488
Woodruff Pond
Goodnow Mtn
Lookout
Goodnow Pond
Goodnow
Hudson River
Newcomb
Hedgehog Hill
Beaver Brook
Howard Hill
DH
Zack Pond
4 WD
North Branch
Joseph Mtn
Goodnow Flowage
Br
River
Wolf Creek
Little Vanderwhacker Bk
Brook
Kays Hill
Sixth Lake Mtn
Vanderwhacker Mountain
Lookout
Fifth Lake
Lakes
Eighth Lake
Chain Lake
Polaris Mtn
Hudson
Vanderwhacker
River
Moose Pond
Buck Mtn
Aiden Lair
Beaver Mtn
Cedar Mtn
Split Rock Pond
Gaging Sta
River
Beaver Pds
HUDSON AND DELAWARE
Balfour Lake
Grassy Pond
Stillwater
Thumb Pond
Mink Pond
Pine L
Boreas Mtn
Baldface Mtn
Dunk Hill
Rankin Pond
Gaging Sta
Corner Pond
Huntley Pond
Hudson
Bullhead Pond
River
Indian
Bullhead Pond
Pine Mtn
Venison Mtn
Bad Luck Mountain
Pine Mtn
OK Slip Pond
ESSEX CO
HAMILTON CO
Creek
Pudding Hollow
Lake Francis
Ross Pond
Starbuck Mountain
Abanakee
Big Bad Luck Pond
Dutton Mtn
Bell Mtn
Black Mtn
Deer
Beaver Meadow Br
Lake
Raquette Brook
526
Prier Pond
Davis Mtn
Center

and cross-cutting configurations can be explained on the basis of an immiscible Fe-Ti oxide liquid. A third possibility exists which is a combination of the foregoing alternatives, i.e., magnetite-ilmenite could begin to precipitate relatively early in the history of the complex but continued fractionation might still result in liquid immiscibility at a later stage.

Arguments against liquid immiscibility at Sanford Lake have commonly focussed on the low concentrations of apatite in these rocks. Lindsley (1991), however, points out that P_2O_5 and apatite do not necessarily partition into the immiscible oxide melt. Moreover, P_2O_5 may not be a direct cause of liquid immiscibility but, rather, may play an indirect role in keeping the magma molten until Fe-Ti-O networks in the melt can no longer coexist with the silicate networks and immiscibility occurs. Because of this late magmatic association, apatite and immiscibility would appear to be more directly connected than may actually be the case.

Field Stop

The stop at Sanford Lake takes advantage of excellent relationships exhibited in boulders on either side of the Calamity Brook Road at the gated entrance to the Cheney Pond Road. Over three dozen large, fresh boulders from the mines provide outstanding exposure of anorthosite, gabbro, ore-bearing gabbro, and massive ore crosscutting anorthosite. Several boulders containing both anorthosite and ore exhibit what appear to be coeval and pillowing relationships between the two phases. In other instances massive ore and ore-bearing gabbro cross-cut anorthosite. A number of boulders show irregular oxide-rich veins, some of which clearly contain separated fractions of oxide and silicate minerals. Many of the ore boulders contain xenoliths and enclaves of anorthosite and xenocrysts of andesine some of which are black due to oxide inclusions. Several boulders of Marcy-type anorthosite are present as are some sheared, hornblende-bearing gabbroic anorthosite.

Relationships seen in these boulders demonstrate that the iron-titanium-oxide ore derives from the gabbros and bears intrusive relationships to the anorthosite. Polished slabs show the oxide phases to intimately penetrate and disrupt the anorthosite on a scale smaller than the grain size of magnetite and ilmenite in adjoining ore. This suggests that the oxide melt intruded as a liquid and this observation, together with evidence of liquid immiscibility in similar rocks, lead McLelland *et al.* (1993) to suggest that most, if not all, of the Sanford Lake ores were emplaced as immiscible liquids.

STOP A-4: OLIVINE METAGABBRO IN ROADCUTS ON BLUE RIDGE HIGHWAY

Description modified from Bohlen, McLelland, Valley, and Chiarenzelli (1992).

Steep roadcuts on either side of the highway expose good examples of Adirondack olivine metagabbro. The rock consists of round, ~.25 cm coronas of red biotite and brown hornblende coronas on oxides set in a garnetiferous matrix of green, spinel-clouded plagioclase and subophitic pyroxenes. Olivine is not abundant in this outcrop although it is widespread throughout most of this relatively large body.

A large variety of olivine metagabbros in the Adirondacks, ranging from Mg-rich to Fe-rich, are exposed throughout the region but are especially abundant in proximity to bodies of anorthosite. The southern and eastern margins of the Marcy massif are especially rich in olivine metagabbro. McLelland (1986) has suggested that these bodies are representative of the magmas ponded at the crust-mantle interface that gave rise to the parental magmas of the anorthosite. The bodies not exposed at the surface are interpreted to be late plutons that ascended, without ponding, after the major mass of Anorhosite-Mangerite-Charnokite-Granite (AMCG) suite had risen and provided crustal pathways. This suggestion is consistent with geochronological data indicating that the gabbros are contemporaneous with the AMCG suite..

Coronas developed in olivine metagabbros have been of petrologic interest for over 100 years. McLelland and P.R. Whitney investigated these features in the 1970s and 1980s.

STOP A-5: SARANAC LAKE - METANORTHOSITE

Description modified from Bohlen, McLelland, Valley, and Chiarenzelli (1992) and from Abruzzi (1978).

The coarse grained andesine rock here is typical, in both composition and igneous texture, of the most voluminous member of the anorthosite series. This exposure contrast with the gabbroic (noritic in part) metanorthosite of the "border facies" in having ,10 percent mafic minerals and in being coarser-grained. Buddington (1939) interpreted the finer-grained border facies to be a relatively chilled sample of the parent magma.

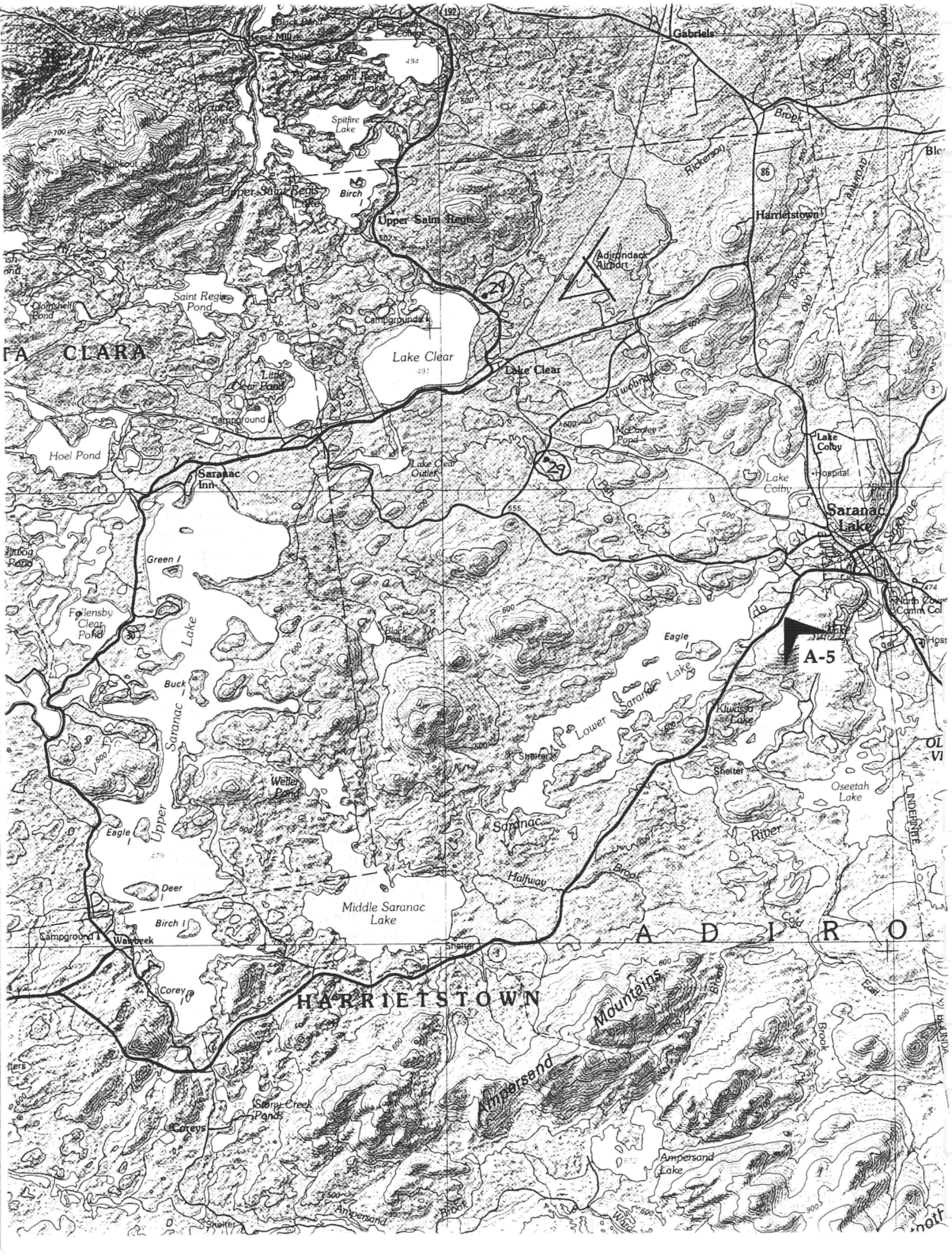
The primary minerals are andesine (An_{40}-$An5_0$ and locally antiperthitic), augite, hypersthene, ilmenite, magnetite, and apatite. Amounts and relative proportions of the mafic minerals vary considerably. Quartz is a normative mineral (up to 5%) but is uncommon in thin section. The andesine occurs in two forms, as bluish-gray megacrysts dusted with extremely fine iron-titanium oxides, and as a clear, finer grained recrystallized groundmass. Metamorphic minerals include garnet, secondary clinopyroxene, amphibole, and less commonly biotite, clinozoisite, and scapolite.

Zircons extracted from anorthosite near this location are small and clear, and yield ages of 1050 Ma (McLelland *et al.*, 1990). Silver (1969) originally interpreted these low-uranium, "soccer-ball" type of zircon as metamorphic, and McLelland and Chiarenzelli (1990) have corroborated this by demonstrating an emplacement age of 1130-1135 Ma for the anorthosite. In addition, the cores of air abraded zircons yield minimum ages >1113 Ma and baddelyite in these rocks give minimum ages of >1087 Ma. It is common to find 1050 Ma zircon growing on baddelyite grains. All of these observations converge upon the conclusion that the 1050 Ma zircon ages date granulite facies metamorphism.

The oxygen isotopic composition of the Marcy anorthosite is 2.5 permil heavier than other "normal" anorthosites. This anomaly was ascribed by Taylor (1969) to exchange with pervasive ^{18}O-enriched C-O-H fluids during regional metamorphism. However, Morrison and Valley (1988a) have shown that the ^{18}O enrichment is a magmatic feature that was acquired before the anorthosite intruded the crust at shallow levels. Values of $\delta^{18}O$ in the metanorthocite in the NW lobe of the Marcy Massif are extremely homogenous ($\delta^{18}O = 9.3\pm0.2$), which in conjunction with the preservation of magmatic features (Davis, 1969), indicates that the oxygen isotopic composition reflect magmatic values rather than exchange with metamorphic fluids.

In anorthosites of the Saranac Lake area coronas have developed around pyroxenes and Fe, Ti-oxides. The necessary reactants responsible for the development of the coronas seem to be orthopyroxene, plagioclase, and Fe, Ti-oxides (ilmenite and magnetite). Depending on the compositions of the ferromagnesian phases, quartz can be either a reactant or a product (McLelland and Whitney, 1977). Abruzzi (1978) described three types of corona.

Type 1: The cores of these coronas are generally composed of orthopyroxene, but in a few cases the cores consist mainly of clinopyroxene grains, suggesting that the primary orthopyroxene



Gabriels
Upper Saint Regis
Upper Saint Regis Lake
Spitfire Lake
Spectacle Ponds
Birch I
Harrietstown
Adirondack Airport
Saint Regis Pond
Clamshell Pond
CLARA
Lake Clear
Lake Clear
Little Clear Pond
Campground
Hoel Pond
Saranac Inn
McCauley Pond
Lake Colby
Hospital
Saranac Lake
Green I
Follensby Clear Pond
Buck I
Upper Saranac Lake
Black Pond
Eagle I
Lower Saranac Lake
Kiwassa Lake
A-5
Weller Pond
Oseetah Lake
Saranac River
Halfway Brook
Eagle I
Deer I
Birch I
Middle Saranac Lake
Campground
Wawbeek
Shelter
A D I R O
Cold Brook
Corey I
HARRIETSTOWN
Mountains
Ampersand
Stony Creek Ponds
Coreys
Ampersand Lake
Ampersand Brook
Shelter

has been exhausted. Some of the clinopyroxene grains contain exsolution lamellae of orthopyroxene. The grains of orthopyroxene that remain are highly altered and barely distinguishable. Thus, the cores of the coronas are composed of an aggregate of ortho- and clinopyroxenes. This core is generally surrounded by a shell of clinopyroxene, which in turn is surrounded by a narrow rim of garnet. The rims of products, clinopyroxene and garnet, separate the reactants, orthopyroxene and plagioclase.

Type 2: In some of the thin sections studied the pyroxenes are surrounded by a rim, of variable thickness, of an amphibole which is pleochroic from green to brown. In these rocks the occurrence of garnet is, to a large extent, limited to coronas around Fe, Ti-oxides. There is, in some cases, a small amount of garnet adjacent to the amphibole, but in general the development of the amphibole is more pronounced where the garnet rims are undeveloped.

Type 3: Coronas are also well formed around the Fe, Ti-oxides, lending credibility to the theory (McLelland and Whitney, 1977) that the oxides are necessary reactants in the garnet-producing reaction. In most cases a very well-developed, though narrow, rim of garnet surrounds the oxide phase.

Recent studies of corona textures in similar rocks by Johnson and Carlson (1990), Carlson and Johnson (1991), and Joesten (1986) suggest that a simple interpretation of the history of these textures may be unwise.

STOP A-6: CASCADE SLIDE, MARBLE XENOLITH IN ANORTHOSITE

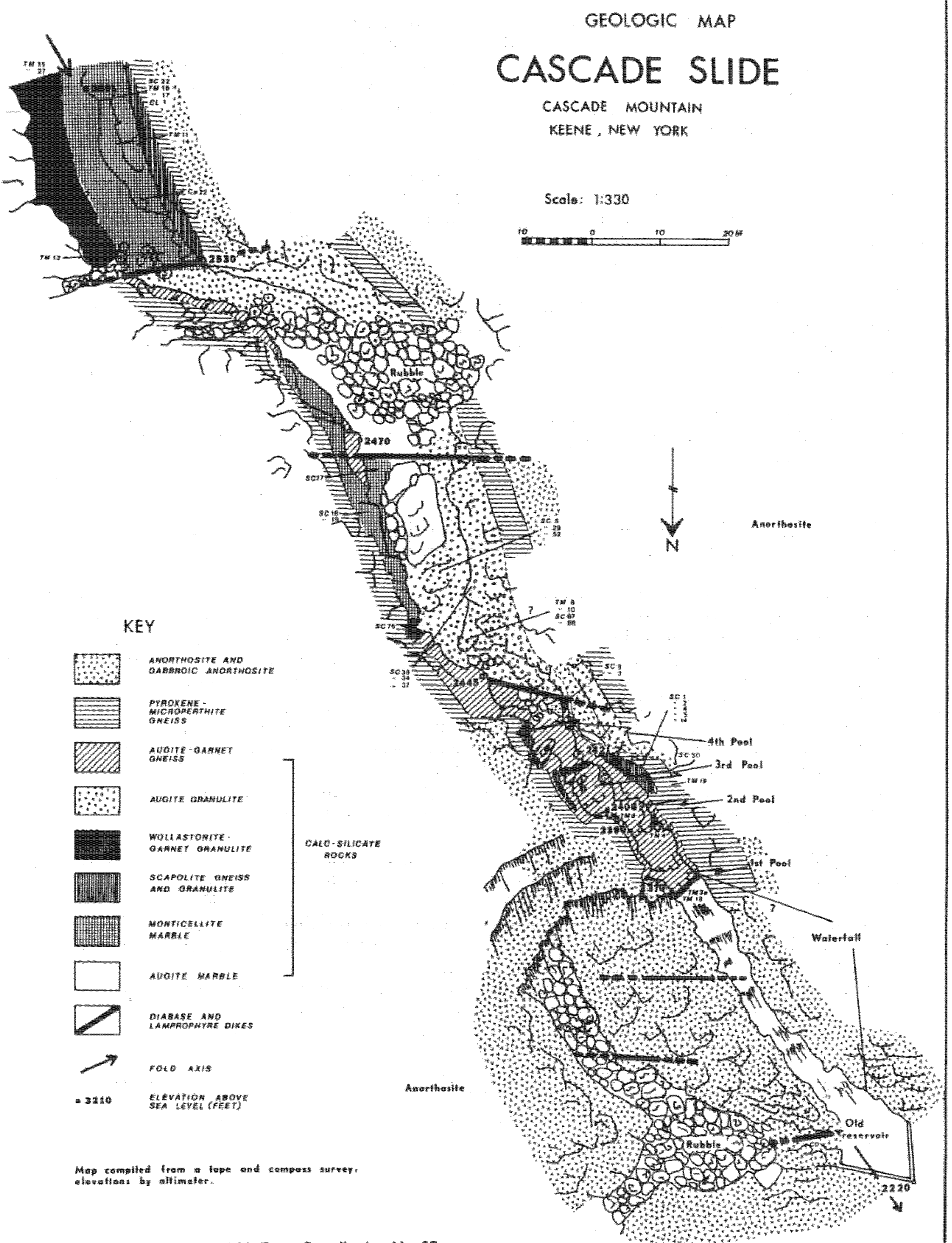
Description from Bohlen, McLelland, Valley, and Chiarenzelli (1992).

Walk south up talus slope to remains of a dam at the base of a waterfall. From this point, climb the steep gully to the east of the falls. Use extreme caution! This is a very steep climb for about 60 m, and there are many loose rocks. When you have attained enough elevation to be above the cliff, traverse to the west to the stream that supplies the waterfall. In the stream bed above the falls there are several xenoliths and schlieren of marble ± calcsilicate, surrounded by anorthosite. The largest of these bodies measures approximately 30 x 200 m in exposure, is compositionally zoned, and contains several unusual minerals. Most notably, the xenolith contains sanidinite facies index minerals wollastonite, monticellite (Mo_{92-89}), and akermanite as well as cuspodine, harkerite, and wilkeite (Kemp 1920; Baillieul 1976; Tracy *et al.* 1978; Valley and Essene 1980b). Other minerals present include garnet (Grs_{80-18}, And_{80-15}), spinel (Mg_{73}), calcite, forsterite (Fo_{92}), magnetite, clinopyroxene scapolite (Me_{78-50}), quartz, and titanite.

Field relations, deformation and geochronology make it clear that these marble bodies were entrained within the anorthositic magma before the peak of granulite facies metamorphism. The exact timing of intrusion vs. regional metamorphism is still a matter of debate. We strongly favor pre-metamorphic rather than deep syn-metamorphic intrusion as originally proposed by Valley and O'Neill (1982), but in either case it is certain that both anorthosite and marble experienced the pressures and temperatures of granulite facies metamorphism (Valley and Essene 1980b). Thus, the mineralogy of these bodies may be used to study the P-T fluid conditions of granulite facies metamorphism. The origin of these minerals, which we believe was at low P and high T, is irrelevant in this regard because of the pervasive nature of the granulite overprint.

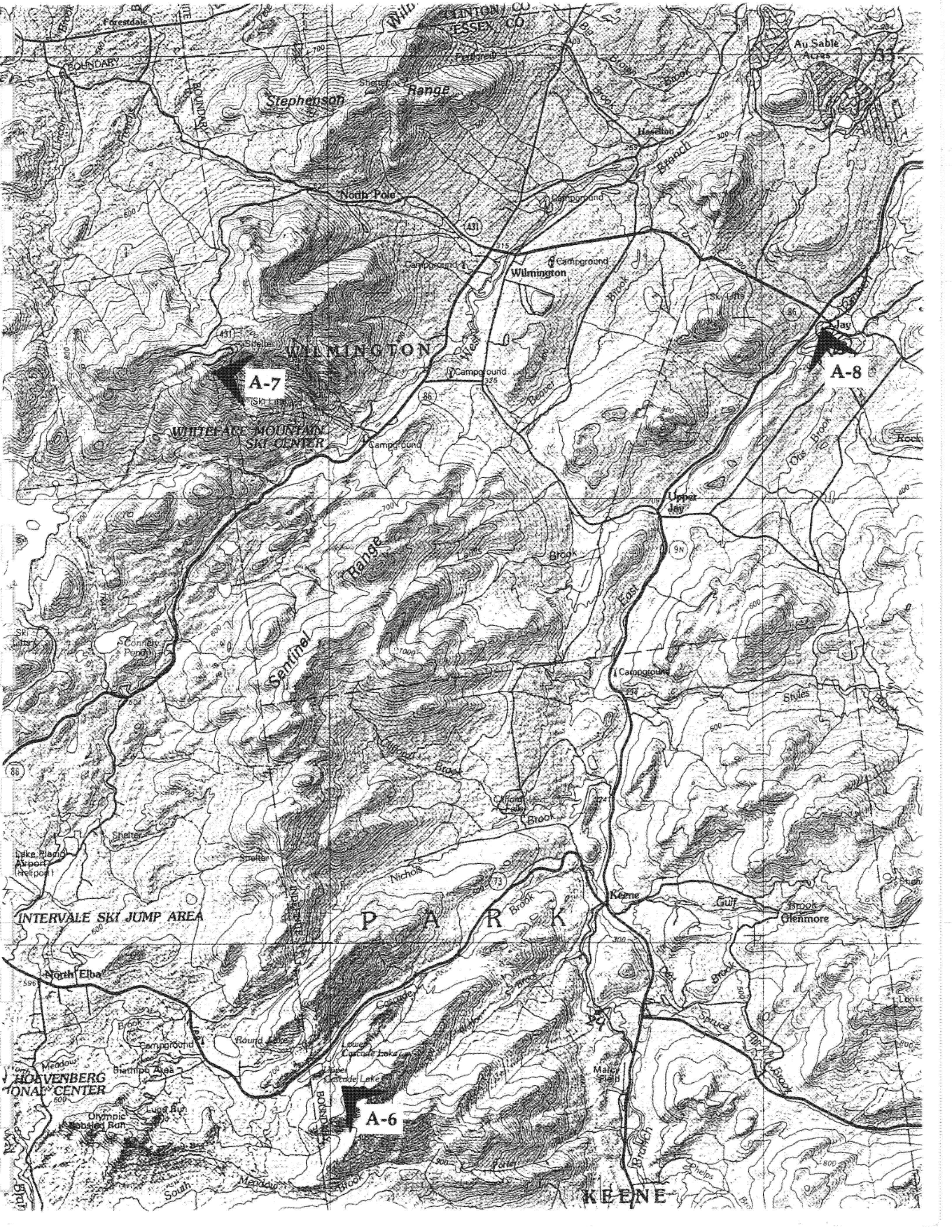
Solid-solid mineral reactions at Cascade Slide indicate that P and T attained at least 7.4 kbar and 750 °C, respectively (Valley and Essene, 1980b; Bohlen *et al.*, 1985). Valley and Essene (1980b) describe assemblages of akermanite + monticellite + wollastonite with equilibrium metamorphic textures as well as symplectic intergrowths of wollastonite and monticellite. At these

PLATE 1



Map by Thomas A. Baillieul, 1976. From Contribution No. 27, Geology Dept., Univ. Mass.-Amherst.

TAB 75



A-7
A-8
A-6
WILMINGTON
WHITEFACE MOUNTAIN
SKI CENTER
Wilmington
North Pole
Haselton
Au Sable
Acres
Jay
Upper
Jay
Stephenson
Range
Sentinel
Range
CLINTON CO
ESSEX CO
Forestdale
BOUNDARY
Campground
Shelter
Ski Lifts
Connery
Pond
Lake Placid
Airport
(Heliport)
INTERVALE SKI JUMP AREA
P A R K
Keene
Glenmore
North Elba
Round Lake
Lower
Cascade Lake
Upper
Cascade Lake
Marcy
Field
Clifford
Falls
Biathlon Area
Olympic
Bobsled Run
Luge Run
HOEVENBERG
ONAL CENTER
KEENE
Styles
Brook
Gulf
Spruce
Nichols
Cascade
Clifford
Beaver
Otis
East
West
Branch
Phelps
Porter
South
Meadow
431
86
9N
73

temperatures and pressures, the presence of wollastonite, monticellite or akermanite requires that log f_{CO_2} be ≤ 4.35, ≤ 3.32, or ≤ 2.5 respectively.

Further evidence that granulite facies fluid infiltration has not been important at Cascade Slide comes from oxygen isotopes (Valley and O'Neil, 1984; Valley 1985). Any fluids (H_2O or CO_2) passing through the xenolith would first have passed through the surrounding anorthosite ($\delta^{18}O = 9.7$). Subsequent exchange with the calcsilicates ($\delta^{18}O = 17.6$ to 26.1) would tend to homogenize this large premetamorphic difference with the result that $\delta^{18}O$ in the xenolith would be reduced. The highest values of $\delta^{18}O$ (26.1) in monticellite marble are thus very restrictive to theories of fluid infiltration and require fluid/rock < 0.1.

Three lines of evidence argue against the presence of fluid during the granulite facies metamorphism at Cascade Slide: 1) Assemblages of monticellite + forsterite + diopside + calcite + spinel plot in the fluid-absent field, including that if a fluid had existed, $P_{H_2O} + P_{CO_2} \leq 0.4$ kbar. 2) The large gradients in buffered values of f_{CO_2} across the body and the fragile nature of the buffering assemblages would have been erased by CO_2 infiltration. even by quantities as low as CO_2/rock = 0.001. 3) The preservation of high $\delta^{18}O$ in the core of the xenolith and the sharp gradients of up to 18 permil/15 m would all have been homogenized if either H_2O or CO_2 had infiltrated the xenolith in quantities greater that fluid/rock = 0.1. These results are all consistent with the polymetamorphic history proposed by Valley (1985).

Monticellite has also been found at Westin Mines (5 km to the E of Cascade Slide) where magnetite skarn replaces marble at the contact of the anorthosite massif (Valley and Graham, 1991). This locality is on private property and will not be visited. Magnetites from marble at this deposit were the first to be analyzed for oxygen isotope ratio by ion microprobe with accuracy of ±1‰ (1σ). This analysis yields spatial resolution as small as 2 µm and has reduced sample size by 11 orders of magnitude relative to conventional techniques, permitting new studies of oxygen diffusion, fluid exchange, and Adirondack cooling rate.

STOP A-7: TOP OF WHITEFACE MOUNTAIN

Description modified from Bohlen, McLelland, Valley, and Chiarenzelli (1992).

The rocks at the summit of Whiteface Mountain represent the type locality for the Whiteface type, or facies, of anorthosite. This is typically finer-grained, garnetiferous, and somewhat more gabbroic than the Marcy facies. In general the Whiteface facies exhibits a chalky-white color and commonly contains significant quantities of black hornblende.

As originally defined, the Whiteface facies was considered to be the border facies of the anorthosite massif with the coarse, less gabbroic Marcy facies restricted to the core. As previously noted, the actual distribution is not nearly so regular and examples of Whiteface facies are found throughout the massif and range in composition from gabbroic anorthosite to anorthosite. The important thing to retain from the nomenclature is that the Whiteface type is a clearly intrusive rock, and its more gabbroic varieties represent a good starting material for other rock types in the anorthosite massif. As discussed previously, Whiteface anorthositic gabbro was probably emplaced as a plagioclase-rich crystal mush that had evolved from gabbroic precursors fractionated at the crust-mantle boundary.

STOP A-8: EXPOSURES IN THE EAST BRANCH OF THE AUSABLE RIVER BY THE COVERED BRIDGE IN JAY

Description modified from Bohlen, McLelland, Valley, and Chiarenzelli (1992).

Upon parking in southernmost area on the southeast side of the river, walk a few tens of feet farther south along the paved road. At the sharp bend in road there is a low, polished outcrop that exposes two exceptional examples of rafts of coarse, blue-gray andesine anorthosite of the Marcy facies. The more southerly of these is enveloped in medium-grained gabbro similar to the xenolith-bearing, gabbroic facies on Giant Mountain. The gabbro is, in turn, surrounded by a fine-grained anorthosite. The northern raft of Marcy anorthosite lacks the rim of gabbro and is directly surrounded by the white, fine-grained anorthosite. The northern raft also contains a giant pyroxene whose edges show subophitic relationships with plagioclase.

Within the river there occurs a wide area of water-swept exposures of white, fine-grained, and highly layered anorthosite containing a few remnant blue-gray andesines as well as a few blocks of subophitic gabbro. The exposures are disrupted by two types of dikes: 1) late unmetamorphosed (Phanerozoic?) diabase paralleling the river, and 2) irregular, pyroxene-rich dikes and veins that trend mainly N-S and E-W but show other orientations and right angle turns as well. Some of the dikes are 15-20 cm wide but most fall into the 2-5 cm range. Both sharp and gradational contacts exist. Several of the dikes intrude along zones of mafic mylonite that may be of the same composition as the dikes themselves.

Mineralogically the dikes consist of coarse, emerald-green clinopyroxene and Fe, Ti-oxide. Some dikes also contain small quantities of plagioclase, garnet, and apatite. The apatite-free dikes contain very Mg-rich clinopyroxene (X_{Mg} ~80) while the apatite-bearing ones are less magnesium (X_{Mg} ~65). The key to understanding these dikes is to note that some exhibit comb structure defined by pyroxene and plagioclase crystals growing perpendicular to dike walls. Thus texture is diagnostic of growth from a fluid and provides compelling evidence that the dikes were intruded as liquid-rich magma. Preliminary experimental work by D. Lindsley (personal communication) indicates that representative dike material reaches its liquidus at 1200 °C (max).

It is suggested that these pyroxene-rich, and sometimes apatite-bearing, dikes represent immiscible silicate fractions complementary to magnetite-ilmenite deposits. Emplacement may have occurred when jostling about of essentially congealed plagioclase cumulates resulted in fracturing and the development of large blocks whose shifting provided passageways along which pyroxene-oxide magmas could be filter pressed. Depending on the batch of magma tapped, the intruding material would vary from Mg-rich to Fe-rich with apatite. Some dikes may have been cumulate-rich.

Dikes of the sort exposed in the river at Jay are common throughout the Marcy anorthosite massif, and their presence indicates a substantial, but small, amount of mafic material - and possibly mafic cumulate - in the massif. Invariably these dikes consist of green clinopyroxene and Fe, Ti-oxide, and it is common to find the silicate and oxide phases physically separated within the same vein. It is possible that, where this occurs, it reflects liquid immiscibility operating on late-stage interstitial fractions which are filter-pressed into veins.

STOP A-9: WILLSBORO WOLLASTONITE MINE

Description modified from Bohlen, McLelland, Valley, and Chiarenzelli (1992).

The Willsboro mine (now abandoned) is situated near the eastern end of a belt of metasedimentary rocks that can be traced west and then southwest for almost 15 km, close to or at the contact of the Westport anorthosite dome, to the Lewis mine at its western end.

The lowermost rock unit at the Willsboro mine is anorthositic gneiss of the Westport Dome. It contains over 90% plagioclase (An_{50}), with large (>5 cm) dark bluish plagioclase megacrysts in a lighter bluish-gray matrix. At the mine, the anorthosite is overlain locally by a mafic gneiss a few tens of meters thick; this unit is absent further west. Above the mafic gneiss is the wollastonite ore, a coarse-grained (1-5 cm) rock. In spite of tight isoclinal folding, the foliation is subdued and is defined by oriented wollastonite grains and layers and streaks of dark minerals.

