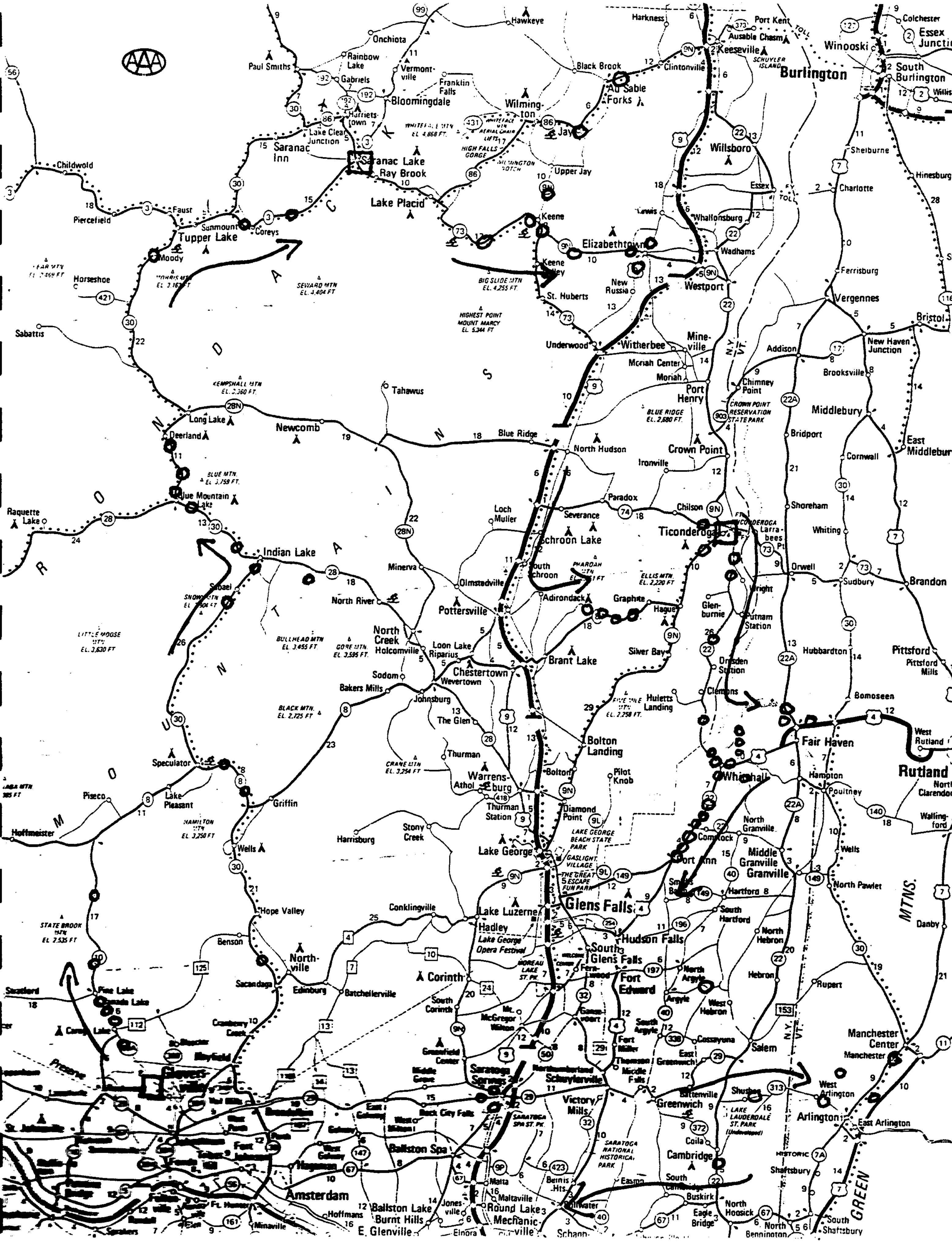
AAA
Paul Smiths
Onchiota
Rainbow Lake
Gabriels
Vermontville
Franklin Falls
Bloomingdale
Hawkeye
Harkness
Port Kent
Ausable Chasm
Keeseville
Clintonville
Black Brook
Au Sable Forks
Wilmington
Jay
Upper Jay
Harrietstown
Lake Clear Junction
Saranac Inn
Saranac Lake
Ray Brook
Lake Placid
High Falls Gorge
Childwold
Piercefield
Faust
Sunmount
Tupper Lake
Coreys
Moody
Horseshoe
Sabattis
Seward Mtn El. 4,404 Ft
Big Slide Mtn El. 4,255 Ft
Highest Point Mount Marcy El. 5,344 Ft
Keene
Keene Valley
St. Huberts
Elizabethtown
New Russia
Lewis
Whallonsburg
Wadhams
Westport
Willsboro
Essex
Winooski
Essex Junction
Colchester
South Burlington
Burlington
Shelburne
Hinesburg
Charlotte
Ferrisburg
Vergennes
Bristol
New Haven Junction
Addison
Brooksville
Middlebury
East Middlebury
Cornwall
Bridport
Shoreham
Whiting
Orwell
Sudbury
Brandon
Underwood
Witherbee
Mineville
Moriah Center
Moriah
Port Henry
Chimney Point
Crown Point Reservation State Park
Blue Ridge El. 2,680 Ft.
Crown Point
Ironville
Tahawus
Kempshall Mtn El. 2,360 Ft.
Long Lake
Newcomb
Blue Ridge
North Hudson
Deerland
Blue Mtn. El. 3,759 Ft.
Blue Mountain Lake
Raquette Lake
Indian Lake
Sabael
Snowy Mtn
Minerva
Olmstedville
North River
Loch Muller
Severance
Paradox
Schroon Lake
South Schroon
Pharoah Mtn
Ellis Mtn. El. 2,220 Ft
Chilson
Ticonderoga
Larrabees Pt
Adirondack
Graphite
Hague
Glenburnie
Putnam Station
Wright
Pottersville
North Creek
Little Moose Mtn. El. 3,630 Ft
Bullhead Mtn El. 3,455 Ft
Gore Mtn. El. 3,595 Ft.
Holcombville
Loon Lake
Riparius
Chestertown
Wevertown
Brant Lake
Silver Bay
Dresden Station
Hubbardton
Pittsford
Pittsford Mills
Sodom
Bakers Mills
Johnsburg
The Glen
Black Mtn. El. 2,725 Ft
Five Mile Mtn. El. 2,258 Ft.
Huletts Landing
Clemons
Bomoseen
Fair Haven
West Rutland
Rutland
Speculator
Thurman
Crane Mtn El. 3,254 Ft
Warrensburg
Athol
Thurman Station
Bolton
Bolton Landing
Pilot Knob
Whitehall
Hampton
Poultney
North Clarendon
Piseco
Lake Pleasant
Griffin
Hamilton Mtn El. 3,250 Ft
Hoffmeister
Wells
Harrisburg
Stony Creek
Diamond Point
Lake George Beach State Park
Gaslight Village
The Great Escape Fun Park
Lake George
Comstock
North Granville
Middle Granville
Granville
Wallingford
Fort Ann
Smiths Basin
Hartford
North Pawlet
Hope Valley
Conklingville
Lake Luzerne
Hadley
Lake George Opera Festival
Glens Falls
Hudson Falls
South Hartford
North Hebron
Hebron
State Brook Mtn El. 2,535 Ft
Benson
Northville
Sacandaga
Edinburg
Batchellerville
Corinth
Moreau Lake St. Pk.
South Glens Falls
Fenwood
Fort Edward
North Argyle
Argyle
West Hebron
Rupert
Danby
Green Mtns.
Pine Lake
Canada Lake
Caroga Lake
Cranberry Creek
Mayfield
Gloversville
South Corinth
Mt. McGregor
Wilton
Gansevoort
South Argyle
Fort Miller
Cossayuna
Greenfield Center
Saratoga Springs
Northumberland
Schuylerville
Thomson
Middle Falls
East Greenwich
Salem
Manchester Center
Manchester
West Arlington
Arlington
East Arlington
Middle Grove
Rock City Falls
Saratoga Spa St. Pk.
Victory Mills
Greenwich
Battenville
Shushan
Lake Lauderdale St. Park (Undeveloped)
East Galway
West Milton
Galway
Saratoga National Historical Park
Coila
Cambridge
Historic
Shaftsbury
Ballston Spa
Hagaman
Amsterdam
Malta
Maltaville
Round Lake
Mechanicville
Bemis Hts.
Stillwater
Easton
South Cambridge
Buskirk
Eagle Bridge
North Hoosick
North Bennington
South Shaftsbury
Ballston Lake
Burnt Hills
Jonesville
Hoffmans
E. Glenville
Elnora
Schaghticoke
St. Johnsville
Fort Johnson
Fonda
Tribes Hill
Fort Hunter
Minaville
Galway

CONTENTS

PETROLOGY, STRUCTURE AND REGIONAL GEOLOGY OF THE SOUTHERN ADIRONDACK MASSIF

MAY 4 - 7, 1995

This field trip attempts to integrate petrology, structural geology, stratigraphy, historical geology, regional geology, and common sense to better understand and appreciate the geology of the Adirondack region. I hope there will be something for everyone, enough different kinds of geology that you can get your hooks into something that interests you.

The trip will focus on the hard rocks and their petrologic and structural evolution. We will concentrate on middle to late Proterozoic rocks of the Adirondack massif, which is a good representative of the Grenville tectonic province of the Laurentian shield. On Sunday we will examine the western edge of the Ordovician Appalachian orogen, in the thrust faults of the Taconic range east of the Adirondacks. We will evaluate whether the Paleozoic deformation affected Adirondack rocks.

We will camp three nights. Thursday night will be spent at Hampton Beach campground on Sacandaga Lake north of Gloversville, on the south margin of the Adirondack region. Friday night will be at the KOA campground at Whiteface Mountain, north of Lake Placid. Saturday night will be at Rogers Rock campground on Lake George near Ticonderoga. We will arrive home in Newark very late on Sunday night.

Bring your very warmest and very driest clothing and outdoor gear. I'll bet it still gets mighty cold at night, and there are oceans of mud everywhere. Field boots required. Hammer, hand lens required.

This guidebook is organized in two sections - explanatory papers, and road logs/stop descriptions. We will use previously published trips. The initial paper, by Whitney, is an overview of the Adirondacks as a whole. Following that are three papers on the Southern Adirondacks by Jim McLelland, emphasizing the structure, geochronology/history, and petrology of that region. There are many more stops than we can possibly visit, but at least we have a choice. The actual stops visited will depend on time, energy, newness of the geology, and other factors.

I urge you before the trip to color some of the important maps, especially those on pages 33, 34, 35, 58 and 60.

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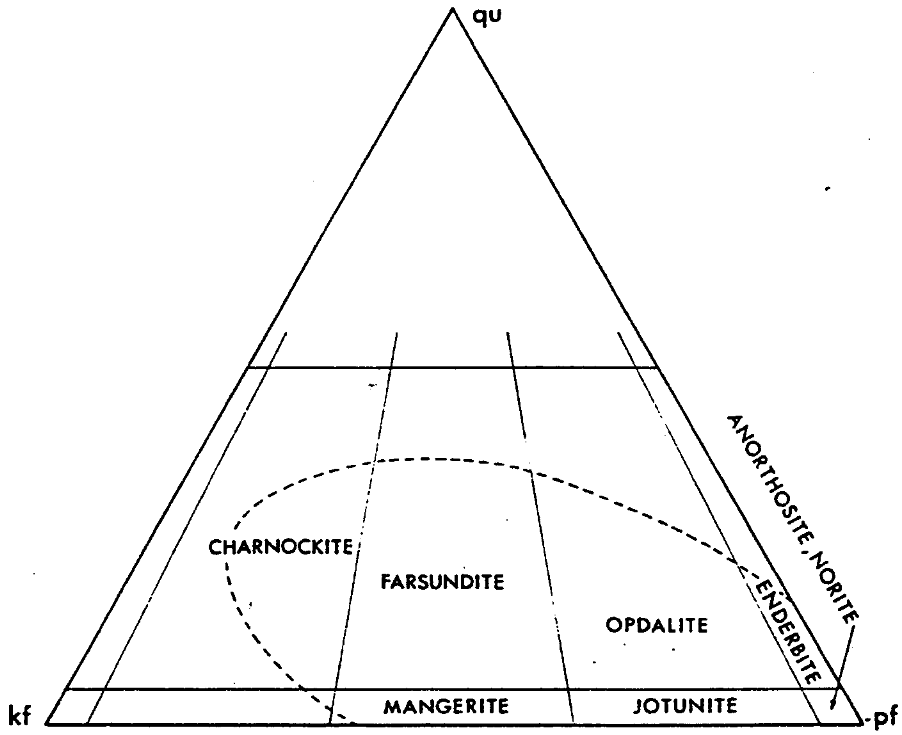
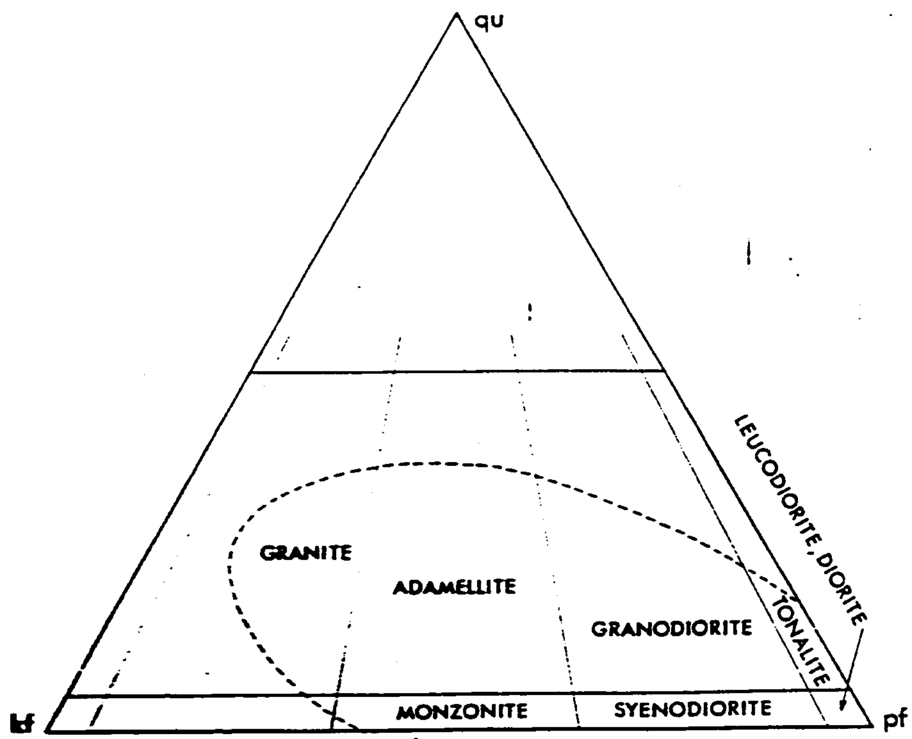
QAP Nomenclature based on Mafic Minerals



orthopyroxene



hornblende or biotite

GENERAL GEOLOGY OF

THE ADIRONDACKS

28th Intl. Geol. Congress, 1989

IGC FIELD TRIP T164:

THE ADIRONDACK MOUNTAINS: A SECTION OF DEEP PROTEROZOIC CRUST

Philip R. Whitney[1], Steven R. Bohlen[2], James D. Carl[3], William deLorraine[4], Yngvar W. Isachsen[1], James M. McLelland[5], James F. Olmsted[6], and John W. Valley[7]
with contributions by
Ian Cartwright[7], Jean Morrison[7], Paul W. Ollila[8], and Bruce Selleck[5]

OVERVIEW OF ADIRONDACK GEOLOGY

INTRODUCTION

The Adirondack Mountains of northern New York State are underlain by Middle Proterozoic (Neohelikian) rocks of the Grenville Province, exposed in a breached Cenozoic dome. This trip consists of a traverse from upper amphibolite-facies metavolcanic and metasedimentary rocks in the northwest lowlands, southeastward across a major zone of high ductile strain, into granulite-facies plutonic rocks of the Adirondack highlands, which record depths of 25-30 km in a doubly-thickened continental crust between 1.1 and 1.0 Ga. This guidebook is divided into two major sections. The first is an overview of Adirondack geology, with sections on regional setting, stratigraphy, igneous rocks, metamorphism, structure, geochronolgy, stable isotopes, economic geology, and neotectonics, and a speculative outline of the geologic history of the region. The second section is a road log, with extended descriptions for those stops that are the subject of current or recent research. There are 38 numbered stops on the planned route, in addition to 8 lettered alternate stops that will be visited if time permits.

[1]Geological Survey, New York State Museum, Albany, New York

[2]U.S Geological Survey, Menlo Park, California

[3]Department of Geology, State University of New York College at Potsdam, Potsdam, New York

[4]Zinc Corporation of America, Balmat, New York

[5]Department of Geology, Colgate University, Hamilton, New York

[6]Department of Earth and Environmental Sciences, State University of New York College at Plattsburgh, Plattsburgh, New York

[7]Department of Geology and Geophysics, University of Wisconsin, Madison, Wisconsin

[8]Luvak, Inc., Sherburne, Massachusetts

REGIONAL SETTING

The Grenville Province of eastern North America (Figs. 1,2) comprises a zone, several hundred km wide, of polydeformed Proterozoic (Helikian) metamorphic rocks. It stretches

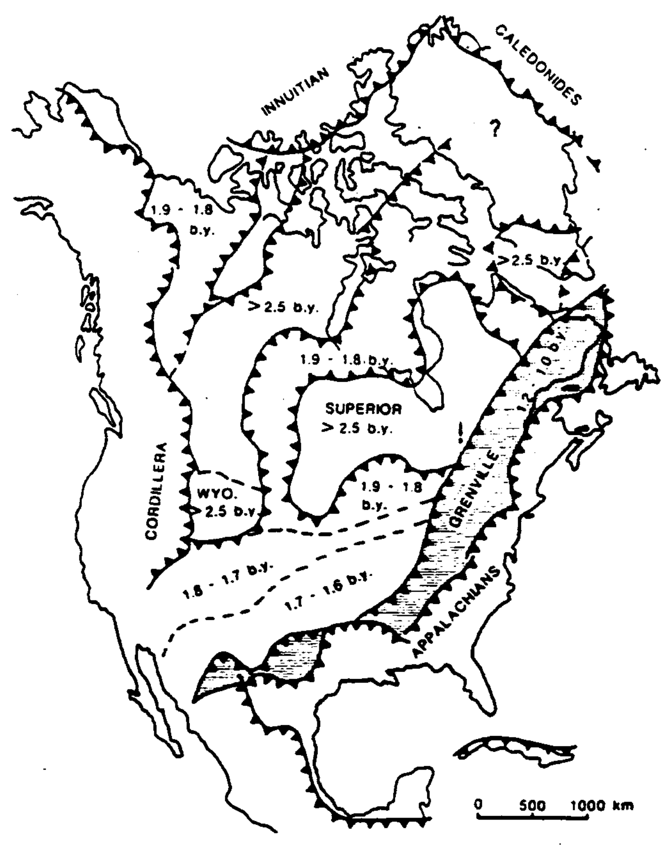


FIGURE 1 Major precambrian orogenic belts and age provinces of North America, adapted from an unpublished map by Paul Hoffman of the Geological Survey of Canada.

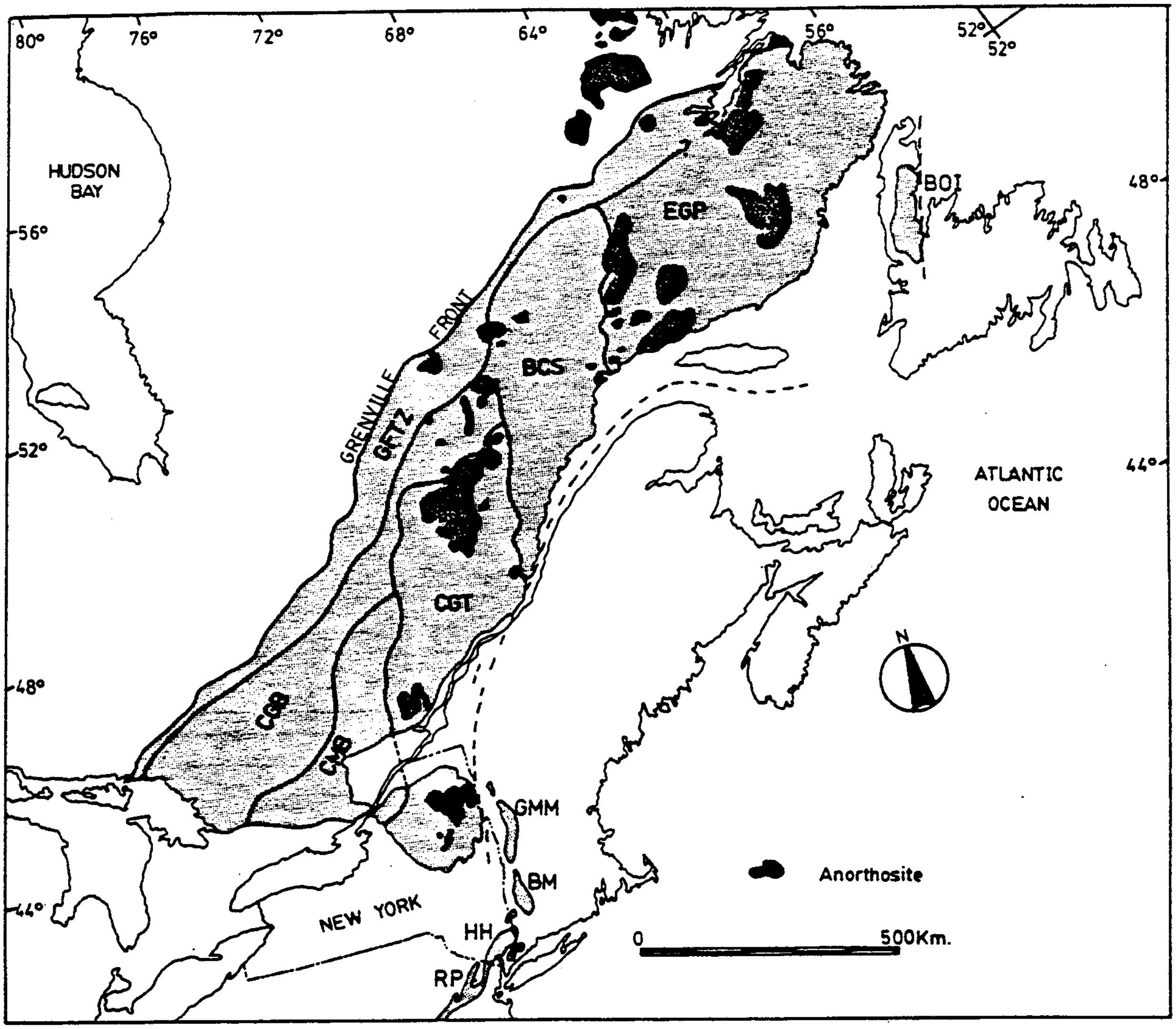


FIGURE 2 The Grenville Province of northeastern North America (after Wynne-Edwards, 1972). Shaded areas are 1.2-1.0 Ga rocks; black is anorthosite. GFTZ, Grenville Front Tectonic Zone; CGB, Central Gneiss Belt; CMB, Central Metasedimentary Belt; CGT, Central Granulite Terrain; BCS, Baie Comeau Segment; EGP, Eastern Grenville Province; BOI, Bay of Islands Complex; GMM, Green Mountain Massif; BM, Berkshire Massif; HH, Hudson Highlands; RP, Reading Prong.

from southern Labrador to the eastern shore of Lake Huron, and thence in the subsurface southwestward to the Llano Uplift in Texas. Blocks of Grenville age rocks are also found in northern Mexico. The Grenville Province is characterized by radiometric ages in the range 1.45 to 1.0 Ga, although in the northeastern part, close to the Grenville Front (Fig. 2), these ages are overprinted on older rocks. It is bounded on the northwest by the Superior (>2.5 Ga) and Southern (1.9-1.8 Ga) provinces of the Canadian Shield along the Grenville Front; on the southeast it is covered by Paleozoic rocks, with the exception of numerous, partially allocthonous exposures within the Appalachian orogen (Bay of Islands, Green Mountain and Berkshire massifs, Hudson Highlands-Reading Prong, Blue Ridge Anticlinorium, inter alia).

The Adirondack Mountains lie just west of the Appalachians on the North American craton They are a breached Cenozoic dome that expose rocks of the Grenville province through a window in lower Paleozoic sedimentary rocks. The Proterozoic exposures in the Adirondacks are connected to the Grenville of Ontario by narrow zone known as the Frontenac Axis (Figs 2,3). The Adirondack uplift has a diameter of 200 km and a structural relief of about 2 km. Maximum topographic relief reaches 1200 m in the High Peaks zone. Figure 4, a mosaic of satellite photographs, contrasts the

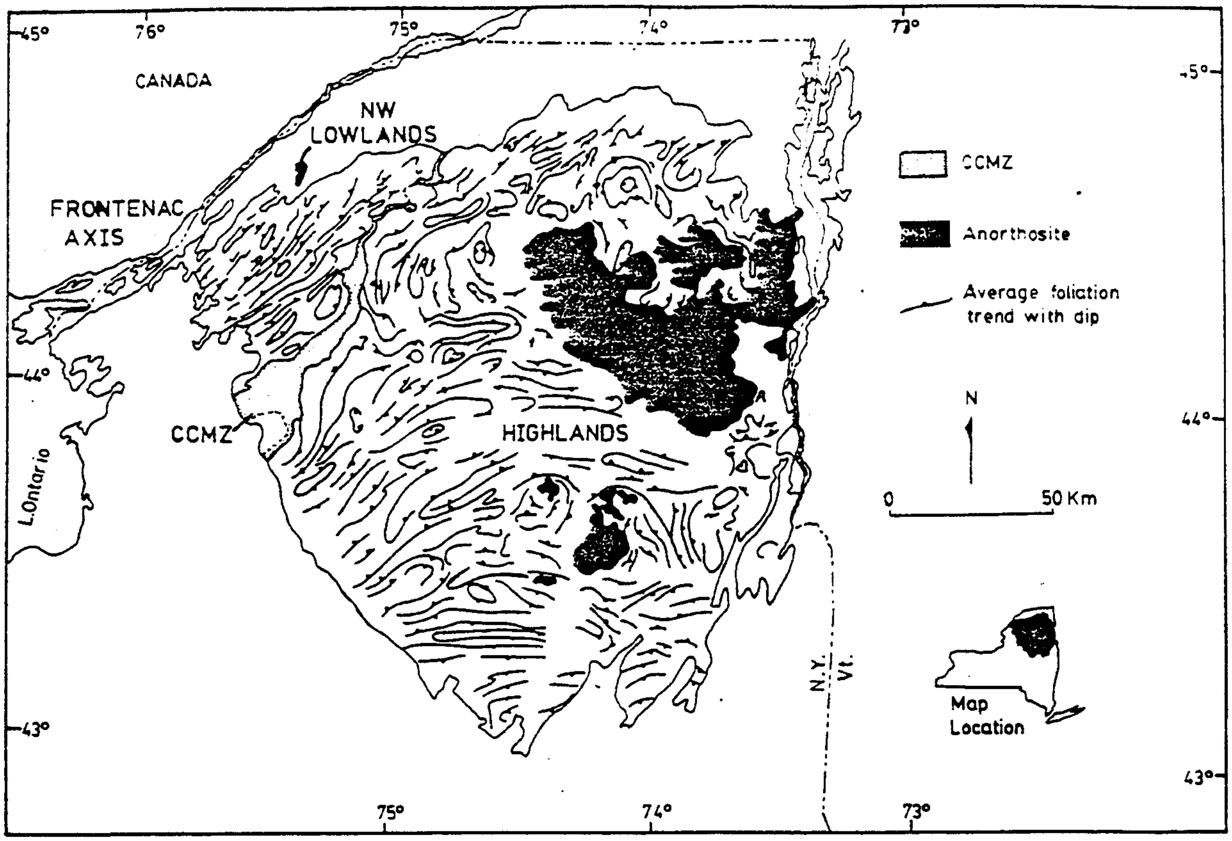


FIGURE 3 Map showing the subdivisions of the Adirondacks and location of major anorthosite bodies. CCMZ, Carthage-Colton Mylonite Zone.

mountainous topography of the Adirondack dome with the relatively flat surrounding terrane underlain by slightly deformed, unmetamorphosed Paleozoic rocks.

The Adirondacks are customarily divided into two regions (Fig. 3); the Adirondack highlands, which resemble the Central Granulite Terrane of the Grenville Province (Wynne-Edwards, 1972), and the northwest Adirondack lowlands, which correspond to the Central Metasedimentary Belt (Fig. 2). The lowlands and highlands are separated by an irregular, NE-trending, zone of high ductile strain, commonly called the Carthage-Colton mylonite zone (Geraghty and others, 1981). The dominantly metasedimentary rocks of the northwest lowlands are of upper amphibolite facies, except in the northeast near Colton and within a few km of the Carthage-Colton zone, where granulite facies assemblages are found. The highlands have a large proportion of metaigneous rocks, and are at granulite facies throughout. The High Peaks zone, in the central to northeastern part of the highlands, is underlain by a large (about 3500 km^2), composite body of of metanorthosite (the Marcy massif); smaller bodies of metanorthosite and related rocks are scattered throughout the highlands (Figs. 3,7).