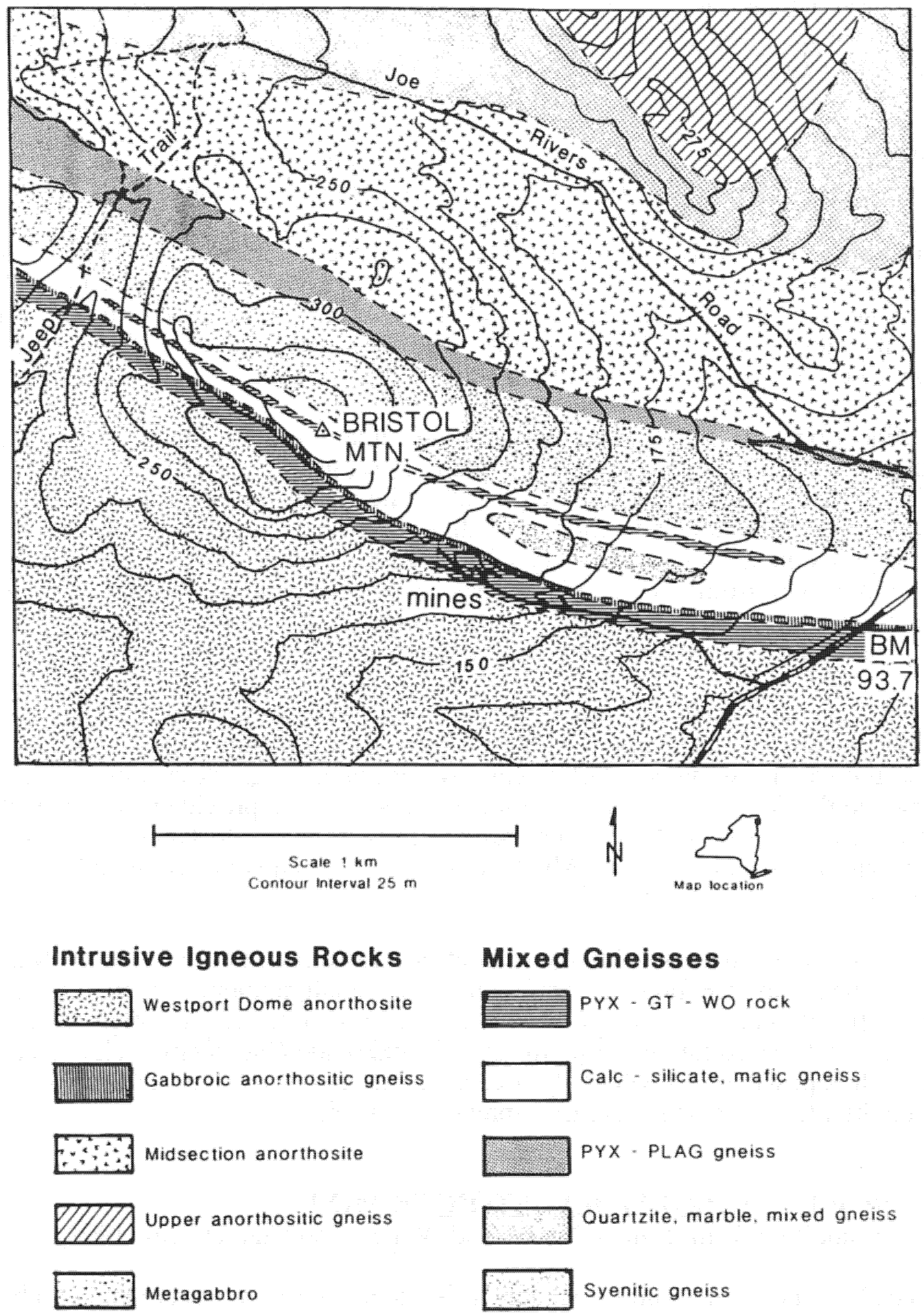
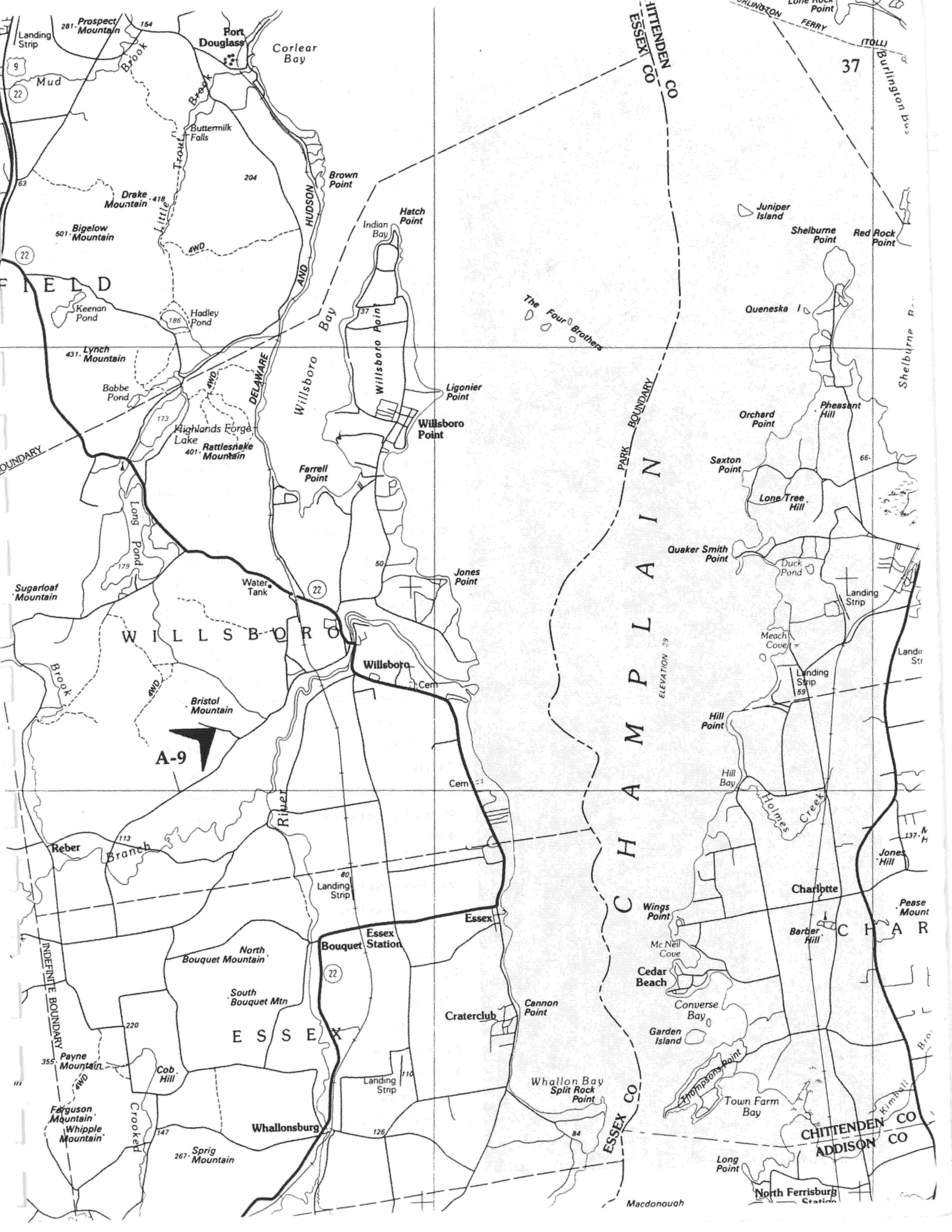


Figure A-5. Geologic map of the vicinity of the Willsboro wollastonite mine. Modified from Olmsted and Ollila (1988).


Landing Strip
281 Prospect Mountain
154
Fort Douglass
Corlear Bay
Mud
Brook
Buttermilk Falls
Trout
Little
204
Drake Mountain 418
Bigelow Mountain 501
4WD
Brown Point
HUDSON
AND
DELAWARE
Hatch Point
Indian Bay
ESSEX CO
CHITTENDEN CO
BURLINGTON FERRY
(TOLL)
Burlington Bay
Juniper Island
Shelburne Point
Red Rock Point
Lone Rock Point
FIELD
Keenan Pond
Hadley Pond
186
Willsboro Bay
37
Willsboro Point
The Four Brothers
Queneska I
Shelburne
Lynch Mountain 431
Babbe Pond
173
Highlands Forge Lake
Rattlesnake Mountain 401
BOUNDARY
Ligonier Point
Willsboro Point
PARK BOUNDARY
Orchard Point
Pheasant Hill
Farrell Point
Saxton Point
66
Lone Tree Hill
Long Pond
179
Sugarloaf Mountain
Water Tank
22
50
Jones Point
Quaker Smith Point
Duck Pond
Landing Strip
WILLSBORO
Willsboro
Cem
Meach Cove
ELEVATION 95
LAKE CHAMPLAIN
Landing Strip
59
Brook
Bristol Mountain
A-9
Hill Point
Hill Bay
Holmes Creek
Cem
River
Reber
113
Branch
137
Jones Hill
80
Landing Strip
Charlotte
Essex
Essex Station
Bouquet
Wings Point
Pease Mount
Barber Hill
CHAR
North Bouquet Mountain
Mc Neil Cove
Cedar Beach
South Bouquet Mtn
Converse Bay
Craterclub
Cannon Point
220
ESSEX
Garden Island
INDEFINITE BOUNDARY
355 Payne Mountain
Cob Hill
Thompsons Point
110
Landing Strip
Whallon Bay
Split Rock Point
Town Farm Bay
Kimball
Ferguson Mountain
Whipple Mountain
Crooked
147
Whallonsburg
126
84
ESSEX CO
CHITTENDEN CO
ADDISON CO
267 Sprig Mountain
Long Point
North Ferrisburg Station
Macdonough

Celebrating over 40 years as the foremost producer and marketer of cost performance wollastonite and chemically modified wollastonite products. NYCO markets these products globally to producers of plastics, elastomers, automotive polymer composites, electrical components, coatings, construction, friction and a variety of other industries.

To meet such widely different requirements, NYCO produces a range of products derived from the high quality ores mined in the Adirondack Region of New York State. NYCO's mineral resources are sufficient to meet future market requirements for decades to come – sustaining the company's reputation for quality, reliable delivery and dependable service.

NYCO's plants in Willsboro, Essex County, New York are modern facilities. Plant #1 was completely modernized and its capacity increased to 120,000 tons per annum to meet growing worldwide demand.

Plant #2 applies chemistry to the surface of wollastonite and other minerals to produce second generation performance materials. Plant #2 was built in 1989 and is a state-of-the-art facility.

Integrated into these two plants are a modern pilot plant, quality assurance laboratories, programmable logic controls and the practice of Statistical Process Control.

NYCO Minerals, Inc. is a proud subsidiary of Fording Coal Holdings Inc. Company, a wholly owned subsidiary of Canadian Pacific Ltd. Through this ownership, NYCO has access to strong financial backing and enjoys a wealth of industrial expertise and management skills.

NYCO's growth has been through product innovation. Product development is a partnering between NYCO's personnel and our customers. Our basic philosophy is to exchange information to identify the best product, the best service, the best package. From partnering, we have learned and will continue to learn about our products and services and how to improve them. We will only continue to be successful if our customers are successful. We are involved and committed to help our customers make better, more economical products.

Please call, fax or write us for additional technical information and sample(s) on our products. Our technical staff is on hand to consult with you on your particular product/application requirements.

High Aspect Ratio

GRADES

NYAD G®
NYAD G® SPECIAL
NYAD G® MAR
G-40 WOLLASTOKUP®
G-40 WOLLASTOCOAT®
NYAD® FP
NYCOR® R
G-RRIM™ WOLLASTOKUP®
G-RRIM™ WOLLASTOCOAT®

The high aspect ratio grades are produced by special attrition milling and air classifying. These products are produced to maintain the integrity of their particle size and aspect ratios. They typically have an aspect ratio ranging from 20:1 to 10:1. These grades are used in the following applications:

PLASTICS/ELASTOMERS

- molded plastic castings
- air conditioner covers
- car distributor caps
- pot handles
- commutators
- automotive coil tops
- toaster end plates
- automotive fascias & bumpers
- automotive door panels & fenders
- automotive grills
- automotive side claddings
- rocker panels
- air dams
- adhesives
- sealants
- hot water heater temperature controls
- circuit boards
- switch knobs
- O rings
- electric terminal blocks
- square D circuit breakers
- electrical plugs
- electrical connectors
- electrical switch boxes
- fan shroud assemblies

COATINGS

- roof coatings
- block fillers
- floor patching
- latex ceiling paints
- asphaltic emulsion coatings
- textured & stucco coatings
- swimming pool coatings

CONSTRUCTION

- structural tile bodies
- textured glazes
- calcium silicate board
- fire resistant board
- ceiling tiles
- roof slates
- corrugated roofs
- joint compounds

FRICTION

- brake pads and linings
- clutch facings
- clutch plates
- appliance friction disks
- brake blocks
- wear strips

MISCELLANEOUS

- welding rod fluxes
- pultrusion bars
- refractories
- molten metal transports

Milled

GRADES

NYAD® 200
NYAD® 325
NYAD® 400
325 WOLLASTOKUP®
325 WOLLASTOCOAT®
400 WOLLASTOKUP®
400 WOLLASTOCOAT®

The milled grades are produced on ball or pebble mills and ground to mesh size specifications. These grades consist of a 200 mesh, 325 mesh and a 400 mesh. Their aspect ratios typically range from 7:1 to 3:1. The milled grades find broad use in the following applications:

PLASTICS/ELASTOMERS

- elastomers
- rubber reinforced traffic markers
- brake pistons
- potting compounds
- O rings
- gaskets
- fluoroelastomer seals
- microwave cookware
- walking grids
- manhole grates
- cog belt pulleys
- commutator pulleys
- countertops

COATINGS

- oil & water based paints
- block fillers
- ceiling tile paints
- powder coatings
- anti-skid coatings
- architectural coatings

CERAMICS

- bodies
- glazes
- frits
- enamels
- colours
- ceramic tiles
- artware
- dinnerware
- fluxes
- sanitaryware
- low-loss electrical insulating materials

CONSTRUCTION

- wall board
- ceiling tiles

FRICTION

- extruder grinding wheel linings

MISCELLANEOUS

- metallurgical
- continuous casting powders
- mild abrasives
- adhesives
- bowling ball cores

Fine Particle Size

GRADES

NYAD® 475
10 WOLLASTOKUP®
NYAD® 1250
10 WOLLASTOCOAT®

The fine particle size grades are produced by milling and air classifying to a top and medium size specification. These grades are available as a 475 mesh and a 1250 mesh product. The aspect ratios are typically 3:1. They are used in the following applications:

PLASTICS

- paintable & plateable wheel covers
- automotive hubcaps
- automotive door handles
- refrigerator ice water handles
- vehicle window cranks
- shower handles
- car door latches
- wheelchair hubcaps
- chain saw housings
- lamp housings
- mirror housings

COATINGS

- industrial and marine coatings
- latex & oil primers
- powder coatings
- pipe coatings
- coil coatings
- intumescent coatings

MISCELLANEOUS

- dental impressions
- sealers

nyco MINERALS, INC.

NYCO® Minerals, Inc.
Willsboro, New York, USA 12996-0368
Telephone: 518-963-4262 Telefax: 518-963-4187 Telex: 957014
NYCO® Minerals, Inc. Europe
Ordrupvej 24, P.O. Box 88
DK-2920 Charlottenlund, Denmark
Telephone: 31643370 Telefax: 31643710

IN-201-92-1

The ore consists of only three minerals: wollastonite, grandite garnet (Grs_{10-90}), and Di-Hd clinopyroxene. Thick, skarn-like masses of grossular and pyroxene are found along the edges or occasionally within the unit. The mineralogy and geologic setting favor an origin by contact metamorphism of siliceous carbonate rocks at the time of anorthosite intrusion. The fact that calcite and quartz are ordinarily absent in this high-variance assemblage suggests that metasomatism played a major role in the ore forming process (Buddington, 1939; DeRudder, 1962). Prior to 1982 it was widely accepted that the anorthosite intruded at depths of 20-25 km during Grenville regional metamorphism.

Above the wollastonite are thin units of anorthositic, granitic, mafic and calcsilicate gneisses, and amphibolites. In some of these exposures we find masses of nearly pure garnet rock suggesting repetition of the ore unit. Observe that wollastonite weathers rapidly and is rarely seen at the surface, locally leaving garnetite or diopsidite "skarns" as the sole evidence of its presence.

The uppermost unit is a thick, sill-like body of gabbroic anorthosite gneiss . This extends the entire length of the metasedimentary belt and differs from the anorthosite of the Westport Dome both in composition and texture. The contacts of this unit with metasedimentary rocks are not well exposed here, but elsewhere they are marked by pyroxene-rich skarns. As we return to the mine we will examine a metagabbro unit and its contacts with enclosing metasedimentary rocks.

Valley and O'Neil (1982, 1984) and Valley (1985) have reported anomalously low $\delta^{18}O$ (-1.3 to 3.1; up to 20 permil lower than typical Adirondack marbles) in the wollastonite ore within 125 m of the anorthosite contacts, as well as extremely sharp $\delta^{18}O$ gradients between the wollastonite and the surrounding rocks. The low $\delta^{18}O$ values cannot be explained solely by devolatilization reactions (Valley, 1986) and result from deep circulation of heated meteoric waters along fractures at the time of anorthosite intrusion. Because such fluids would be at hydrostatic pressure, they should not penetrate a ductile metamorphic environment where fluids are at lithostatic pressure. This suggests that skarn formation occurred at shallow depths (<10 km) relative to granulite facies metamorphism. These stable isotope data were the first quantitative evidence for shallow intrusion of anorthosite within the Grenville Province. The origin of the sharp gradients is enigmatic; they may represent pre-granulite facies faults or shears, but their preservation through granulite facies metamorphism indicates that there was no significant fluid movement across strike during the regional metamorphism.

<u>STOP A-10: WOOLEN MILL GABBRO</u>

Description modified from Bohlen, McLelland, Valley, and Chiarenzelli (1992).

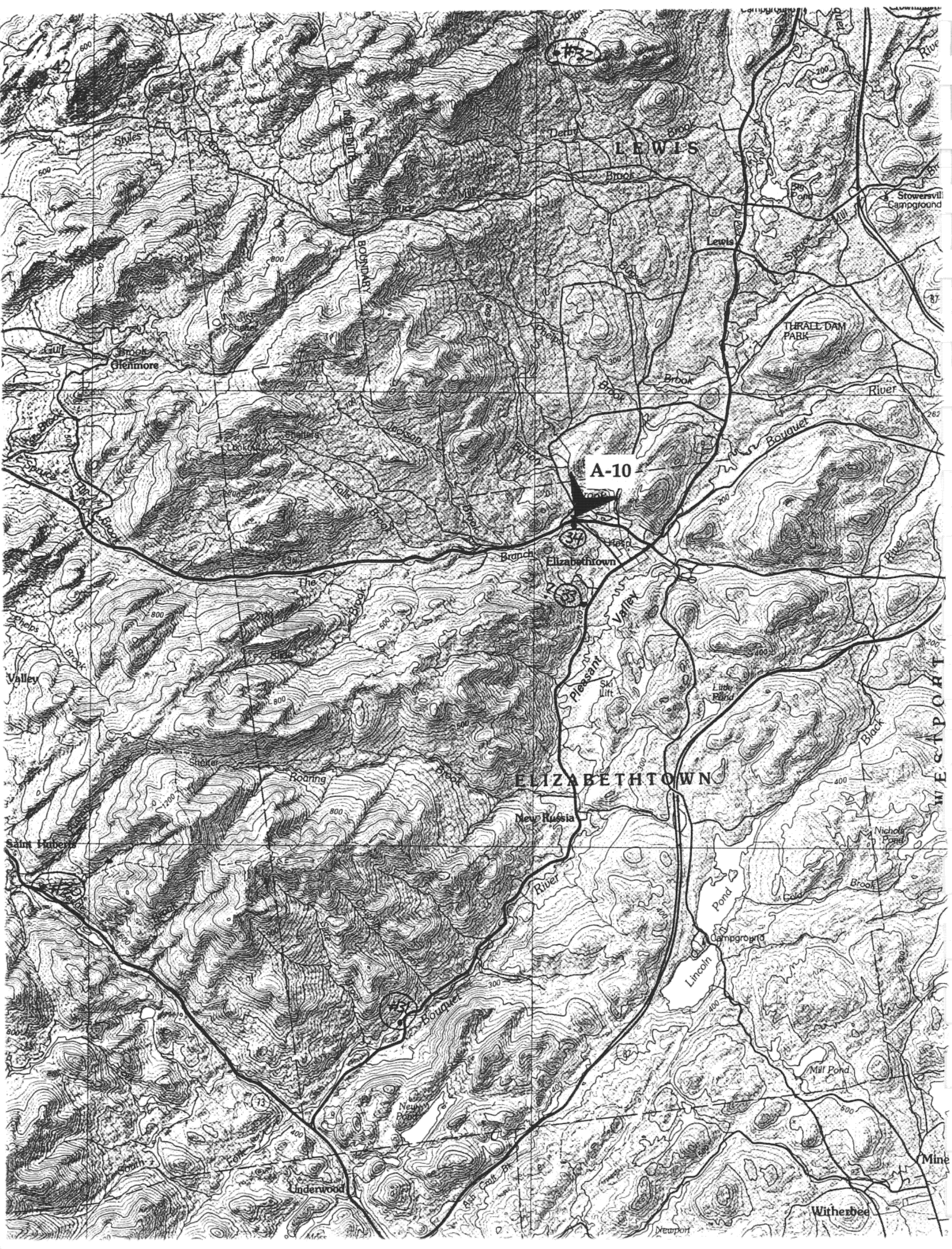
Park on the right side of the road opposite high roadcut on left. The cut shows metanorthosite intruded by a dark, fine-grained rock, first described by Kemp and Ruedemann (1910) as the "Woolen Mill Gabbro". It is a clinopyroxene-garnet-oligoclase granulite with considerable opaque oxides and apatite, and minor K feldspar and quartz. It contains a few large, uncrushed andesine xenocrysts, probably derived from the host anorthosite. The texture is that of a granulite, but the xenocrysts have apparently escaped recrystallization or grain-size reduction, even along their margins. This peculiar situation may be explained by static recrystallization of an initially fine-grained intrusive rock. The composition of rock is that of a somewhat K_2O rich (1.20 wt%) ferrogabbro of the type common in the Adirondack Highlands, especially near magnetite-ilmenite concentrations. It also is found associated with anorthosite at Split Rock Falls and near Elizabethtown, and is commonly present as disrupting material in block structure. Woolen Mill gabbro may represent gabbroic anorthosite magma enriched in mafic components by separation of cumulus plagioclase as suggested by mixing calculations (Ashwal 1978).
This is the type locality for deWaard's (1965) clinopyroxene-almandine subfacies of the granulite facies.

Cross the road and examine the outcrops in the stream bed. At the west end of the stream exposures, Woolen Mill gabbro clearly crosscuts anorthosite, and veins and dikes of the gabbro extend into the anorthosite. Within the anorthosite there is well-developed "block structure" where several types of anorthosite have undergone brittle fracture before being intruded by thin dikes or veins of mafic as well as felsic material. Some of these veins are identical to the mafic granulite in the roadcut (and at the west end of the stream exposure) and are part of the anorthosite suite. Some of the disrupting material is anorthositic gabbro more commonly associated with the anorthosite as on Giant Mountain or Lake Clear. In addition, a variable amount of granitic material is present in many of the veins as revealed by straining. The relationships here suggest formation of a plagioclase-rich cumulate, which was then fractured and intruded by a later mafic differentiate. This apparently brittle behavior supports a relatively shallow depth of intrusion. Notice also the very large (up to 10 cm) giant orthopyroxenes that occur in the anorthosite, especially near the contact with Woolen Mill gabbro.

The anorthosite in the stream bed contains the characteristic post-metamorphic alteration assemblages of calcite $\pm$ chlorite $\pm$ sericite that are commonly seen as late-stage, hairline vein fillings or as alteration products of Fe-Mg silicates throughout the Adirondacks (Buddington,1939; Morrison and Valley, 1988b). Average values of $\delta^{18}O$ and $\delta^{18}C$ for calcite are +12.6 and -2.2 permil, respectively, which suggests that the alteration fluids were deep seated in origin and exchanged with igneous as well as metasedimentary rocks. These veins are related to the formation of at least some high-density, CO_2-rich fluid inclusions and the temperatures of alteration are estimated at 300 °-500 °C (Morrison and Valley, 1991, 1988b).

The retrograde fluids that have infiltrated the anorthosite to precipitate calcite have not significantly altered its oxygen isotopic composition. Values of $\Delta_{(calcite\text{-}plagioclase)}$ range from 0.96 to 6.6, indicating that the isotopic composition of the alteration minerals was controlled primarily by the hydrothermal fluid and that the $\delta^{18}O$ of the host rock remained largely unchanged due to low fluid/rock ratios.

Values of $\delta^{18}O$ (plag) for the "blocks" and their host anorthosite at this outcrop range from +8.5 to +9.3. In general, the metanorthosites in the NE part of the massif are somewhat more isotopically heterogeneous than those in the northwestern part of the massif, but they show the same roughly 2.5 permil enrichment in $\delta^{18}O$ relative to "normal" anorthosites worldwide (Morrison and Valley, 1988a).



A-10
LEWIS
ELIZABETHTOWN
WESTPORT
Lewis
Elizabethtown
New Russia
Glenmore
Saint Huberts
Underwood
Witherbee
Stowersville Campground
THRALL DAM PARK
Lincoln Pond
Campground
Little Pond
Mill Pond
Nichols Pond
Big Pond
Ski Lift
Pleasant Valley
Bouquet River
Black River
Spruce Mill Brook
Jackson Brook
Roaring Brook
Phelps Brook
Derby Brook
Shelter
Lookout
Valley
BOUNDARY
Newport

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GEOLOGY OF THE CONNECTICUT RIVER VALLEY
in the vicinity of Amherst and Smith Colleges

E.S.Belt, J.T.Cheney, T.A.Harms
Department of Geology
Amherst College
Amherst, MA 01002

Introduction

The geology of the Connecticut Valley involves formations from three distinct periods of geologic time. Each is represented by degree of lithification and metamorphism. (1) Precambrian and Early and Middle Paleozoic rocks are highly metamorphosed and intruded by igneous rock. (2) Triassic and Jurassic rocks are weakly to strongly lithified sedimentary rocks, although some basalt dikes and lava flows are also present. (3) Late Pleistocene (Wisconsinian) and Holocene aged strata are unconsolidated.

The generalized bedrock geology of the Connecticut Valley is illustrated in Figure 1, which is based on mapping by Jahns (1951) and Willard (1951) amd recently up-dated by Zen (1983). In the center of Figure 1, Mesozoic sedimentary and volcanic rocks are found beneath a cover of Pleistocene and Holocene unconsolidated sediments. To the east of the Mesozoic lowlands lie the Precambrian-cored Pelham Dome, south of which lies the Belchertown Pluton (Devonian age) and east of which lies the Monson gneiss (Ordovician age). To the west of the Mesozoic lowlands lie metamorphosed pelites and quartzites generally of Devonian age, although cores of anticlines show Ordovician rocks of higher grades of metamorphism. The Mesozoic, Pleistocene and Recent geology form the focus of this field trip.

The Mesozoic events recorded in the area begin in the Late Triassic when a half-graben formed. The major controlling fault lay on the east side of the basin, and was active until early Jurassic time. Similar grabens formed a string of basins that extended from Maritime Canada and the Bay of Fundy, through central southern New England, New Jersey, southeastern Pennsylvania, central Maryland, Virginia, North Carolina, South Carolina, and eastern Georgia. All of these listric-fault bounded basins resulted from extensional tectonics that accompanied the opening of the Atlantic Ocean basin. Similar basins of this age have been discovered beneath the Continental Shelf off Nova Scotia and New England as well as in northwestern Africa (for a recent summary, see Lorenz, 1988). We shall examine the facies and discuss the sedimentary events of the Mesozoic basin in our area.

Conglomerates of Late Triassic to Early Jurassic age are found adjacent to the Border Fault (Figure 1). The pebbles in these conglomerates can be correlated to units in the present-day Pelham Hills even though they were eroded from mountains in that area 200 million years ago. Border conglomerates pass westward into pebbly sandstones and shales; this lateral facies change from margin to center of the depositional basin can be found within all the sedimentary sequences in the depositional basin. Most of the Triassic/Jurassic strata resulted from alluvial fan and fluvial depositional processes, although some resulted from lacustrine sedimentation. The fluvial sediment became oxidized to a characteristic red color after burial. Periodically, large lakes occupied the valley, and left behind organic-rich dark gray shales that are interbedded with the red river and fan deposits. The dark-gray lake deposits also interfinger laterally with the red alluvial deposits. Dinosaur footprints are best preserved in the lake-margin deposits because the sediment

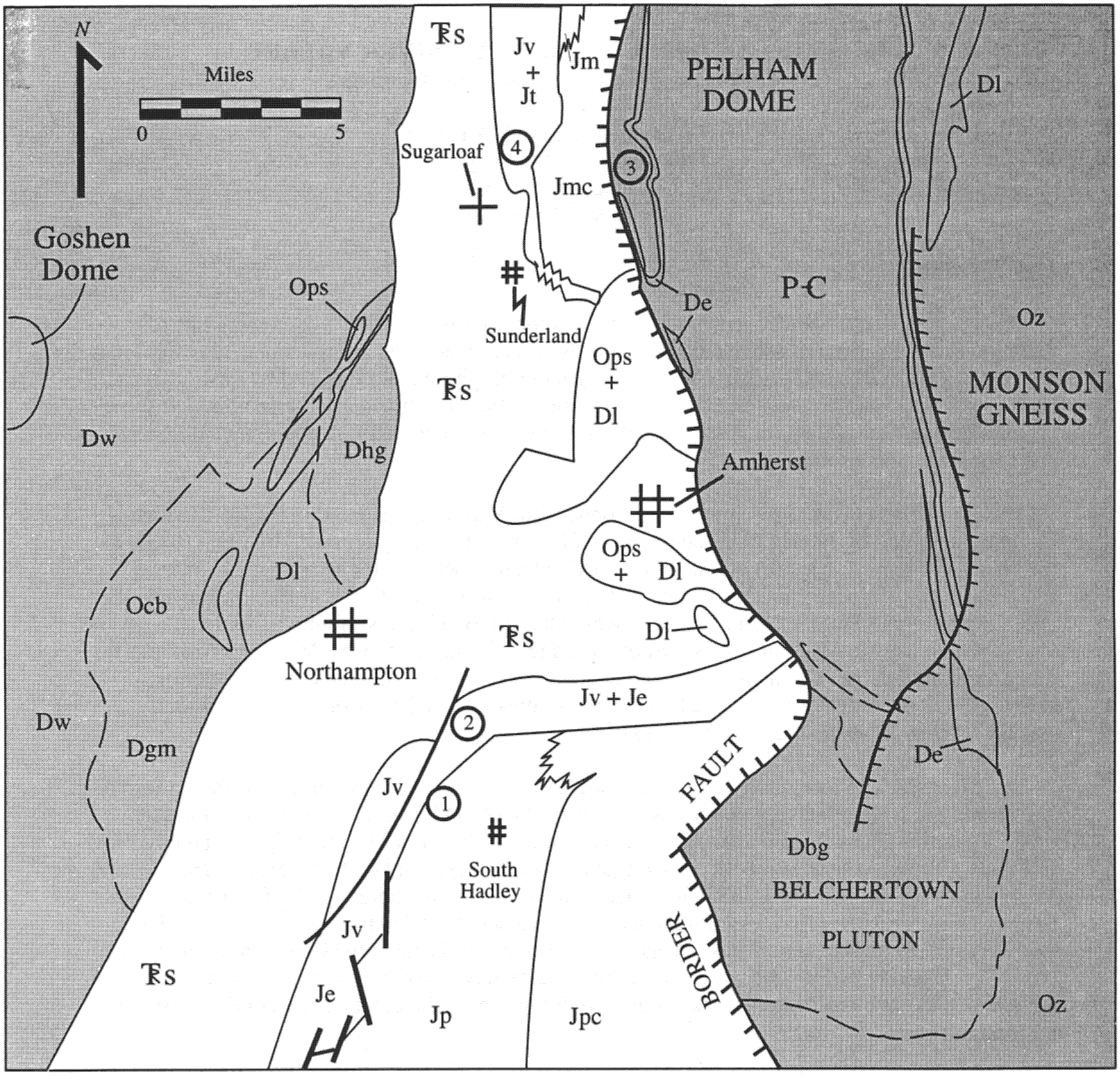

Figure 1. Bedrock geologic map in the vicinity of Amherst and Smith Colleges. Areas screen gray are Precambrian and Paleozoic metamorphic rocks that form basement for the Mesozoic cycle of sediments. The symbols in that area include: **Ops**, Partridge Fm.; **Oz**, Monson gneiss; **Dl**, Littleton Fm; **Dw**, Waits River Fm; **Dbs**, Belchertown complex. In the area of Mesozoic rock: **TR-s** is Sugarloaf Arkose; **TR-n** is New Haven Arkose; **J-p** is Portland Formation; **J-mc** and **J-pc** are conglomerate facies of the Mt. Toby Fm. and Portland Fm., respectively. **J-v** are volcanics belonging to the Holyoke, Deerfield, Hampden basalts as well as the Hitchcock volcanics and the Granby tuff. **J-e** (East Berlin Fm.) and **J-t** (Turner's Falls Fm.) are but two of the sedimentary rock units interbedded with these volcanics. Stop numbers are shown in circles.

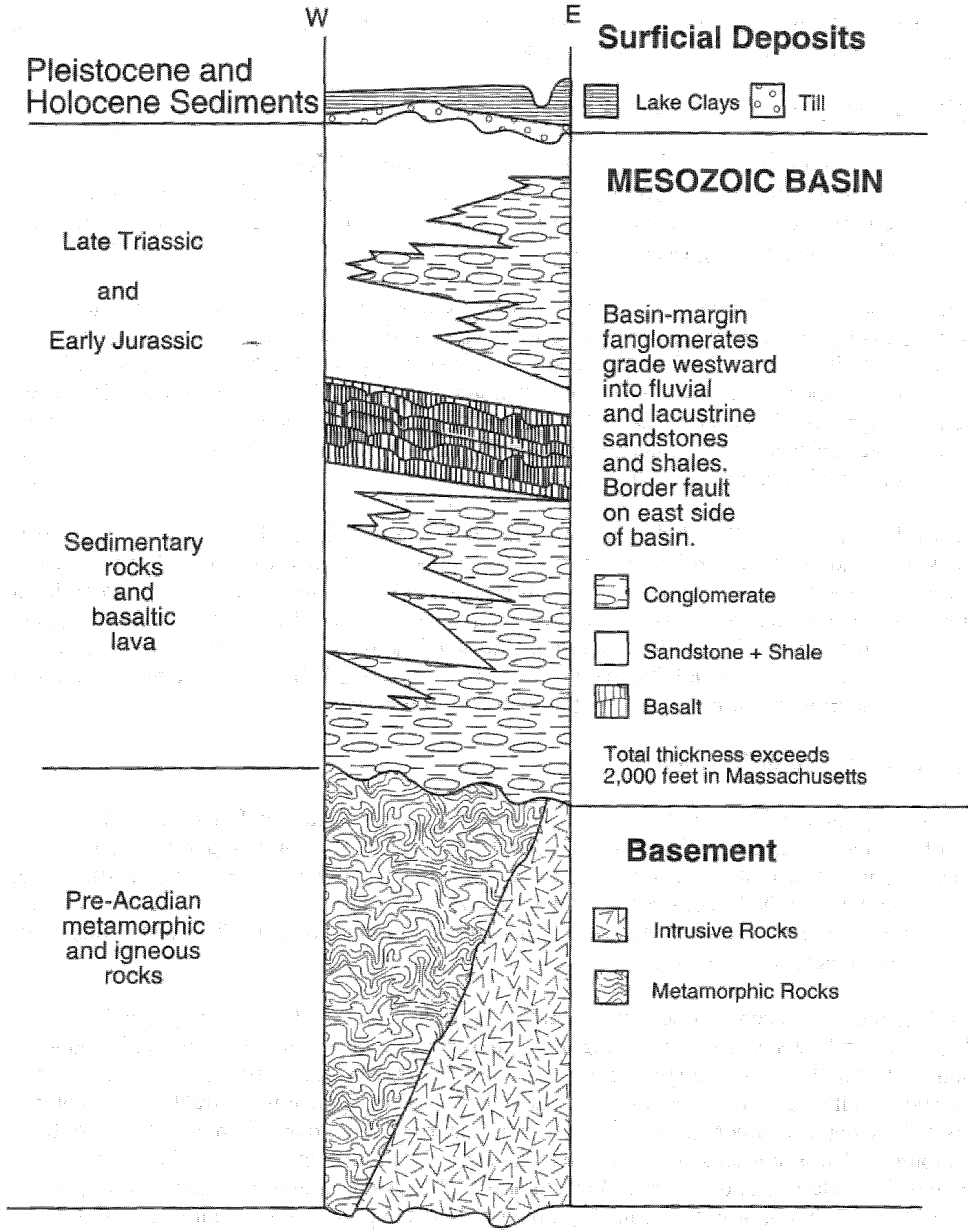

Figure 2. Generalized geologic section of rocks and sediments in the area shown in Figure 1. Total thickness of Mesozoic strata is highly variable, being very thin, less than 100 ft. preserved above basement in the region of Amherst; it is thicker, in excess of 2000 ft., just south of the Holyoke Range. In the Hartford area, nearly 20,000 ft. of Mesozoic sediment has been extrapolated from surface outcrops.

there consisted of fine sand, silt and clay and the footprints could become quickly covered by fresh sediments; preservation in some cases is excellent.

STOP B-1: Mt. Tom (Smith's Ferry)

Dinosaur footprint locality is along U.S. Route 5, 1 mile north of the city of Holyoke, Mass., and about 1 mile south of the Old Smith's ferry; a state sign on the east side of Route 5 indicates the parking area for the site. The strata bearing the footprints crop out between the highway and the railroad tracks below, dipping toward the river.

The Lower Jurassic Longmeadow Sandstone exposed here consists of fine-grained sandstone, siltstone, and claystone, mainly brown and gray. The footprints, all badly weathered, numbered about 134 in 1970. Ostrom (1972) described three different types of tracks; all are three-toed tracks made by bipedal dinosaurs; most were walking westward. The sediments were deposited along the margin of a flood-plain lake (Hubert and others, 1976). Fossils in the underlying dark-gray lacustrine shales include fish and bivalved crustacea (McDonald, 1982; that article describes other dinosaur-track localities in the Connecticut Valley).

Edward Hitchcock (1793-1864, third president of Amherst College, 1845-1854), was the first to recognize dinosaur tracks in North America. His collection of trackways from 35 miles along the Connecticut River is housed in the Pratt Museum at Amherst College. It is the largest collection of dinosaur tracks in the world. Hitchcock was first shown such tracks by a farmer in 1835, and he compared them with plaster-of-paris casts he made of the foot marks of living birds. From these comparisons he concluded that the dinosaur tracks were made by very large birds. Not a bad decision considering modern thinking on the subject of dinosaurs!