This trip will consist of a NW-to-SE transect across the Adirondacks. The first two and one half days will be spent in the northwest lowlands and Carthage-Colton zone, followed by three days in the central highlands including the high peaks, and the final two days in the southern and southeastern highlands. Figure 4 shows the route which the trip will follow.

STRATIGRAPHY AND SEDIMENTARY ENVIRONMENT

Northwest Lowlands

Many efforts to subdivide the metasedimentary sequence in the Northwest Adirondack lowlands have resulted in the generalized lithic column shown in Figure 5. Formations and subunits shown in the column are believed to have widespread distribution and to be useful for regional correlations.

At the base of the column are pink leucogneisses, with thin amphibolite layers, that appear on the map (Fig. 6) as 13 or 14 prominent ovoid bodies. The rocks within these bodies will be referred to henceforth as the Hyde School Gneiss (deLorraine and Carl, 1988). This comprises most of the unit

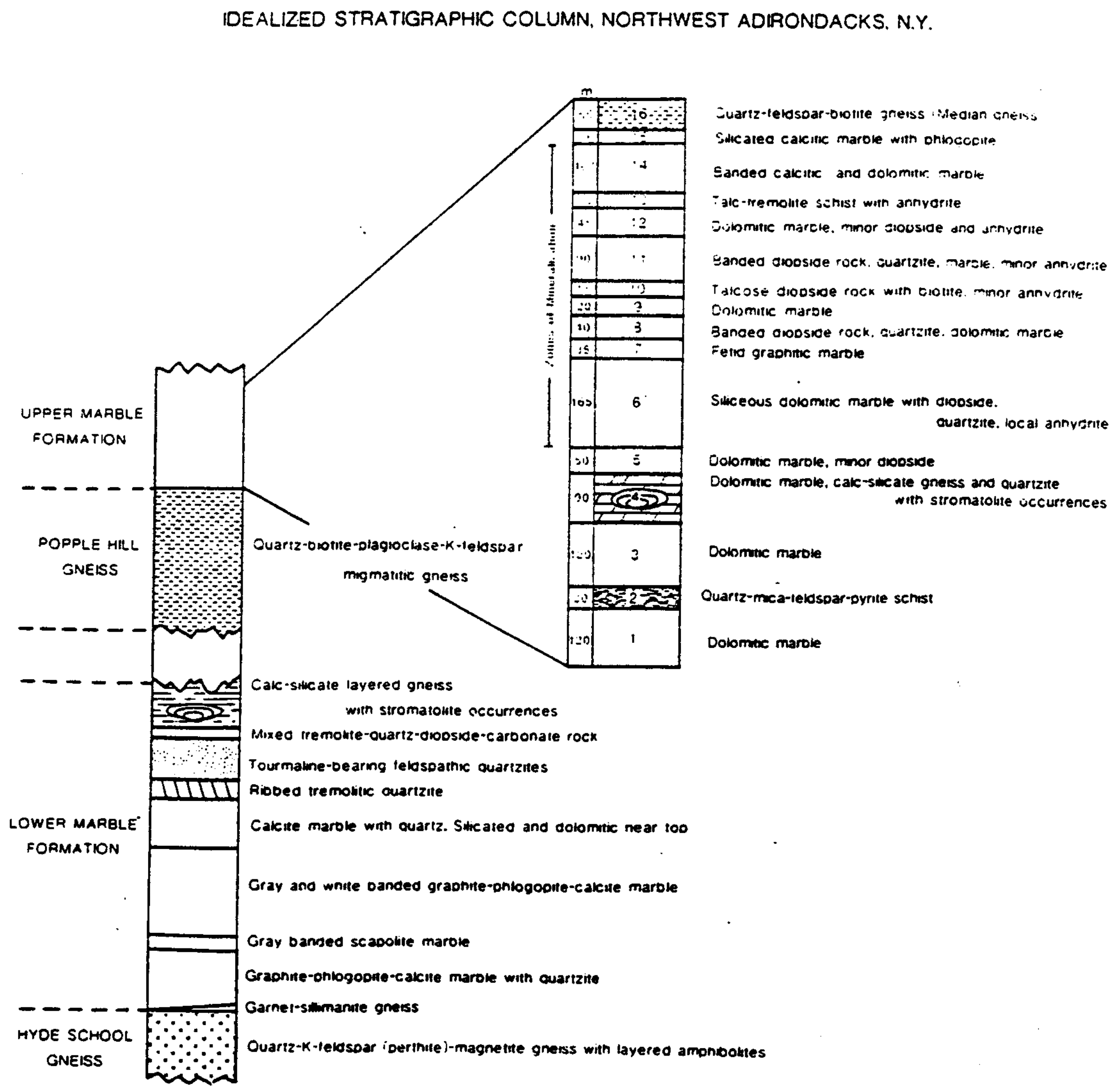


FIGURE 5 Proposed stratigraphic column for the northwest Adirondack lowlands. Undeformed contacts have not been described between the Lower Marble Formation and the Popple Hill Gneiss. The thickness of the upper members of the Lower Marble Formation is unknown. Members of the Upper Marble Formation are numbered according to Brown and Engel (1965); typical thicknesses in meters are shown at the left of the column (after deLorraine and Carl, 1986).

referred to as the Alexandria Bay Gneiss by Wiener and others (1984), but the latter name is misleading in that rocks of the type section may be significantly older than those in most of the other exposures (Chiarenzelli and others, 1987; also see discussion of Stop A). The leucogneisses have a generally rhyolitic composition (Table 1, Column A), although a tonalite-trondhjemite facies is locally present, and thin, conformable amphibolite layers are common. These rocks have been interpreted as metamorphosed ash-flow tuffs (Carl and Van Diver, 1975). Formational status for the Hyde School Gneiss has been proposed by Carl and others (in prep.). Overlying the leucogneisses is a thin, discontinuous garnet-sillimanite gneiss that, in turn, is overlain by the Lower Marble Formation. The lower part of the Lower Marble is dominated by graphite-phlogopite-calcite marbles. The upper part of this unit contains dolomites, diopsidic marbles, quartzites, biotite-quartz-feldspar gneisses, and calcsilicate gneisses, many of which are enriched in granular tourmaline. Similar marbles, quartzites, and tourmaline-bearing gneisses, northwest of the Thousand Islands in Ontario, are tentatively correlated with the Lower Marble Formation (deLorraine and Carl 1988).

The migmatitic Popple Hill Gneiss (biotite-quartz-plagioclase gneiss with varying amounts

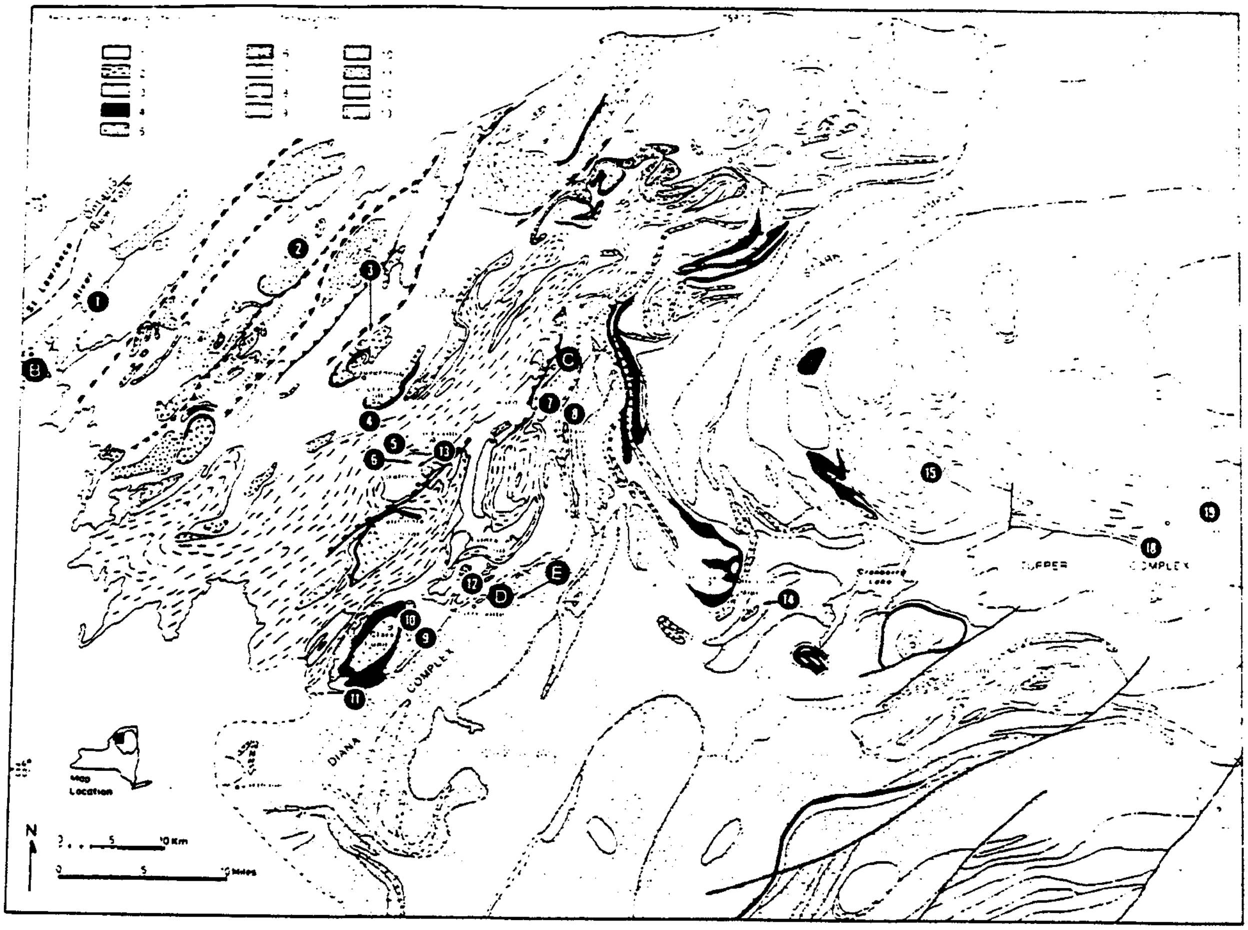


FIGURE 6 Generalized geological map of the northwest lowlands and the northwestern part of the Adirondack highlands. Numbers and letters designate stops. Rock Units: 1. Upper Marble (lowlands). 2. Popple Hill Gneiss (lowlands, includes numerous layers and lenses of granitic gneisses); undifferentiated biotite-quartz-feldspar gneisses (highlands). 3. Lower Marble (lowlands); undifferentiated metasedimentary rocks (highlands). 4. Aluminous gneisses. 5. Hyde School Gneiss (lowlands). 6. Metagabbro and amphibolite. 7. Leucogranitic gneisses (highlands). 8. Granitic gneisses. 9. Mangeritic and charnockitic gneisses. 10. Metanorthosite. 11. "Hermon" granitic gneisses (lowlands; not shown within Popple Hill Gneiss). 12. Antwerp-Rossie granitoids (lowlands). 13. Granitic gneisses of the Alexandria Bay area (lowlands).

of K feldspar) overlies the Lower Marble; the contact may be tectonic in part. This unit, spelled "Poplar Hill" by Wiener and others (1984), and formerly called the Major Paragneiss (Engel and Engel, 1958), has the geochemical characteristics of dacitic volcanics (Carl 1988). Pelitic and semi-pelitic lithologies are present but subordinate except near the base and top. Above the Popple Hill Gneiss is a second carbonate section, the Upper Marble Formation. This unit is dominated by well-layered dolomites and thinly bedded, quartzose (cherty) calcsilicate gneisses and diopsidic dolomites, with local occurrences of bedded anhydrite in the subsurface. Several of the sixteen mappable subunits (Fig. 5 and Brown and Engel, 1956) contain stromatolites (Isachsen and Landing, 1983). The Upper Marble is host to the zinc deposits of the Balmat-Edwards district.

The northwest Adirondack stratigraphy is a record of sedimentation in a mildly extensional intracontinental environment. Basal felsic volcanics represented by the Hyde School Gneiss are overlain by the lower of two carbonate sections, the Lower Marble Formation. The carbonates, evaporites and clastics of this unit record deposition in a relatively stable tectonic regime, perhaps in a shallow epicontinental sea with coastal sabkhas. Carbonate deposition was interrupted by dacitic volcanism (Popple Hill Gneiss), accompanied by minor reworking and pelitic sedimentation. At the top of the column, the Upper Marble Formation records a return to shallow water, stromatolitic cherty dolomite and evaporite deposition.

Highlands

Compared to the northwest lowlands, the stratigraphic picture in the Adirondack highlands (Figs. 7,8) is relatively obscure. Metasedimentary rocks of the highlands all appear to belong to shallow water, epicontinental or shelf-type sequences. Quartzites and metapelites dominate within the southern Adirondacks and give way to increasingly carbonate-rich rocks to the north

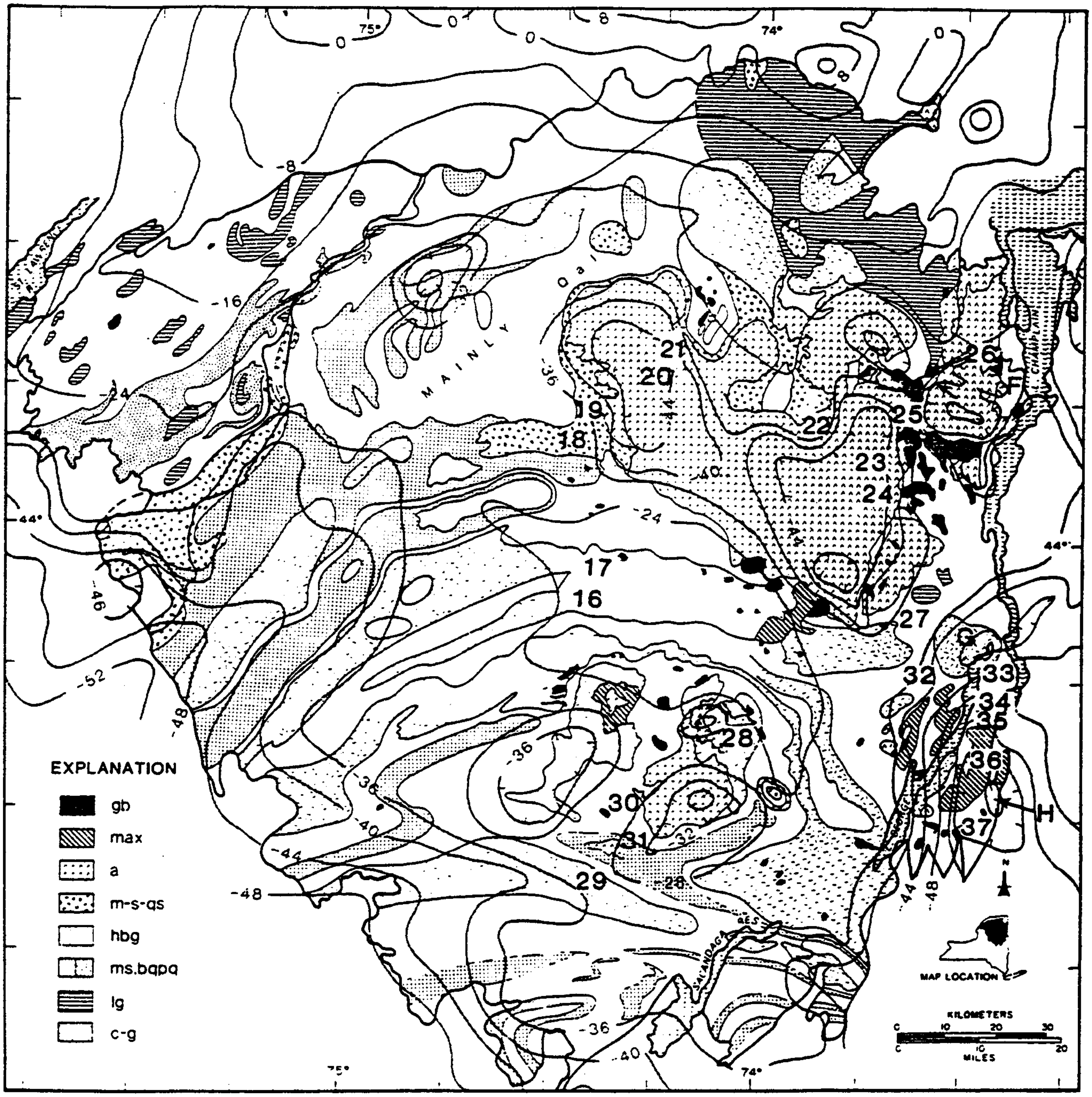


FIGURE 7 Generalized geological map of the Adirondack highlands. Gray contours: Bouguer gravity anomaly in mgal. Numbers and letters designate stops. Legend: gb, olivine metagabbro; max, interlayered anorthositic and mangeritic rocks; a, metanorthosite; m-s-qs, mangerite-syenite-quartz syenite; hbg, hornblende granitic gneiss; ms, undifferentiated metasedimentary rocks; bqpg, biotite-quartz-plagioclase gneisses; lg, leucogranitic gneisses; c-g, charnockitic and granitic gneisses.

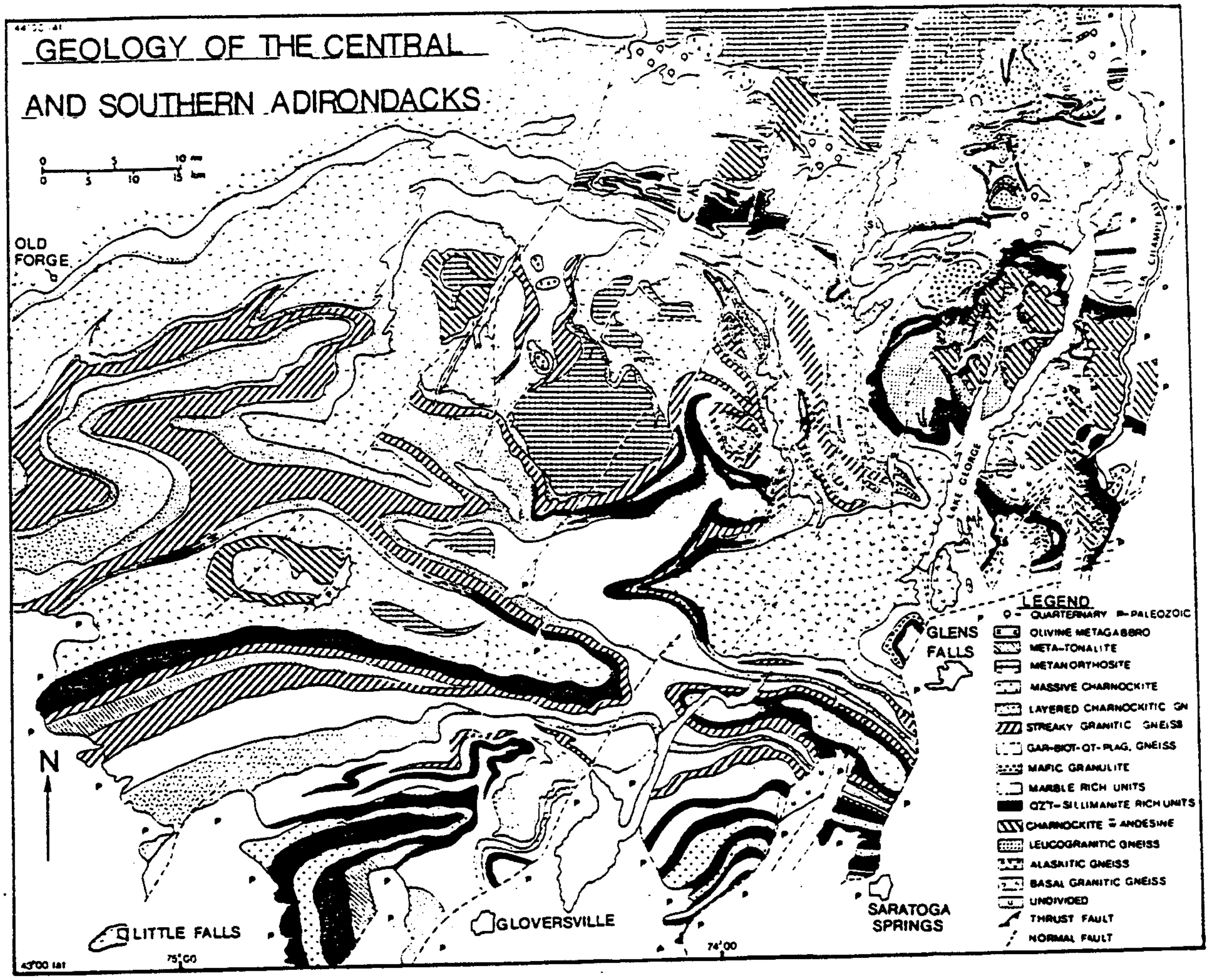


FIGURE 8 Geologic map of the south-central Adirondack highlands.

and west. In the west-central and northeastern highlands, there is considerable evidence that the protoliths of the metasedimentary rocks were deposited in a hypersaline environment, :ther restricted shallow marine or continental (lacustrine or playa). In general the metasedimentary sequence appears to be thickest and richest in clastics in the south and thinnest in the northwest. A metavolcanic component may become increasingly important towards the north and east.

Stacking order in the metasedimentary and metavolcanic gneisses is uncertain, as is the origin of the layering. Clearly some of the layering may be interpreted as primary stratification, i.e. quartzites and marbles. However, in many other instances, equally pronounced layering may be the result of tectonism; straight gneisses (Davidson and others 1982) are present locally in the southeastern highlands. In general, tectonic processes have resulted in profound internal lithic changes and disruption of the original stacking order, as well as transposition and reorientation of originally discordant lithic units into parallelism. These layered sequences include numerous quartzofeldspathi gneisses of uncertain ancestry and relative age; some of these are probably intrusive ir origin but are now conformable with the enclosing rocks. No unequivocal facing criteria have yet been recognized in the highlands (McLelland and Isachsen, 1980). Despite these problems, lithic sequences hav been defined and traced in the field for considerable distances, and their continuit makes possible the mapping of local and regional structures throughout the area.

Detailed descriptions of the layered rock are given in Wiener and others (1984) and McLelland and Isachsen (1980). Here we focu on the broader aspects of the lithic sequen and their implications for the structural framework of the Adirondacks.

Following Wiener and others (1984), McLelland and Isachsen (1985) distinguish a set of basal gneisses referred to as the Piseco Group. These quartzofeldspathic gneisses include mangeritic, charnockitic, hornblende granitic and alaskitic varieties.

In general, compositional layering is poorly defined but is locally manifested by mafic streaks and layers. Two principal subdivisions have been recognized, the lower of these being the Pharaoh Mountain Gneiss consisting of charnockitic and granitic facies which grade into one another along strike. Above the Pharaoh Mountain is the Brant Lake Gneiss, which consists of pink or gray, fine-to medium-grained equigranular, leucogranitic to quartz dioritic gneisses with local amphibolite interlayers. Rounded zircon grains suggest that this unit was reworked and partially water laid. Ubiquitous magnetite imparts a distinctive high aeromagnetic signature to these rocks. Locally, magnetite occurs in 0.5-1 m thick conformable layers, often accompanied by pyroxene-garnet skarns and albite-rich gneiss. These rocks, as well as leucogneisses within the structurally overlying Lake George Group, are similar in many respects to the Hyde School Gneiss of the Northwest Lowlands, where they have been interpreted as metamorphosed rhyolitic and dacitic ash-flow tuffs (Carl and VanDiver, 1975).

Recently, U-Pb zircon ages (Chiarenzelli et al., 1987) have cast some doubt on the basal position of the Piseco Group and the correlation of members assigned to it with the Hyde School Gneiss, as proposed by Wiener and others (1984). In the first place, the age of the Highlands units lies between 1160 and 1130 Ma, while the Hyde School Gneiss yields an age close to 1300 Ma. In addition, pink leucogranitic gneiss at the type locality (Piseco Lake) gives an age of 1150$\pm$5 Ma, but supposedly overlying metasedimentary rocks appear to be continuous with metapelites thirty kilometers to the south at Canada Lake, where they are intruded by a roughly 1300 Ma tonalite gneiss. Thus the Piseco Group appears to be partly intrusive and partly volcanogenic, and its status as a stratigraphic unit is dubious.

Apparently lying above the Piseco Group, and preserved in synclinal structures, is a complexly folded sequence of layered metasedimentary and metaigneous rocks. Typically these consist of marbles, calcsilicate rocks, quartzites, garnetiferous biotite-quartz-feldspar gneiss with varying amounts of graphite and sillimanite, leucogneisses, amphibolites, and a variety of granitic and charnockitic gneisses of possibly intrusive origin. This assemblage of layered rocks has been named the Lake George Group by Wiener and others (1984).

Uncertainty exists concerning the order of succession of units within the layered sequence, and the situation is further complicated by structural complexities, primary facies changes, lack of continuous outcrop, and large distances between well studied areas. Therefore, lithic sequences must be defined and utilized locally with great care taken in any attempts at regional correlation. In general, the Lake George Group consists of a lower marble-rich horizon overlain by a streaky, pink granitic gneiss which is followed upward by another marble-rich unit. Structurally above these units there occurs a sequence of quartzofeldspathic and garnet-biotite-quartz-feldspar gneisses. In the southern and southeastern Adirondacks the carbonate-rich units appear to undergo along-strike facies changes into quartzites and sillimanite-garnet-biotite metapelites (Fig. 8). However, the quartzite-rich sequence of the southernmost Adirondacks (Fig. 8) may include units not present to the north (Wiener and others, 1984). Toward the west-central and northeastern highlands, calcsilicate rocks with metaevaporitic affinities become increasingly abundant, and quartzofeldspathic leucogneisses are a major component of the layered sequence. If these changes can be clearly shown to be lateral, correlation across much of the Adirondack highlands may be possible.

IGNEOUS ROCKS

Northwest Lowlands

Metaigneous rocks in the northwest Adirondacks range from gabbros and diorites through monzonites to hornblende and biotite granites and syenites. Metagabbro within the northwest Adirondacks occurs as oval- to lens-shaped masses, generally several km across. The gabbros are older than the pegmatites and assorted granitoids that transect them, but igneous textures are locally preserved in their interiors. Relationship, if any, to the anorthosite suite of the Adirondack highlands and to the diorites and other igneous rocks associated spatially with the metagabbros has yet to be established.

The granitoid rocks are an important but enigmatic group, and we are in the early stages of sorting them out. These gneisses are generally conformable, sheet-like bodies. Evidence of intrusive origin includes aplite and pegmatite veins, apophyses extending into adjacent metasediments, xenoliths, and possible zoned K-feldspar phenocrysts. These features, however, are often obscured by foliation and grain size reduction, and by disruption during folding, particularly in the older and smaller intrusions. Buddington's (1939) Antwerp granite, the diorites and granitoids near Rossie, and the Huckleberry

Mountain granite just east of the Beaver Creek lineament (Brown, 1983 and in prep., see Fig. 6), are examples of small plutons and sills intrusive into surrounding metasediments. To the northwest, larger granitoid bodies are intusive into quartzites and gneisses in the Thousand Islands (Fig. 6 and Stop A).

Inequigranular biotite granitic gneisses characterized by abundant K-feldspar megacrysts are widespread throughout the northwest Adirondacks. These sheet or sill-like bodies, in various stages of grain-size reduction, are well exposed near Hermon village. Evidence for intrusive origin for "granite of the Hermon type" (Buddington, 1939) includes occurrences at various stratigraphic levels including both the Lower and Upper Marble Formations and as numerous thick layers and lenses within the Popple Hill Gneiss. Of great interest are the megacrysts. Some are euhedral, zoned, and at angles to the gneissic foliation; others are mylonitized, and the rock is best described as augen gneiss. It is uncertain whether these megacrysts are phenocrysts or porphyroblasts. Preliminary work on these rocks suggests a wide span in age and possibly diverse modes of origin.

The failure to distinguish metasedimentary from metaigneous rocks has led to problems of correlation. Buddington (1939) originally described rocks near Rossie Village (Figs. 4,6) as diorites intruded in sill- or lens-like form, but Lewis (1969), without elaboration, regarded them as metasediments and correlative with Popple Hill Gneiss. Wiener et al. (1984) gave them formational status as the "Pleasant Lake Gneiss" at the top of the northwest Adirondack stratigraphic column. These authors also correlated the intrusive granitic rocks in the Thousand Islands (Stop A) with the Hyde School Gneiss (Stop 2). Geochemical and other arguments against such correlations are presented by Carl, deLorraine and others (in prep.), and recent geochronologic data (Chiarenzelli and others, 1987) cast doubt upon any correlation of the Hyde School with the Thousand Islands granitoids. Such divergent opinions about correlations across distances of a few km do not bode well for regional correlations of specific units with similar rocks in either the Central Metasedimentary Belt of Ontario or the Adirondack highlands.