STOP B-2: Summit of Mt. Holyoke

From this point can be seen: to the east the Pelham Hills, underlain by Paleozoic and Precambrian metamorphic and igneous rocks; to the north Mesozoic strata found beneath the Connecticut Valley with its river (the stronger basaltic and conglomeratic rock forming the higher hills); and to the west the Berkshire Hills, underlain by Paleozoic and in the distance Precambrian metamorphic and igneous rocks (see Fig. 1). The purpose of this stop is to discuss the Mesozoic and more recent geology of the area.

Pre-Pleistocene erosion produced the overall relief of the area; the Pleistocene glaciation modified the landscape slightly, rounding the topography and depositing drumlins and ground moraine. During the waning phases of the ice, about 15,000 b.p., Lake Hitchcock formed in the Connecticut Valley (Figure 4), behind a morainal dam that had formed in central Connecticut, near Rocky Hill. Coarse sediment deposited by streams entering the lake and formed deltas; the three deltas found in Massachusetts are of considerable size. Counting varves in the lake-bottom sediments from Hartford northwards, shows that the lake lasted for approximately 4,000 years (Ashley, 1972); when it drained, about 10,650 years b.p., the present-day stream network began to develop.

When the lake drained and the Connecticut River flowed again, the huge Montague delta blocked it from flowing south through the town of Miller's Falls (about 25 miles north of stop #2). Instead the river turned west, it carved a deep gorge through the Mesozoic sediments and lava near the present town of Turners Falls, producing the best natural exposure of the Mesozoic in the entire

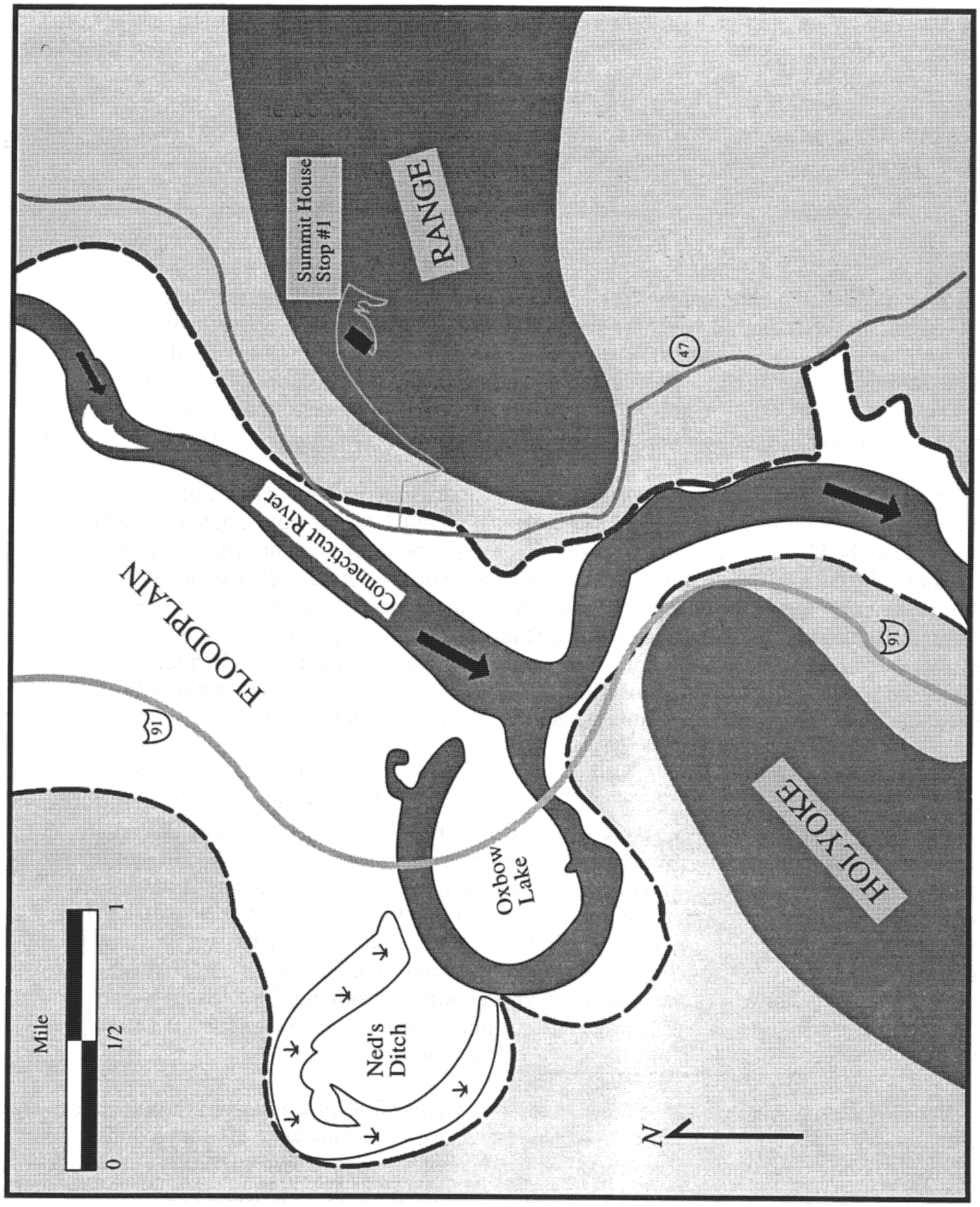

Figure 3. Map showing Connecticut River channel and flood plain relationships to the Holyoke Range. Ned's Ditch is the earliest oxbow still visible in the flood plain. Oxbow Lake was produced by the 1840 cut-off of a large gooseneck bend in the river. Illustrations of this bend prior to the cut-off can be studied (Belt, 1989; also Thomas Cole's 1836 painting of the Oxbow). The gooseneck was called an oxbow even before the cut-off. After the cutoff, this feature becomes the type Oxbow Lake.

Connecticut Valley basin. The river then flowed southward and on through the water gap in the Holyoke Range visible below just southwest of this stop.

North of the water gap, the river constructed a wide floodplain, but to the south, it is confined in its own gorge. Thus the water gap produced a local baselevel that was of pre-glacial origin. The river may have been superposed from an old erosion surface or "peneplain". Hovever, previously existing faults, formed by the folding of the Mesozoic strata where the massive lava flow "hinges" and swings abruptly from west to southwest, may have assisted the river in cutting the water gap. As a consequence of the local base level upriver from the water gap, the river has a gentler gradient north than south of the mountain; upriver the Connecticut River was freer to wander laterally. And wander it did, producing many generations of meander bends and cut-offs of which two are shown in Figure 3. The present Oxbow resulted from the cut off (1840 flood) of the gooseneck meander bend of the river; Hitchcock and his wife watched the gooseneck breached by the floodwaters. Ora Hitchcock illustrated this event as she did so many other matters of geologic importance in her husband's *Final Report on the Geology of Massachusetts* (Hitchcock, 1841).

The basalt around the Prospect House (also known as the Holyoke House or the Summit House) shows columnar jointing and locally some amygdaloidal textures, both of which were amply described by Edward Hitchcock (1833, 1841). Baked sedimentary rock occur beneath but not above the basalt. Hence the rock was of extrusive origin. Hitchcock also noted the grooves and scratches on top of the basalt. We now ascribe these grooves to Pleistocene glacial effects, although Hitchcock (1833) argued that they resulted from icebergs that floated in a huge lake that he postulated had existed throughout New England. Erratic boulders found on top of the Holyoke Range, he suggested, were also brought there by icebergs. We know they resulted from Pleistocene glaciers - 150 m.y. younger than the strata of the Holyoke Range.

Historians have argued that Hitchcock, a devout Congregationalist as well as geologist, let religion influence his geologic reasoning. They argue that Hitchcock believed that all the effects we now know to be of glacial origin were simply the result of Noah's Flood. It can be pointed out however, that Hitchcock had a reasonable geological point, given what was known at the time about glaciers and their effects. Agassiz had proposed in 1840 that a vast ice sheet covered all of New England at some unknown time in the past, but that the only modern evidence used by Agassiz came from the Alpine or valley glaciers of Switzerland. Hitchcock disagreed with Agassiz' hypothesis because it was not necessary to postulate a huge ice sheet, larger than any known Alpine glacier when lake deposits were well known from the Connecticut Valley (named Lake Hitchcock, Figure 4, by B.K. Emerson). Continental ice sheets had not been reported by Agassiz, nor were they known by Hitchcock at that time.

Beneath the river and glacial sediments, the Connecticut Valley was carved through relatively easily eroded sedimentary rock of Mesozoic age. These sediments were deposited during Late Triassic and Early Jurassic time within a half-graben whose main operating fault lay to the east, at the base of the Pelham Hills. This Border Fault runs from Vermont to Long Island Sound, and had a throw of at least 5 miles, and because it was active all during sedimentation, the younger beds tilt eastward more gently than the older beds. The fluvial, alluvial fan and lacustrine sediments that accumulated in the basin indicate a fluctuation of climate from semi-arid to sub-humid. The lakes that periodically formed have been correlated from just south of here to New Haven, CT. Some of these lakes were of sufficient depth to accommodate turbidity current deposits. Most lake deposits indicate playa and shallow water conditions, however. These have mud cracked units and in some cases micrite layers. Volcanic eruptions were not uncommon during Late Triassic to Early Jurassic time. Most of the lavas were extruded on land, although in one case, pillow structures indicating sub-aqueous extrusion, have been found. Note the basalt beneath the Prospect House.

STOP B-3: Border Fault of the Mesozoic Basin at Mt. Toby

This stop is 8.5 miles north of the center of Amherst Town, on Route 63 east of Sunderland. Outcrops of metamorphic rock are found along the east side of the road. To the west of the railroad track, Roaring Brook has formed a gorge and waterfall. The base of Mt. Toby is a fault scarp: the Mesozoic Basin Border Fault. Outcrops of Mt. Toby conglomerate and depositional breccia are found in the gorge. The maximum throw on the Border Fault has been estimated in Connecticut to be approximately 5 miles. The dip of the fault there also flattens at depth (D.U.Wise, pers. communic. 1986) towards the west, a feature common in the listric style of faults. In the Mt. Toby area, and south to the Holyoke Range, fault steps are found which bring the basement up close to the ground level (Chandler, 1979).

STOP B-4: Varved clays of Lake Hitchcock

This stop is in a ravine on the east flank of Pocumtuck Range (North Sugarloaf Mountain), Town of Deerfield. En route to the stop, which is 1.7 miles north of the west abutment of the Connecticut River bridge at the village of Sunderland, one may note the river terraces south of Sunderland and the Upper Triassic conglomeratic sandstone just west of the bridge at the foot of the South Sugarloaf Mountain.

Pleistocene Lake Hitchcock clays are exposed here in a narrow gorge cut by a stream flowing eastward from the Pocumtuck Range into the Connecticut River. Note the varves, and the distorted beds with pebbles in them. The one distorted bed here includes 12 varves. Its origin is obscure: it may be a result of floating ice that dropped the pebbles, but then why the distorted varves? The distorted varves may have resulted from earthquakes (similar distorted varves are ascribed to earthquakes in Quebec), but then why the pebbles? The proximity to the Pocumtuck "range" may offer a clue. These distorted beds likely resulted from subaqueous slumping, wherein nearshore pebbles and sand were entrained by the slump.

The varved deposits in this part of the Connecticut Valley are nearly 200 feet thick. The 4,000 varves in the lake deposit have been precisely correlated as far as central Connecticut (Schafer and Hartshorn, 1965). This is much like the science of dendrochronology: varve thickness is proportional to length of time of ice-free water each summer with the winter's accumulation simply from suspended clays, just as tree-ring thickness is proportional to length of summer growing season plus a tiny bit of growth during the winter. Indian legends relating to the lake imply a cataclysmic event had occurred wherein all the water drained from the area within a short period of time. One possible cataclysm would have been the topping of the morainal dam at Rock Hill, CT by an unusually high rainfall event. Ice floes may have clogged the natural spillway in bedrock. Similar ice jams caused extra flooding during the 1936 flood event, the worst flooding ever recorded on the Connecticut River.

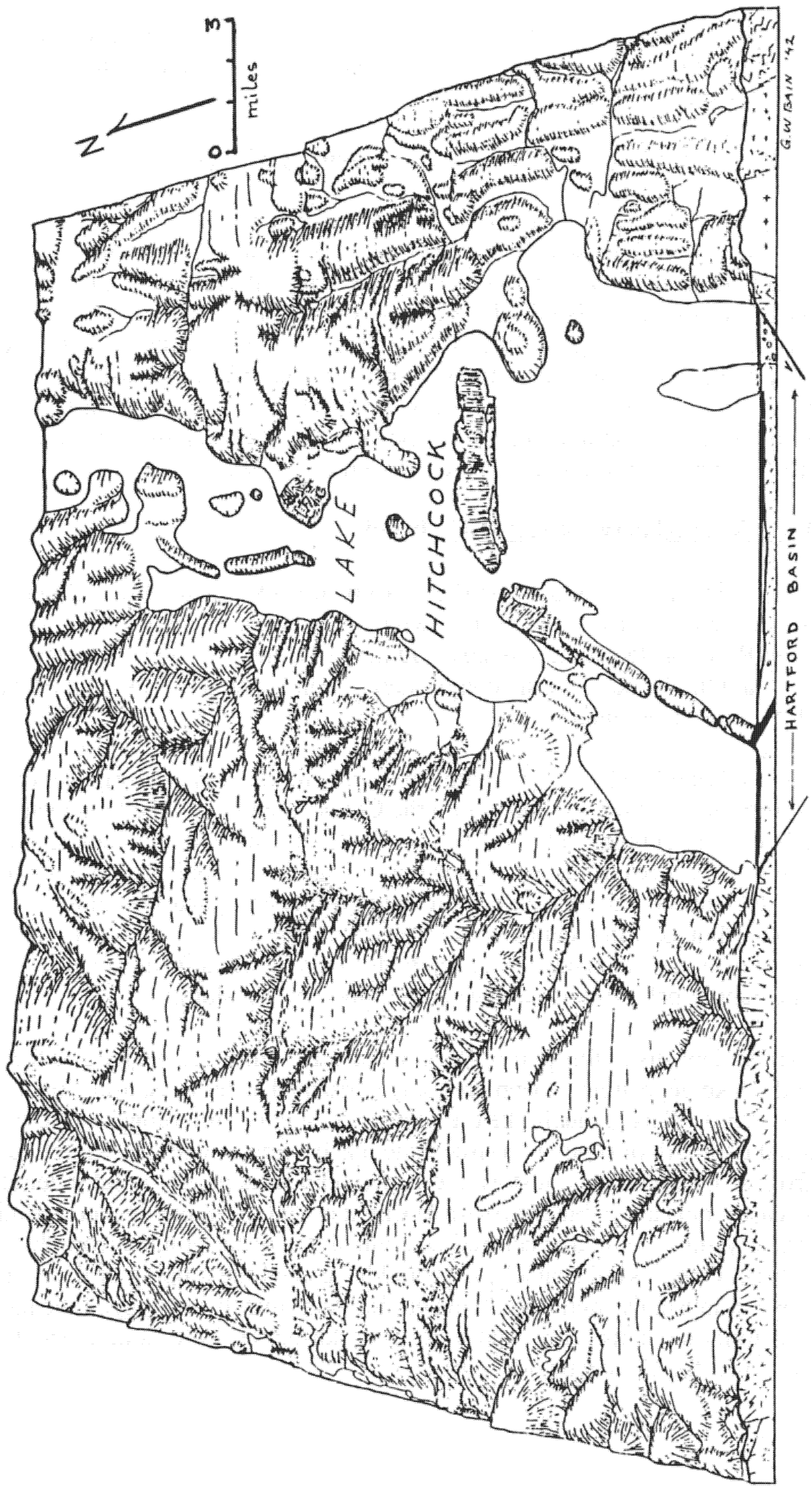

Figure 4. Reconstruction of Lake Hitchcock shortly before its catastropic draining from an illustration prepared by George W. Bain of Amherst College over 50 years ago.

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The Acadian Metamorphic High, South Central Massachusetts

Modified from Thomson, Peterson, Berry, and Barriero (1992).

Introduction

The Acadian metamorphic high of southern New England contains the most intense Acadian metamorphism in the Appalachian orogen. Regional metamorphic zones for central Massachusetts, based on assemblages within pelitic schists, are shown in Figure 1 (Tracy, 1975, 1978; Tracy *et al.*, 1976; Robinson *et al.*, 1978, 1982, 1986a). A broad area in the middle of the metamorphic high is in Zone VI, characterized by the assemblage sillimanite + K-feldspar + garnet + cordierite. Metamorphosed igneous rocks in Zone VI include assemblages of orthopyroxene + augite + plagioclase and orthopyroxene + K-feldspar, indicative of the lower granulite facies (Hollocher, 1985). Calc-silicate rocks in this zone include the assemblages plagioclase + quartz + calcic scapolite (A.B. Thompson in Goldsmith and Newton, 1977) and wollastonite + anorthite (Berry, 1989, 1991; Robinson, 1991). Leucocratic layers in migmatitic gneisses represent a diverse suite of fluid-absent partial melts that formed over a range of P-T conditions (Robinson *et al.*, 1982; Tracy and Dietsch, 1982; Thomson, 1992; Thomson *et al.*, 1992).

Recent field guides for this area give a more regional perspective of the high grade region (Robinson *et al.*, 1982, 1986a, 1989). This field trip will examine two localities in Zone VI of regional metamorphism exhibiting high-grade pelitic rocks and their partial melts (Fig. 1).

Regional Setting

The rocks visited on this trip lie within the Merrimack synclinorium (Fig. 1), one of two major tectonostratigraphic features of the New England Appalachians, regionally deformed and metamorphosed during the Devonian Acadian orogeny (Hall and Robinson, 1982). The Bronson Hill anticlinorium to the west is partly defined by a series of structural culminations, exposing Late Proterozoic and Late Ordovician basement (Thompson *et al.*, 1968; Tucker and Robinson, 1990). The Merrimack synclinorium to the east contains highly deformed and metamorphosed Silurian - Devonian sedimentary rocks with minor volcanics (Hall and Robinson, 1982).

In central Massachussetts, the Acadian orogeny has been divided into three generalized stages of deformation: the nappe stage, backfold stage, and dome stage (Robinson, 1979; Berry, 1987, 1989). Westward transport of material during the nappe stage produced large-scale fold- and thrust-nappes resulting in an inverted metamorphic gradient in some places (Thompson *et al.*, 1968; Robinson, 1979; Thompson, 1985; Elbert, 1988; Robinson *et al.*, 1988, 1991). During the backfold stage, the tectonostratigraphy along the eastern margin of the Bronson Hill anticlinorium was overturned to the east, accompanied by emplacement of some gneiss bodies in the Bronson Hill anticlinorium (Robinson, 1979; Robinson and Hall, 1980; Robinson *et al.*, 1982; Thompson, 1985; Peterson, 1992). Continued upward as well as lateral transport of basement gneisses during the culminating dome stage produced the final geometry of the gneiss domes of the Bronson Hill anticlinorium (Robinson, 1963, 1979).

Metamorphism

Differences in the character of metamorphic facies and in the metamorphic history are observed between the Merrimack synclinorium and Bronson Hill anticlinorium (Fig. 2). Within the Merrimack synclinorium in Massachusetts, high temperatures and moderate pressures were achieved through early high temperature/low pressure Buchan-style metamorphism followed by heating to peak thermal conditions and then compression during slow cooling (Schumacher *et al.*, 1989, 1990; Robinson *et al.*, 1986b, 1989; Thomson, 1989, 1992). In contrast, rocks from the Bronson Hill anticlinorium appear to have followed a clockwise P-T path involving compression, followed by cooling and decompression (Robinson *et al.*, 1982; Schumacher *et al.*, 1989, 1990; Schumacher, 1990; Tracy *et al.*, 1991). However, interpretation of the metamorphic history of this belt is complicated by a Late Paleozoic metamorphic overprint in certain parts of the anticlinorium (Robinson and Tucker, 1991; Robinson *et al.*,1992).

Evidence for this counterclockwise P-T path in the Merrimack synclinorium has been found in independent petrologic investigations of pelitic rocks, calc-silicate rocks, partial melts, and fluid inclusion studies (Tracy and Dietsch, 1982; Robinson *et al.*, 1982, 1986a, 1988, 1989; Schumacher *et al.*, 1989; Berry, 1989, 1991; Thomson, 1989, 1992; Winslow *et al.*, 1991). The metamorphic features preserved in Zone VI of the Merrimack synclinorium can be grouped informally into early metamorphic features, peak metamorphic features, and retrograde features. These features are breifly summarized below. A more detailed discussion may be found in Thomson *et al.* (1992).

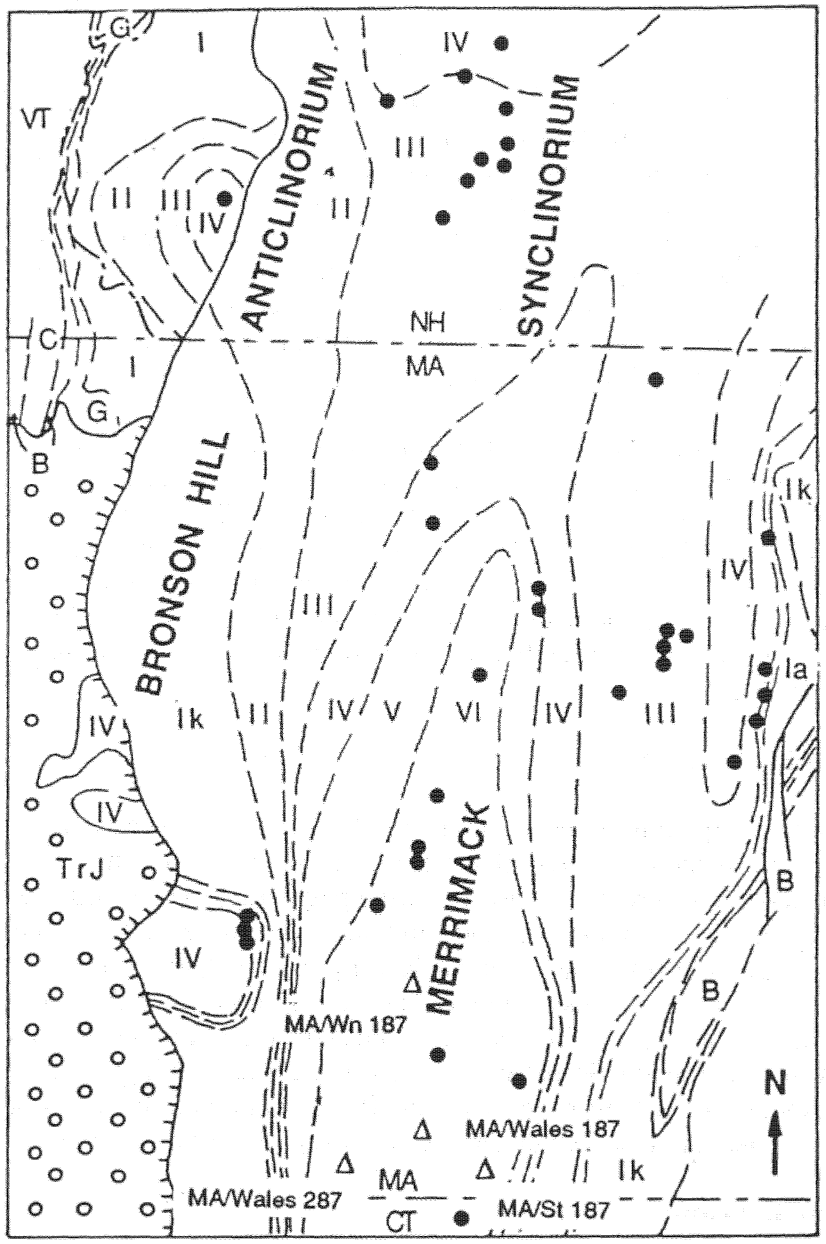
Early Metamorphic Features

An early low-pressure high-temperature history that involved partial melting is suggested by the occurrence of sillimanite pseudomorphs after andalusite in pelitic schists and gneisses (Fig. 1) and by mineral reactions in cordierite±garnet-bearing pegmatites (Tracy and Dietsch, 1982; Thomson, 1992). Analysis of both melt- and solid-state mineral reactions in the pegmatites suggest that the cordierite formed earlier and at lower pressures than the peak of metamorphism (Thomson, 1992). Some pegmatites contain cordierite that has broken down to aggregates of garnet + sillimanite + quartz, including samples collected from our second stop in the granulite facies region. A more complete discussion of cordierite±garnet-bearing pegmatites, including the melt reactions responsible for their genesis and the P-T conditions followed through the later stages of metamorphism is presented in Thomson (1992) and Thomson *et al.* (1992).

Peak Metamorphic Features

Garnets that belong to the prograde assemblages of Zone VI and parts of Zone V are characterized by homogeneous interiors with rims which differ in composition only where adjacent to other ferromagnesian minerals such as biotite or cordierite (Hess, 1969, 1971; Tracy *et al.*, 1976; Robinson *et al.*, 1982; Tracy, 1982). It has been suggested that prograde metamorphism and partial melting at temperatures of 675 to 730°C flushed the system of metamorphic fluids so that later localized retrograde ion exchange reactions occurred in the absence of significant amounts of metamorphic fluid (Robinson *et al.*, 1986a). Partial melting of pelitic schists and gneisses of appropriate bulk composition resulted in the formation of leucocratic garnetiferous melt segregations, the genesis of which is discussed in Thomson (1992) and Thomson *et al.* (1992).

Peak granulite-facies metamorphic conditions of pelitic schists and gneisses have been estimated using a variety of thermobarometric techniques applied to garnet and cordierite core compositions and isolated matrix biotite. The calculated pressures and temperatures have been estimated at up to 740°C and 6.5 kbar (Robinson *et al.*, 1986a, 1989; Thomson, 1989, 1992; Peterson and Thomson, 1991; Peterson, 1992a,b). The univariant assemblage anorthite +



Metamorphic Zones and Features:

VI	Garnet-Cordierite-Sillimanite-K-feldspar
V	Sillimanite-K-feldspar
IV	Sillimanite-Muscovite-K-feldspar
III	Sillimanite-Muscovite
II	Sillimanite-Staurolite
I_k	Kyanite-Staurolite
I_a	Andalusite-Staurolite
G	Garnet
B	Biotite
C	Chlorite
TrJ	Mesozoic Sedimentary and Volcanic Rocks
●	Sillimanite Pseudomorphs after Andalusite

Figure 1. Generalized map of metamorphic zones in central Massachusetts and southern New Hampshire based on pelitic rock assemblages (from Robinson *et al.*, 1982). Sample numbers are localities of rocks that have been dated using U-Pb systematics (see Thomson *et al.*, 1992).

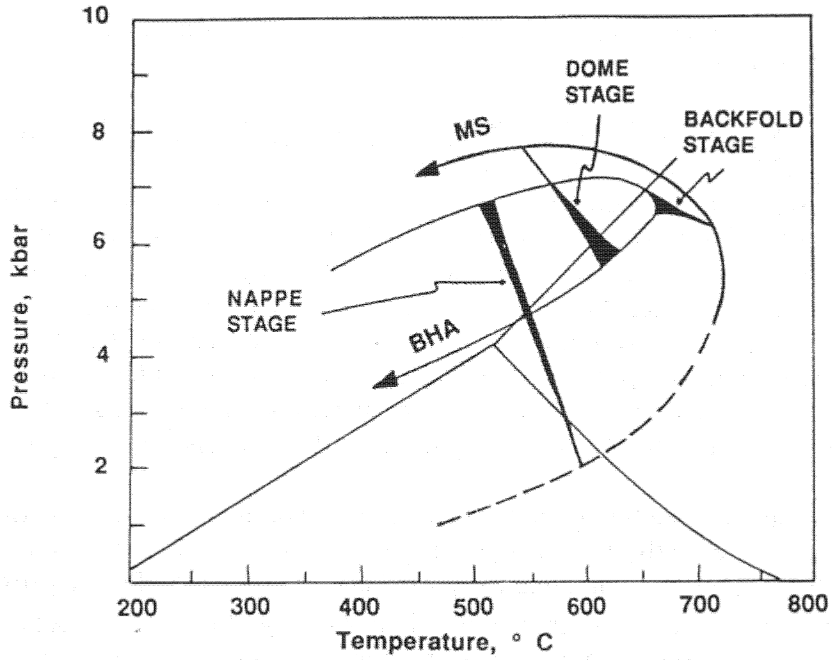


Figure 2. Integrated contrasting P-T paths fro the Keene dome of the Bronson Hill anticlinorium (BHA) and Merrimack synclinorium (MS) (from Schumacher *et al.*, 1989). Thick ends of tie lines lie on trajectory of the BHA; thin ends on trajectory of the MS.

wollastonite + grossular + quartz from a rock south of Holland, Massachusetts (Berry, 1989; Robinson, 1991) requires temperatures in excess of 730°C assuming 5 kbar, or in excess of 760°C assuming 6 kbar (Berman, 1988). The univariant assemblage corundum + garnet + sillimanite + spinel in a rock near the Holland-Sturbridge line in southern Massachusetts, provides a temperature estimate of 750°C assuming 5 kbar, or 800°C assuming 6 kbar (for further details, see STOP 7 in Thomson *et al.*, 1992).

Retrograde Features

On the whole, peak metamorphic features in Zone VI are remarkably well preserved. Retrograde effects such as chloritized biotite or sericitized feldspar are rare as compared, for example, to rocks in lower grade zones. The most common post-peak retrograde effects appear to be ion exchange reactions that took place at lower temperatures, affecting compositions of garnet, biotite, and cordierite near mutual grain boundaries (Tracy *et al.*, 1976; Robinson *et al.*, 1986a, 1989; Peterson and Thomson, 1991; Peterson, 1992; Thomson, 1992).

Petrographically discernable retrograde reaction textures in Zone VI include: fine-grained cummingtonite along orthopyroxene grain boundaries (Hollocher, 1985; Robinson *et al.*, 1986a, 1989); pale green biotite on edges and fractures of cordierite and/or garnet near large K-feldspar grains within partial melts (Tracy and Dietsch, 1982; Robinson *et al.*, 1989; Thomson, 1989, 1992); garnet + sillimanite + quartz aggregates entirely within large pegmatite cordierite grains (Tracy and Dietsch, 1982; Robinson *et al.*, 1989; Thomson, 1989, 1992); muscovite-bearing pegmatites and granitic dikes (Robinson *et al.*, 1989; Berry, 1989; Thomson, 1992; Peterson, 1992); and secondary epidote veins (Robinson *et al.*, 1989). While most of the small-scale retrograde features are related to post-Acadian cooling, some features, particularly the latter two, could be late Paleozoic in age.

STOP B-5: RANGELEY FORMATION AND LEUCOCRATIC GARNETIFEROUS MELT SEGREGATION (Wales Quadrangle).

This stop provides the opportunity to collect fresh samples of Qtz-Pl-Kfs-Sil-Bt-Grt±Crd gneiss of the Lower Silurian Rangeley Formation. In addition, the locality has a spectacular exposure of a leucocratic garnetiferous melt segregation thought to be the result of fluid-absent biotite-dehydration melting reactions of the surrounding gneisses. The mineral assemblage within the segregation is dominated by garnet, orthoclase and quartz. The surrounding gneiss, on the other hand, contains a greater proportion of hydrous minerals, particularly biotite, and lesser amounts of quartzofeldspathic minerals. The garnets within the concordant to slightly discordant segregation are coarser grained (2 cm) than garnets in the surrounding host gneiss.

Samples collected from a tiny roadbed outcrop (since obliterated) near the intersection of Hitchcock and McBride roads, are vein-type migmatites with similarities to the rocks observed on the pipeline. The samples contain euhedral garnet crystals ($Alm_{69.3}Prp_{26.1}Sps_{4.8}Grs_{2.8}$) to 4 cm in diameter set within a coarse-grained matrix consisting of quartz, orthoclase and sillimanite. In addition, the samples contain abundant sillimanite pseudomorphs after andalusite up to 3 cm in diameter and 9 cm in length.

Samples of pelitic schist and and adjacent melt segregation from a site just west of Mt. Hitchcock Road on the pipeline were analyzed in detail by electron microprobe. Biotite analyses reveal that the most Fe- and Ti-rich biotites are those present in the leucocratic segregation as inclusions in K-feldspar (X_{Mg} = 0.498 - 0.521; Ti/11 oxygens = 0.292 - 0.335). The most magnesian biotites are isolated matrix grains within the leucocratic segregations (X_{Mg} = 0.552 - 0.573; Ti/11 oxygens = 0.258 - 0.283). Gneiss matrix biotites have an intermediate X_{Mg} (X_{Mg} = 0.534 - 0.544; Ti/ 11 oxygens = 0.251 - 0.267). The biotite inclusions in K-feldspar may represent compositions that existed in the gneisses just as melting to form the segregations began. The compositions of garnet grains within the melt segregation ($Alm_{70.7}Prp_{25.6}Sps_{1.9}Grs_{2.3}$) differ slightly from that of the associated gneiss ($Alm_{69.7}Prp_{25.2}Sps_{2.3}Grs_{3.7}$), although this may not be true for all melt segregations and gneisses. Garnets in both the melt segregations and the gneisses are homogeneous except where adjacent to biotite and cordierite. When compared to average garnet core compositions, garnet compositions adjacent to biotite are characterized by higher X_{Fe}, higher almandine and spessartine content, similar to slightly lower grossular content, and lower pyrope content.

Thermobarometric calculations based on garnet core and isolated matrix biotite compositions of the gneiss and adjacent melt segregation suggest that the melt segregations formed under conditions similar to the peak granulite facies metamorphism recorded in the host gneisses. The garnet-biotite calibration of Thompson (1976) yields peak metamorphic temperatures of 700 - 710°C. Minimum pressure estimates based on garnet compositions and the calibration of Tracy *et al.* (1976) are 6.2 - 6.3 kbar. Temperatures calculated using the Fe- and Ti-rich biotite inclusions in K-feldspar within the segregation together with garnet core compositions range from 750 - 780°C. These would represent maximum temperatures for leucocratic garnet melt segregation formation if it is assumed that the biotite inclusions did not re-equilibrate and that they are, in fact, in equilibrium with garnet core compositions.