The domical bodies of Hyde School Gneiss were long considered to be igneous rocks; Buddington (1929, 1939) described them as "phacoliths", concordant minor intrusions occupying the crests or troughs of folds. This view was commonly held until the early 1960's, when Engel and Engel (1963) suggested that they are metasomatically altered quartzites. Lewis (1969) regards these gneisses as a widespread stratigraphic unit. Detailed mapping of amphibolite layers and new petrographic and geochemical data reveal a lithic sequence and a genetic relationship between major and minor folds (Carl and VanDiver, 1975). The amphibolite layers are not xenoliths but concordant layers folded together with the surrounding rocks. The dome-like structure and elliptical map pattern may result from interference of late open folds with early isoclinal folds (Foose and Carl, 1977); alternatively they may be a manifestation of curvilinear fold hinges (see discussion of structure). These observations, as well as the geochemical similarity between these rocks and certain compositionally zoned volcanics of the western United States (Carl and VanDiver 1975), lead us to believe that the protolith of the Hyde School Gneiss was volcaniclastic rather than intrusive, and that the several bodies all occupy the same stratigraphic position.

Highlands

The intrusive origin of many rocks of the Adirondack highlands can be demonstrated locally by relict igneous texture, crosscutting relationships, and the presence of rotated xenoliths. Where these features are not present, large-scale compositional homogeneity, lack of prominent compositional layering, relatively coarse, equigranular textures, and similarity to known intrusives are often used to infer an intrusive origin. Intrusive rocks in the highlands include the following types: Metanorthosite and gabbroic anorthosite gneiss; jotunitic, mangeritic, and quartz syenitic gneisses; charnockitic and granitic gneisses; and olivine metagabbros. Quartz dioritic gneisses are present locally in the southern and southeastern areas.

Metamorphosed anorthositic rocks underlie much of the central and northern Adirondacks and are best exposed in the Marcy Massif (Figs. 3,6,7). Smaller domical bodies include the Snowy Mountain and Oregon domes (Figs. 8,12). Generally, these bodies consist of coarse-grained andesine anorthosite to anorthositic gabbro, with the former being dominant. Gabbroic anorthosite gneiss is concentrated near the massif margins. Late differentiates include ferrodiorites and ferromonzodiorites (jotunites), ferrogabbros, and ultramafic sheets or dikes rich in Fe-Ti oxides, Fe-rich pyroxenes, and apatite. A late, crosscutting oxide-pyroxene rock with minor plagioclase has been mined for ilmenite and magnetite at Sanford Lake near the southern margin of the Marcy Massif; it may be a late differentiate or a remobilized cumulate

(Ashwal, 1978).

Simmons (1964) showed that the major anorthosite bodies of the Adirondacks are associated with strong negative gravity anomalies (Fig. 7). Aeromagnetic surveys (Zeitz and Gilbert, 1981) show strong negative magnetic anomalies also associated with the large anorthosites. The presence of strong negative gravity and magnetic anomalies to the east of Lake George (Fig. 7) is consistent with the presence of one or more large anorthositic masses at shallows depths in this region. In the northwestern highlands gravity lows extend to the north and west of the Marcy Massif (Fig. 7) and suggest the presence of subsurface anorthosite as far northwest as the Carthage-Colton zone (Buddington, 1969). Small bodies of anorthositic rocks occur at the surface close to the CCMZ south of Russell (Fig. 6) and near Carthage, as well as locally within the Diana Complex (Hargraves, 1969). Simmons (1964) interpreted the gravity data from the Marcy Massif as indicating that the shape of that body is a 3-5 km thick slab with at least two deep extensions that may be either feeder pipes or diapiric roots. However, Morse (1969, 1982) has stressed that these interpretations depend critically on average densities which are difficult to estimate, and therefore other shapes are not excluded. More detailed gravity work by Mann and Revetta (1979) in the northeastern part of the Marcy Massif suggests a multidomical structure of the anorthosite in that area. In addition to the large, possibly composite, anorthosite intrusions, smaller stratiform to lensoid anorthosites are commonly present in the layered metasedimentary rocks of the central and northeastern highlands (Isachsen and others 1975, Beddoe 1981).

Metamorphosed orthopyroxene-bearing mangerite and quartz syenite are commonly present at the margins of the large anorthosite bodies (Figs. 6,7,8). These rocks, which locally crosscut the anorthosite, form a partial envelope around the Marcy Massif and completely surround the Snowy Mountain body. Blue-gray andesine xenocrysts, evidently derived from the anorthosite, are common in these rocks close to anorthosite contacts and are occasionally found up to 10 km from the nearest exposed anorthosite. Rapakivi textures are locally present within quartz syenitic gneisses of the Stark anticline and Diana Complex (Buddington 1939).

The mangerites have been variously interpreted as post-anorthosite intrusives (Buddington 1939); differentiates from a common granodioritic magma that also produced anorthosite (deWaard 1969); and as contact anatectic melts (Isachsen 1969). Both field evidence (Buddington 1939) and trace element patterns (Simmons and Hanson 1978; Ashwal and Siefert 1980) appear to rule out models involving consanguinity with the anorthosites. The presence of mafic mangerite next to the anorthosite, possibly due to mixing of quartz mangeritic magma and mafic differentiates of the anorthosite suite, as well as local permeation of the anorthosite by mangerite and the presence of andesine xenocrysts in the mangerite, taken together suggest that the mangerite and anorthosite are coeval.

Granitic gneisses in the Adirondack highlands (the Piseco Group in part) are of two principal types. One is a charnockitic gneiss with hornblende and two pyroxenes; this rock has a characteristic olive-gray color and weathers to a maple-sugar brown. The other is a pink to white hornblende granitic gneiss, without pyroxenes but locally containing biotite. In the field, these two rocks show gradational relationships, and geochemically (Table 1, columns H and I, and Whitney, 1986) they appear to be a single differentiated suite. The more mafic charnockites overlap compositionally with the quartz syenites, and the entire mangerite-quartz syenite-charnockite-granite suite may be comagmatic. Pink leucogranitic gneisses (the alaskites of Buddington and Leonard, 1962) may be felsic late differentiates of the charnockite-granite suite; they are often difficult to distinguish from the (metavolcanic ?) leucogneisses among the layered rocks of the Lake George Group.

Extensive, sill-like bodies of quartz dioritic to tonalitic gneiss have been reported from the Lake George region (McLelland 1986a) and the southern Adirondacks (McLelland and Isachsen 1980). In the field these rocks resemble charnockites and mangerites, but they are low in potassium (Table 1, Col. R), and their feldspars are almost wholly plagioclase (An_{30-40}). They invariably contain thin amphibolite layers that have been dismembered into angular boudins. Quartz content averages 15-25%.

Numerous bodies of olivine metagabbro and metatroctolite are scattered throughout the eastern and southern Adirondacks; these rocks are scarce to absent in the western highlands. The greatest concentration of metagabbros as well as the largest bodies are found along the eastern and southern margins of the Marcy anorthosite massif. Several of the larger bodies show a pronounced igneous layering. In a few locations, these rocks are seen to crosscut the anorthosites, but they may be only slightly younger. It is possible that these rocks may be intrusions into upper- or mid-crustal regions of olivine tholeiite magmas associated with anorthosite genesis in the upper mantle or lower crust (Emslie, 1978). However, the metagabbros are relatively iron-rich (Table 1, Cols. P,Q) compared to undifferentiated tholeiites and thus have evidently undergone significant fractionation prior to emplacement. They differ from mafic members of the anorthosite suite (Table 1, Cols. N,O) in that the latter are usually quartz-normative, while the olivine metagabbros are strongly silica-undersaturated.

Pegmatitic and granitic dikes, both deformed and undeformed, are scattered throughout the region but are most common in the southeast. An Rb/Sr muscovite age of 963±40 Ma (B. Giletti, written communication to YWI) obtained from a late undeformed pegmatite from the southeasternmost Adirondacks, suggests that these are young pegmatites associated with late Middle Proterozoic uplift and cooling of the Adirondack metamorphic terrane. However, Putman and Sullivan (1979) have shown that the composition of some of these dikes is consistent with an origin at high (7 kbar) pressure, after cessation of deformation but before major uplift had occurred.

METAMORPHISM

Pressure-Temperature-Time

Peak metamorphic temperatures as determined by feldspar and oxide thermometry are shown in Figure 9 (after Bohlen and others, 1985). Determination of metamorphic temperatures involves reintegration of albite exsolved from alkali feldspar and ilmenite (oxidized ulvospinel component) exsolved from magnetite. Details of this technique are given in Bohlen and Essene (1977). The overall metamorphic temperature regime as inferred from oxides and feldspars is supported by numerous other independent thermometers and by silicate, silicate-carbonate and silicate-sulfide equilibria. For example, results from calcite-dolomite (Goldsmith and Newton, 1969), calcite-graphite (Valley and O'Neil, 1981) and garnet-clinopyroxene (Ellis and Green, 1979) thermometers generally corroborate the results obtained from oxides and feldspars. In addition, occurrences of such assemblages as phlogopite-calcite-quartz, tremolite-calcite-quartz, grossular-rich garnet-quartz, akermanite-wollastonite-monticellite, and Fe-rich orthopyroxene-clinopyroxene throughout the Adirondacks provide useful limits on metamorphic temperatures. For example, the coexistence of grossularitic garnet and quartz, combined with the general absence of coexisting wollastonite and plagioclase, as reported by Valley and Essene (1980a,b), constrains metamorphic temperatures below 850^{0}C. Similarly, occurrences of Fe-rich orthopyroxene (Fs_{85}-Fs_{95})-clinopyroxene pairs in charnockites throughout the Adirondack highlands (Davis, 1969; Jaffe and others, 1978; Bohlen and Boettcher, 1981) and absence of metamorphic ferropigeonite in the same or similar rocks, limit metamorphic temperatures to below 825^{0}C (Lindsley, 1983). Such limits are important because they also constrain the maximum temperature in any pervasive, regional metamorphic event involving these rocks prior to the main Grenville metamorphism. The general agreement of various thermometers and phase equilibria confirms the results of oxide and feldspar thermometry, as well as providing evidence that the inferred temperatures record peak, or near-peak, metamorphic temperatures.

Estimates of metamorphic pressures in the Adirondacks have encompassed the range from 2-12 kbar. However, recent calibrations of numerous independent barometers have greatly reduced this uncertainty. Figure 10 shows the pressures of Adirondack metamorphism recorded by the rocks at or near the thermal maximum as deduced from eight barometers, including garnet-rutile-sillimanite-ilmenite-quartz (Bohlen and others, 1983b), ferrosilite-fayalite-quartz (Bohlen and Boettcher, 1981), sphalerite-pyrrhotite-pyrite (Brown and others, 1978), fayalite-anorthite-garnet-quartz (Bohlen and others, 1983a), garnet-sillimanite-plagioclase-quartz (Newton and Haselton, 1981, as applied by numerous workers), kyanite-sillimanite (Richardson and others, 1968), and the stability of akermanite (Valley and Essene, 1980b). Figure 10 includes for comparison the orthopyroxene and garnet-clinopyroxene "isograds" of deWaard (1969). Most of these barometers are insensitive to temperature; knowledge of temperature to within ±50^{0}C generally allows determinations of pressure to within ±0.5 kbar. The wide variety of well-calibrated barometers that apply in both peraluminous and metaluminous rocks yield a remarkably coherent picture of the pressure of metamorphism at the thermal maximum. Pressure increases from 6.0-6.5 kbar in the area around Gouverneur in the northwest lowlands to 6.5-7.0 kbar along the Highlands-Lowlands boundary, approaching 7.5 kbar near Colton and reaching 7.5-8.0 kbar throughout the central Adirondack highlands. The accuracy of these metamorphic pressures is estimated to be ±0.5 kbar, with most of the uncertainty resulting from inaccuracies in the solution parameters of garnet and pyroxenes. Inasmuch as the present crustal thickness in the Adirondacks is on the order of 35 km (Katz, 1955), these pressures, corresponding to depths of 25-30 km, indicate a nearly double crustal thickness (60-65 km) at the time of the Grenville orogeny (ca. 1100-1050 Ma), comparable to the modern Tibetan Plateau or

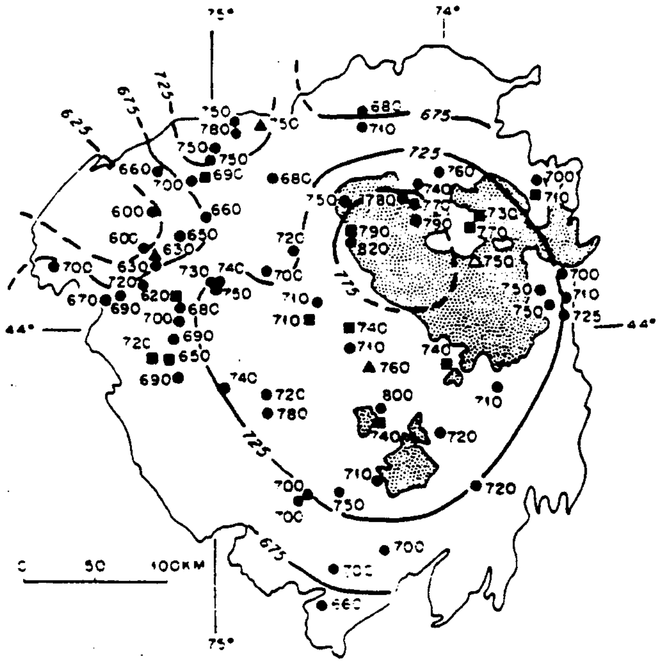


FIGURE 9 Metamorphic temperatures, in °C, after Bohlen and others (1985). Temperatures from coexisting feldspars (filled circles); magnetite-ilmenite (squares); calcite-dolomite (filled triangles); and akermanite (open triangle). Stippled area: anorthosite.

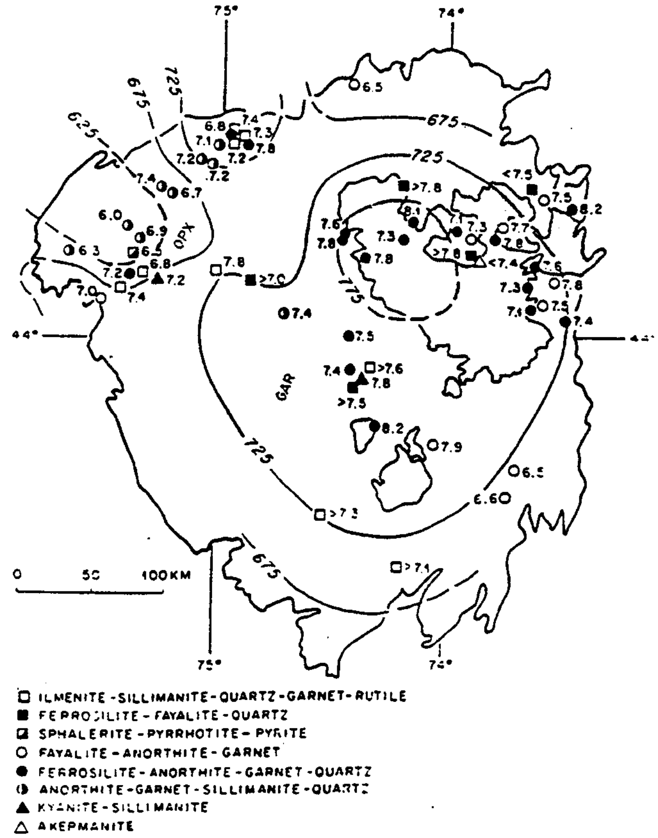


FIGURE 10 Metamorphic pressures, after Bohlen and others (1985). Isograds (not precisely located), orthopyroxene in metamafites (OPX) and garnet in quartzofeldspathic gneisses (GAR), after deWaard (1969). Contours are temperatures from Figure 9.

the Andean arc.

The pressure-temperature-time path of regional granulite facies metamorphism in the Adirondacks as deduced by Bohlen (1987) is shown in Figure 11. A similar path was independently estimated by Lamb (1987) from measurements of CO_2 density in fluid inclusions that are too high for the PT conditions of regional metamorphism and which therefore require a cooling and uplift path that is concave toward the temperature axis. Although a complete discussion of how the path in Figure 11 was constructed is beyond the scope of this review, the path can be inferred from compositional variations in rims of garnet and pseudomorphs of one mineral after another, especially sillimanite after andalusite as recently recognized by Bohlen (unpublished data). Sillimanite is the sole Al_2SiO_5 phase in Adirondack rocks, with two exceptions: one locality in the highlands (Boone 1978) and one in the lowlands (Wiener 1981), at both of which kyanite occurs with sillimanite in texturally ambiguous relationships. As discussed by Bohlen (1987) the type of P-T-time path exhibited in Figure 11 is characteristic of many regional granulite terranes and indicates that magmatic heating before and during compressional tectonism is an important feature of granulite development. A possible tectonic setting likely to generate such a P-T-time path is an Andean-type subduction zone. With advances in geochronology, the time coordinates of the path in Figure 11 will be quantified.

Metamorphic Fluids

In the Adirondacks, as elsewhere, the composition and role of metamorphic fluids has been a matter of great debate. The genesis or chemical modification of nearly every rock in the Adirondacks has, at one time or another, been attributed to pervasive infiltration by large amounts of fluid during regional metamorphism. However, within the last decade, the transition from amphibolite to granulite facies has been shown to be directly related to decreasing activity of water, as well as to increasing temperature. Available evidence now supports either magmatic or pre-metamorphic processes, or both rather than massive metasomatism as the explanation for many of the characteristic features of granulite-

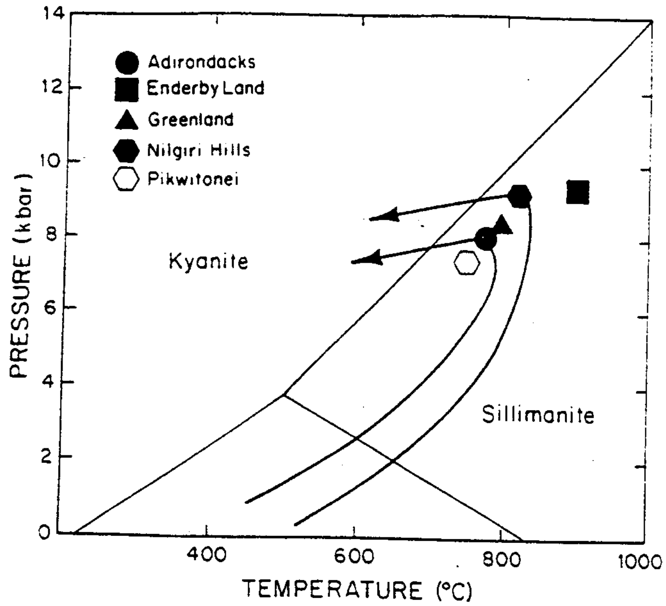


FIGURE 11 Pressure-Temperature-Time path for Adirondack highlands, after Bohlen (1987).

facies metamorphic terranes.

At least three periods of fluid activity can now be recognized. First, hydrothermal activity occurred at the contacts of shallow plutons, including anorthosite, before regional metamorphism. Then, widespread partial melting and some localized fluid migration took place during regional metamorphism. Finally, small quantities of fluid formed retrograde minerals after the peak of regional metamorphism.

Fluid conditions during regional metamorphism in the Adirondacks were heterogeneous and locally variable (see review by Valley and others, in press). Quantitative estimates of water activity show that values in the granulite facies were 0.2 $\pm$ 0.1, but the transition to low aH_2O is not smooth. Low aH_2O is also estimated for the migmatitic Popple Hill Gneiss (uppermost amphibolite facies), and no correlation is seen between aH_2O and either distance to the orthopyroxene isograd, or metamorphic temperature. Fluid pressure was significantly below lithostatic pressure in many granulite facies gneisses, including some charnockites and calc-silicates, indicating that metamorphism was fluid-absent; i.e there was no free fluid phase. Fluid-absent metamorphism has resulted from prior partial melting in some rocks and from pre-granulite contact metamorphism in others. In contrast, total fluid (H_2O + CO_2) pressure was close to lithostatic pressure in many thick marble units of the granulite facies. Oxygen fugacity varied from values near the hematite + magnetite buffer in some iron-rich metasedimentary rocks, to substantially below the quartz + fayalite + magnetite buffer in orthogneisses. The anorthosite suite is, in general, more oxidized than associated granitic gneisses (Valley and others, 1989). Low values of fO_2 and fH_2O demonstrate fluid-absence in some gneisses (Lamb and Valley, 1984, 1985).

F and Cl are important substitutions in micas and amphiboles. In iron-poor marbles, F substitutes for OH up to 96 mole percent in phlogopite and 82 percent in tremolite (Valley and others, 1982). Grossular rims on altered wollastonite (Stop 10) contain up to 0.7 wt. percent F (Valley and others, 1983). Fluorine solid solution significantly extends the stability of phlogopite and tremolite at granulite facies conditions. Chlorine substitution is greatest in more iron-rich compositions, especially orthogneisses. Hornblende rims on oxides in metanorthosites contain up to 3.2 wt. percent Cl, which Morrison (1988) attributes to a residual magmatic origin.

Polymetamorphism is documented at skarn zones adjacent to anorthosite where large quantities of hydrothermal fluid (with significant meteoric water) were channeled during shallow contact metamorphism. Wollastonite-bearing skarns were preserved through subsequent granulite facies recrystallization by CO_2-poor and generally fluid-absent conditions in these zones.

Several lines of evidence support the conclusion that CO_2 infiltration has not been an important process in the Adirondacks and that low water activities were the result of partial melting reactions plus, locally, an earlier high T, low P contact metamorphism. 1) Widespread textural evidence exists for partial melting, in the form of migmatites, pyroxene-bearing "sweats", and localized magmatic intrusion. 2) Stable isotope signatures indicate that many rocks, including anorthosite, were not infiltrated by significant quantities of H_2O-CO_2 fluid. This is seen in the preservation of pre-metamorphic oxygen and carbon isotopic compositions and sharp isotopic gradients (Valley and O'Neil, 1984). 3) In many rocks phase equilibria show either low fCO_2; fluid-absent conditions; or sharp gradients in XCO_2, fCO_2, or fO_2. 4) Low $\delta^{13}C$ graphites (-20 to -27.3) are preserved in many non-carbonate rocks (Weis and others, 1981).

Fluid inclusions containing high density (up to 1.15) CO_2 have been found in many Adirondack rock types. These inclusions satisfy the criteria that are typically applied in other granulite facies terranes for samples of peak metamorphic fluids. However, the Adirondack CO_2-rich inclusions have been found in low-fCO_2 rocks, including wollastonite skarn and charnockite, indicating that the inclusions were trapped after the peak of granulite metamorphism (Lamb and others, 1987). The densities of these retrograde inclusions define a PT path that is concave toward temperature (Lamb and others, 1987). Cathodoluminescence reveals microscopic, late, veins of calcite and other retrograde minerals occupying brittle fractures in orthogneisses, including anorthosite (Morrison and Valley, 1988b). These veins are believed to be the pathways for trace quantities of 300-500^0 C fluids which fed the fluid inclusions.

STRUCTURE

Northwest Lowlands

Four to five phases of folding are recognized in the northwest Adirondacks (Wiener and others, 1984; deLorraine 1979). The first phase resulted in the development of isoclinal folds with a prominent axial planar foliation. This is the regional foliation; it developed at upper amphibolite facies. The scale of early isoclinal folding and the extent to which first phase isoclinal folds influence regional map patterns by repetition of stratigraphic units is uncertain. Migmatization in gneisses of appropriate composition occurred prior to or during first phase folding, as shown by second phase folds in migmatite veins (e.g. at Stop 5). Second phase folds are isoclinal in style and overturned to the southeast. Second phase fold axial surfaces trend NE-SW producing the prominent regional grain of the northwest Adirondacks. Variability in trends and plunges of second phase fold axes is related in part to "porpoising" of the hinges within the axial surfaces. Important structural features of major second phase isoclinal folds are revealed by mine workings in the Sylvia Lake syncline. These include axial plane shears and thrust faults, large fold hinge tectonic fish, tectonic slides that excise sections of stratigraphy, and curvilinear (sheath) fold hinges. A good example of a sheath fold in marble is exposed at the Wight talc mine (Stop 13B). Second phase folds refold an earlier foliation and schistosity.

Third phase folds are less prominent than second phase folds and tend to be upright to moderately overturned to the southeast. Axes of third phase folds are approximately coaxial with the second phase folds so that interference between the two produces hook-shaped interference patterns (Figs. 6,12). Fourth phase folds are gentle, open, and trend to the northwest (Foose 1974; Wiener and others 1984; deLorraine 1979). Previously t modified regional dome and basin pattern o the northwest Adirondacks (Fig. 6) was ascribed to the interference of NE trendin second and third phase folds with NW tren fourth phase folds. An alternative interpretation is that the ovoid outcrops o Hyde School Gneiss are apical exposures of curvilinear, "porpoising" second phase hin In this context the "domes" cored by Hyde School Gneiss are viewed as F2 regional tongues or sheath folds that project upwar through the erosional surface rather than a fold interference structures.

Metamorphism and deformation were accompanied by plutonism ranging from gabbr to granodioritic to granitic and syenitic. There is some evidence that suggests that metagabbros intruded prior to or during ea phase folding. Tabular bodies of equigranu and megacrystic Hermon granite gneiss are folded by second phase isoclinal folds. Foliation in the granite gneiss is either axial planar to, or folded by, the second folds. Near the Carthage-Colton zone, syen of the Diana Complex intruded before or du second phase isoclinal folding, and a pronounced mylonitic fabric in the rock is axial planar to the F2 folds (Wiener 1981). Other intrusive bodies such as the Rossie a Antwerp granitoids are also deformed by sec phase folds. Late, relatively undeformed granites and pegmatites are minor.

First and second phases of folding were accompanied by mylonitization in a ductile regime (Hudson and others 1986; Wiener and others 1984; Brown, 1988). Mylonites occur pelitic rocks at the base of the Popple Hi gneiss in a belt from Antwerp to Hermon (Ambers and Hudson, 1985). It is uncertain whether they mark a major thrust contact ne the base of the Popple Hill, or merely rec a greater susceptibility of pelitic rocks t grain size reduction. Regional mapping has shown that parts of the stratigraphic succession are missing locally, in a way th is consistent with thrusting. Foose (1974) interpreted the Lower Marble- Popple Hill gneiss contact as a folded tectonic slide. I so, the relative age of thrusting is constrained by second phase folds that fol the mylonite. In general, mylonites in the Adirondack lowlands away from the Carthage-Colton zone are more common than is sugges in the literature.

The Elm Creek slide at Edwards (Stop C) i zone of major dislocation and offset. Here t entire section of Popple Hill gneiss was excised along a ductile shear zone identifi by juxtaposition of Upper and Lower marbles separated only by mylonitized slivers of

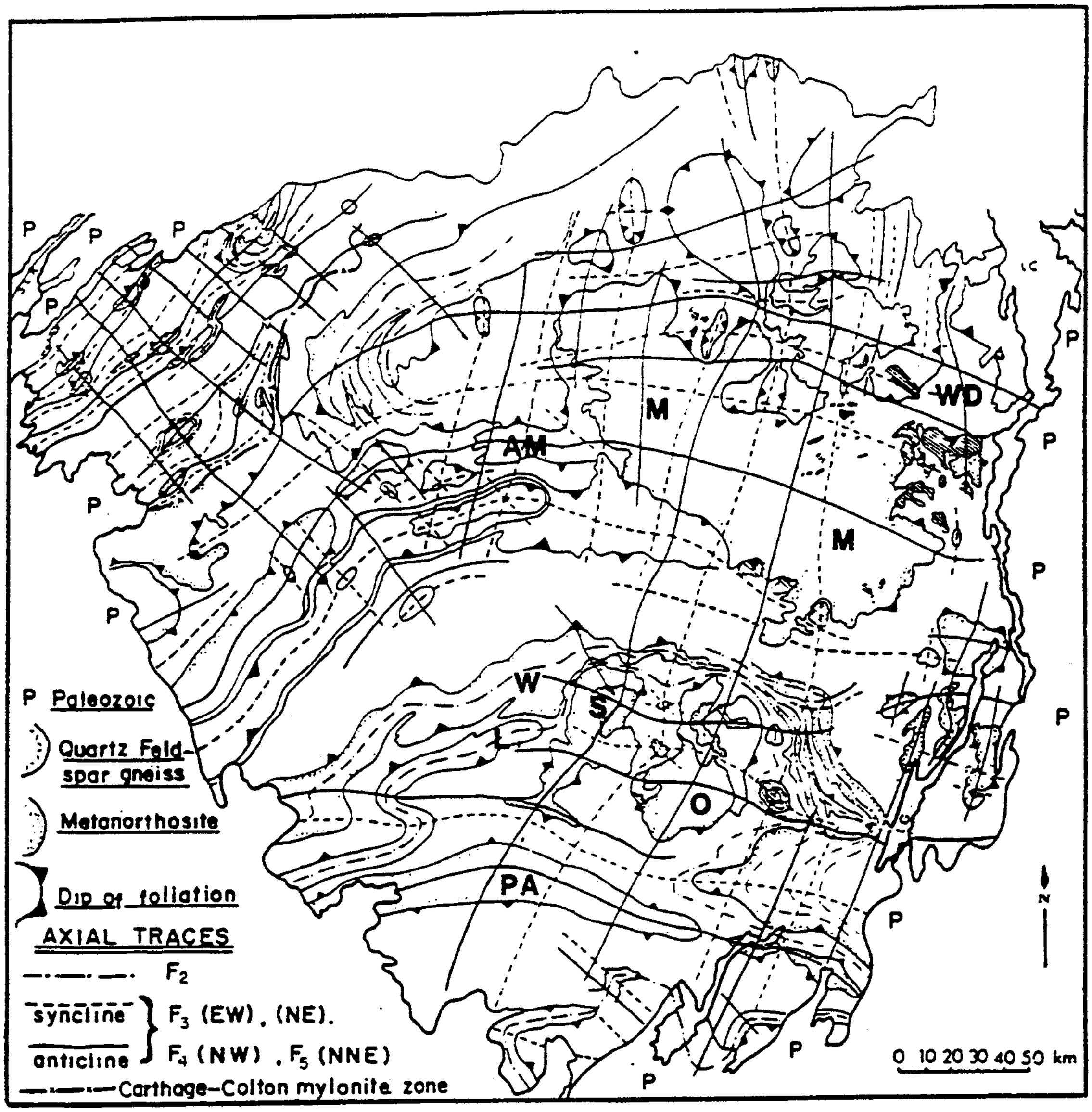


FIGURE 12 Generalized structural map of the Adirondacks. Fold generations (F_4 excepted) may not be correlatable from highlands to lowlands. Am, Arab Mountain anticline; L, Little Moose Mountain syncline; M, Marcy Massif; O, Oregon Dome; PA, Piseco anticline; S, Snowy Mountain Dome; W, Wakely Mountain nappe; WD, Westport Dome.

granite. It was this structural juxtaposition of marbles that led Foose (1974) and Wiener and others (1984) to propose a stratigraphic column with only one marble unit.