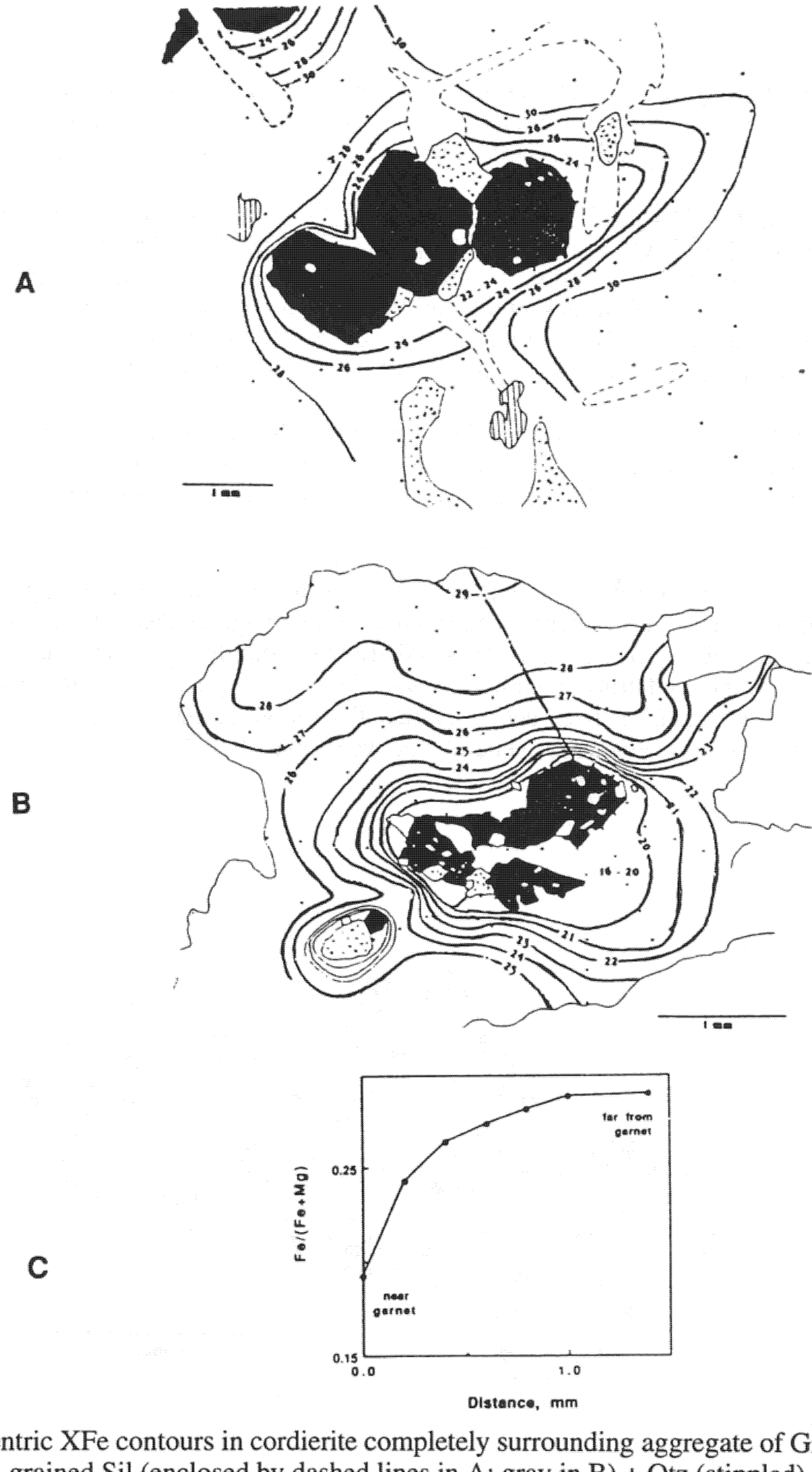


Figure 3. Concentric XFe contours in cordierite completely surrounding aggregate of Grt (black) + fine-grained Sil (enclosed by dashed lines in A; gray in B) + Qtz (stippled). Red-brown biotite inclusions are shown in striped pattern. Analysis points in cordierite shown as small black dots: (A) Sample Sturb2 collected from STOP 2; (B) Sample P11-1. Traverse of cordierite compositions along black line in (B) is shown in (C) (from Thomson, 1992).

A sample (MA/Wales 287, Thomson *et al.*, 1992) of coarse-grained sillimanite-orthoclase-garnet-biotite schist was collected at this locality. Monazites from this sample give a concordant age of 363±1 Ma showing that the partial melting and metamorphism here is Devoniian.

STOP B-6: CORDIERITE±GARNET-BEARING PEGMATITE, SCHISTS, AND GRANULITES OF THE SILURIAN(?) PAXTON FM (Southbridge Quad).

This stop will allow you to collect pristine samples of cordierite±garnet-bearing pegmatite thought to be the result of fluid-absent, biotite-dehydration melting reactions of the surrounding gneisses. The host rock at this locality consits of quartz-sillimanite-garnet-cordierite schists and interlayered biotite and calc-silicate granulites of the Silurian(?) Paxton Formation.

The pegmatite consists of the assemblage Qtz-Pl-Kfs-Sil-Bt-Crd-Grt. Cordierite within the pegmatite is blue to dark lavender in color and up to 8 cm across. Some samples from this locality show large dark patches up to 15 cm across that appear to be a pinite alteration of original cordierite.

Samples from this locality show evidence of cordierite breaking down to aggregates of garnet + sillimanite + quartz. The cordierite in the vicinity of these aggregates is zoned, particularly within 1000 μm of the aggregate (Fig. 3). Cordierite far from the aggregate has an XFe of 0.30 - 0.31. The X_{Fe} of the cordierite decreases dramatically as the aggregates are approached and becomes as low as 0.22 directly adjacnt to garnet. Garnet within the aggregates is also strongly zoned (Fig. 4). Pyrope contents are generally between 26 and 28%.

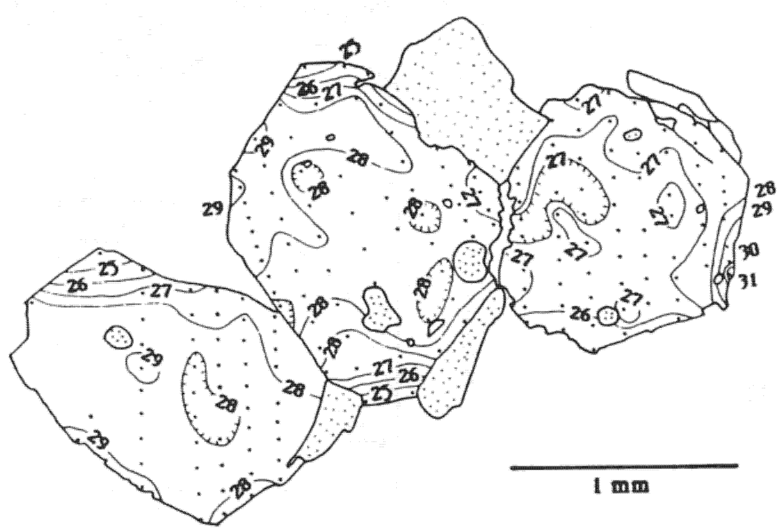


Figure 4. Contours of mole percent of pyrope in garnet in aggregate entirely within large cordierite from sample Sturb2 (STOP 2). Qtz (stippled), Crd and Grt (unpatterned). Analysis points shown as small black dots. See Figure 3c.

Conditions of the onset of cordierite breakdown were estimated at 762°C and 5 kbar using the aggregate garnet "core" compostion and the cordierite composition far from the aggregate. The last recorded conditions of cordierite breakdown have been estimated at 617°C and 6 kbar based on the compositions of the aggregate garnet rim and the adjacent cordierite. All calculations used the calibrations of Thompson (1976) and Bhattacharya (1986). Conditions of the onset of cordierite breakdown recorded in a number of samples from the pipeline are between 760 and 800°C, 5 - 5.5 kbar, similar to those recorded at this locality. However, the sample from this locality records "final" conditions of breakdown that suggest lower pressures and higher temperatures than those recorded in the west. The last recorded conditions of cordierite breakdown are between 570 and 590°C and 6.6 - 6.7 kbar in samples 10.5 miles to the west. This data may suggest that, during the late stages of cordierite breakdown, temperatures were slightly lower and pressures slightly higher in the west than in the east.

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Regional Geology of Western and Central New England

Modified from Robinson (1986) and Thompson, Cheney, and Robinson (1986).

Regional Stratigraphic and Structural Setting

The bedrock geology of western and central New England is exposed in three synclinorial belts separated by two major anticlinoria (Figures C-1, C-2). From west to east these are the Middlebury synclinorium, the Berkshire-Green Mountain anticlinorium, the Connecticut Valley synclinorium, the Bronson Hill anticlinorium, and the Merrimack synclinorium. In Massachusetts the east margin of the Connecticut Valley synclinorium and the west margin of the Bronson Hill anticlinorium are unconformably overlain by Upper Triassic-Lower Jurassic continental sedimentary rocks and basaltic volcanics localized along a major west-dipping listric normal fault. In Vermont the Green Mountian anticlinorium is cut by the 100 m.y. old Cuttingsville Syenite Complex and the Connecticut Valley synclinorium is cut by the 120 m.y. old Ascutney Complex.

The final forms of the synclinoria and anticlinoria were produced by Acadian (Devonian) folding, but the detailed configuration of rock types is a product of both Taconian (Ordovician) and Acadian deformation and facies relationshops of rocks ranging in age from Upper Precambrian to Lower or Middle Devonian.

The Middlebury synclinorium is entirely underlain by North American Grenvillian basement. This is unconformably overlain locally by Eocambrian clastic and carbonate rocks of the Dalton Formation and the by the continuous clean quartz sandstone of the Lower Cambrian Cheshire Quartzite. The Cheshire is the substrate upon which was deposited the extensive North American carbonate bank sequence of dolomites and limestones with subordinate sandstones and shales that ranges in age from Lower Cambrian through uppermost Lower Ordovician, and which was deposited on the then south coast of North America (Laurentia) for a strike length of thousands of kilometers. The carbonate bank sequence was block-faulted and locally eroded even down to basement before being overlain unconformably by a Middle Ordovician carbonaceous flysch sequence with local calcitic limestone at the base. The Middle Ordovician flysch was then tectonically overridden by the lowest and westernmost of the Taconic allockthons, consisting of a starved shale facies of strata from Eocambrian through Middle Ordovician age. We know that the Taconic facies represents a continental slope rise sequence deposited oceanward of the carbonate bank, because it contains local limestone conglomerates and carbonate-cemented clean quartz sands that had their source in the carbonate bank environment or shoreward of it. We also know that the earliest emplacement of the allochthon was during Ordovician graptolite zones 12 and 13, because fossils of these ages can be found in the matrix of a "blocks-in-shale" unit at the top of the Middle Ordovician flysch, containing lithified blocks of nearly all units of the Taconic facies. Unfortunately this "simple" picture is greatly complicated by imbricate thrusting after rocks of the Taconic facies and flysch had been metamorphosed to slate, and by the consequent production of younger tectonic mélanges.

We know that the Taconic facies was deposited east of the west margins of the Grenvillian basement in the Green Mountian and Berkshire anticlinoria, because the base of the Middlebury synclinorium sequence rests unconformably on the Grenvillian in these locations. However, the exact source of the Taconic facies east of this point is much debated. In Vermont, Karabinos (1986), Thompson *et al.* (1986) and Downie *et al.* (1986) have described a major shear zone within the green Mountains between two basement types along which are located fault slivers and tight infolds of Eocambrian to Cambrian rocks of Taconic affinities. They suggest that this

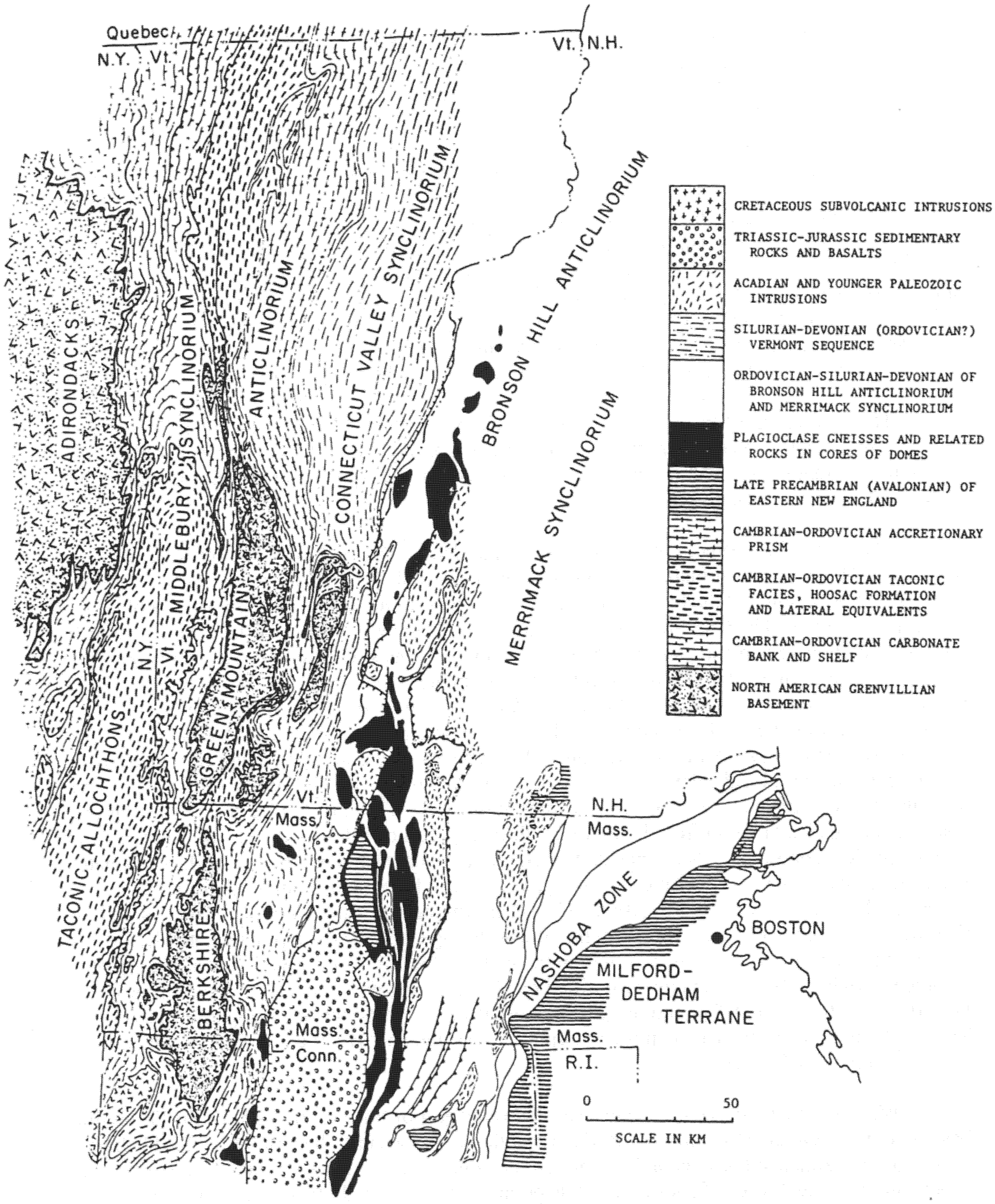
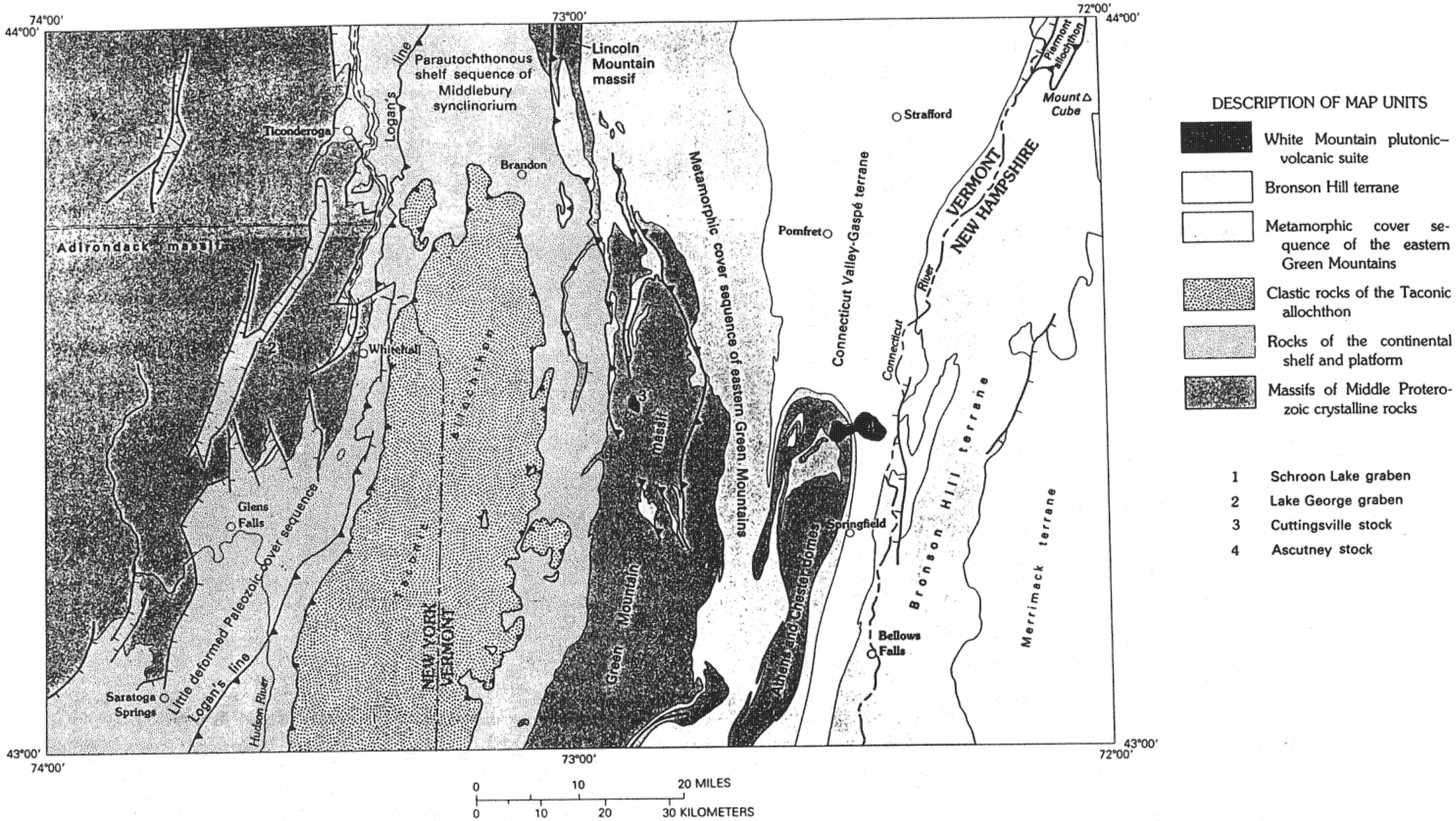


Fig. C- 1. Generalized geologic map of northwestern and central New England showing major stratigraphic and structural features.

synclinal infold may pass entirely beneath the Grenvillian rocks of the western part of the Green Mountains and reappear as tectonic windows within the deepest exposed part of the Chester gneiss dome still farther east. In this model the Eocambrian-Cambrian strata deposited unconformably on the Grenvilian basement along the east flank of the Green Mountain anticlinorium and in the outer part of the Chester dome were deposited farther from North America than the Taconic facies. In Massachusetts, Stanley and Ratcliffe (1985) have determined that the Hoosac Formation, which rests on the Grenvillian basement on the east side of the Berkshires, was originally deposited west of the Taconic facies. The Hoosac is in fault contact with still more easterly facies along the Whitcomb summit thrust, which is regarded in Massachusetts as the source region of the Taconic allochthons. Stanley and Ratcliffe suggest that the Whitcomb summit thrust traces along the east margin of the Hoosac Formation in Vermont and is the source region of the Taconic rocks there also, in contradiction to the opinions of Karabinos, Thompson, and Downie.

Most workers thoughout the region are in agreement that the Cambrian-Lower Ordovician strata east of the Hoosac Formation in Vermont and Massachusetts belong to a complex accretionary wedge of continental slope and oceanic strata that was caught in an east-dipping subduction zone between North America and an eastern plate during Taconian closing of Iapetus (Stanley and Ratcliffe, 1985). Evidence, best exposed in adjacent Quebec, suggests that one of the earliest events in this closing was the obduction of a then young ophiolite sequence onto the North American continental margin, probably in earliest Ordovician time. One of the distinct problems region is how metamorphosed volcanics in tectonic contact with fragments of this ophiolite were subsequently brought rapidly back near the surface to cool within the Middle Ordovician.

The pre-Silurian rocks of the Bronson Hill anticlinorium, consisting of five major lithic associations, are considered to represent the eastern plate that was overriding the accretionary wedge (Robinson and Hall, 1980; Hall and Robinson, 1982; Zen *et al.* 1983). The oldest exposed rocks are microcline gneissed, quartzites and related rocks in the interior of the Pelham gneiss dome that are interpreted to be a series of metamorphosed rhyolitic volcanics and interbedded sediment of late Precambrian age and show lithic affinities to rocks in southeastern Connecticut interpreted to be part of the Avalon plate. These are overlain by a complicated series of calc-alkaline plagioclase gneisses and amphibolites that may include layered mafic and felsic volcanics, mafic dikes, and gabbroic, tonalitic, and granitic intrusions. These rocks may be as old as late Precambrian, although a popular view is that they represent the Ordovician roots of an island arc. Several of the gneiss domes expose massive granitoid intrusive rocks of the Oliverian series that give apparently reliable Middle Ordovician ages, although discordant intrusive contact relations with the overlying Ammonoosuc Volcanics can rarely be demonstrated. Furthermore, local basal conglomerate and quartzite in Ammonoosuc Volcanics overlying Plagioclase gneisses hint of an unconformity at this position. The Ammonoosuc Volcanics have been well characterized as arc volcanics ranging in composition from tholeiitic basalt through andesite and dacite to K-rich rhyolite that give evidence of derivation from melting of mantle or basalt protoliths without significant contributions from continental crust (Aleinikoff, 1977; Schumacher, 1981a, 1983; Leo, 1985). The underlying Oliverian granitic rocks, in contrast, show evidence of a continental heritage (Zartman and Leo, 1984) which has recently been identified as probably of Grenvillian age (Barbara Barreiro, personal communication, 1986), supporting earlier suggestions (see Zen *et al.*, 1983) that North American basement may underlie the Bronson Hill belt at depth. Above the Ammonoosuc Volcanics is the Middle Ordovician Partridge Formation dominated by metamorphosed sulfidic black shale with subordinate arc-type volcanics very like those of the Ammonoosuc (Hollocher, 1985). The Partridge has been traced discontinuously northward into northwestern Maine where it contains Middle Ordovician graptolites.



Map showing major lithologic—tectonic units of the Glens Falls quadrangle. (Terrane is used in this map for a mass of related rocks, not necessarily for a distinct accretionary unit in the plate tectonic sense.)

Figure C-2. Major lithologic -- tectonic units of the Glens Falls 1x2° Quadrangle (Thompson, McClelland, and Rankin, 1990)

The plagioclase gneisses of the "eastern basement" extend considerably west of the present position of the Bronson Hill anticlinorium and appear from beneath younger cover in a series of gneiss domes in western Massachusetts and Connecticut. In the Bristol dome of Connecticut, Stanley (Hatch *et al.*, 1984) has recognized a tectonic contact where the plagioclase gneisses of the "eastern basement" appear to be thrust above probable uppermost Lower Ordovician strata of the accretionary wedge in his proposed "Bristol thrust". In the Chester dome of Vermont it is possible that the plagioclase gneisses of the Barnard Volcanics also represent "eastern basement" that is thrust over the Moretown Formation of the accretionary wedge in a thrust like the Bristol thrust (see Figure C-4). If so, then the entire thickness of the Taconian accretionary wedge may be exposed in tectonically thinned fashion on the flanks of the Chester dome.

The status of the Ammonoosuc Volcanics and Partridge Formation with respect to the proposed Bristol thrust is somethat equivocal. These units or equivalents form a clear upper part of the sequence on the eastern side, but there is some evidence to suggest that they may also have been deposited on the west side, thus healing over the suture. It is curious in this light to note that the fossil age of the Partridge Formation is virtually identical to the youngest fossil age obtained in the matrix of the "blocks-in-shale" unit beneath the Taconic allochthons.

The Silurian-Devonian cover sequence in the western and central parts of the Connecticut Valley synclinorium has been and continues to be plagued with uncertainties. On the west, a thin sequence of conglomerate and calcareous rocks is traceable into Quebec where there are Silurian fossils, implying Lower Devonian age for most of the overlying shales, siltstones and calcarious granulites of the Goshen, Northfield, Gile Mountain and Waits River formations. On the east limb of the synclinorium fossil control is excellent, showing that the basal Clough Quartzite is Lower Silurian (late Llandovery), the calcarious Fitch Formation ranges up into the uppermost Silurian (Pridoli) and that the black shales and sandstones of the Littleton Formation range into the uppermost Lower Devonian (late Emsian). A structural solution to a stratigraphic dilemma in westen Massachusetts implies that the lower part of the Littleton Formation is thrust westward for many 10's of kilometers on the proposed Whately thrust over the Gile Mountain and Waits River formations and implies that these units are at least as young as Lower Devonian (Robinson *et al.*, 1984). In apparent contradiction to this, two graptolite localities found in the last two years, one in Vermont, one in Quebec, indicate a Middle Ordovician age for some of the carbonate-bearing strata (Bothner and Finney, 1986). To further confuse the issue, newly collected plant fossils from an older locality in Quebec about twenty miles along strike from the graptolites indicate a late Lower Devonian age.

Eastward from the Bronson Hill anticlinorium into the Merrimack synclinorium, the thin Silurian section thickens rather abruptly across a "tectonic hinge" into the thick shale and clastic sequence of the Merrimack trough, the post-Taconian sedimentary basin that closed during the initial stages of the Acadian orogeny. Although fossil control in western Maine, New Hampshire and central Massachusetts is poor, lithic correlations with the central Maine and thence to relatively fossil-rich Aroostook County are good and lead to a fairly clear picture of sedimentary history. The early and middle Silurian history involved clastic sedimentation with a source in the eroding Taconian orogen to the west. The upper Silurian is dominated by calcareous clastic rocks with a possible contribution from Upper Silurian volcanoes known to have existed to the east. The Lower Devonian saw a sedimentary source reversal with clastics being shed westward from tectonic lands to the east (Hall *et al.*, 1976), forming deltaic complexes and a marine clastic wedge that prograded westward across the site of the Bronson Hill anticlinorium and beyond.

Eastward from the line of the proposed Bristol thrust there is little or no evidence for severe Taconian deformation or metamorphism, and all major structural features are Acadian or younger. The tightly controlled Silurian-Devonian stratigraphic sequence of the Bronson Hill anticlinorium has allowed delineation of a series of early west-directed fold nappes. More recent delineation of the sequence in the Merrimack synclinorium has made it possible to show that the fold nappes are truncated by slightly younger thrust nappes (P.J. Thompson, 1985; Elbert, 1986) which locally have carried rocks of the synclinorium sequence over the top of the anticlinorium onto its present west limb. The early recumbent folds and thrusts, including both those of the Acadian and those of the Taconian west of the Bristol have been further folded in the later phases of the Acadian orogeny, including the gravitationally induced formation of a series of complex gneiss domes. These domes dominate the Bronson Hill anticlinorium, but are also abundant in the adjacent Connecticut Valley synclinorium and contain a variety of buoyant core rocks ranging from Grenvillian granitic rocks to Ordovician plutons.

Regional Distribution of Metamorphic Zones

Metamorphic zones in the field trip area (Figure C-3) were produced during at least two complex metamorphic cycles, the Taconian and the Acadian. The abundance of pelitic schists throughout the region has made it possible to use the Barrovian sequence of isograd minerals with some modification over a wide area. However, the detailed meaning of these isograds can vary from place to place. For example, the kyanite zone areas in northwestern Vermont (as at Mount Grant) mainly contain kyanite in pelites of high Al content, commonly associated with chlorite and chloritiod. This is indicative of a much lower metamorphic intensity than the appearance of kyanite in pelites of low Al content, as, for example, in the Littleton Formation of western New Hampshire and central Massachusetts where kyanite appears only upon breakdown of the assemblage staurolite + chlorite + muscovite to produce biotite + kyanite. The metamorphic zones shown on Figure C-3 are based on the most intense metamorphism known in Phanerozoic rocks in these areas. The heavy dashed line separates areas where the most intense metamorphism was Taconian to the west from those where it was Acadian to the east. This boundary has been very difficult to delineate and is based mainly on detailed K/Ar studies of metamorphic minerals (Lanphere and Albee, 1974; Sutter, Ratcliffe, and Mukasa, 1985; Lanphere, Laird, and Albee, 1983; Albee, 1974; Laird, Lanphere, and Albee, 1984) across areas where petrologic distinctions may be difficult or impossible to detect. West of the line there are some examples of what appear to have been Acadian retrograding of higher grade Taconian assemblages. East of the line, for a short distance, there are local examples of low grade relict assemblages from the Taconian overprinted by higher grade Acadian assemblages (see for example Rosenfeld, 1968; Karabinos, 1984) though this seems to be limited to rocks that belonged to the Taconian accretionary prism. The difficulty and uncertainty of this line are pointed out by the very latest K/Ar work on the peak metamorphic kyanite zone minerals at Mt. Grant which seems to suggest an Acadian age for this zone, in contradiction to Figure C-3. Detailed studies indicate the Taconian metamorphism was itself polyfacial, particularly as shown by detailed amphibole zoning in northern Vermont and also by relict eclogites in southwestern Massachusetts (Harwood, 1979; Maggs, Cheney, and Spear, 1986).

Texturally zoned garnets, with two distinct growth stages, from the Pinney Hollow pelites of the Athens dome have been interpreted by Rosenfeld (1968) as "tectonometamorphic angular unconformities". He suggested that the outer zone represents an Acadian overgrowth of a pre-existing Taconian core, because similar textures were not found in nearby Silurian-Devonian rocks. Thompson, Tracy, Lyttle and Thompson (1977) have shown that such discontinuities can

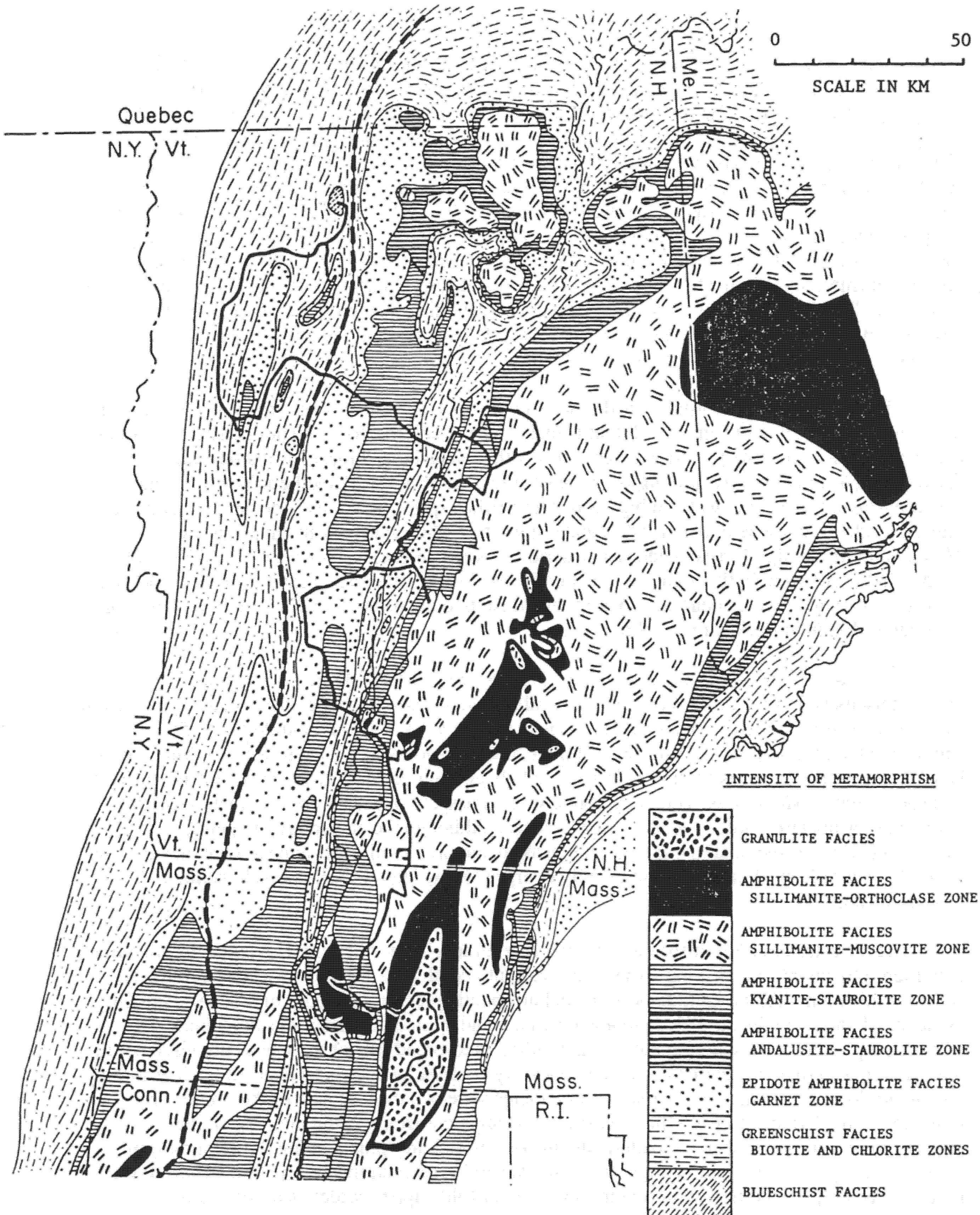


Fig. C-3. Generalized metamorphic map of northwestern and central New England showing distribution of peak metamorphic assemblages of Acadian and Taconian age. Heavy dashed line separates areas to the east where peak assemblages are Acadian from areas to the west where they are mainly Taconian.

develop via garnet-consuming reactions during a single prograde event. Nonetheless, Karabinos (1984a), using textural and chemical data, has shown that a pervasive retrograde event separated two prograde stages of garnet growth in similar rocks from Jamaica, Vermont (southwest of the Chester dome). Similarly zoned garnet has been reported in the northern Chester dome by Downie (1982) and at Gassetts, Vermont. Discontinuities in garnet growth could also have resulted from momentary cooling due to the Acadian emplacement of hot nappes into a cooler environment, but before thermal relaxation. Laird and Albee (1981a, 1981b) have given evidence for polymetamorphism in the eastern cover rocks of the Green Mountain massif. Specifically, they have interpreted amphibole zoning patterns, from amphibolites in the Cambrian Pinney Hollow Formation, as indicative of two phases of medium-pressure Ordovician metamorphism as well as medium-pressure and a later lower-pressure Devonian metamorphic phases. Pinney Hollow amphibolites from the outer mantle of the Athens dome apparently record only the two Acadian phases (Laird and Albee, 1981a, 1981b).

The Acadian metamorphism is divided into distinct western and eastern highs separated by the Connecticut Valley low grade belt that is mainly in the chlorite zone. In northeastern Vermont the western high is of low-pressure facies series with andalusite and local sillimanite zones surrounding a series of cross-cutting Acadian granites. In central and southern Vermont, and western Massachusetts the western high is of intermediate-pressure facies series with kyanite at the highest grade except where there is sillimanite in western Massachusetts and adjacent Connecticut. This metamorphic high is associated with a series of Acadian gneiss domes, and local isograds are subparallel to foliation in the domes as most typically illustrated in the Chester-Athens dome. By contrast, Acadian isograds bear no apparent relationship to exposures of Grenville basement along the Green Mountian Berkshire anticlinorium.

The eastern metamorphic high in New Hampshire, western Maine, and central Massachusetts is dominated by a broad area of sillimanite-muscovite zone rocks locally punctuated by areas with sillimanite-orthoclase assemblages. Recent work (Lux and Guidotti, 1985; Thomson and Guidotti, 1986) indicates that the sillimanite-orthoclase zone in western Maine may be related to contact effects of the late Paleozoic Sebago Pluton. In southwestern New Hampshire and south-central Massachusetts there are nine small areas and one large area characterized by sillimanite-orthoclase-garnet-cordierite assemblages. Although the characterizing pelitic assemblage is the same in all these areas, the peak metamorphic minerals appear to be late tectonic in southwestern New Hampshire, but heavily overprinted by later deformations in souther Massachusetts.