Late NE- trending faults and lineaments separate the Precambrian rocks into panels that differ somewhat in stratigraphy and structural trend (Brown 1973, 1980). Stratigraphic differences reflect variations in thickness of strata from panel to panel and different erosional levels across panels. There is Paleozoic offset along some of these faults, while Proterozoic intrusives lie along traces of others (Brown, 1988).

Carthage-Colton Zone

The Carthage-Colton Mylonite Zone (CCMZ) has been described and discussed by Geraghty and others (1981) and Isachsen (1985). It is one of several major NE-trending discontinuities in the southern part of the Grenville Province (Davidson and others, 1982), although it is unique among these in having a NW rather than SE dip. It forms the 110 km border between the Adirondack highlands and the northwest lowlands, and extends both NE and SW beneath Paleozoic cover; the dip is NW and variable, averaging about 45^0. The thickness of the CCMZ ranges from 3 m to 5 km. It consists of anastomosing domains of very high strain that enclose lenticular bodies of less strained rock. Field recognition of mylonitized rock is based on grain size reduction and, except in granofelses, enhanced foliation or lamination. These characteristics are best shown in quartzites and granitoids - especially megacrystic varieties (s-c mylonites) - and, to a lesser degree, in anorthositic metagabbro, metapelite, amphibolite and calcsilicate rock. Marbles rarely show mylonitic fabric. Ultramylonites are uncommon, narrow, and localized. Dynamic recrystallization at upper amphibolite to granulite facies conditions (630^0-760^0C; total P ~6-7 kbars) indicates that the depth of ductile faulting was on the order of 20 km. Subsequent reactivation of this zone is locally apparent. Cartwright and others (in prep.) have studied small cross-cutting shear zones and find that deformation correlates with increased $\delta^{18}O$, growth of secondary biotite, and resetting of Fe-Ti oxide thermometers (see Stop 7). Two mylonitization events have also been recognized by Heyn, Weathers and Bird (1987).

Fabric asymmetries within the CCMZ, considered statistically, indicate a prevailing NW (hanging wall down) sense of transport. The amount of strain cannot yet be estimated, but the configuration itself suggests a deep-crustal analog of Cordilleran-type metamorphic core complexes, with mid-crustal rocks dragged downward to lower crustal levels along a detachment surface. Such extensional tectonics may have accompanied lateral spreading of the doubly-thickened Grenville crust after cessation of orogenic compression ca. 1.1-1.05 Ga.

Highlands

The structural framework of the highlands is characterized by large folds, including an early set of E-W isoclines that may represent partially exposed sheath folds. Because of the existence of an earlier foliation those isoclines are designated F_2. Within the central and southern Adirondacks F_1 folds have been recognized only as minor intrafolial folds coaxial with F_2. In most of the Adirondacks the early foliation is axial planar to F_1 and is folded by minor and major F_2 folds.

The large, regional fold nappes of the Adirondacks belong to the F_2 generation (Fig. 12). These are isoclinal and either recumbent or reclined. Many have exceptionally large dimensions, the largest being the Canada Lake nappe, Wakely Mountain nappe, and the associated Little Moose Mountain syncline (Fig. 12, McLelland and Isachsen, 1980). These folds are accompanied by a strong axial planar foliation which cannot be distinguished from the F_1 foliation except in F_2 hinge areas. Within the southern and central Adirondacks, F_2 axial traces are curvilinear, averaging approximately E-W (Fig. 12), while in the northwest highlands the axial traces swing towards the NE. In both areas F_2 fold axes trend from EW to WNW. Kinematic indicators suggest dominantly SE-over-NW movement.

F_3 folds in the southern and central highlands are open and upright with E-W axes and axial traces (Fig. 12), but become tight or even isoclinal with a northeasterly trend in the northwest lowlands, if indeed the early fold phases in the highlands and lowlands can be correlated. The F_3 folds have very large dimensions, as exemplified by the Piseco anticline in the southern highlands. These folds are accompanied by a weak, locally developed, axial plane foliation. Interference of F_2 and F_3 folds forms hook-shaped outcrop patterns.

F_4 folds trend NW and are open and upright, and are best developed in the northwest lowlands (Fig. 12) where they form basin and dome interference patterns with F_3 folds. F_5 folds are best developed in the eastern highlands where they are open, upright, trend NNE and form basin and dome interference patterns with F_3 folds (Fig. 12). With the probable exception of F4, correlation of highlands fold phases with those in the northwest lowlands is uncertain at best, and much work remains to be done to decipher the relative timing of structural events in the two regions.

Most Adirondack rocks exhibit strong lineations including mineral grains, fold hinges, rods, streaks, and ribbons. The ribbons, an example of which will be seen at Stop 29, generally consist of quartz and feldspar; they are believed to be elongation lineations formed in response to regional rotational strain (McLelland, 1984). Their evolution may be traced from originally megacrystic quartzofeldspathic gneiss into mylonitic ribbon gneisses with ribbons oriented parallel to the maximum elongation direction of the finite strain ellipsoid. The ribbons contain asymmetric feldspar augen whose tails serve as kinematic indicators, with most showing a SE-over-NW sense of displacement.

Throughout the southern Adirondacks ribbon lineations are parallel to F_2 axes, suggesting that these folds have been rotated into their present orientations by the rotational strain responsible for the lineation. Sheath folds, most easily recognized within the calcsilicate units, have tube axes parallel to F_2 and are consistent with this hypothesis (McLelland 1984, 1986b). In the southern Adirondacks ribbon lineations also parallel F_3 axes.

Berry (1960), working in the area northwest of Whitehall, mapped several extensive zones of mylonite and low angle faults interpreted as thrusts, the largest of which occurs between Lake George and Lake Champlain (Fig. 12). Recent investigations by McLelland (1986b) demonstrate that the charnockitic upper sheet truncates layering in underlying metasediments and that the mylonite zone that marks the contact is folded, being preserved in F_3 synclines and breached by F_3 anticlines. Shear sense indicators suggest SE-over-NW thrusting. Similar mylonitic zones are found in many locations throughout the Adirondack highlands, and are especially common in the southeastern region. Yet to be studied in detail, these zones may mark additional thrusts or tectonic slides (Anderson and others, 1983).

Brittle Structure

Superimposed on the Adirondack topographic grain that is controlled by differential erosion, is a stronger pattern of linear valleys eroded along NNE to NE fracture system (Figs. 4,13,14). A study of LANDSAT images for eastern North America shows the Adirondacks and the rest of the Grenville Province to be the most pervasively-fractured terrane in that region.

Individual fracture systems exceed 100 km in length within the Adirondacks, and some extend southward beneath Paleozoic strata across New York and into Pennsylvania for total lengths exceeding 400 km (Fig. 13). Their extension far outside the dome shows that they are a part of a pre-domal fracture system that was reactivated and accentuated during the Tertiary to Recent Adirondack uplift (Isachsen 1985).

The eastern Adirondacks is a block-faulted terrane of Proterozoic metamorphic rocks containing more than 100 high-angle faults and topographic lineaments that extend both north and south into Paleozoic sedimentary rocks. Elevations are stepped down rather abruptly to the east across a succession of fault blocks going from more than 900 m above sea level to lower than 130 m below sea level in the floor of Lake Champlain.

Although high-angle faults account for most of the NNE linear features in the eastern Adirondacks, similar linear valleys near the center of the dome generally show no displacement of mapped geologic or aeromagnetic units. Most such features, where studied on the ground, were found to be steeply-dipping "zero displacement crackle zones" (ZDCZ) rather than faults (Fig. 14). These are zones, up to a few tens of meters wide, of intensely fractured rock that differ from faults in not showing throughgoing shear planes or visible offsets along fractures, an from joint zones in that the fractures have a more diverse array of directions and are curved. Isachsen and others (1983) interpret these crackle zones, one of which we will examine at Stop 24, as tensional features tha formed originally during late stages of unloading following the Grenville crustal doubling, and that were reactivated much late by crustal stretching at shallower levels ove a rising dome. Crackle zones may be part of a vertical continuum in a tensional regime that would produce narrow fissures at the surface and ZDCZ's at deeper levels.

The time of initiation of the northeasterly trending fracture system is not known. Isachsen (1976a), adopting Dewey and Burke's (1973) Tibetan Plateau model for the Grenvill province, suggested that the fracture system (which parallels the Grenville Province trend originated at the surface of a "Grenville Plateau" in response to post-orogenic spreading of double-thickened crust. The fracture system would then propagate downward into the isostatically rising crust as it passed through the brittle/ductile transition zone.

Bosworth and Putman (1986) have suggested that the NNE-NE trending faults have experienced several periods of movement from Precambrian to Recent. The time(s) of resurgent movement cannot yet be ascertained, but earliest Cambrian (Iapetan opening), Middle-late Ordovician (Taconian), and Mesozoic are possibilities. Some evidence exists for Tertiary and Recent activity (Isachsen, 1985), as summarized under Neotectonics (below) and the discussion for Stop 38).

GEOCHRONOLOGY

Numerous attempts throughout the past 20 years to apply radiometric dating methods to Adirondack rocks have yielded a mass of data, some of them conflicting, probably because of extensive resetting of isotopic ages by granulite facies metamorphism. U/Pb zircon methods appear to be least susceptible to thi type of error, and in the following summary w rely heavily upon zircon chronology, and in particular on the work of Silver (1969) and Chiarenzelli and others (1987). Where older work is cited in the following discussion, th ages have been recalculated using the decay constants recommended by Steiger and Jager (1977).

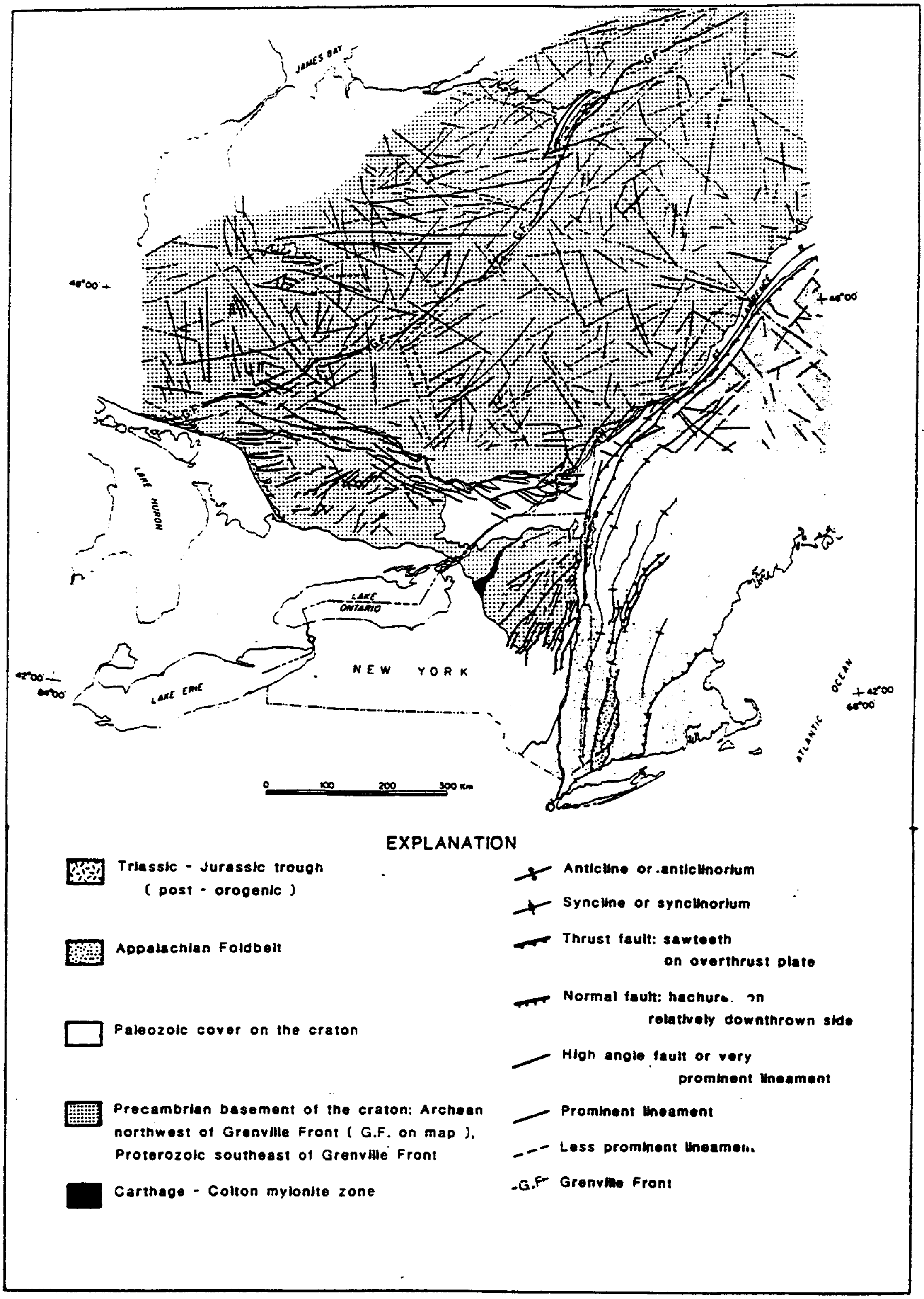


FIGURE 13 Brittle structure map of the southwestern Grenville Province. Based on da- from Geological Survey of Canada and U.S. Geological Survey.