The western margin of the eastern high is complex. In northern New Hampshire this high is bordered by andalusite-bearing rocks. Further south conditions were close to those of the Al-silicate triple point. At the latitude of Bellows Falls pelitic rocks at high structural levels contain evidence of early andalusite with subsequent kyanite and sillimanite, and rocks containing sillimanite pseudomorphs after andalusite are rather widespread east of the gneiss domes, even in the sillimanite-orthoclase-garnet-cordierite zones. By contrast, pelitic rocks close to the interior of the gneiss domes were in a kyanite-sillimanite facies series on the east side of the domes or remained in the kyanite zone on the west side. It is apparent from a wide range of studies completed or in progress that the pattern and history of metamorphism along the west margin of the easten Acadian metamorphic high, including several preserved metamorphic overhangs, was related to a complex thermal evolution involving early fold nappes, widespread syntectonic plutonic sheets, later thrust nappes, and still later gneiss domes. The pattern in central Massachusetts is further complicated by the thermal anomaly of the late Acadian (380 Ma)

Belchertown Intrusive Complex, which heated local strata above ambient kyanite zone conditions during Acadian dome-stage deformation.

Taconian Metamorphism in the Context of Plate Tectonics

Taconian metamorphism in western New England appears to have been limited to the region where the eastern North American continental margin, its cover, and the related accretionary wedge were subducted beneath the eastern plate (Stanley and Ratcliffe, 1985). The metamorphism was obviously polyfacial with the earliest high-P, low-T phase extremely difficult to study, but the details have been and will continue to be difficult to decipher, particularly because the highest grade rocks were heavily overprinted during the Acadian. In the uppermost (*i.e.* Middle Ordovician) strata overlying the eastern part of the accretionary wedge and in the Middle Ordovician rocks of the eastern plate, there is little or no evidence of any Ordovician metamorphism more intense than the Acadian metamorphism in immediately overlying Silurian-Devonian strata, which are locally at chlorite grade. However, in some pre-Middle Ordovician strata there is evidence of older metamorphism. This is particularly true in the Late Precambrian rocks of the Pelham Dome (Robinson *et al.*, 1975; Roll, 1986) where there is evidence of a pre-Acadian metamorphism of medium-pressure granulite facies, the age of which is presently unknown.

Acadian Metamorphism in the Context of Plate Tectonics

Interpretation of Acadian metamorphism in the context of plate tectonics is made difficult by the lack of evidence for a pre-existing oceanic tract near the locus of most intense deformation and metamorphism between the Bronson Hill anticlinorium and the region around Boston where late Precambrian and Cambrian strata remain little metamorphosed. There is, however, abundant evidence for the development of a Silurian marine basin, the Merrimack trough, filled with a thick clastic sequence that was covered over by Lower Devonian flysch at the beginning of the Acadian orogeny (Hall *et al.*, 1976). A model favored by some (Chamberlain, 1986; Spear, 1986; Robinson *et al.*, 1986) is that the Merrimack trough was formed by late Ordovician-Silurian crustal extension parallel to the west margin of a single eastern (Avalon) plate in the back-arc region of the previous Middle Ordovician volcanic arc system. Crustal thinning provided a region for high heat flow into the bottom of the sedimentary pile, a favorable environment for the emplacement of a variety of mafic to felsic plutons, and an appropriate setting for early low pressure (Buchan) phases of metamorphosm and partial metling. It is suggested that this earler phase was overprinted by higher pressure assemblages caused by tectonic burial by thrust sheets from the east and, at least in central Massachusetts, that the rocks with those peak metamorphic assemblages were subsequently involved in imbricate thrusting with development of mylonites, that carried part of the tectonic assemblage westward over the present site of the gneiss domes. Indeed, the Bronson Hill domes may have been localized near the east edge of thick Bronson Hill basement, itself possibly overlying Grenvillian basement, where the basement has been most heavily loaded by thrust sheets from the east. In such a model the Connecticut Valley metamorphic low may represent an H_2O-rich thermal blanket or wedge separating the effects of tectonically superimposed hot rocks from the east, from the effects of isotherms moving upward perpendicular to foliation planes, as shown by isograd patterns surrounding the Vermont line of Gneiss domes (compare Figures C-1and C-3).

Massif-Cover Relationships

As mapped by Karabinos (1986), the western shelf sequence south of Clarendon, Vermont rests unconformably on Grenville basement, although some minor thrust faults are present. North of Clarendon, however, basement and unconformably overlying Proterozoic Z to Cambrian basal clastics of the eastern Vermont sequence have been thrust westward over the western shelf sequence. To the south, this major shear zone along the western boundary of the northern Green Mountain massif apparently passes eastward onto the interior of the massif. Downie, Thompson, and Slack (1986) have described septa of cover rocks, similar to those in the eastern Vermont sequence, that are teconically juxtaposed with basement in the vicinity of this shear zone.

At least two major Paleozoic events, involving crustal shortening and metamorphism, in both the basement and its eastern cover sequence, have been recognized by Downie, Thompson, and Slack (1986; see also Karabinos, 1984, 1984a, 1985, 1986). The earlier, Taconian, event involved west-southwest movement and relatively ductile deformation as manifest by overturned to recumbent folding and the basement/cover thrust faulting. The later, Acadian, event involved upright to slightly overturned folding and may also have involved thrusting along new or pre-existing faults.

Kyanite-staurolite zone mineral assemblages in once higher grade pelitic Grenville basement, pre-date the earliest deformation recognized in both the basement and eastern cover sequence rocks. The occurrence of these kyanite-staurolite assemblages suggests that the initial Paleozoic metamorphism occurred at relatively deep crustal levels. This early metamorphism conceivably records tectonic loading from the east at the beginning of the Taconian orogeny. The rocks were then transported to shallower crustal levels prior to or during the Acadian orogeny and metamorphosed at conditions producing garnet-and biotite-zone assemblages in the eastern cover rocks and extensive biotite-zone retrograding of the basement rocks.

The Chester Dome Area

The essentially homoclinal sequence of deformed, garnet-zone Paleozoic metasediments and metavolcanics on the east flank of the Green Mountain massif is complicated, in southeastern Vermont, by the Chester and Athens gneiss domes. These are part of a belt of structural domes that extends from central Vermont to Connecticut. This belt, located just west of the Connecticut axis, is parallel to the somewhat analogous belt of gneiss domes that constitutes the Bronson Hill anticlinorium (Thompson *et al.*, 1968).

Thompson, Rosenfeld, and Downie (1986) point out that the Chester and Athens domes are comparable in lithic types, structural features and metamorphic history to certain of the deeper zones of the Alps, specifically, the lower Pennine nappes of the Val d'Ossola and Lepontine regions and in the Taüern window. The doming probably occurred during the later stages of metamorphism accompanying the Acadian orogeny. It was preceded by at least three stages of highly ductile deformation; the first of these was probably Taconian and the last probably Acadian. All three events prior to doming involved basement rocks in isoclinal folding that was probably recumbent and possibly some thrust faulring.

The domes provide a view of some of the lowest tectonic levels of the crystalline core of the Appalachians. They can be regarded as consisting of a core, an inner mantle and an outer mantle. Basement rocks in the cores contain a variety of quartzo-feldspathic gneisses, amphibolites (some with 5-7 cm chlorite clots after garnet), and graphite-schists associated with marble and calc-silicate. These core rocks are very likely reworked Grenvillian rocks, similar to those now exposed in the southwestern Adirondacks. Septa of cover rocks in the inner mantle resemble, in protolith, the septa of eastern cover rocks seen along the major shear zone in the central Green Mountains some 20 km to the west. These probably represent the root zones for the movements that resulted in the emplacement of the Taconic allochthons. The outer mantle contains metamorphosed sedimentary and volcanic rocks, including metamorphosed ultramafic rocks, more indicative of an oceanic environment. Alternative interpretations of some of these relations are being formulated by Ratcliffe and coworkers (Ratcliffe, 1994; Armstrong and Ratcliffe, 1994).

All rocks in the vicinity of the domes bear evidence of strong internal shear. Carbonate rocks and carbonaceous pelites appear to have been particularly susceptable. However, discrete surfaces of dislocation cannot be identified with certainty , nor can they be disproved. The extent of shearing in the rocks is well illustrated by the occurrence of "rolled garnets" in the inner mantle of the Chester dome that recorded a minimum of 720° of rotation during syn-kinematic growth (see the pioneering work of Rosenfeld, 1968, for method and description). Another interpretation of these garnets was recently given by Vance and Holland (1993).

Several detailed petrologic studies conducted in southeastern Vermont (*e.g.* Downie, 1982; Downie *et al.*, 1986; Karabinos, 1984, 1984a, 1985, 1986; Rosenfeld, 1968, 1972; Thompson, Lyttle and Thompson, 1977; and Thompson, Tracy, Lyttle and Thompson, 1977) have shown that porphyroblasts of staurolite and kyanite and the rims of garnet porphyroblasts are undeformed and overprint the deformational fabrics in the rocks. Accordingly, maximum metamorphic grade was attained in this area subsequent to, or near the end of dome emplacement. The concentric pattern of isograds around the domes is a manifestation of deeper crustal conditions brought near the surface by the rise of hot rocks in the domes rather than folding of the isograds, that is the domes acted as "heat pipes". The widespread occurrence of retrograde hydration and exchange reactions documented by these workers, attests to the slow cooling of hot rocks carried to a lower temperature crustal environment by the rise of the domes. The apparent pre-dome configurations of metamorphic zones in southeastern Vermont may serve to constrain partially the thickness and/or western extent of eastern derived thrust/nappe rock packages during the Acadian orogeny.

Figure C-4. Generalized geologic map for part E of field trip in western New Hampshire and eastern Vermont with route indicated. State line is west bank of Connecticut River. Horizontally ruled areas are Grenvillian basement of North America. Areas outlined in heavy black are gneisses in cores of domes of the Bronson Hill anticlinorium and the Barnard Gneiss of Vermont, a potential candidate for "eastern basement" above the Taconian accretionary prism. Gray shading indicates areas of probable Silurian-Devonian strata and intrusive rocks.

STOP C-1: BROCKWAYS MILLS

Description modified from Boxwell and Laird (1987).

Streamcuts here are Standing Pond Volcanics in contact with rocks of the Waits River Formation to the west (upstream) and rocks of the Gile Mountain Formation to the east (downstream). The contact with garnet grade calcareous schist of the Waits River Formation is very sharp and well-exposed. The contact between Standing Pond Volcanics and garnet grade, micaceous schist of the Gile Mountain Formation is not exposed.

Directly east of the contact with the Waits River Formation is a spectacular example of fasicular schist or "garbenschiefer". Splays of amphibole cover the rock, some of which emanate from two-inch diameter garnet porphyroblasts and radiate in 360°. Some amphibole grains within individual fascicles appear curved also. Within garnet porphyroblasts concentric and sigmoidal inclusion trails are visible. These patterns have been described by Rosenfeld (1968, 1972) and are used to interpret the rotational directions of the rocks during deformation. Please do not hammer on the excellent exposure of these garnets.

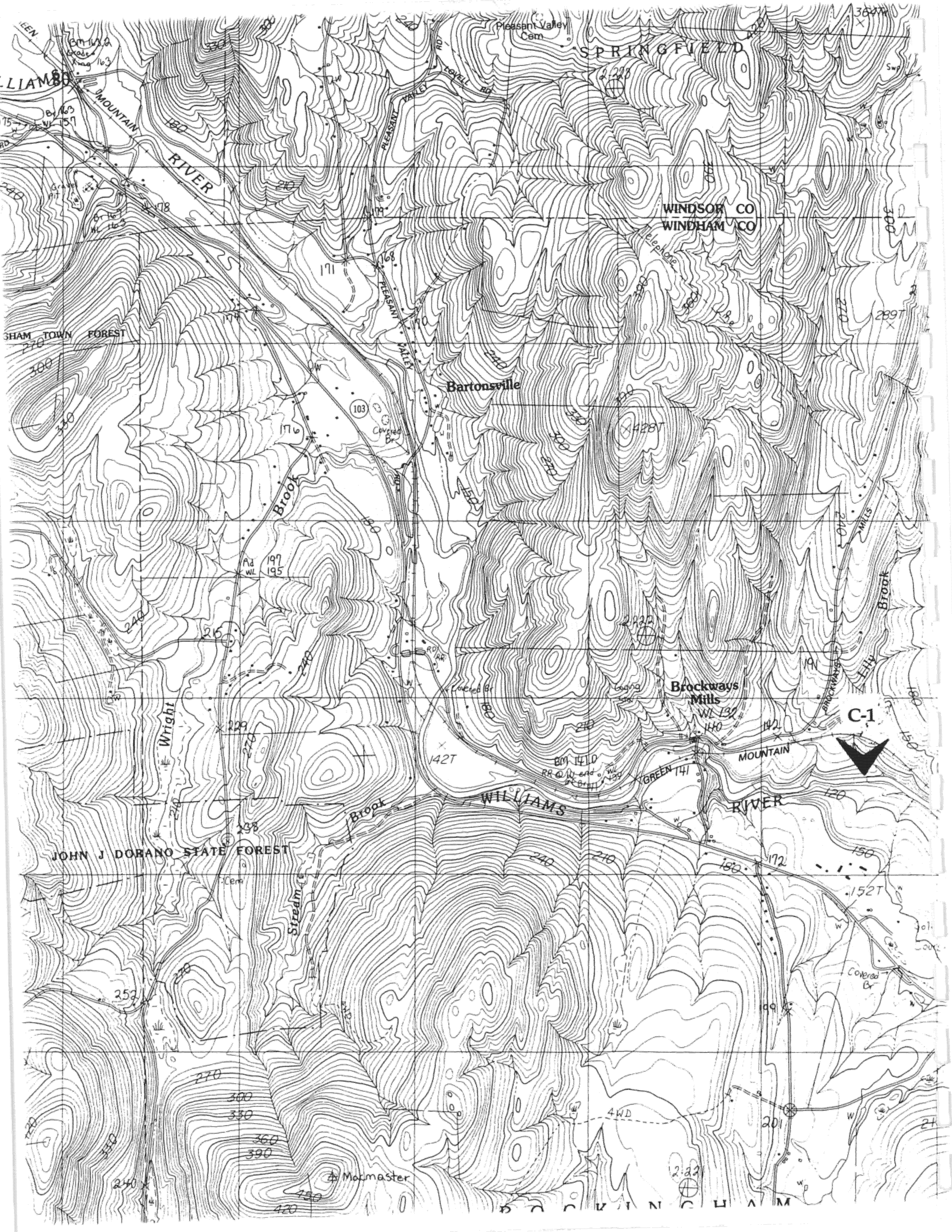
Downstream from the fasicular schists are other well defined layers of Standing Pond Volcanics. Some layers are very micaceous and appear similar to the garbenschiefer yet contain no garnets. Farther downstream massive, laminated, green-gray and white weathering amphibolites, some of which contain rusty pits where carbonate has weathered out, are present. Near the falls is a layer which weathers orange and is very light on the fresh surface. Plagioclase crystals are easily visible and predominate in this felsic rock.

Downstream from the falls is a layer of mafic rock which is laminated black and white. On the eastern side of the pool (downstream) are micaceous schists of the Gile Mountain Formation. Medium-grained garnet knobs are present in this rock.

Microprobe analyses of a sample of laminated, dark green-gray weathering amphibolite that does not contain garnet allowed classification of the amphibole as hornblende and the plagioclase as oligoclase (An_{17} - An_{20}) which is consistent with epidote-amphibolite facies zone classification.

Many of the laminae within layers at this stop appear to be cut by an S2 schistosity. F2 folds plunge moderately to the northeast, and are generally open folds. The long axes of amphibole grains lie parallel to the plane of S2 and in some localities the long axes of amphibole grains are parallel to axes of F2 folds. The curved fasicles of amphibole and rotated garnet grains are interpreted as indicating that F2 and M2 were contemporaneous.

Primary layering is visible in micaceous schists of the Waits River Formation upstream from the fasicular schist. Graded beds within the layers indicate that rocks to the east stratigraphically overlie those to the west. This is consistent with the interpretations of Fisher and Karabinos (1980) in which they conclude that the Gile Mountain Formation overlies the Waits River Formation.



Pleasant Valley Cem
SPRINGFIELD
MOUNTAIN
RIVER
PLEASANT VALLEY RD
LOVELL RD
WINDSOR CO
WINDHAM CO
Telephone
HAM TOWN FOREST
Bartonsville
103
Covered Br
Brook
Wright
Brockways Mills
C-1
MOUNTAIN
WILLIAMS
RIVER
GREEN
BROCKWAYS MILLS
Lily Brook
JOHN J DORANO STATE FOREST
Stream
Brook
4WD
Macmaster
428T
289T
142T
152T

STOP C-2: THE CHESTER SOAPSTONE QUARRY

Description modified from Thompson, Cheney, and Robinson (1986).

The ultramafic body exposed here occurs in the Cambrian Ottauquechee Formation in the outer mantle of the Chester Dome. This stop provides an opportunity to examine and collect material from one of the classic blackwall sequences described in the pioneering work of Hess (1933) and Phillips and Hess (1936) on the formation of serpentinite and the origin of the blackwalls. The metasomatic reaction zone at this quarry at the original boundary between serpentinite and country rock is also the source of four new biopyriboles described by Veblen (1976), Veblen and Burnham (1975), Veblen, Buseck, and Burnham (1977), Veblen and Burnham (1978), Veblen, Buseck, and Burnham (1978a, 1978b).

A house has been constructed on the edge of the Vermont Mineral Products Quarry. Permission from the owners should be obtained before visiting this site. Walk to the northeast end of the quarry to the only remaining outcrop of serpentinite. Proceed around the eastern rim of the pond, passing on the west side of the house, to an exposure of the blackwall, south of the pond outlet stream. Finally continue east to the dump where fun should be had by all.

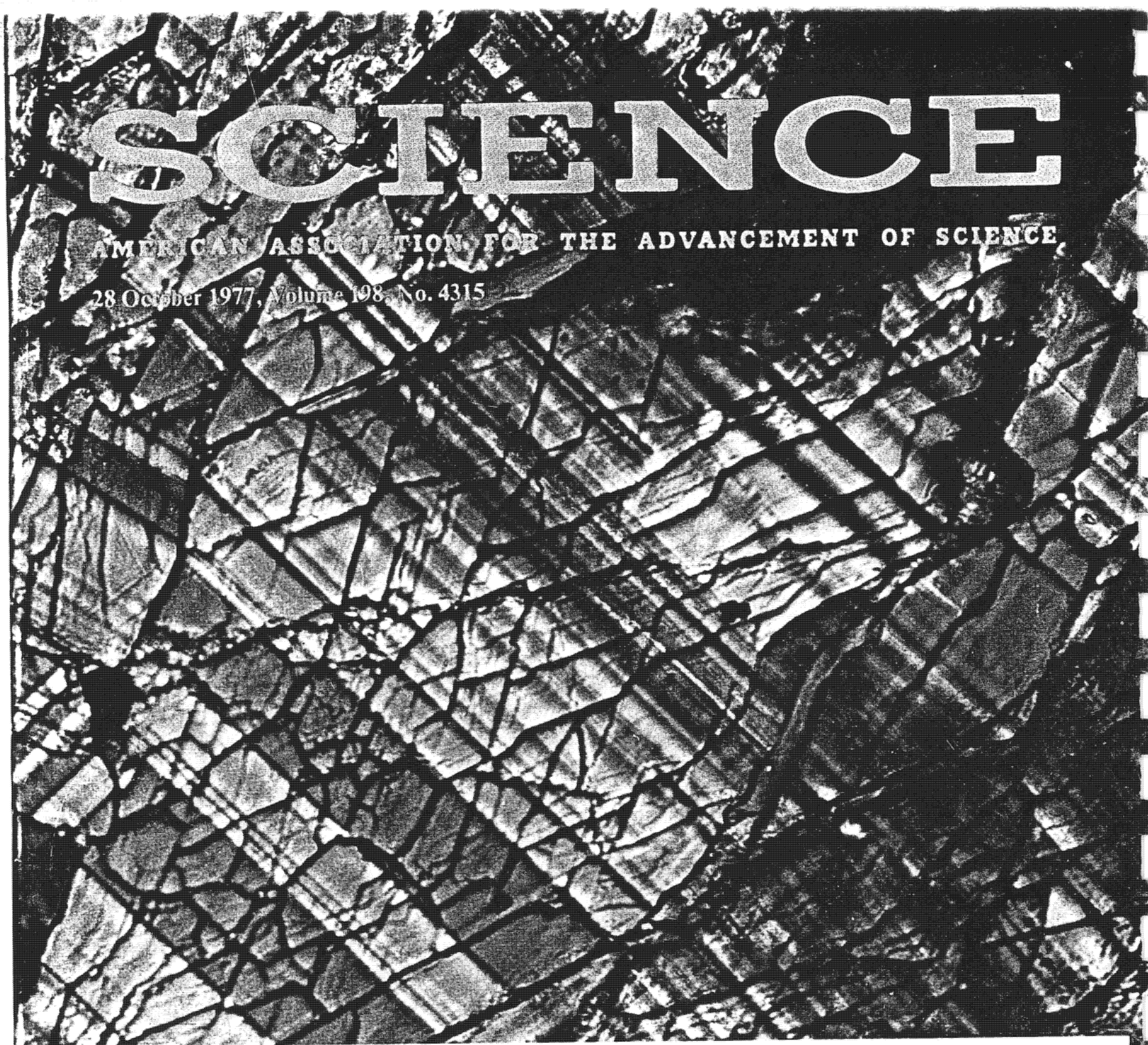
Geologic Setting of New England Ultramafic Bodies

The Chester ultramafic bodies occur in a belt of alpine peridotites extending, along the axis of the Appalachian orogenic belt, from Alabama to Newfoundland. As described by Sanford (1982), the ultramafic bodies occur in virtually all pre-Silurian units, including remobilized Precambrian gneisses. Accordingly, they are generally considered to have been emplaced in the Ordovician. In order of decreasing abundance, the host rocks include: greenschist, amphibolite, pelites, metasandstones, quartzite, and marble. The ultramafic rocks of New England and Quebec were likely emplaced in the solid state at low temperature into, at best, weakly metamorphosed sediments near the onset of the Taconian orogeny. Their emplacement is probably related to the obduction of a young ophiolite sequence onto the North American margin.

Although the timing of serpentinization is much debated, most bodies in the lowest grade sediments are highly serpentinized. Hence serpentinization clearly preceded the formation of the metasomatic reaction zones, which developed during the Acadian orogeny.

Metasomatic Reaction Zones

As described by Hess (1933), "practically the whole mass of the ultrabasic has been altered to talc and carbonate, leaving only occasional serpentine relics. A sheath of biotite schist or biotite actinolite schist from several inches to a few feet thick completely surrounds the deposit. This sheath represents approximately the original contact of the ultrabasic rock with the country rock." Hess added in a footnote that the biotite rock at the original interface with the country rock is called "blackwall" as are the chlorite rocks which surround the normal (lower temperature) type of talc deposits. Although not well exposed at the time, Phillips and Hess managed to describe quite accurately the reaction zone in their Figure 5 (Figure C-5) and now observable on the south side of the pond, beyond the outlet. Their more detailed description implies that locally, at least within a "horse" of altered country rock, no longer present, within the soapstone, a zone of chlorite separated a tourmaline-rich biotite rock and the actinolite. This chlorite zone is probably the source of the magnetite- (up to 1.5 cm) and pyrite- (up to 3 cm) chlorite rocks common in many museums. With luck, samples representative of the various reaction zones can be obtained from



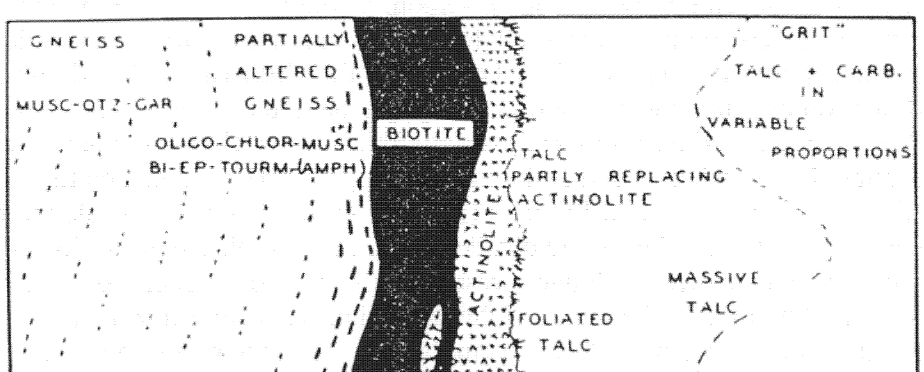


Figure C-5. Field sketch of mineral zoing at the Vermont Mineral Products Co. quarry, near Chester, VT. From Phillips and Hess (1936).

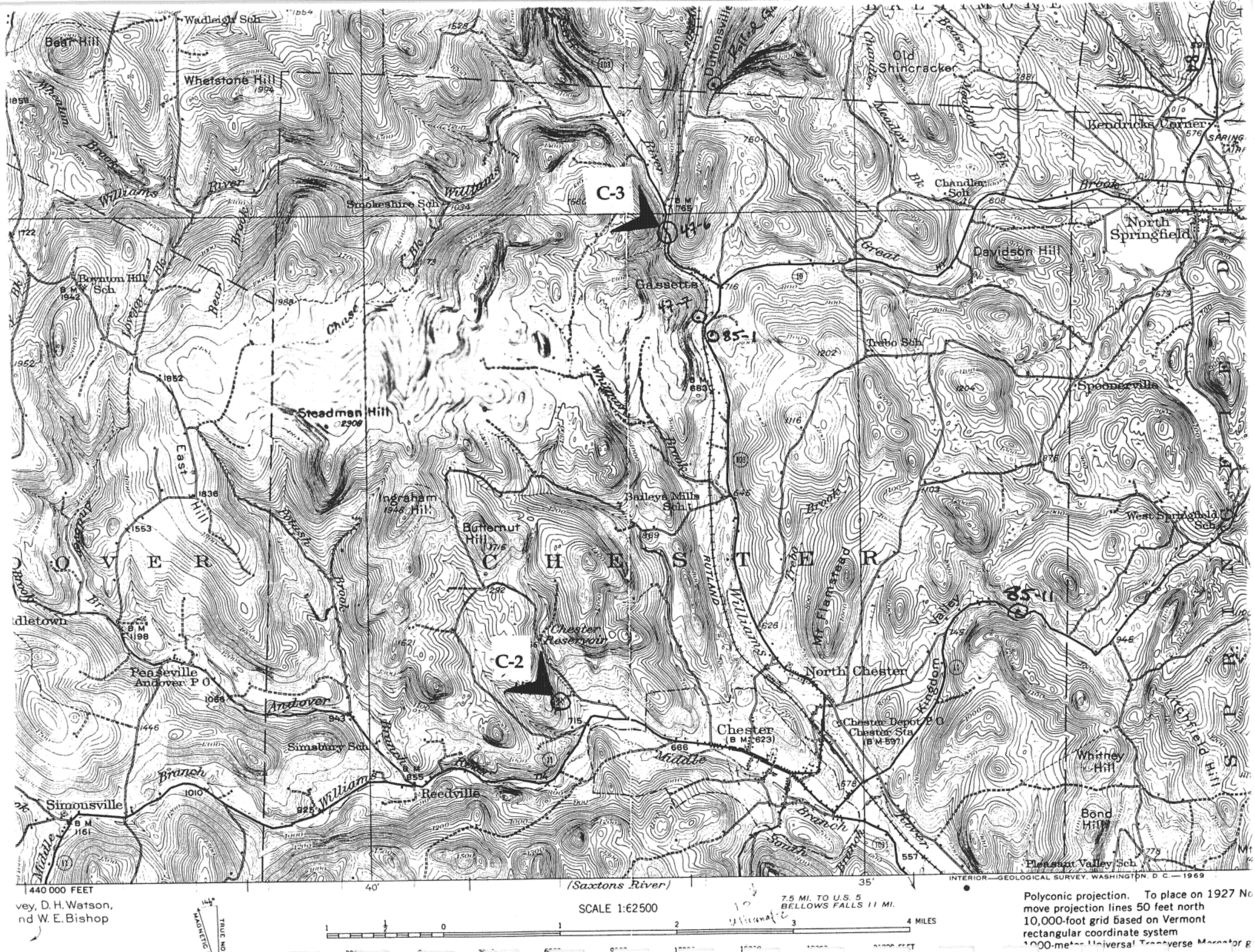
the dump south of the house. However, samples of the pyrite- and magnetite-bearing chlorite rocks have long been scarce and seem to have come from the dump now serving as a foundation for the house.

The metasomatic nature of the reaction zones was surmised by Phillips and Hess (1936) who attributed their formation to the mass transfer primarily of silica into the ultramafic and magnesia and iron into the country rock. They also noted the apparent gain of alumina and water by the "blackwall". More recently, Brady (1977) has argued on a theoretical basis that the monomineralic nature of the reaction zones is a consequence of diffusion-imposed chemical potential gradients of several components, principally silica and magnesia. He also concluded that the talc + magnesite zone forms in response to a diffusion-imposed gradient in the chemical potential of CO_2 and that the original host-serpentinite contact is most likely near the interface between the biotite (or chlorite, if present) and actinolite zones (compare with Figure C-5). These conclusions have been largely substantiated, and quantified, by Sanford's (1982) detailed microprobe/field study of four other quarries in Vermont and Massachusetts. In addition Sanford provided an exceedingly detailed description of textural and mineral compositional variations in typical reaction zones as a function of metamorphic grade.

Triple-Chain Silicates

Although David Veblen (personal communication, June 27, 1986) claims to have lugged all appropriate material back to Harvard's Peabody Museum, it may be possible to obtain samples of material containing the four Mg-Fe chain silicates first identified by Veblen and Burnham (1975, 1978, and references above) in samples from the south (and only remaining) dump. These new biopyriboles occur with anthophyllite, talc and cummingtonite, between the chlorite and actinolite zones. The new minerals include the triple chain silicates (b=27Å) jimthompsonite (orthorhombic, space group *Pbca*) and clinojimthompsonite (C2/*c*), and chesterite, the orthorhombic mineral (space group $A2_1ma$) containing mixed double and triple chains (b=45Å). An unnamed monoclinic mineral (space group A2/*m*, A*m*, or A2) is presumed to be the analog of chesterite. However, the occurrence of this mineral as very fine grained intergrowths in chesterite has precluded its confirmation. The new minerals are chemically and structurally intermediate between talc and anthophyllite. The ideal composition for jimthompsonite and clinojimthompsonite is $(Mg, Fe)_{10}Si_{12}O_{32}(OH)_4$, and that of chesterite is $(Mg, Fe)_{17}Si_{20}O_{54}(OH)_6$. Although the physical and optical properties of these minerals are close to those of the Fe-Mg amphiboles, the cleavage angles (37.8° and 44.8°) are distinctive. Because these minerals commonly occur in (010) intergrowths, crystals with b near the plane of the section have a spectacular appearance under crossed polars (see the color photograph on the Oct. 28, 1977 cover of *Science*).

In detail, all of the material used by Veblen and co-workers is from a single block of blackwall found on the quarry dump. Similar material may have been reported in place on the northern wall of the quarry. The generalized zoning sequence is chlorite / fibrous talc / jimthompsonite + clinojimthompsonite + chesterite + the unnamed mineral + anthophyllite + cummingtonite / anthophyllite + cummingtonite + chesterite + the unnamed mineral + actinolite / actinolite + massive talc. Because anthophyllite has been extensively replaced by fibrous talc plus all four of the new minerals, they are considered part of a retrograde reaction sequence from anthophyllite to talc.


C-3
C-2
47-6
47-7
85-1
85-11
Wadleigh Sch
Bear Hill
Whetstone Hill
Old Shincracker
Kendricks Corner
Chandler Sch
North Springfield
Davidson Hill
Smokeshire Sch
Williams River
Boynton Hill Sch
Gassetts
Trebo Sch
Spoonerville
Steadman Hill
Ingraham Hill
Butternut Hill
Baileys Mills Sch
West Springfield Sch
O V E R
C H E S T E R
Chester Reservoir
Peaseville Andover P O
North Chester
Mt Flamstead
Chester
Chester Depot P O
Chester Sta
Simsbury Sch
Reedville
Simonsville
Whitney Hill
Bond Hill
Pleasant Valley Sch
Litchfield Hill
(Saxtons River)
INTERIOR—GEOLOGICAL SURVEY, WASHINGTON, D. C.—1969
440 000 FEET
vey, D. H. Watson,
nd W. E. Bishop
SCALE 1:62500
7.5 MI. TO U.S. 5
BELLOWS FALLS 11 MI.
4 MILES
Polyconic projection. To place on 1927 No
move projection lines 50 feet north
10,000-foot grid based on Vermont
rectangular coordinate system

STOP C-3: THE WILLIAMS QUARRY AT GASSETTS, VERMONT: CALC-SILICATE REACTION ZONES

Description modified from Thompson, Cheney, and Robinson (1986). This quarry is currently owned and operated by All Stone Corp., PO Box 440, Ascutney, VT 05030 (802-674-2371). Permission to visit must be obtained from the owners.

The interbedded calc-silicate rocks and phengite-bearing quartzo-feldspathic gneisses exposed in the Williams quarry are from the upper part of the Proterozoic Z to Cambrian Tyson Formation. These rocks together with the disconformably overlying high-alumina pelites of the Hoosac Formation belong to the inner mantle sequence of the Chester dome. Here the inner mantle rocks occur along the western side of an infold of Lower Paleozoic rocks into the Grenville gneisses that form the core of the dome. This crescent-shaped septum is similar in many respects to Alpine mulde (Figure C-2). In fact, the phengite gneisses exposed in the quarry are very similar to the mulde gneisses quarried extensively at Val Maggia, Switzerland. However, these gneisses cannot be distinguished from the basement gneisses because they are completely sandwiched between calc-silicates of the Tyson Formation.

A.B. Thompson (1975) has described in detail the nature and reaction history of the calc-silicates occurring in the thinly laminated carbonate-pelite sequence of the Tyson Formation. According to Thompson (1975), the original mineralogy of the calc-mica schists has been modified due to the development of reaction zones. Thicker "pelitic" layers of the schists resemble the enclosing gneiss and consist of quartz (55%), alkali feldspar (10%), plagioclase (5%), phengite (10%), biotite (15%), and accessories (5%) including epidote, calcite, titanite, apatite, and rare graphite and pyrrhotite. The marble layers consist primarily of coarse calcite with irregular diopside, commonly adjacent to rounded quartz grains, concentrated in their centers. Less com-monly, small actinolite crystals, also in contact with rounded quartz, occur within the recrystallyzed calcite. The occurrence of the diopside and actinolite in the coarse calcite suggests that subordinate amounts of dolomite, relative to quartz and calcite, were present in the protolith. Dolomite in various associations with calcite, actinolite, diopside, biotite, microcline, and quartz occurs in the Tyson carbonate units elsewhere in the Chester dome (*e.g.* Cavendish Gorge). The reaction zones between the "pelite" and the metacarbonate contain a variety of associations of biotite, cilnozoisite, microcline, actinolite, diopside, calcite and quartz. Minor amounts of titanite, apatite, graphite, pyrrhotite and chalcopyrite also occur in these zones. In detail the nature of the zones depends upon the relative thickness of the carbonate and pelite layers, and perhaps upon the extent of initial gradation between lithologies. A characteristic intergrowth of clinozoisite and biotite, commonly with microcline and quartz, occurs between thicker pelite layers and thin carbonate layers. The minerals comprising the intergrowth are generally randomly oriented but confined to distinct layers. The intergrowths characterizing reaction zones between thin pelitic and thicker carbonate layers consist of coarse diopside (0.5 cm), containing inclusions of actinolite and calcite, and randomly oriented blades of actinolite. The bladed actinolite contains inclusions of diopside and calcite.

A complex reaction history, involving the elimination of dolomite from the carbonates and "muscovite" from the pelitic interlayers, has been developed by Thompson (1975) to account for the many textural and compositional variations of these rocks many of which are not reviewed here. Although this reaction history will not be reviewed, it is interesting to consider that many aspects of the reaction zones between the thicker pelite and carbonate layers, characterized by the zoisite-biotite intergrowths, may result from the diffusion of primarily calcium. The possibility of mass transfer has been considered by Thompson (1975) and discussed in some detail by Brady (1977) for similar rocks in northern Vermont.

STOP C-4: THE GASSETTS SCHIST

Description modified from Thompson Cheney, and Robinson (1986).

The schist at Gassetts, as well as at other localities in the Chester dome area, such as Black River, Hawks Mountain, and Star Hill contains kyanite-biotite zone mineral assemblages that reflect a wide range in alumina content. The detailed study of these rocks has provided a nearly complete description of of AKNa and AKFM Facies Types (Thompson, 1957, 1961; Thompson and Thompson, 1976). Moreover, the detailed consideration of components extraneous to the AKFM projection has permitted mapping of the AFM plane with respect to saturating CaO, Na_2O, Fe_2O_3, and TiO_2 phases (Thompson, 1972).

Two size fractions of white mica occur here, as originally described at Glebe Mountain, Vermont by Rosenfeld (1956) and subsequently observed world-wide, including Pizzo Forno, Switzerland. The fine-grained mica that you cannot visually resolve into discrete flakes in hand specimen is paragonite; the coarser grained mica is muscovite. The composition gap between gravimetric analyses of these micas (from ~Pg_{38} to Pg_{83} in the muscovite and paragonite respectively, see Table C-1) is as narrow as observed anywhere. However, repeated microprobe analyses of muscovite from this locality are consistently lower in Na and Ca than the gravimetric analyses (compare Tables 1 and 2 in Thompson, Lyttle, and Thompson, 1977). This analytical discrepancy may reflect "perthitic" exsolution of paragonite in the muscovite. Hence the gravimetric analyses may provide an integrated composition reflecting the maximum metamorphic conditions whereas the microprobe may be measuring cooler, retrograde compositions.

A third white mica, margarite, also occurs in these rocks. Margarite occurs only as inclusions in garnet here (see also Downie, 1982), however, matrix margarite has been reported in similar rocks from western Massachusetts (Cheney, 1980).

Although not observed at this locality, similar high-alumina rocks in the vicinity of Star Hill have zoned muscovite porphyroblasts (phengitic cores and "normal" rims) that have overgrown matrix phengite in the foliation (Downie, 1982). These complex white-mica assemblages also occur in the Gassetts-like pelites of the Hoosac Formation in western Massachesetts (Cheney, 1980, 1986). Cheney (1986) has shown that the evaluation of these assemblages, utilizing the methods outlined by Thompson (1980, 1982a, 1982b) and Spear and Selverstone (1983), may be useful in constraining the P-T paths of the rocks in which they occur.

Mineral Assemblages

The long road cut on the east side of the Williams River contains a continuous sequence of Gassetts Schist, a part of the Hoosac Formation. The high alumina assemblage containing kyanite + biotite + muscovite + paragonite + staurolite + garnet + quartz with titanohematite and minor epidote, apatite and rare calcite, occurs at the north end of the outcrop, although the best samples are usually found in the rubble across the road. The alumina content of the rocks generally decreases to the south along the cut and the southernmost portion of the road cut contains the assemblage quartz + plagioclase + muscovite + biotite + garnet ± calcite, tourmaline, and magnetite - with or without graphite.

South of this road cut, after a covered interval at the position of the upper Tyson carbonates are exposures of gneisses that are metamorphosed graywackes of the Tyson Formation and/or strongly modified core gneisses of the Chester Dome. These rocks contain quartz + microcline +

plagioclase + biotite + muscovite, with minor calcite, epidote, sphene, tourmaline, apatite, and less common pyrite, graphite and magnetite. Of interest is that garnet does not coexist with microcline in the presence of biotite +magnetite + graphite.

In the middle of the road cut, the following muscovite + quartz- bearing assemblages occur sequentially from south to north: plagioclase + paragonite + staurolite + garnet + biotite; paragonite + staurolite + garnet + biotite; kyanite + paragonite + staurolite + garnet + biotite. Kyanite and plagioclase have not yet been observed together in any sample. Moreover, both hematite and magnetite, although extensively altered and exsolved, occur in the AKFM assemblage muscovite + paragonite + staurolite + garnet + biotite + quartz. Chlorite occurs in many of these assemblages and may be primary where interlayered with biotite and muscovite in the groundmass. More commonly, however, chlorite occurs as selvages around garnet or intergrown with biotite. This variety of mineral assemblages allows the nearly complete description of kyanite + biotite zone AKNa and AFM facies types as well as the mapping of the AFM plane with respect to saturating Na_2O, Fe_2O_3, and TiO_2 phases (Thompson, 1972, as shown on Figure C-6). As developed by Thompson (1979), and shown in Figure C-7 the muscovite coexisting with various three-phase AFM assemblages will have a specified amount of phengite component and the muscovite becomes more phengitic in three-phase AFM assemblages containing progressively less-aluminous phases.

AFM Phase Relations

In the high-alumina pelite, more than three AFM phases (garnet-staurolite-biotite-kyanite) apparently coexist with muscovite + paragonite + quartz and possibly primary chlorite. This suggests as do the representative analyses of Table C-2 and the crossing tie lines and ranges in the compositions of the minerals, plotted in Figure C-8 that additional components, *e.g.* MnO, CaO, and/or Fe_2O_3 stabilize one or more of the AFM phases, and/or that local assemblages represented by three-phase triangles reflect equilibration at a range of metamorphic conditions.

The utilization of garnet-rim compositions in Figure C-8 assumes that only the outermost part of the garnet was involved in the last recorded reactions. The relatively high modal ratios of garnet to biotite and and of staurolite to biotite enhances the possibility that biotite compositions may have been altered during cooling by exchange reactions (Tracy *et al.*, 1976) The garnets are "nomally" zoned with Ca- and Mn-enriched cores (up to 17 mole % grossular and 5.5 mole % spessartine). Detailed microprobe analyses of core garnet, discussed more fully below, show that Mg increases and that Ca and Mn decrease from core to rim, whereas Fe is distributed in an irregular manner. The staurolites from different assemblages have different alumina contents and contain up to 2.25 wt% ZnO (see Thompson, Lyttle, and Thompson, 1977, Table 4). In biotite both alumina and Fe/(Fe+Mg) contents depend upon the compositions of adjacent phases. In particular, where associated with chlorite, the biotite is distinctly more iron-rich than where adjacent to staurolite and/or kyanite. In fact, these chlorite + biotite tie lines cross staurolite + biotite tie lines for pairs from both the high-intermediate-alumina schists. Moreover, the tie lines between coexisting chlorite and biotite, associated with anhedral staurolite cross the tie lines between kyanite and biotite (Figure C-8). These phenomena may be due to retrograde reactions or a very small scale of prograde equilibration, For example, the biotite from the three phase assemblages of garnet-staurolite-biotite and staurolite-kyanite-biotite differs by 6% in Fe/(Fe+Mg) for phases within a distance of one centimeter. Thus the scale of equilibration is extremely limited, even within single microprobe samples, and different, although compatible, three-phase AFM assemblages may apparently occur on a centimeter scale.

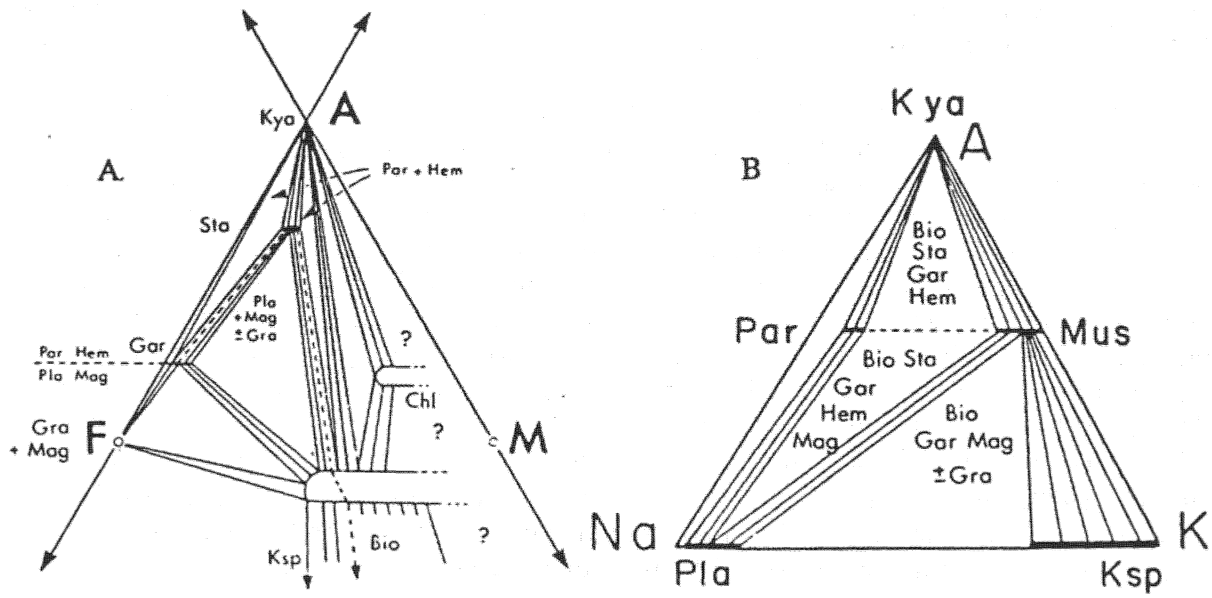


Figure C -6. (A) Schematic AFM projection of the observed facies type showing distribution of saturating phases. (B) Schematic ANaK diagram for the observed facies type. The associations of AFM and saturating ferric phases or graphite are shown. From Thompson, Lyttle and Thompson (1977).

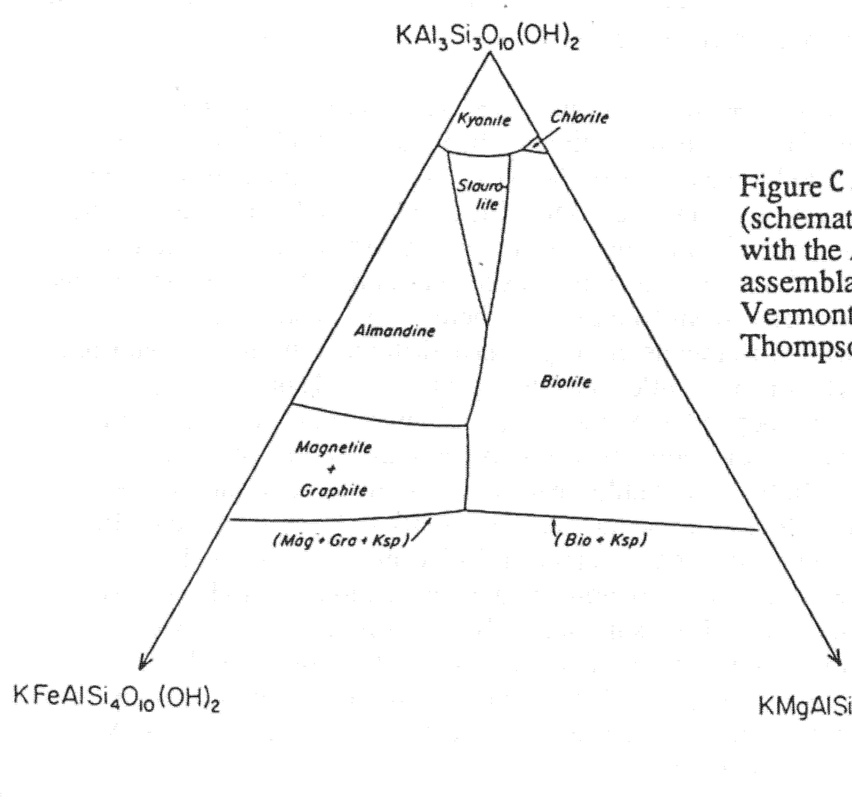


Figure C -7. Muscovites (schematic) coexisting with the AFM assemblages at Gassetts, Vermont, U.S.A. From Thompson (1979).

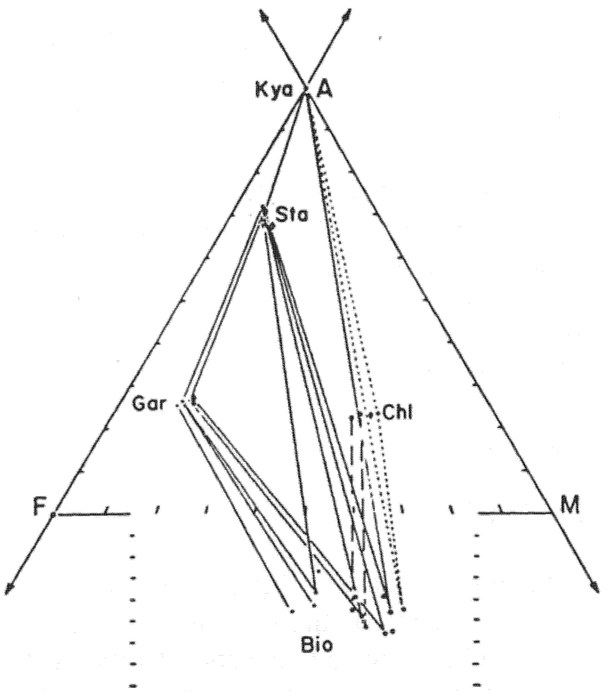


Figure C-8. AFM projection from KAl_3O_5, SiO_2, and H_2O showing gravimetric analyses (open diamonds) and microprobe analyses (other symbols) of minerals from the Gassetts Schist. The other symbols re - present analyses of AFM phases coexisting with quartz and ANaK phases as follows: solid circles with muscovite-paragonite-kyanite, open squares with muscovite-paragonite only, solid triangles with muscovite-paragonite-plagioclase, crosses with muscovite-plagioclase only. Dotted lines connect compositions of the possible prograde three-phase assemblage chlorite-biotite-kyanite. Dashed lines connect comp - ositions of possible retrograde chlorite and biotite, usually associated with irregular staurolite. From Thompson, Lyttle and Thompson (1977).

Table C-1. Gravimetric mineral analyses from Gassetts, Vermont. See Thompson, Lyttle and Thompson (1977) for sources of data. Parag Corrected = Paragonite corrected for muscovite.

	MUSC	PARAG	PARAG CORRECTED FOR MUSC	TOURM	STAR	STAR	GARNET	CHTE RIM ON GARNET
SiO2	48.30	45.18	44.51	35.63	28.73	28.22	37.63	24.99
TiO2	0.24	0.41	0.49	0.62	1.39	0.61	0.18	0.00
Al2O3	32.98	39.19	40.31	31.35	49.89	49.86	21.38	22.25
Fe2O3	2.33	0.96	0.70	0.00	4.16	4.00	0.54	3.92
FeO	0.27	0.29	0.30	7.47	10.88	11.44	32.04	16.69
MnO	0.00	0.00	0.00	0.00	0.12	0.06	0.95	0.03
MgO	0.30	0.10	0.10	7.90	2.99	2.73	4.53	19.67
CaO	0.13	0.96	1.10	0.85	0.00	0.00	2.75	0.00
Na2O	3.11	5.57	6.00	1.92	0.01	0.00	0.00	0.00
K2O	7.46	2.16	1.20	0.08	0.02	0.00	0.00	0.04
Be2O3				10.28				
P2O5				.06				
H2O -					0.22	0.90	0.17	0.52
H2O +	4.72	5.22	5.3	3.84	1.53	2.21	0.30	12.31
Total	99.85	100.06	100.0	100.0	99.94	100.03	100.48	100.26
ANHYDROUS FORMULAS								
#OX/FORMULA	11	11	11	29	23	23	12	14
Si	3.191	2.931	2.880	5.88	3.978	3.965	2.986	2.553
Ti	0.012	0.020	0.024	0.08	0.145	0.064	0.011	0.000
Al	2.568	2.996	3.074	6.10	8.141	8.256	1.999	2.679
Fe3+	0.116	0.047	0.034	0.00	0.433	0.423	0.032	0.301
Fe2+	0.015	0.016	0.016	1.03	1.260	1.344	2.126	1.426
Mn	0.000	0.000	0.000	0.00	0.014	0.007	0.064	0.003
Mg	0.030	0.010	0.010	1.94	0.617	0.572	0.536	2.995
Ca	0.009	0.067	0.076	0.15	0.000	0.000	0.234	0.000
Na	0.398	0.701	0.753	0.61	0.003	0.000	0.000	0.000
K	0.629	0.179	0.099	0.02	0.004	0.000	0.000	0.005
Be				2.19				
P				.05				
ATOM RATIOS								
Fe2/Fe2+Mg	0.333	0.636	0.615	0.347	0.671	0.701	0.799	0.323
Fe3/Fe2+Fe3	0.885	0.741	0.680	0.000	0.256	0.239	0.015	0.174
Na/Na+Ca+K	0.38	0.74	0.81					

The crossing tie lines and textural relationships between chlorite + garnet (*e.g.* chlorite rims) and chlorite + biotite + staurolite can be partially, but not completely, reconciled in the context of local variations in a_{H_2O}, at constant T and P as shown by Figure C-9.

As illustrated by Figure C-10, the assemblages garnet + staurolite + biotite, staurolite + biotite + kyanite, garnet + staurolite + kyanite, and chlorite + biotite + kyanite indicate P-T-a_{H_2O} conditions below the (Fe-Mg) and (Na-K) discontinuous reaction(s):

Staurolite + Muscovite = Garnet + Biotite + Kyanite
Paragonite + Quartz = Muscovite + Albite + Kyanite + H_2O

respecitvely, and above the Fe-Mg discontinuous reaction

Staurolite + Chlorite + Muscovite = Biotite + Kyanite.

Additional components or decreased a_{H_2O} will displace these Fe-Mg reactions in a predictable manner as discussed by A.B. Thompson (1976a, 1976b). In this context a variety of commonly used geothermometric and geobarometric techniques, including a consideration of various experimental and theoretical continuous and discontinuous reactions, oxygen and carbon isotone fractionation, piezoelastic effects of quartz inclusions in garnet, and various solvi (muscovite-paragonite, ilmenite-hematite and calcite-dolomite), yield a large temperature range of 360 °C to 740 °C and pressures of 4-6 kbar. Most of the temperature estimates fall between 535 °C and 620 °C at 6 kbar which agrees well with peak P-T estimates of 600± 20 °C (garnet-biotite geothermometry) and 6± 1 kbar (plagioclase-garnet-kyanite-quartz geobarometry) reported in the Cavendish area (Star Hill-Hawkes Mountain) for similar high alumina schists by Downie (1980, 1982).

Compositional Zoning and Mineral Inclusions in Garnet

A more complete story of the metamorphism of the Gassetts Schist, particularly of its earlier stages, was worked out by a detailed study of the zoned garnets and their included minerals by Thompson, Tracy, Lyttle, and Thompson (1977) from which the following information is abstacted.

Petrographic examination of the solid inclusions in garnet showed the presence of at least two white micas, staurolite, chloritoid, chlorite, epidote, hematite, ilmenite, magnetite, rutile, and quartz. No biotite inclusions were observed, and possible kyanite inclusions were observed only at the extreme rim of the garnet. Subsequent analyses of successive garnet zones and of solid inclusions revealed a systematic distribution.

Detailed microprobe traverses across the garnet permitted contouring of the two-dimensional section in cation percentage of Ca, Mn, Mg, and Fe. Gravimetric analysis of a garnet from another sample of Gassetts Schist (by Jun Ito) indicates less than 1.6 mole % $Ca_3Fe_2^{3+}Si_3O_{12}$ as the only ferric component (see J.B. Thompson, 1957, p.854; Thompson, Lyttle, and Thompson, 1977, Table 1). This is likely to be a maximum value because of abundant ilmenohematite inclusions. Representative microprobe analyses (corresponding to the section along profile C-D in Figure C-11) are given in Table C-2. As can be seen from these analyses, the

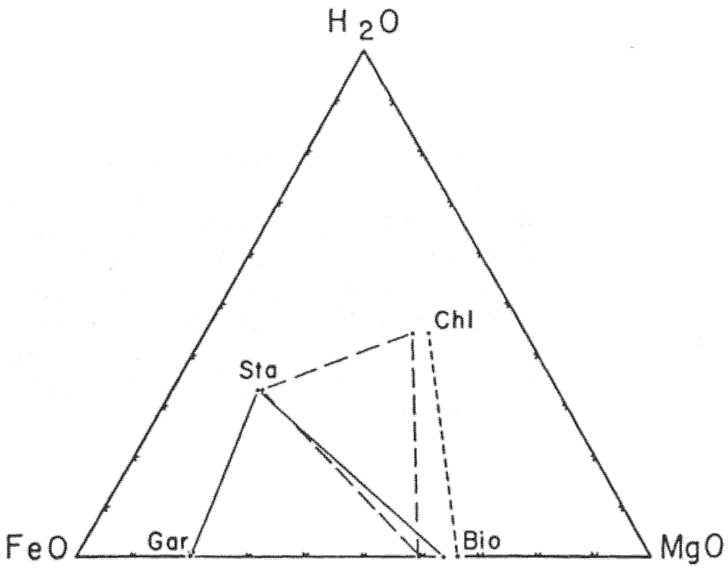


Figure C-9. Projection of the microprobe analyses of phases from the high-alumina pelite onto the plane FeO-MgO-H_2O from $H_4K_2O_3$, Al_2O_3, and SiO_2 (see also Rumble, 1974, p. 374). Solid lines connect compositions of prograde garnet-staurolite-biotite. Dotted lines connect compositions of possible prograde chlorite-biotite (with kyanite). Dashed lines connect compositions of possible retrograde chlorite-biotite-irregular staurolite. From Thompson, Lyttle and Thompson (1977).

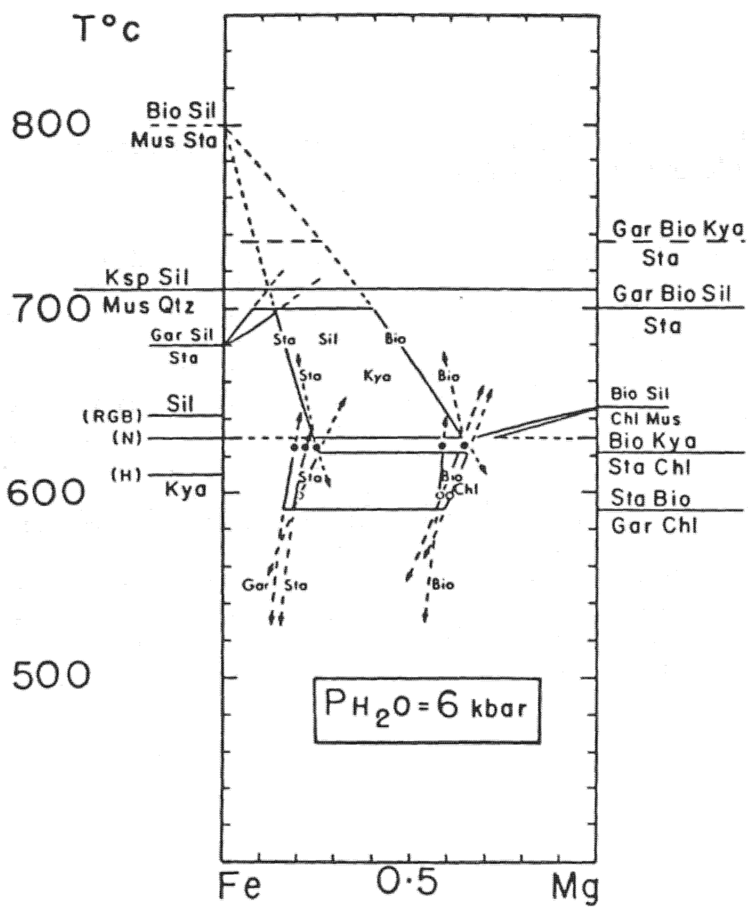


Figure C-10. Approximate T - X(Fe-Mg) projection calculated for $P_{total} = PH_2O =$ 6 kbar for some of the contin - uous and discontinuous reac - tions. Kyanite-sillimanite data are shown from the studies by Newton (1966, N), Richardson, Gilbert and Bell (1969, RGB), and Holdaway (1971, H). Solid circles show the compositions of the prograde assemblages at Gassetts, and open circles show the compositions of possible retrograde assem - blages (see Figure 6). From Thompson, Lyttle and Thompson (1977).

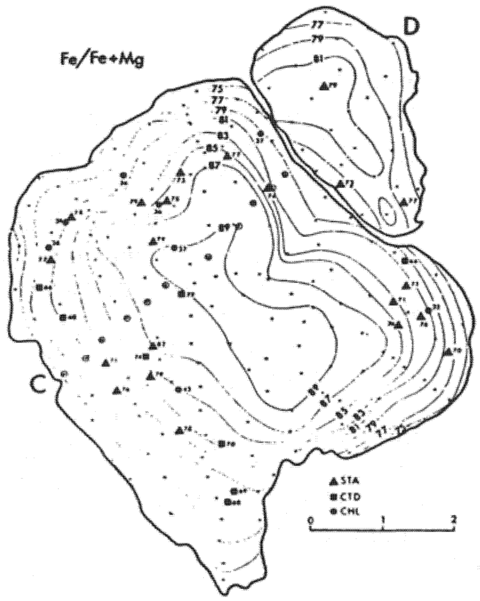


Figure C-11. Contours for X_{Fe} = [100 Fe/(Fe + Mg)] in garnet, together with the same atomic percentage for staurolite, chloritoid, and chlorite inclusions. The spot locations corresponding to the garnet analyses in Table 1 are shown by circled spots along profile C-D. Scale bar is 2 mm long. From Thompson, Tracy, Lyttle and Thompson (1977).

Table C-2. Representative garnet analyses along profile C-D (see circled spot locations in Figure C-11.). From Thompson, Tracy, Lyttle and Thompson (1977).

	G-60	G-59	G-58	G-57	G-56	G-55	G-14	G-45	G-46	G-47	G-48
SiO_2	35.70	36.39	36.54	37.02	37.79	37.46	35.85	37.10	37.89	38.08	37.65
TiO_2	0.04	0.09	0.08	0.07	0.08	0.10	0.06	0.08	0.06	0.02	0.08
Al_2O_3	21.38	20.99	20.72	20.43	20.30	20.44	20.62	20.45	20.43	20.84	20.65
Cr_2O_3	0.02	0.02	0	0	0.04	0.12	0.05	0	0.04	0.03	0
FeO	33.94	33.53	32.74	32.24	32.40	33.29	32.13	33.31	32.76	33.34	34.74
MnO	0.39	0.66	0.94	2.36	1.83	2.07	2.53	2.05	1.86	0.91	0.82
MgO	5.46	4.69	3.60	2.89	2.43	2.47	2.03	2.31	2.54	4.25	4.78
CaO	2.17	3.22	4.80	5.05	5.65	5.32	6.15	5.56	5.49	3.91	2.10
Total	99.08	99.59	99.42	100.06	100.52	101.27	99.42	100.86	101.07	101.38	100.82
Mol% Am	72.5	71 8	70.8	69.7	70.7	71.4	69.6	71.4	70.9	71.2	74.2
Mol% Py	20.8	17.9	13.9	11.1	9.4	9.5	7.8	8.8	9.8	16.2	18.2
Mol% Gr	5.9	8 9	13.3	14.0	15.8	14.6	17.1	15.3	15.2	10.7	5.8
Mol% Sp	0.8	1 4	2.0	5.2	4.1	4.5	5.5	4.5	4.1	1.9	1.8

garnet contains negligable Ti and all iron is considered to be ferrous. The complex distribution of calcium, manganese, and iron in these garnets clearly illustrates the inadequacy of *single radial profiles* across garnet grains for use in deducing compositional histories. The pronounced maxima and minima in cation concentration show an internal structure that may reflect mineralogical differences in initial bedding subsequently rotated during porphyroblast growth. Rosenfeld (1970, sec. 5, p. 25-50) illustrates examples of the deformation of bedding surfaces in rotated garnets. Some of our contours are consistent with the case where the rotation axis is approximately parallel to the plane of the section. Garnets from the Gassetts Schist (Thompson, 1979) and other units nearby (Rosenfeld, 1968, fig. 14-5) have inclusion spirals showing strong rotation during growth. The nature and composition of the solid inclusions bear a systematic relationship to local garnet composition.

As noted above, the macrocrystalline phases define an AFM facies-type characterized by garnet - staurolite - biotite - kyanite (and possible primary chlorite) all with quartz. Local associations of staurolite -garnet - biotite, staurolite - biotite - kyanite, staurolite - garnet - kyanite, and possibly chlorite - biotite - kyanite are observed. The ANaK facies-type at this grade is characterized by muscovite - plagioclase - potassic feldspar.

The presence of chloritoid, margarite, and rutile only as inclusions in the garnet implies that progressive reactions have eliminated these phases from the present "equilibrium" assemblage. The location and compositions of the inclusions relative to adjacent garnet composition and overall garnet zonation may be used to determine the successive continuous and discontinuous reactions undergone by the sample during the regional metamorphism (see also Tracy *et al.*, 1976).

A progressive series of AFM facies types has been deduced from the relative locations and compositions of chlorite, chloritiod, and staurolite inclusion within the garnet (see Figure C-12). The occurrence of chlorite (X_{Fe} = .37) and chloritoid (X_{Fe} = .77) near the core of the garnet (X_{Fe} = .90) suggests the presence of the three-phase assemblage garnet - chlorite - chloritoid at lower grades. Other possible three-phase assemblages at this grade are also shown in Figure C-12. It is possible that the lowest grade facies type recorded by the inclusions is not that of Figure C-12a but that shown in Figure C-12b, where the staurolite may have been part of the adjacent macrocrystalline assemblages but could not coexist with garnet because of possible chloritoid solution to pure Fe-chloritoid.

<u>Implications of reaction history</u>. The so-called "staurolite - in" reaction, separating topologies e and f of Figure C-12:

Garnet + Chlorite = Staurolite + Biotite

resorbs garnet and, if followed by an increase in metamorphic grade, could produce "garnet unconformities" like those described by Rosenfeld (1968). However, depending upon the amount of resorption and the effective volume of equilibration of the garnet, this resorption may be difficult to recognize due to continued garnet growth with increase of metamorphic grade (see Figure 6 of Thompson, Tracey, Lyttle, and Thompson, 1977).

The garnet and its inclusions of chloritoid and staurolite contain measurable amounts of MnO. Manganese stabilizes the four-phase assemblage garnet - chloritoid - staurolite - chlorite to P-T-aH_2O conditions above the AFM reaction:

Chloritoid = Garnet + Staurolite + Chlorite.

With the addition of MnO, this Fe-Mg discontinuous reaction becomes a continuous AFM-Mn reaction that can eliminate chloritoid from the matrix.

Finally, several of the remarkable insights into the reaction history of high-alumina pelites gleaned from the detailed study of one garnet from the Gassetts Schist have subsequently been confirmed by workers in nearby areas. Downie (1982), employing similar methods applied to several samples, has deduced virtually the same reaction sequence for both the Gassetts Schist and Pinney Hollow Formation at the north end of the Chester dome. Cheney (1980, 1986) has also documented a comparable reaction sequence in a continuous belt of high-alumina, Gassetts-like schists from the Hoosac Formation of Western Massachusetts, that range from garnet zone though kyanite-biotite zone. In addition, Karabinos' (1985) detailed work on the chloritoid to staurolite zone transition in Gassetts-like pelites near Jamaica, Vermont has confirmed and expanded upon the AFM-Mn reaction for the elimination of chloritoid as a matrix phase. However, the significance of the textural unconformities observed in garnets from locations thoughout southeastern Vermont remains problematic and much debated. There seem to be three viable, popular, explanations for multiple garnet growth in these rocks: 1) growth of garnet during two separate prograde metamorphic events, as originally postulated by Rosenfeld (1968), 2) prograde Acadian consumption of garnet by the "staurolite-in" reaction followed by additional garnet growth with increasing metamorphic grade, and 3) cooling prior to thermal relaxation associated with nappe emplacement in the Devonian.

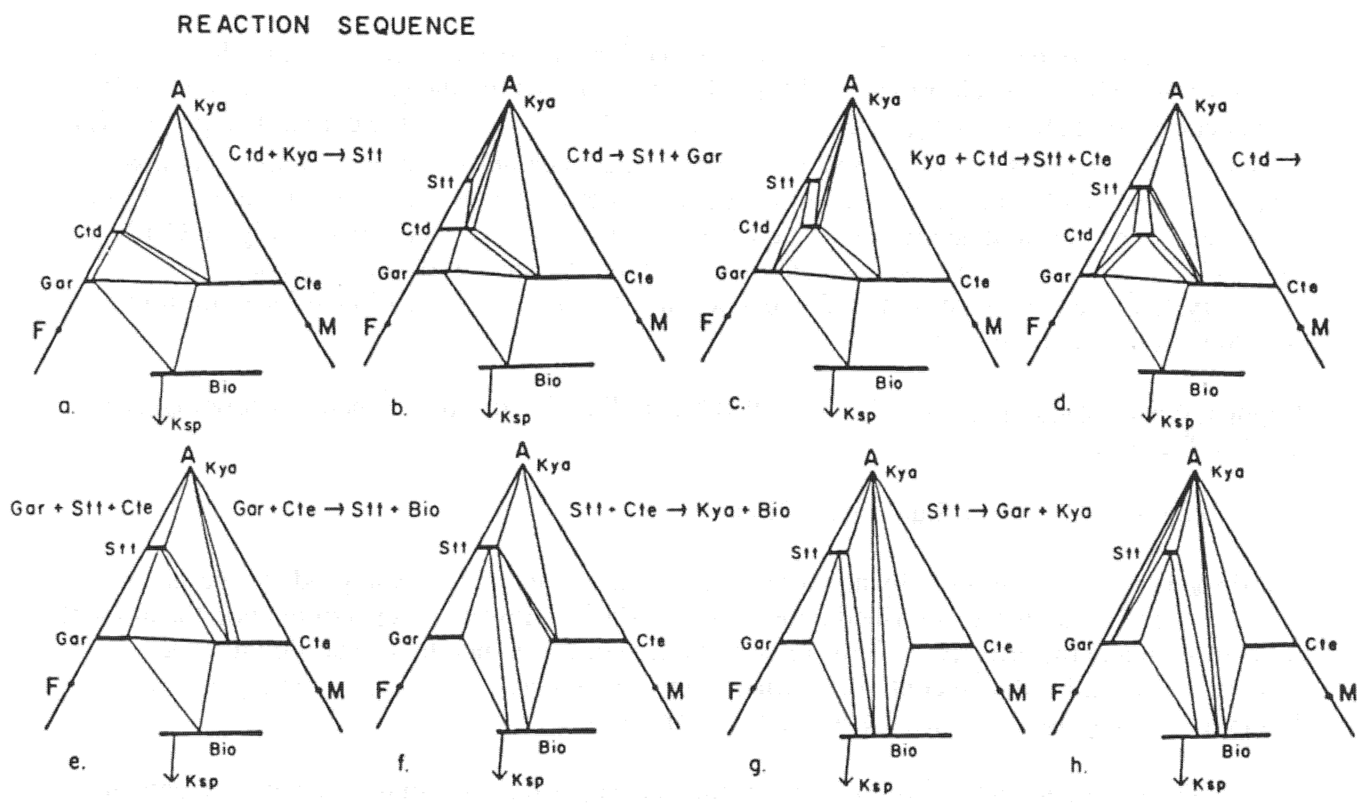


Figure C-12. (a-h) Generalized AFM facies series representing progressive metamorphism of pelitic rocks in the Cavendish area. Two-phase tie lines have been omitted for simplicity. Discontinuous reactions (quartz, muscovite, and H_2O assumed present) describing the change from one facies type to the next are given. Either facies type g or h represents the maximum grade attained (see text for further discussion). From Downie (1982).

R E A D I N G
Sawyer Stand Sch
South Reading
Plymouth Kingdom
Hurricane Hill
Rush Meadow Sch
Felchville Reading P O
Bowen Hill
Robinson Hill
Prospect Hill
COFFEEN CEM
Mt Gilead
Wheeler Sch
Star Hill
Hardy
C-4
Widow Hill
Stockin Sch
Greenbush
Tarbell Hill Rd
Cavendish Center Sch
C A V E N D I S H
Tarbell Hill Sch
Eagleshead
Downers
Gilchrist Sch
Twentymile Stream
Verde Quarry
Proctorsville Potter
47-2
47-5
47-4
Whitesville
Smithville
Proctorsville
Cavendish
Cavendish Gorge
RUTLAND
BLACK RIVER
HAWKS MOUNTAIN
Perkinsville
Quarry Pond
Ball Hill
47-3
Baltimore Sch
B A L T I M O R E
Duttonsville
Old Shincracker
Kendricks Corner
SPRINGFIELD AIRPORT
Smokeshire Sch
Chandler Sch
Williams

STOP C-5: CHLORITOID PHYLLITES OF THE PINNEY HOLLOW FORMATION

Route 100A, Pinney Hollow.

Pinney Hollow is located within the eastern cover sequence of the Green Mountain massif. The large roadcut on the west side of route 100a consists of garnet zone high-alumina phyllites of the Cambrian Pinney Hollow Formation. The common quartz- and muscovite-bearing mineral assemblages in these rocks are chloritoid+chlorite+ paragonite, chlorite+ garnet + paragonite, and chloritoid + chlorite + garnet + paragonite. These mineral assemblages are consistent with the garnet zone AFM topology as shown in Figure C-13. At this locality,the garnet contains chloritoid inclusions and is in turn partially replaced by chlorite. Accessory minerals in these rocks included rutile and/or ilmenite, magnetite, tourmaline, apatite and epidote. Beware the pleochroic brown mineral with straight extinction, "bird'seye" texture and one cleavage for it is "oxy-chlorite". The occurrence of this mineral in Pinney Hollow Formation rocks from this area was verified with the aid of an electron microprobe probe by Mark Wick (1987), an Amherst College student pursuing what turned out to be ficticious or "non-existant biotite+chloritoid phase relations" in the Vermont sequence. As reported by Professor Peter Robinson from the University of Massachusetts, the late professor Francis G. Turner would warn his students that the failure to identify oxy-chlorite as biotite was grounds for failing his petrography course. In addition, Rosenfeld, Thompson, and Zen (1958) reported basal spacings of 9.948 and 9.644 Å for coexisting muscovite and paragonite, respectively, from this locality (designated as Plymouth, Vermont). These basal spacings correspond, approximately, to compositions of Pg_{20} for muscovite and Pg_{93} for paragonite (Zen and Albee, 1964). Other significant work on similar rocks from nearby localities includes that of Hanscom (1973, 1975, 1980) on the "giant chloritoid"(1.5 cm) from Weaver Hill used for the single crystal structural characterization of monoclinic chloritioid (space group C2/c).

STOP C-6: GARNET SCHIST OF THE STOWE FORMATION

US Route 4, Near Bridgwater Corners.

Like the pinney Hollow formation at the previous stop, the Stowe Formation is a Cambrian pelitic unit that contains rocks analogous to those now found in the Taconic clippen of western Vermont. In contrast to the chloritoid phyllites, the schists in the roadcut on the north side of route 4 have a "normal" bulk composition (one that plots below the Gar+ Chte join on an AFM diagram) in that they contain biotite + chlorite+plagioclase garnet in addition to quartz and muscovite. The plagioclase is relativley sodic ($<<An_{23}$) as its index of refraction is obviously less than that of quartz. These rocks also contain magnetite,ilmenite,tourmaline, and apatite. Although numerous quartz ribbons are common throughout the outcrop the garnet-bearing schists are concentrated in layers reflecting original bedding and thus variations in bulk composition. The garnet grains are euhedral but do have thin chlorite rims. Chlorite also occurs in "blobs" as well as defining with muscovite the multiply deformed fabric in these rocks. Biotite occurs only as randomly oriented crosscutting euhedral laths. We know of no detailed work on this locality.

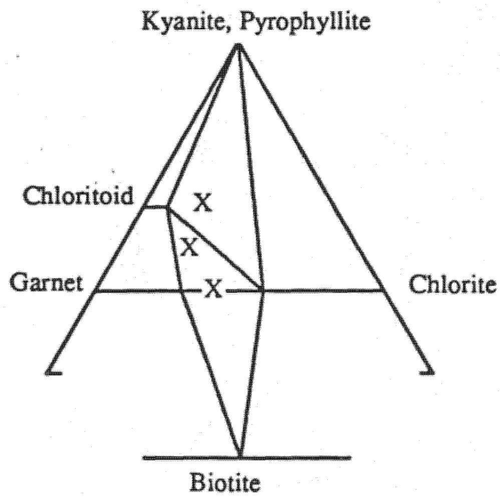
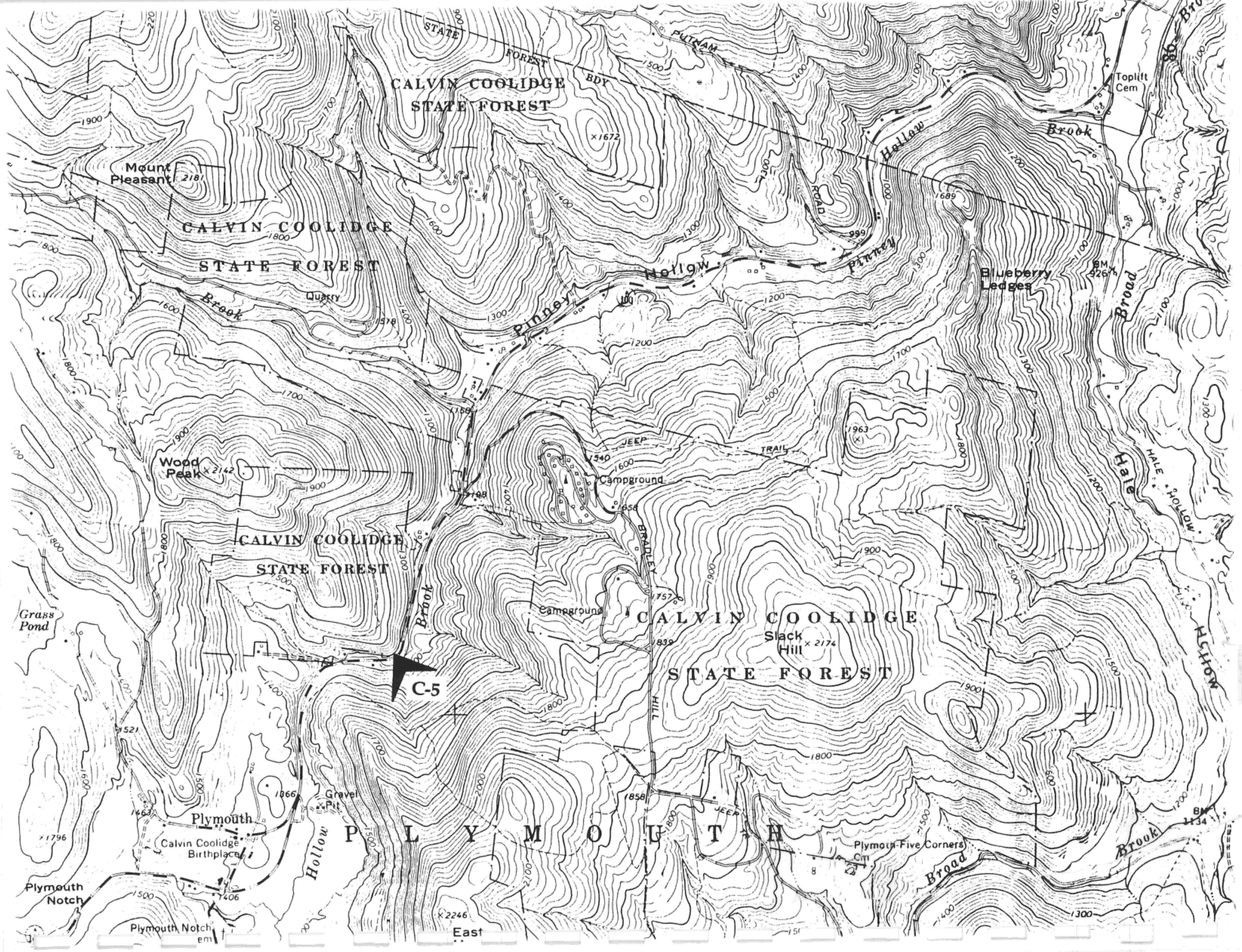
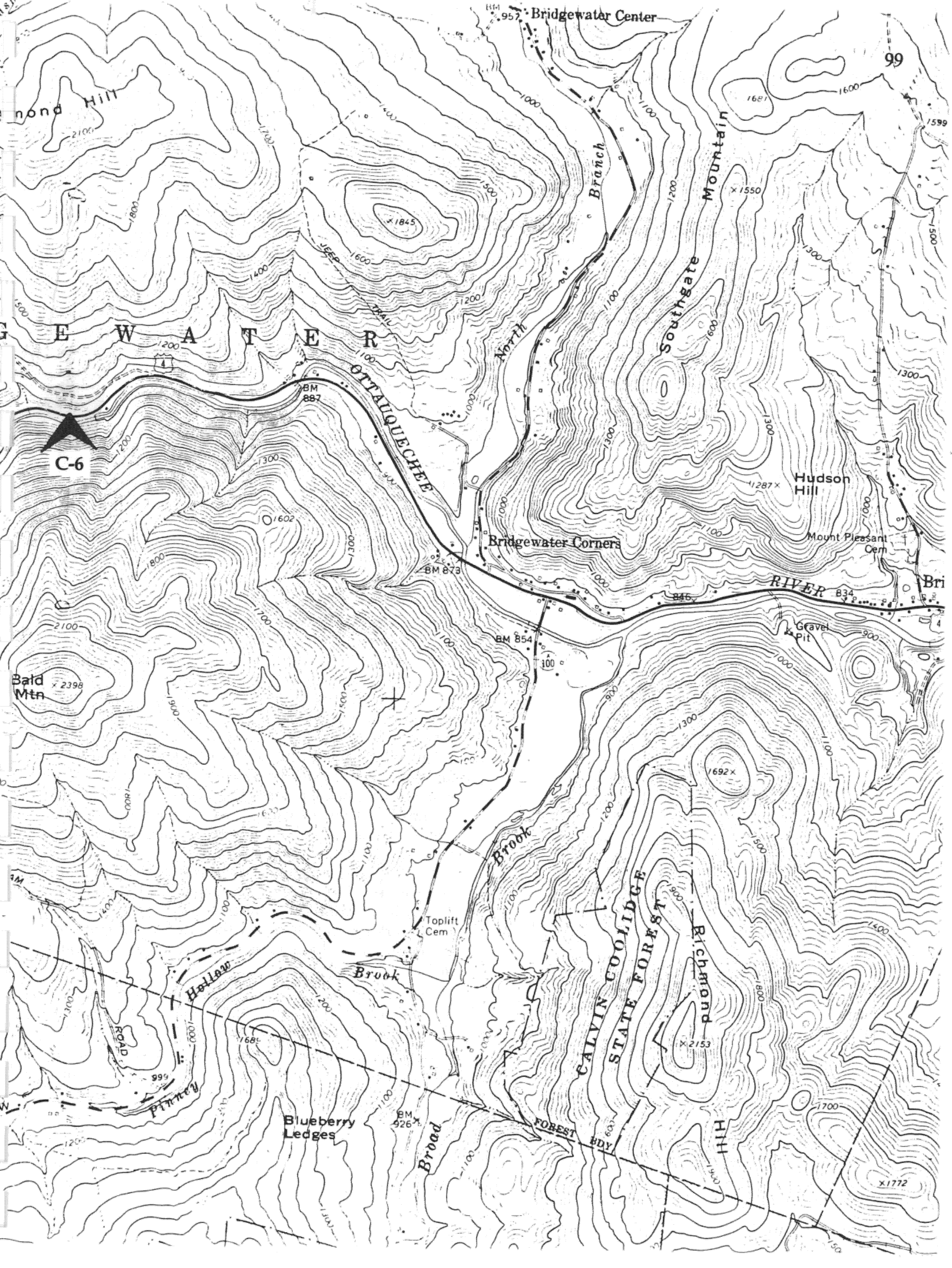


Figure C-13. Schematic AFM projection for garnet-zone mineral assemblages at Pinney Hollow and Weaver Hill.

	W HILL GARNET GRAV	W HILL CHTD GRAV	W HILL CHTD PROBE	D MINE GARNET GRAV	D MINE CHTD GRAV	D MINE CHTE GRAV
SiO2	36.54	24.46	24.88	36.53	23.82	25.19
TiO2	0.15	0.39	0.00	0.10	0.00	0.09
Al2O3	21.74	39.83	39.57	21.42	38.91	22.65
Fe2O3	0.13	1.25	0.00	0.36	2.94	3.34
FeO	31.38	24.54	25.17	31.65	23.30	22.32
MnO	5.02	0.68	0.80	6.43	0.43	0.08
MgO	0.60	1.64	2.08	1.64	3.18	15.04
CaO	3.93	0.00	0.00	1.78	0.00	0.00
H2O +/-	0.95	1.04		0.91	1.16	11.19
H2O +		6.02			6.36	
Total	100.44	99.85	92.50	100.82	100.09	99.90
ANHYDROUS FORMULAS						
#OX/FORMULA	12	6	6	12	6	14
Si	2.973	1.013	1.033	2.967	0.990	2.603
Ti	0.009	0.012	0.000	0.006	0.000	0.007
Al	2.085	1.944	1.936	2.050	1.906	2.759
Fe3+	0.008	0.039	0.000	0.022	0.092	0.260
Fe2+	2.135	0.850	0.874	2.150	0.810	1.929
Mn	0.346	0.024	0.028	0.442	0.015	0.007
Mg	0.073	0.101	0.129	0.199	0.197	2.317
Ca	0.343	0.000	0.000	0.155	0.000	0.000
ATOM RATIOS						
Fe2/Fe2+Mg	0.967	0.894	0.871	0.915	0.804	0.454
Fe3/Fe2+Fe3	0.004	0.044	0.000	0.010	0.102	0.119

Table C-3. Mineral analyses of chloritoid, chlorite, and garnet from Weaver Hill, VT, and the Davis Mine, MA. Grav=gravimetric analysis by Jun Ito, 1958. Probe=microprobe analysis by Roger Hanscom (1973).


CALVIN COOLIDGE
STATE FOREST
STATE FOREST BDY
PUTNAM
Toplift Cem
Brook
Mount Pleasant
2181
Hollow
ROAD
Pinney Hollow
Pinney
Quarry
1578
1672
1689
Blueberry Ledges
BM 926
Broad
Wood Peak
2142
1963
JEEP TRAIL
Campground
BRADLEY HILL
Hale
HALE HOLLOW
Hollow
Slack Hill
2174
Grass Pond
C-5
Gravel Pit
Plymouth
Calvin Coolidge Birthplace
Plymouth Notch
PLYMOUTH
Plymouth Five Corners
Cm
BM 1134
2246
East
1796


Bridgewater Center
Hill
GEWATER
C-6
BM 887
OTTAUQUECHEE
JEEP TRAIL
North Branch
Southgate Mountain
Hudson Hill
Mount Pleasant Cem
Bridgewater Corners
BM 873
RIVER
834
846
Bri
Gravel Pit
BM 854
100
Bald Mtn
2398
1845
1602
1550
1681
1599
1287
1692
Brook
Toplift Cem
Brook
Hollow
ROAD
Pinney
999
1689
Blueberry Ledges
BM 926
Broad
CALVIN COOLIDGE STATE FOREST
FOREST BDY
Richmond Hill
2153
1772
1700

Palermo mine

STOP C-7: THE RUGGLES MINE

Description exerpted from Cameron, Larrabee, McNair, Page, Stewart, and Shainin (1954).

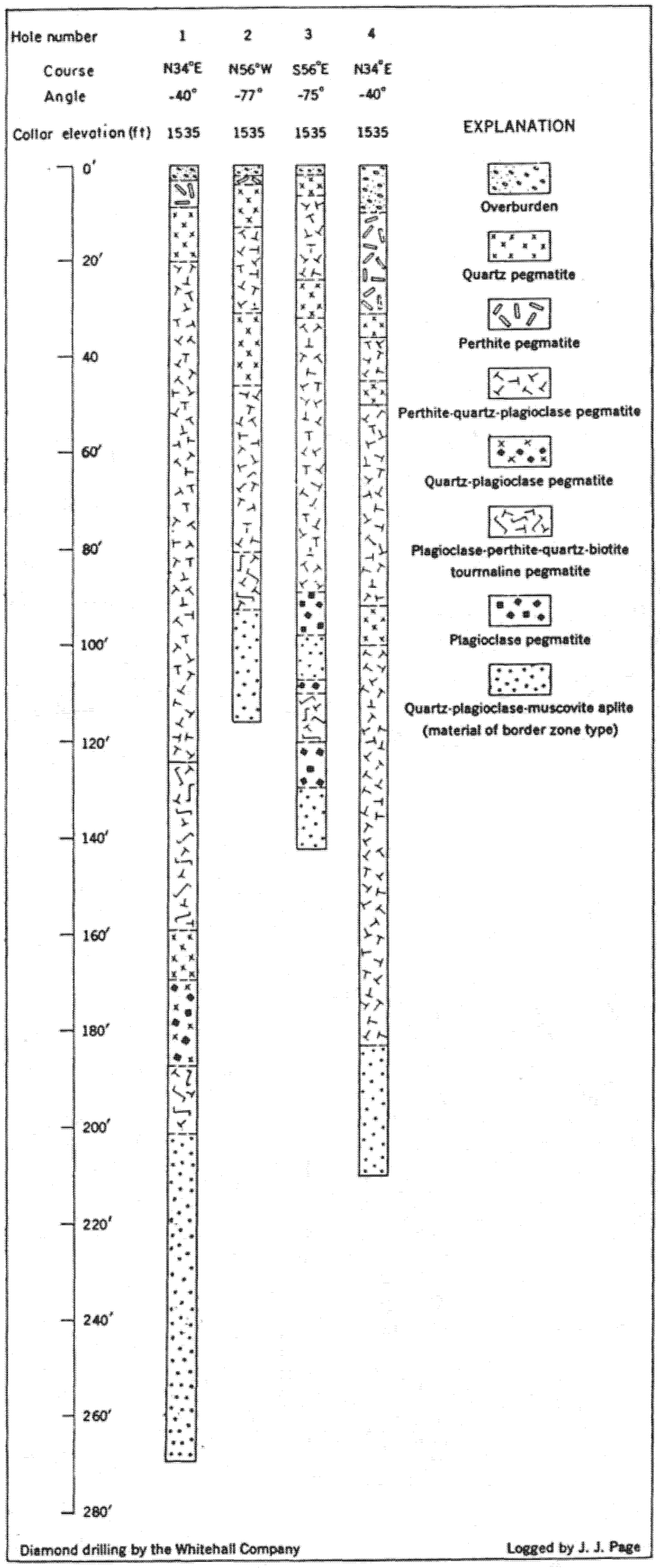


FIGURE 94.—Logs of diamond-drill cores, Ruggles feldspar-mica mine, Grafton, N. H.

RUGGLES FELDSPAR-MICA MINE

The Ruggles feldspar-mica mine (pl. 1, no. 65) is in the town of Grafton on the steep south slope and top of a hill 1.5 miles N. 40° W. of Grafton Center (Cardigan Station). From U. S. Highway 4 at Grafton Center, a graded road leads 1.4 miles westward across a meadow and up the valley of Manfeltree Brook. At this point, a good mine-access road, about 1.5 miles in length, turns north across the brook and extends up a steep hill to the mine.

Commercial production of mica in New Hampshire began at the Ruggles mine in 1803. Operations for mica, and more recently for potash feldspar, have been carried out intermittently to the present time. In 1841, the mine was operated by a Mr. Ingalls of Boston, Mass. (Jackson, 1844, p. 115) who mined about 16 tons of mica worth from $2 to $3 a pound when trimmed. Production about 1840 was only 600 to 700 pounds annually and in 1869 was 26,250 pounds, worth perhaps $60,000 (Hitchcock, 1878, p. 90). Sterrett (1923, p. 143) reports that several men worked the old dumps in 1912, and that mineral rights to the property were owned by Joseph Rogers of Rumney Depot, N. H. At that time, the American Minerals Company was preparing to begin work for feldspar, and the English Mica Co., of New York, had begun working the dumps for scrap mica. The rock was crushed on the dumps and then washed down a 3,200-foot flume to a mill on Manfeltree Brook.

The Whitehall Co., Inc., of New York, present owners and operators, worked the mine for a short time in 1932, and have operated it steadily since 1936. The New Hampshire United Mining Co., Andover, N. H., worked pit 32 (pl. 33) under lease during August and September 1944.

In 1940 Bannerman (p. 3–4) discussed the mineral occurrences of the Ruggles pegmatite. In 1943 (p. 13–16) he described and illustrated the geology of the mine workings and discussed the internal structure of the pegmatite and the distribution of valuable minerals in it.

The underground workings and larger opencuts were mapped in December 1943 and January 1944 by J. J. Page and E. Ellingwood 3d (pls. 33–35, fig. 94). The workings were mapped by Page and F. H. Main in August and September 1944. The maps and sections show the workings as of October 1, 1944. Cores recovered during exploratory diamond drilling in 1940 by the Whitehall Co. were made available for examination.

The Ruggles pegmatite has been worked for 1,550 feet along strike, practically the entire exposed strike length. Most of the workings are on the top and steep south slope of the hill and range in elevation from 1,270 to 1,600 feet. The earliest work was done near the bottom of the slope. In general, each succeeding operation was opened at a slightly higher elevation, and the waste rock was dumped into the earlier cuts. At present, all work is being done near the top of the hill.

WALL ROCKS

The wall rocks of the pegmatite are medium-grained quartz-mica schist, coarse-grained biotite gneiss, and amphibolite of the Littleton formation, and fine- to medium-grained biotite granite. Schist is the most abundant rock type. It is rich in biotite, except near the pegmatite contacts, where muscovite commonly is the more abundant mica, and the schist is heavily tourmalinized. Some layers of schist contain needles of sillimanite, and others contain brownish-black staurolite. Small lenticular bodies of coarse-grained biotite gneiss and amphibolite occur west of the pegmatite. The gneiss seems to have been formed by alteration of the amphibolite. Thin comformable bodies of biotite granite lie adjacent to or in the gneiss. The granite is probably intrusive into the schist and gneiss, but its relations to the pegmatite are unknown.

The average strike of the wall-rock foliation is about N. 25° E., and the average dip is about 65° SE. The bodies of biotite gneiss, amphibolite, and granite have their long axes parallel to the average strike of foliation.

Basic dikes ranging from less than a ¼ inch to 3 feet in thickness cut the pegmatite. They strike about N. 60° E. and dip steeply southeast or are vertical. Most of them fork. They do not appear to extend into the wall rock.

SIZE AND EXTERNAL FORM

The Ruggles pegmatite crops out or has been exposed by mining for 1,640 feet along strike. Its maximum outcrop width is 335 feet, and it ranges from a few feet to at least 160 feet in thickness. The trend of the pegmatite on the surface is about N. 20° E. and the average strike is N. 35° E. The body has an extremely varied southeast dip.

The pegmatite is an irregular lens that has not been deeply eroded. Its apex is between pits A and C. North of pit C the pegmatite plunges northeastward 15°–20°; south of pit C it has an irregular southwestward plunge. (See sec. A–A′, pl. 35.) The north end of the pegmatite resembles roughly an ancient battle-axe in plan. The tabular main part of the pegmatite, in which the major workings are located, forms the shaft of the axe. The blade is formed by a lobelike eastward dipping sheet, which is connected to the main body near the southeast side of pit D.

The pegmatite pinches out at the surface in the open-cut at the north prospect drift, but it widens rapidly south of this working. A few feet east of the north end of pit C, a narrow tongue of schist which also widens southward splits the pegmatite into two parts. East of this schist wedge, the pegmatite forms an eastward-dipping sheet that dips less steeply and is progressively thinner southward. Erosion has exposed the bottom of this sheet on the steep slope south of the top of the hill. A smaller pegmatite body underlies the main east sheet. The lower pegmatite pinches out up dip in a sharp crest.

In pits A and C, west of the schist wedge, the pegmatite has a maximum width of 120 feet on the surface, but it thickens downward to at least 160 feet. At the south end of pit A, the pegmatite plunges southward under schist, although a narrow apophysis of pegmatite extends into the schist. Pegmatite is also capped by schist in the north wall of pit 32, (pl. 33).

South of pit 32, outcrops of schist and marginal zones of pegmatite suggest that here also gently dipping schist covered the pegmatite. Dips along the west wall are steep. Remnants of the schist capping can be seen throughout the southern part of the map area, and the abundance of fine-grained aplite and the numerous apophyses extending southward into the schist at the south end of the pegmatite suggest a plunging structure similar to the one south of pit A.

COMPOSITION AND INTERNAL STRUCTURE OF THE PEGMATITE

Because the Ruggles pegmatite is well exposed it has been possible to map in detail the unusually large number of pegmatite units present in it. The pegmatite shows a well-defined zonal arrangement. The pegmatite units are listed below, in the general order of occurrence inward from the walls:

1. Plagioclase-quartz-muscovite border zone.
2. Plagioclase-quartz pegmatite (wall zone), composed of the following units:
   a. Plagioclase pegmatite.
   b. Cleavelandite pegmatite.
   c. Plagioclase-quartz pegmatite.
3. Muscovite-quartz-plagioclase pegmatite, first intermediate zone.
4. Plagioclase-perthite-quartz-biotite-tourmaline pegmatite, second intermediate zone.
5. Quartz-plagioclase pegmatite, third intermediate zone.
6. Plagioclase-muscovite-quartz-perthite pegmatite, fourth intermediate zone.
7. Perthite-quartz-plagioclase pegmatite, fifth intermediate zone.

8. Graphic granite pegmatite, core of east-dipping sheet.
9. Perthite-quartz pegmatite, sixth intermediate zone of main part of pegmatite.
10. Perthite pegmatite, core of main part of pegmatite.
11. Quartz lenses and veins.
12. Muscovite-plagioclase replacement bodies.
13. Sericitized perthite-plagioclase-quartz replacement body.

Plagioclase-quartz-muscovite border zone.—A discontinuous aplitic border zone commonly is adjacent to the schist walls. It generally is less than 5 feet thick, but has a maximum thickness of at least 40 feet. It is thickest and most persistent along both walls north of pit A and in apophyses extending into schist at the south end of the pegmatite body. It consists essentially of fine-grained equigranular plagioclase (An_7) and quartz with interstitial muscovite. Accessory minerals are garnet, tourmaline, apatite, and microcline. Thin irregular stringers consisting of quartz and plagioclase with scattered muscovite and perthite cut the aplite.

A rock megascopically similar in composition and texture to the border zone occurs in the east crosscut from the north shaft on the east side of pit A, and in drill hole 3. It seems to lie inside the main parts of the pegmatite.

Plagioclase-quartz pegmatite wall zone.—Inside the border zone three units are found in different parts of the mine. All consist of plagioclase, quartz, and muscovite, but they differ markedly in proportion of minerals.

A discontinuous unit of blocky plagioclase occurs inside the aplitic border zone or, where the border zone is absent, against the schist wall. The unit is found chiefly southwest of pit A. Its thickness is commonly less than 3 feet. It consists almost entirely of blocks of white plagioclase, 6 by 8 inches in maximum dimension, with small amounts of quartz and muscovite. Similar material also occurs in drill hole 3 and in the east crosscut from the north shaft. In drill hole 3 it seems to lie between the plagioclase-perthite-quartz-biotite-tourmaline zone and the border zone. In the east crosscut of the north shaft, a plagioclase unit similar in all respects to that found elsewhere in the pegmatite lies between the perthite-quartz-plagioclase zone and the biotite-bearing unit.

The cleavelandite unit is commonly less than 3 feet thick at most places and consists essentially of coarsely bladed cleavelandite with minor quartz, greenish muscovite, and blue-green apatite. The cleavelandite is commonly stained by iron and manganese oxides. The unit shows much the same relationship to the wall rock and the border zone south of pit 32 as the blocky plagioclase unit north of it.

The plagioclase-quartz unit contains plagioclase (An_{4-5}), gray quartz, and minor muscovite and black tourmaline. It is exposed in the eastern sheet-like part of the pegmatite and continues northward to the North Prospect drift. It commonly is inside the border zone but also occurs along the bottom and top of the sheet-like part of the pegmatite.

Muscovite-quartz-plagioclase zone (sheet-mica-bearing).—The muscovite-quartz-plagioclase zone is a discontinuous intermediate zone of the pegmatite. Muscovite is the most abundant mineral, and commonly 50 to 75 percent of the zone consists of large books oriented normal to the margins of the zone. Quartz and plagioclase (An_2) are also essential minerals, and black tourmaline, commonly intergrown with muscovite, and green apatite are accessories. It is best exposed under the schist capping on the 120- and 140-foot levels and on the steeply dipping west wall of the pegmatite. On the 120-foot level, it is thickest on the nose of the pegmatite and extends back along the walls for about 150 feet from the nose itself. The old stope at the south end of the 140-foot level was developed in this zone under the flat-lying schist roof. More recently, mica has been obtained from the zone on the steeply-dipping west wall of the pegmatite. The location of workings down hill from pit 32 suggests a similar localization of the mica-rich zone there. The zone occurs inside the plagioclase unit or against the schist where the plagioclase unit is absent. It is discontinuous where the border zone is thick.

Plagioclase-perthite-quartz-biotite-tourmaline zone.—The plagioclase-perthite-quartz-biotite-tourmaline zone forms a thick capping over and around the central zones at the northeast end of the pegmatite. It thins down dip and southward along strike. Only scattered small bodies of this material were seen south of pit A. It occurs most commonly between the border zone or plagioclase zone and the perthite-quartz-plagioclase zone. Pendants and apparently isolated bodies of biotite-rich material occur in the latter zone.

The mapping of this zone was based on the presence of biotite, and several minor mineralogic variants are included in it. In general, it consists essentially of plagioclase (An_{4-5}), perthite, gray quartz, biotite, and black tourmaline. Muscovite is an abundant though minor constituent. Plagioclase is more abundant than perthite in most exposures and commonly is somewhat stained. In pit A and in drill holes 1 and 3 most of it is green. Quartz occurs as anhedral grains between other minerals. Biotite commonly occurs in strips as much as 3 feet long and 6 inches wide, but less than ½ inch thick. Muscovite is in isolated books or intergrown with strip biotite. Black tourmaline may occur in small tabular bodies which, viewed from a distance, seem to be biotite. Some of the biotite strips pass lengthwise into tourmaline, but individual books of biotite and crystals of tourmaline are also present. Sulfides are locally abundant in the green plagioclase.

Quartz-plagioclase zone.—Irregular quartz grains that average 1 inch across make up at least 75 percent of the quartz-plagioclase zone. The rest is mostly interstitial white plagioclase (An_2) that appears to be fine-grained cleavelandite. Isolated books of muscovite, light-green beryl crystals, and perthite are accessory minerals. This unit lies between the border zone, the muscovite-quartz-plagioclase zone, or the plagioclase unit of the wall zone and the central perthite-rich zones. It is exposed south of the west crosscut from the north shaft and in, and south of, pit 32.

Plagioclase-muscovite-quartz-perthite zone (sheet-mica-bearing).—The plagioclase-muscovite-quartz-perthite zone consists essentially of plagioclase (An_2), muscovite, gray quartz, perthite, and minor tourmaline and biotite. The mica books are large and extend into the adjoining zones. The mica-rich shoots occur at the inner margin of the zone and have been mined in the long drift on the west side of pit A and in the crosscut in pit D. The plagioclase-muscovite-quartz-perthite zone lies between the biotite-rich zone described above and the no. 2 feldspar zone described below. Commonly, it is 3 to 5 feet thick. The mica-rich shoots are discontinuous, and found only along the northwest (footwall) side of the no. 2 feldspar zone in pits A and D. The shoots appear to have definite vertical limits. Widely spaced mica books of large size occur at the outer edge of the no. 2 feldspar zone elsewhere in pit A. These scattered books are believed to be in small isolated segments of the zone.

Perthite-quartz-plagioclase zone.—The perthite-quartz-plagioclase zone, known locally as the no. 2 feldspar zone, is exposed in pits A, C, and D and in the underground workings from them. It has a maximum thickness of 120 feet and completely encloses, or almost so, the large perthite pegmatite lens in the pits. It consists largely of perthite, but gray quartz and plagioclase are abundant and considerable parts of this zone may consist entirely of plagioclase and quartz. Small areas of graphic granite are present and muscovite, in small books, is an abundant accessory mineral. Scattered crystals of green beryl, as much as 5 feet long and 2 feet in diameter, occur in this zone at the margin of the perthite pegmatite in pit A.

Graphic granite zone.—The center or core of the sheet-like eastern part of the Ruggles pegmatite consists of graphic granite. The angular quartz spindles of the intergrowth commonly are less than one-half inch in length. Perthite also occurs in quartz-free crystals.

Quartz-perthite zone.—The quartz-perthite zone is best exposed in the walls of pit 32 but it occurs also in the southeast cuts. Its relation to the perthite pegmatite mined out in pit 32 is uncertain. It consists essentially of milky to light gray quartz which contains large blocks of perthite. One or more of the boundaries of the perthite blocks usually is straight and sharp. Contacts of this unit with the perthite pegmatite probably are gradational, like the contacts with quartz bodies.

Perthite zone.—The large lens of perthite pegmatite mined in pits A, C, and D is at least 400 feet long, 60 feet wide, and 50 feet high. It has been the principal source of no. 1 feldspar. A large part of it has been mined out in pits A and C. It pinches out south of pit A and seems to be thinning northeast of pit D although considerable feldspar is believed to be available under the present workings. A similar smaller lens was reported to have been mined in pit 32. These lenses seem to represent the discontinuous core of the pegmatite, although the relationship of the perthite pegmatite in pit 32 to the quartz-perthite unit is obscure. The perthite zone consists almost entirely of perthite. Plagioclase is almost lacking, except in the perthite intergrowths. Small quartz stringers cut the perthite but are not abundant. Sericite is present locally along fractures in the feldspar.

Quartz bodies.—Gray to milky quartz occurs in irregular lenses associated with the large lenses of perthite pegmatite, in isolated lenticular bodies inside various zones of the pegmatite, and in late cross-cutting veins and stringers. Irregular lenticular bodies of quartz are present adjacent to or enclosed in the perthite zone in pits A, C, and D. One lens in pit A is 200 feet long and extends diagonally under the large perthite lens. Other irregular quartz bodies occur in various positions within the pegmatite but apparently have no definite relation to the enclosing units. Quartz veins up to 4 feet thick cut the central parts of the pegmatite in pits A, C, and D, but apparently do not extend into the schist walls. Many of these veins have strikes perpendicular to the strike of the pegmatite. They have steep south or southeast dips.

The quartz is commonly light gray but is white and milky in the center of the thicker bodies. Small amounts of muscovite, garnet, sulfides, and uranium minerals occur in places. In pits A and C, a fine-grained aggregate of albite is found with the late quartz veins.

Muscovite-plagioclase unit (scrap-mica-bearing).—Adjacent to the perthite pegmatite, irregular muscovite-plagioclase replacement bodies are present. They consist largely of greenish-rum wedge muscovite in closely packed aggregates of diversely oriented books with small amounts of interstitial plagioclase and traces of quartz. The mica books are less than 2 inches across and commonly "A", reeved, or herring-bone. The plagioclase associated with the muscovite seems to be fine-grained cleavelandite.

Sericitized perthite-plagioclase-quartz unit.—A perthite-plagioclase-quartz unit containing an abundance of heavily sericitized perthite is exposed at the southwest corner of pit A. Associated with the sericitized perthite are unaltered perthite blocks, white blocky plagioclase, finer-grained areas of interlocking plagioclase and quartz, and small scaly muscovite crystals in granular smoky quartz. Quartz also occurs as larger irregular bodies. Occasional large rum-colored muscovite books, and isolated, somewhat rounded bodies of aplitic material resembling that of the border zone, are also present.

The outer limits of this unit are difficult to define, because sericitization of the perthite, on which mapping of this unit was based, ranges from minor alteration to almost complete replacement, and because the unit has been produced by the sericitization of parts of several zones. Isolated perthite blocks along the margins of the no. 2 feldspar zone in the west drift of the 140-foot level are sericitized; the main exposures of this unit are in the southwest corner of pit A and on the 120-foot level vertically below it. With decrease in the intensity of alteration, it grades into the no. 2 feldspar zone and also into the aplitic border zone. Muscovite and plagioclase are most abundant near the pegmatite wall. The few large mica books may represent an extension of the marginal muscovite-quartz-plagioclase zone.

On the east side of pit A, in the east crosscut from the north shaft, and in drill hole 3, pegmatite units comparable in composition and texture to those found elsewhere in the dike are exposed, but the units have peculiar structural relations to the other units. No satisfactory explanation of the structural relationships of these units can be offered at present. Possible interpretations include the presence of a second pegmatite intrusion, or the presence of abrupt rolls in the schist wall.

PEGMATITE EAST OF RUGGLES PEGMATITE

A small pegmatite lens is exposed east of the sheetlike projection of the main Ruggles pegmatite. The small pegmatite pinches out up dip in a sharp crest and has no surface connection with the sheetlike projection. It contains three zones. An aplitic border zone forms most of the surface outcrop, but apparently is restricted to the crest of the lens. The zone is mineralogically similar to the aplitic border zone of the main pegmatite. The wall zone contains blocky plagioclase, quartz and muscovite and is exposed only along the footwall. The core consists of almost equal quantities of pinkish perthite and quartz.

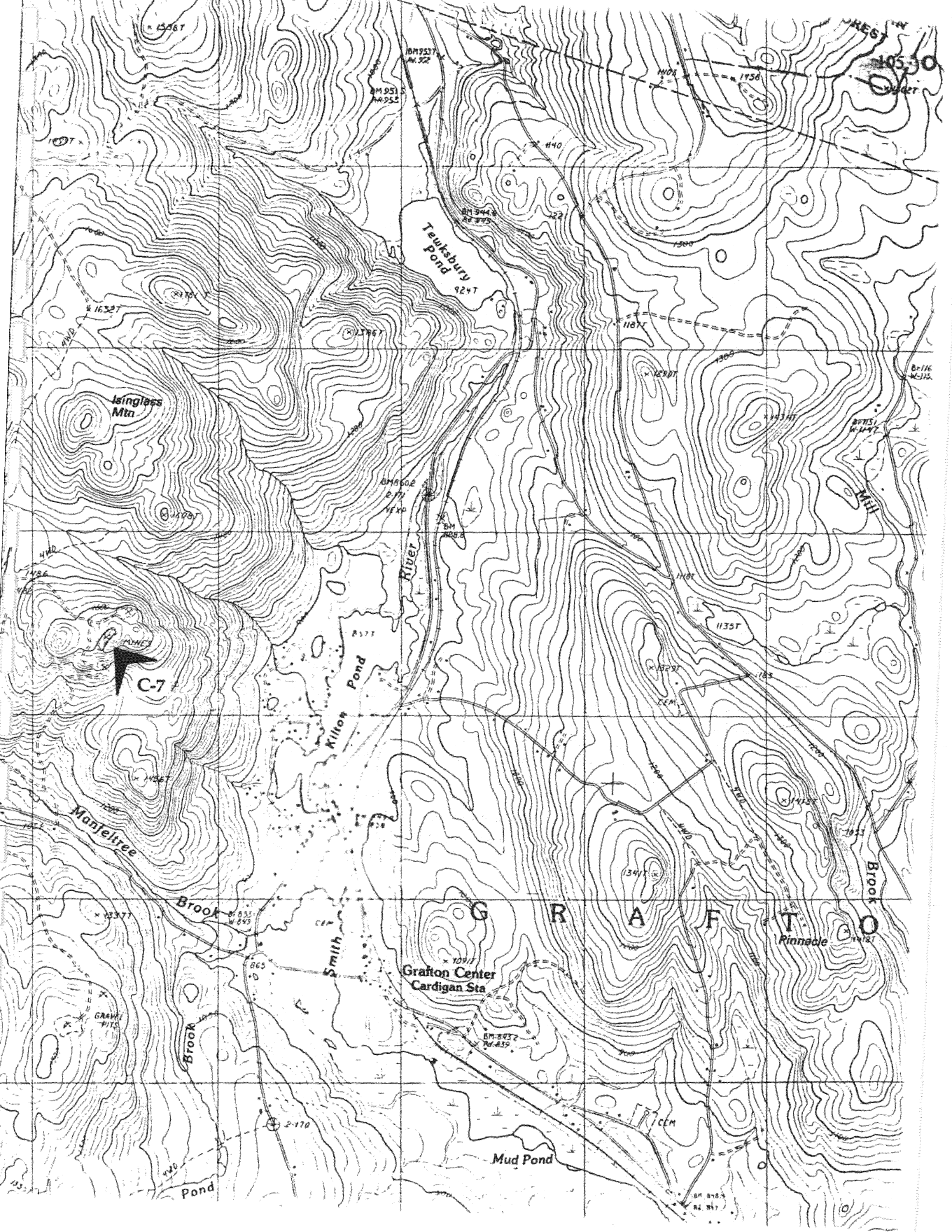
MICA AND FELDSPAR

Sheet mica is obtained from two intermediate zones of the pegmatite. The muscovite-quartz-plagioclase zone near the walls has been the major source of the mica produced by the Whitehall Co., Inc., but considerable production has come also from the mica-rich shoots on the northwest side of the perthite-quartz-plagioclase zone. Mica occurs in books as much as 3 by 5 feet in area and is rum, flat, mostly cloudy and free splitting. It is "soft" and commonly badly cracked and ruled, but a satisfactory percentage of large sheet can be recovered. Many books show iron oxide stains, but these can usually be removed during rifting and trimming. Mica along the west wall on the 140-foot level has a higher content of sheet than that in the eastern part of the zone exposed at this level. In 1944 the recovery of mine-run mica from rock mined in the mica-bearing zones was approximately 12 percent.

The Ruggles mine is operated chiefly for potash feldspar. The feldspar produced is used in soap or scouring agents and must be white and almost completely free from mineral impurities. The central perthite units consist almost entirely of this type of feldspar. Material from the no. 2 feldspar zone is crushed and sized, and the better grades of feldspar are sorted out on picking belts.

Uraninite and its associated alteration products are found in the pegmatite and are of particular interest to collectors. Schaub (1937, p. 156; 1938, p. 334–341) described the occurrence, crystal habit, and composition of uraninite from the mine. Some secondary uranium minerals are also present. These and uraninite are associated with late veins and irregular bodies of quartz, which commonly has a dark smoky color near the small masses of radioactive minerals. Uranium minerals are best exposed at or near the west side of the perthite pegmatite near the south end of pit A and in the roof of the 120-foot level.

The largest reserves of no. 1 feldspar apparently lie under pit D and between pits D and C. The wall between pits A and C and the part of the perthite lens south of pit A contain additional reserves. Much larger reserves of no. 2 feldspar are present. This material would have to be milled to separate the other minerals from the perthite before it could be used in cleansers. The largest reserves of mica seem to lie south of the west wall drift in the 140-foot level and between this working and the flat stope.


Tewksbury Pond
Isinglass Mtn
MINES
C-7
Kilton Pond
River
Manfeltree Brook
Smith Pond
Grafton Center
Cardigan Sta
G R A F T O
Pinnacle
Mill Brook
Brook
GRAVEL PITS
CEM
Mud Pond
Pond
4WD

STOP C-8: I-89 STAUROLITE SCHIST

Exit 15. Montcalm, New Hampshire.

The outcrop is a long road cut on the east side of the north-heading entrance ramp to I-89. Coarse staurolite is developed here in the Littleton Formation on the SW flank of the Mascoma dome (Figure C-14). Some of the staurolite crystals have thin rims of white mica. Notice the very small size of the garnet crystals. This texture, small garnets with large staurolites, is typical of the of schists from Maine to New Hampshire, near the Vermont border and contrasts markedly with the large garnet crystals and smaller staurolite crystals found in the Vermont rocks. The New Hampshire rocks reflect an andalusite-sillimanite (Buchan) sequence whereas Vermont rocks are from a kyaniite-sillimanite (Barrovian) sequence. The two sequences meet in the biotite zone rocks of the "Connecticut Valley Metamorphic low". Armstrong et. al. (1992) have called these the Eastern Acadian (older by ~20 m.y.) and the Western Acadian metamorphic events. An interesting aspect of this rock is it's nearly continuous occurrence over 200 km along strike, reflecting the same tectonic evolution over a large area. The grade goes up and down but the general character of the rock stays the same from north of Littleton NH to South of Putney Vermont.(M.J.Kohn, personal communication). Although we know of no detailed work on this outcrop, it is part of the terrane considered in some detail by Kohn *et al.* (1992). The outcrop also contains a Mesozoic lamprophyre dike.

To get to diopside brachiopods, take exit 15 southbound. follow road south that parallels southbound lane. Park at end of road, follow power line to South ~100 yds, then take path to right to the quarry. Look for curved green brachiopods in the pink + green calc-silicate.

STOP C-9: GILSUM

Description modified from Spear (1992) and Spear and Chamberlain (1986).

This outcrop is Silurian Rangeley Formation (middle member) and contains the highest-grade pelitic assemblages found in the area. The pelitic rocks contain the prograde assemblage garnet + cordierite + biotite + alkali feldspar + sillimanite + plagioclase. Alkali feldspars are large (several cm across) and may not be present in a thin section sized sample. They are very apparent in outcrop, however. Cordierite is commonly observed in greenish clots, which are green because of chlorite alteration.. Fresh cordierite is not easy to find. Several retrograde zones containing chlorite and staurolite can be found at this outcrop. These zones were created as water was introduced into the upper plate (Fall Mountain nappe) sometime after nappe emplacement. Several calc-silicate pods are present in this outcrop. These pods are indicative of the Rangeley schists found in the Fall Mountain nappe.

The P-T data obtained from garnet-biotite thermometry and garnet-plagioclase barometry indicates peak P-T conditions of 740 °C, 3.5 kbar (see Figure C-15). There is abundant textural evidence for the replacement of garnet and cordierite by sillimanite and biotite, which Spear and Chamberlain interpret as the reaction:

$$\text{garnet+cordierite+K-feldspar+}H_2O = \text{sillimanite+biotite}$$

This reaction, plus the garnet zoning produced by this reaction suggests a P-T path of nearly isobaric cooling, as shown in Figure C-15. See Chamberlain (1986), Spear and Chamberlain (1986, stop F5), and Selverstone and Chamberlain (1990) for an alternative interpretation of the reaction texture, P-T path and tectonic significance. This outcrop is located at the intersection of two large F2 and F3 Acadian synclines and Chamberlain (1986) interpreted the metamorphism at this outcrop to have resulted from thermal effects that occurred during downfolding.

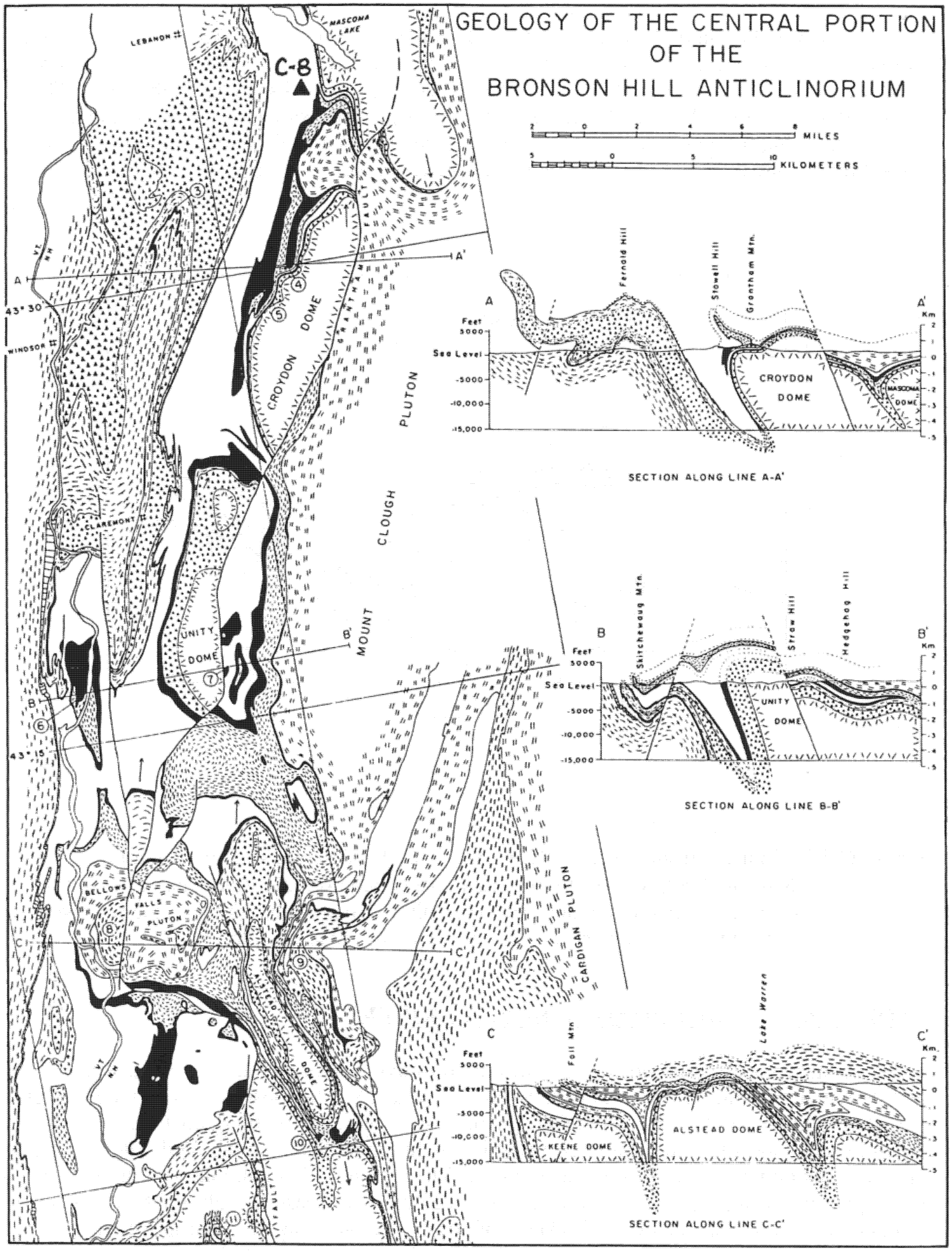


Figure C-14. Northern part of Plate 15-1a of Thompson *et al.* (1968) with cross sections from their Plate 15-1b as modified by Robinson et al. (1991).

Spear and Chamberlain interpret the P-T path to be the result of cooling following thrust emplacement. It is highly significant that the peak pressure experienced by these rocks is only on the order of 3.5 kbar, as contrasted with 6 kbar experienced by rocks nearby to the west. The locations are separated by only 3-4 km, so the difference in pressure requires some amount of tectonic thinning following attainment of peak pressures. Spear and Chamberlain suggest that backsliding along the Chesham pond thrust may be responsible for the tectonic thinning.

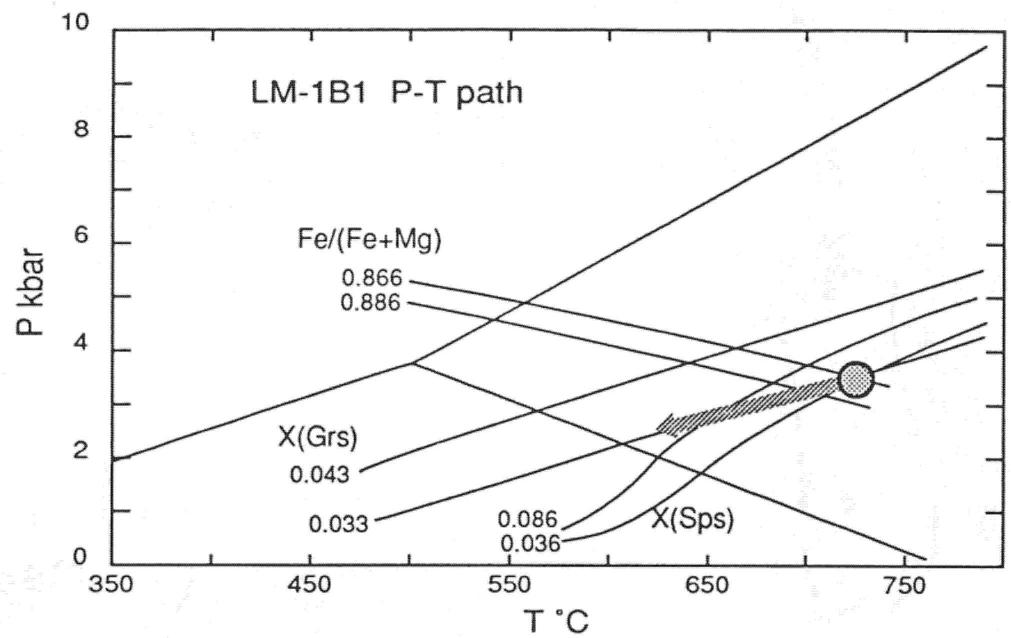
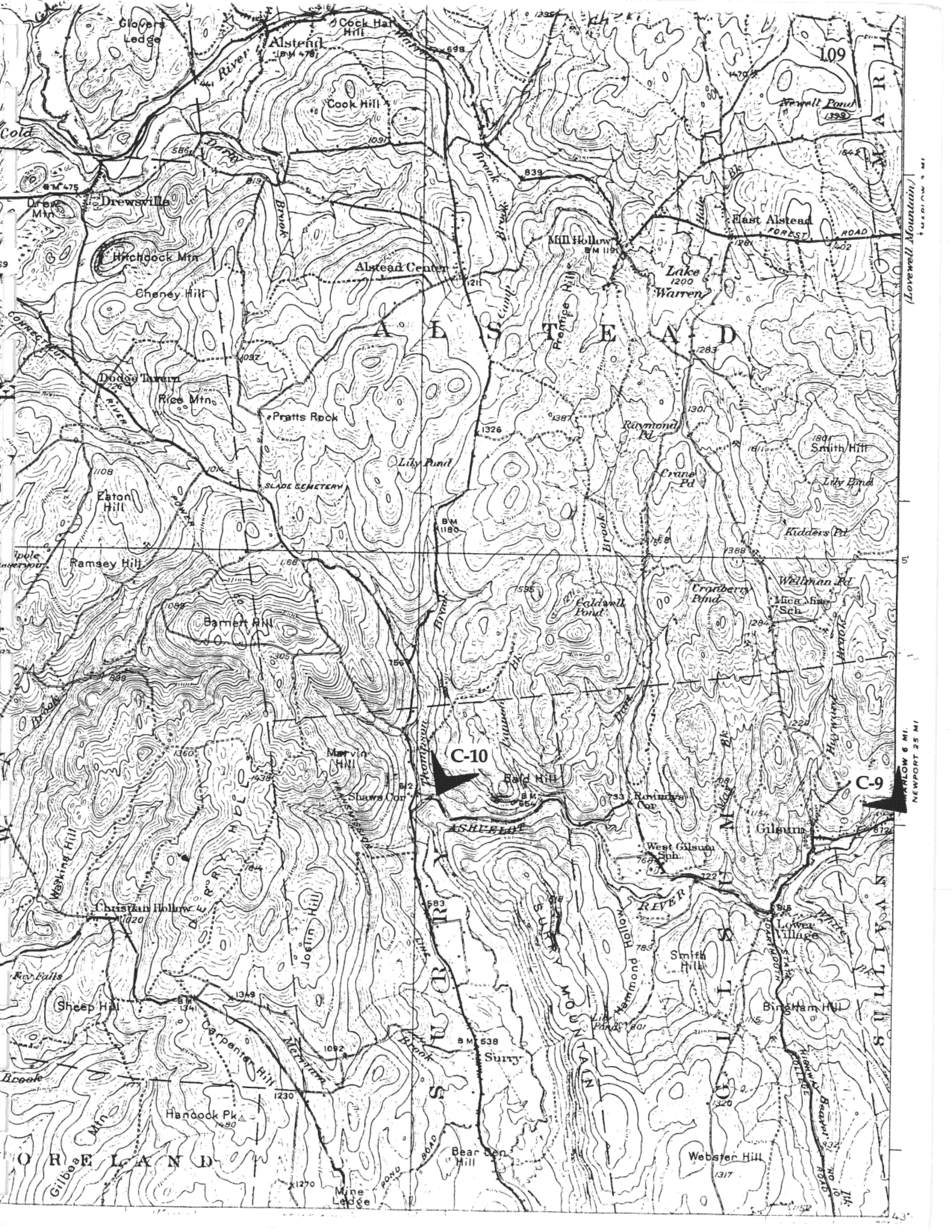


Figure C-15. Modified from Spear (1992, Figure 11). P-T diagram showing contours of Xsps, Xgrs, and Fe/(Fe+Mg) for the assemblage garnet-cordierite-biotite-sillimanite-K- feldspar-quartz-plagioclase. The only P-T path consistent with the observed zoning na dreaction progress is one of nearly isobaric cooling.

STOP C-10: GILSUM ROAD, ASHUELOT RIVER

Outcrops of the Ammonoosuc Volcanics (Ordovician) along the river contain coarse gedrite with exsolution lamellae of anthophyllite. Typical Ammonoosuc rocks on the north end of the Alstead Dome contain the assemblage plagioclase-hornblende-biotite-garnet. However, the outcrops here have the assemblage gedrite-anthophyllite-hornblende-biotite-cordierite-pyrrhotite due to a bulk composition that is depleted in Ca. The presence of amphibole + cordierite suggests lower pressures than those of the Vermont rocks nearby to the west. An ultramafic pod can be found here with talc and chlorite. We know of no detailed work on this outcrop, although similar rocks from the Ammonoosuc Volcanics to the south have been studied extensively by Robinson and Schumacher (e.g. Robinson and Jaffe, 1969; Schumacher and Robinson, 1987; Schumacher, 1988).



Glovers Ledge
Cock Hat Hill
Alstead
Cook Hill
Cold
River
Drewsville
Drew Mtn
Hitchcock Mtn
Cheney Hill
Alstead Center
Mill Hollow
East Alstead
Lake Warren
Newell Pond
Loveweil Mountain
109
A L S T E A D
Prentice Hill
Dodge Tavern
Rice Mtn
Pratts Rock
Lily Pond
Raymond Pd
Crane Pd
Smith Hill
Kidders Pd
Slade Cemetery
Eaton Hill
Ramsey Hill
Cranberry Pond
Wellman Pd
Mica Mine Sch
Caldwell Pond
Barnett Hill
Thompson Brook
C-10
C-9
Marvin Hill
Bald Hill
Shaws Cor
Ashuelot
Gilsum
West Gilsum Sch
River
Lower Village
Christian Hollow
Watkins Hill
Joslin Hill
Surry Mountain
Hammond Hollow
Smith Hill
Bingham Hill
Sheep Hill
Carpenter Hill
Surry
Hancock Pk
Gilboa Mtn
Bear Den Hill
Webster Hill
Mine Ledge
O R E L A N D
Sullivan
Marlow 6 Mi.
Newport 25 Mi.

STOP C-11: BELLOWS FALLS

Description modified from Spear and Chamberlain (1986) and Spear (1992).

The rocks exposed in the river bed are the gray schist of the Rangeley Formation of the upper plate of the Fall Mountain nappe and the Bellows Falls pluton. The contact between these units is well exposed. The uppermost occurrence of pluton is a sharp contact. The contact is in the river bed (usually under water) and here has the appearance of a steeply dipping fault. Proceeding down into the pluton are lenses and pods of schist, which locally have been nearly completely assimilated. Large garnets (1-2 cm radius) can be observed in the pluton in places. These are not igneous garnets but rather the relics of assimilated schist. Dikes of pegmatite cut the pluton but nowhere have we observed plutonic rock intruding the schist. Therefore, Spear and Chamberlain interpret the contact to be a shear zone with pieces of upper plate schist incorporated into the pluton.

The pluton contains plagioclase + K-feldspar + biotite + quartz. A zircon from this locality was analyzed by T. M. Harrison (of UCLA) on the SHRIMP in Canberra, Australia and he obtained an age of 407 ± 5 Ma, which is interpreted to be the crystallization age of the pluton (in Kohn et al, 1992).

Locally there are boudins of amphibolite and calc-silicate. The large porphyroblasts here are sillimanite pseudomorphs after andalusite ("turkey track rocks"). In a few places these sillimanites are folded with the foliation. Assemblages obseved in the gray schist include quartz + plagioclase +garnet + biotite + sillimanite ± spinel (only as inclusions within sillimanite) ± chlorite (retrograde) ± staurolite (retrograde) ± K-feldspar (only as inclusions within garnet) + muscovite + ilmenite.

Conspicuous in some samples is a white selvage surrounding the sillimanite porphyroblasts. This selvage is composed of late muscovite ± staurolite, produced during retrogression of the upper plate. Retrogression is common in the rocks here. The first stage occurred by the reaction

$$\text{garnet} + \text{K-feldspar} + H_2O = \text{sillimanite} + \text{biotite}$$

and is responsible for the consumption of garnet in many samples. Then the reaction

$$\text{sillimanite} + \text{K-feldspar} + H_2O = \text{muscovite} + \text{quartz}$$

is crossed, which converts all K-feldspar to muscovite and produces the selvages of muscovite around sillimanite. Once muscovite replaces K-feldspar, the reaction

$$\text{sillimanite} + \text{biotite} = \text{garnet} + \text{muscovite}$$

proceeds, which results in a second generation of small, idioblastic garnet. In addition, the reaction

$$\text{sillimanite} + \text{biotite} + \text{garnet} + H_2O = \text{staurolite}$$

has also occurred locally, producing small staurolites in selvages around sillimanite and around some early garnets. The fluids responsible for the retrogression presumably came from dewatering of structurally lower rocks following nappe emplacement or perhaps may have come from dewatering of the pluton during crystallization. The retrogression was not pervasive, however, as some samples contain pristine sillimanites.

The P-T path of the rocks from this locality was illustrated be Spear et al (1990) and is reproduced here in Figure C-16. Early metamorphism was low pressure, high temperature into the sillimanite + K-feldspar zone at a pressure of 3-4 kbar. The rocks experienced loading following peak metamorphism up to pressures of 5-6 kbar and cooling to 525 °C. The loading is interpreted to have resulted from the emplacement of another nappe structurally above the Fall Mountain nappe (now eroded away) and the cooling is interpreted to have resulted from the emplacement of the Fall Mountain nappe onto the cooler rocks of the Skitchewaug nappe.

The identity of the nappe that is higher than the Fall Mountain nappe is in question, because it does not crop out anywhere. However, there is evidence that the nappe is the Chesham Pond nappe (named here) that is comprised of still higher grade rocks (garnet + cordierite zone).

One further point to make about timing of metamorphism and thrust emplacement. Rocks of the Fall Mountain nappe were at 700-750 °C at their peak temperature whereas rocks of the Skitchewaug nappe were at 550 °C at their thermal peak. Therefore, the Skitchewaug nappe could not have been immediately below this outcrop throughout its history (or else it would be at the same temperature as this rock). Spear's interpretation is that we have a series of in sequence thrusts with the Fall Mountain nappe being first loaded from above by another nappe, then the pair riding to the west along on the Bellows Falls pluton and then picking up the Skitchewaug nappe when the Fall Mountain nappe had cooled somewhat. The lower contact of the Bellows Falls pluton also assimilates part of the Skitchewaug nappe, so the shortening must have occurred within the pluton itself.

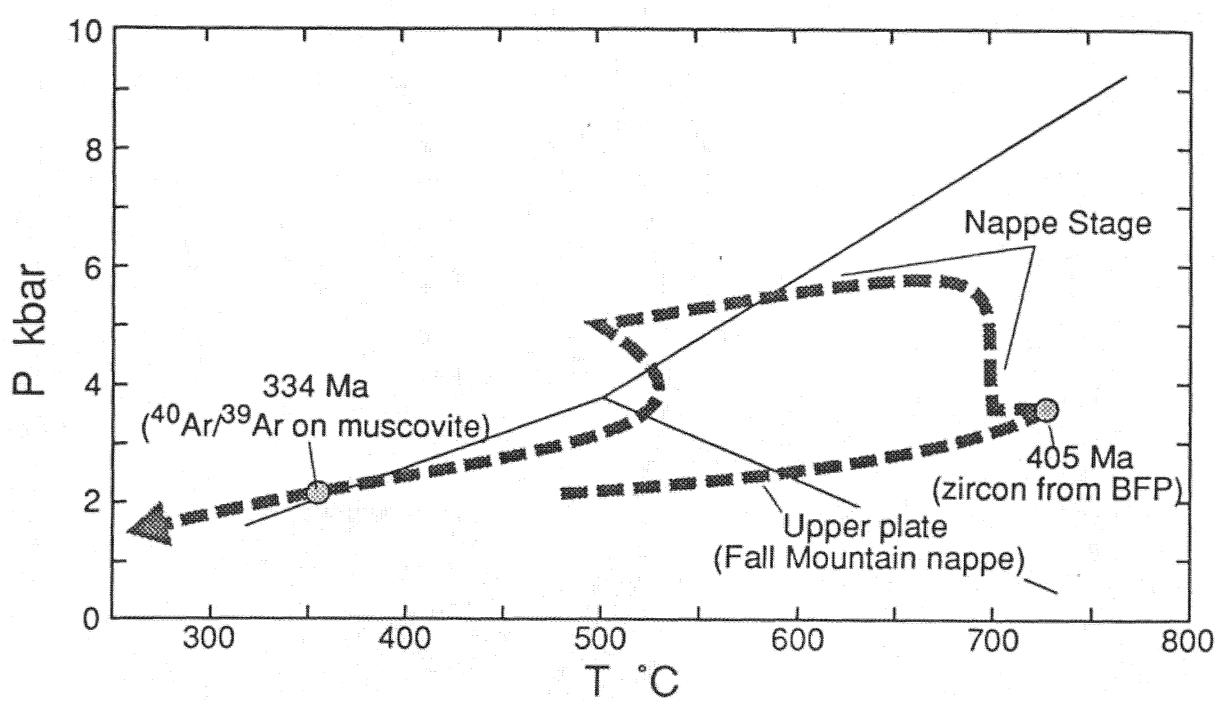
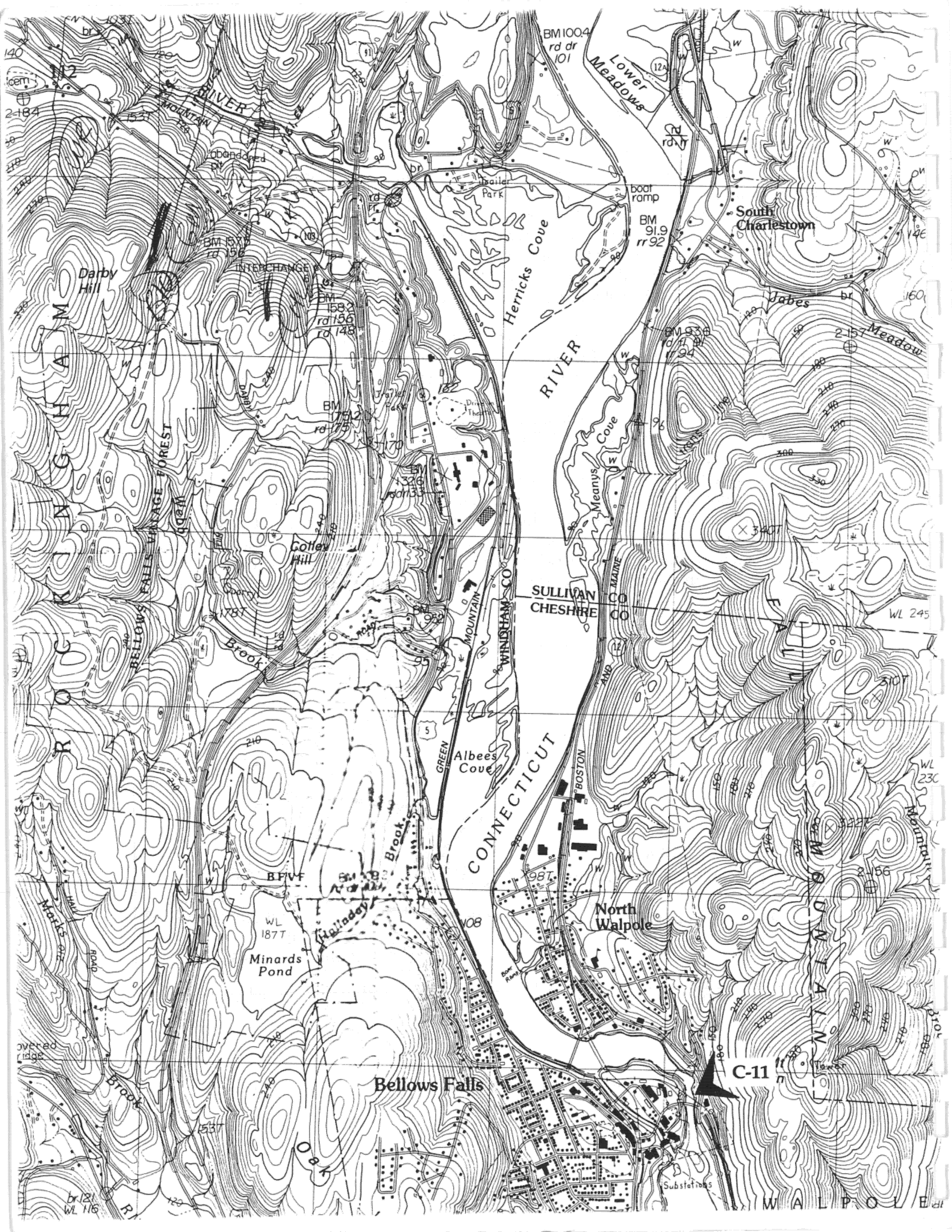


Figure C-16. P-T path for rocks from the Fall Mountain nappe (Spear *et al.*, 1990)



Lower Meadows
Trailer Park
boat ramp
South Charlestown
INTERCHANGE 6
Darby Hill
Herricks Cove
Jabes
Meadow
RIVER
Meanys Cove
Colley Hill
SULLIVAN CO
CHESHIRE CO
WINDHAM CO
GREEN MOUNTAIN
BOSTON AND MAINE
Albees Cove
CONNECTICUT
Brook
Minards Pond
North Walpole
Bellows Falls
C-11
Substations
WALPOLE
ROCKINGHAM
FALL MOUNTAIN

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