The oldest age yet obtained from an Adirondack rock is 1415±6 Ma for a leucogranitic gneiss which intrudes quartzites and calcsilicates near Alexandria Bay (Stop A) in the northwest lowlands (Chiarenzelli an- others 1987). This gneiss resembles the P- School gneiss which forms the lowermost lithostructural unit in the lowlands. Howe-

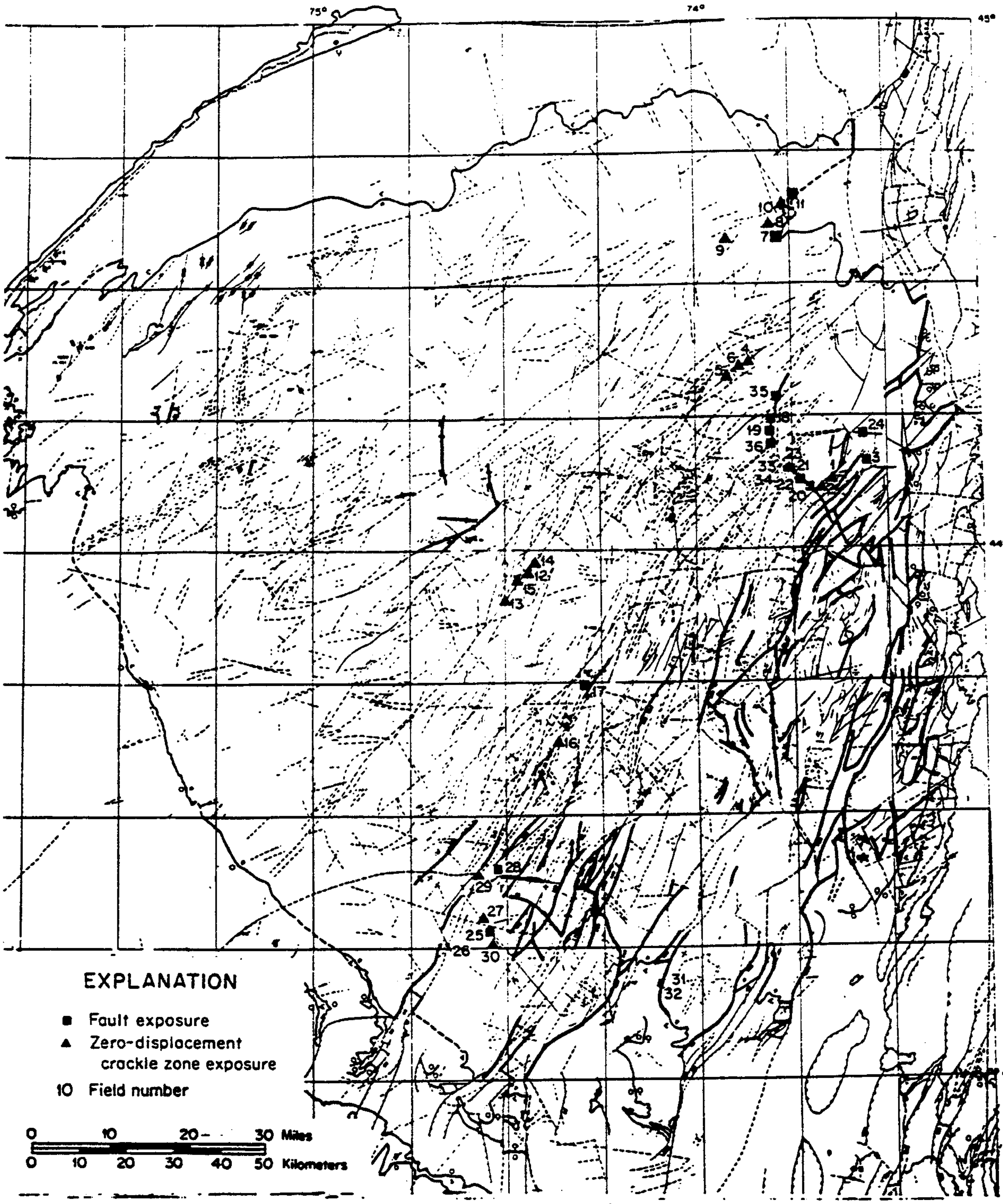


FIGURE 14 Brittle structure map of the Adirondacks (after Isachsen and McKendree, 1977). Heavy lines are high-angle faults for which displacement, movement sense, or breccia localities are known. Dashed lines are linear valleys (lineaments) of unknown origin. Numbered sites have been studied in detail; squares are faults, triangles are zero-displacement crackle zones. Locality 1 is Split Rock Falls (Stop 25).

the latter yields a maximum age of not over 1284 ± 7 Ma (Chiarenzelli and others, 1987), and appears to be conformable rather than intrusive. Thus either the lowlands record at least two cycles of sedimentation and igneous activity, or the rocks at Alexandria Bay represent an exotic terrane unrelated to the rest of the lowlands. Also in the lowlands, two units of Popple Hill gneiss yield Rb/Sr whole rock ages of 1265±25 and 1297±41 Ma

(Grant and others 1984). A mylonitic zone at the contact of the Popple Hill and Lower Marble has a (metamorphic?) age of 1154+19 Ma (Rb/Sr WR, Hudson and others 1986). The Diana igneous complex, which is situated at the highlands/lowlands boundary (Fig. 6) and has been deformed by the Carthage-Colton mylonite zone, has a U/Pb zircon age of 1153±4 Ma (Grant and others, 1986). To date, no well-constrained ages of less than about 1150 Ma have been reported for the northwest lowlands.

In the highlands, the oldest rocks yet documented are quartz dioritic to tonalitic gneisses in the southeastern region, which have a U/Pb zircon age in the vicinity of 1300 Ma (Chiarenzelli and others 1987). These rocks intrude the metasedimentary sequence and contain rotated, foliated xenoliths, suggesting a prior metamorphic event in this part of the highlands. Major igneous activity took place in the highlands over the interval 1160-1130 Ma. Chiarenzelli and others (1987) obtained several zircon ages in this range for rocks of the mangerite-quartz syenite suite, and Hills and Isachsen (1975) report a Rb/Sr whole rock age of 1167+10 Ma for the mangeritic envelope of the Snowy Mountain Dome. These ages are identical within error to that of the petrologically similar Diana Complex, and are only slightly younger than those cited by Carmichael and others (1987) for the Rockport leucogranitic gneiss (1173±4 Ma) and the Gananoque syenite (1162±3 Ma), both of which lie along the Frontenac arch just north of the St. Lawrence River. Charnockitic, granitic and leucogranitic gneisses of the highlands display a wider range of ages, from 1150 to 1050 Ma (Chiarenzelli and others, 1987; Silver, 1969), with a general tendency for the more leucocratic (and Zr-poor) rocks to record younger ages. It is not clear from the isotopic data alone whether these data represent a range of emplacement ages, or varying degrees of resetting and/or metamorphic zircon growth during metamorphism and anatexis.

Silver (1969) obtained three U/Pb zircon ages in the range of 1074-1054 Ma on noritic and pegmatitic facies of the Marcy anorthosite massif, and a zircon concentrate from magnetite-ilmenite ore at Sanford Lake (Fig. 4) gave an age of 1005±10 Ma. Chiarenzelli and others (1987) report zircon ages of 1054±20 and 1052+21 Ma from two samples of metanorthosite near Saranac Lake (Fig. 4), and 1005+5 Ma from an oxide rich gabbroic facies of the anorthosite in the same area. Zircons from the anorthositic rocks typically are equidimensional and multi-faceted with a pale hyacinth color, low uranium content, absence of zoning, and few solid inclusions. This contrasts with zircons from the mangeritic and charnockitic rocks, which are large, elongate, doubly terminated, finely zoned and contain numerous inclusions. Based upon these morphological differences, as well as relationships to the rock fabric, Silver (1969) concluded that zircons of the charnockitic rocks are typically igneous and primary, while those in the anorthositic rocks are of metamorphic origin. The presence of exclusively metamorphic zircons in the anorthositic rocks may relate to the fact that the parent magma was relatively mafic (Emslie 1978, 1985; Morse 1982) and hence undersaturated with zircon (Watson and Harrison, 1983). Under such conditions, zirconium can enter the lattices of pyroxenes (Ewart, 1981) and oxides (Lattard 1987) in significant amounts. Subsequent to cooling this Zr remained as impurities in the mafic silicates and oxides until granulite facies temperatures and metamorphic reactions resulted in exsolution and formation of metamorphic zircon.

Thus the ages of the highlands igneous suite (anorthosite-mangerite-charnockite-granite) may represent either a series of intrusive events spread over a time interval of over 100 Ma (approximately 1150-1050), or a single event at around 1150 Ma, followed by varying degrees of resetting by metamorphism and anatexis during the Grenville (Ottawan) orogeny (ca 1100-1050 Ma), as well as growth of new, metamorphic zircons in the anorthositic rocks. The latter interpretation, while by no means conclusive, is preferred on the basis of the zircon morphology as well as field and geochemical evidence (McLelland 1986; Whitney 1986; McLelland and Whitney 1987) which indicates that these igneous rocks constitute a single bimodal suite. The widely quoted age for the anorthosite of 1288±36 Ma obtained from whole rock and mineral Sm/Nd data by Ashwal and Wooden (1983), is stated by the authors to be oldest age allowed by the data. The well constrained ages on the mangeritic rocks, combined with the field evidence for their contemporaneity with the anorthosite, suggest that the actual age of the latter is closer to 1150 Ma.

Recent work by Mezger and others (1988) and Rawnsley and others (1987) has demonstrated the efficacy of U/Pb systematics in garnets and sphenes in determining the timing of Adirondack metamorphism. Although studies are in their very early stages, some results are available. Sphenes taken from marbles in the northwest lowlands yield ages of 1130-1150 Ma. Those from marbles in the highlands yield ages of 1024-1035 Ma. Garnets from paragneisses in the lowlands give ages of 1152 Ma (near Harrisville) and 1138 Ma (near Edwards)

(Mezger and others, in preparation). The results summarized above suggest that metamorphism in the Adirondack lowlands occurred at around 1150-1130 Ma. It is not known if the younger ages of sphene in the highlands indicates a later metamorphic event, or simply protracted residence of highlands rocks in the lower crust at temperatures above that for closure of U and Pb diffusion in sphene.

STABLE ISOTOPES

Background

Analysis of the stable isotope ratios of oxygen, carbon, hydrogen and sulfur is a powerful tool for elucidation of fluid histories, especially when applied in conjunction with phase equilibria studies (e.g. Valley, 1986b). Stable isotope fractionations among coexisting phases are frequently employed to estimate temperature of equilibration. A number of Adirondack marbles have been analyzed in this manner with assemblages of three or four minerals giving several independent estimates of temperature (Valley and O'Neil, 1984). The concordance among these estimates indicates that all metamorphic minerals except K-feldspar have closely preserved their peak metamorphic values of $\delta^{18}O$ and $\delta^{13}C$. However, Valley and O'Neil (1984) conclude that the accuracy of careful petrologic thermometry (e.g. Bohlen and others, 1985) is greater than that of the stable isotope systems for the Adirondacks because (a) isotopic thermometers have reduced sensitivity at high temperatures, (b) most experimental calibrations of fractionation vs. temperature fail to incorporate the effects of solid solution, and (c) the small but inevitable effects of post-metamorphic isotopic exchange in a slowly cooled terrane are ignored.

The greatest value of stable isotope studies in the Adirondacks has been in the construction of fluid budgets. In well constrained areas, mass-balance considerations allow estimation of fluid/rock ratios and identification of fluid sources. An interesting aspect of these isotopic studies is that, unlike studies of mineral equilibria that preserve information relative to a small portion of the P-T-time path, stable isotopic compositions may preserve the earlier history of the rock. In the Adirondacks, this has made it possible to see through the granulite facies regional metamorphism and to detect an earlier contact metamorphic event (Valley and O'Neil, 1982, 1984; Valley 1985).

Three processes determine the whole rock oxygen or carbon isotope ratio of a metamorphic rock: (a) sedimentary and diagenetic factors which influence the pre-metamorphic composition; (b) metamorphic volatilization; and (c) the infiltration of externally derived fluids (Valley 1986). In certain well defined Adirondack localities, it is possible to estimate the premetamorphic composition and amounts of volatilization. As discussed below, the role of infiltrating fluids (including magmas) can then be evaluated.

Pre-metamorphic isotopic compositions

For many Adirondack samples the most important and most difficult factor to estimate is the isotopic composition of the pre-metamorphic rock. For a metasedimentary rock, this requires consideration of variability at the time of sedimentation, diagenesis, and early metamorphism. The uncertainty can best be evaluated for large data bases. For instance, a compilation of values for unmetamorphosed, Grenville-age limestones (Veizer and Hoefs, 1976) shows almost exactly the same average value and 10-15 permil range in $\delta^{18}O$ and $\delta^{13}C$ as in the high grade Adirondack marbles. A small shift towards lower values in $\delta^{13}C$ of Adirondack calcites results from high temperature exchange with low $\delta^{13}C$ graphite (Valley and O'Neil, 1981). Thus, this range of values may result from differences in pre-metamorphic composition without necessity of additional modifying processes.

The percentage of carbonate minerals in Adirondack marbles and calcsilicates varies from 1 to 100 percent. The mean value is 70% in 75 calcite marbles and 95% in 9 dolomite marbles analyzed by Valley and O'Neil (1984). In calcite marbles, a positive statistical correlation exists between $\delta^{18}O$ and the percentage of carbonate, suggesting that a major control of variability in pre-metamorphic $\delta^{18}O$ values was mixing of high $\delta^{18}O$ marine carbonates and low $\delta^{18}O$ clastic material in the sedimentary environment (Valley and O'Neil, 1984).

The dolomite marbles are particularly interesting because they are nearly pure carbonate. The low percentage of silicates indicates little sediment mixing and a minimum of metamorphic volatilization. The massive nature of many Adirondack dolomite marbles and the small amounts of volatilization (which limits reaction enhancement of rock permeability, Fyfe and others 1978; Rumble and Spear 1983) tends to reduce the importance of infiltration in dolomitic relative to calcitic marbles. Thus, isotopic variability may be expected to be less among dolomites. This is seen by comparison of nine dolomites ($\delta^{18}O$ = 21.0 to 23.8 and $\delta^{13}C$ = -0.3 to 2.9) and 75 calcite marbles ($\delta^{18}O$ = 12.3 to 27.2 and $\delta^{13}C$ = -7.2 to 5.0; Valley, 1986).

Further analysis will determine if it is justified to use deviations from the average dolomite ($\delta^{18}O$ = 22.1) as a means of calculating the volumes of infiltrating fluid. Such an approach is frequently applied successfully in contact aureoles where unmetamorphosed isotopic composition may easily be determined, but its reliability may be questioned in high grade, highly deformed terranes (Valley, 1986). Nevertheless, an intriguing possibility is suggested by comparison of data from the Balmat zinc mine (Stop 13A; $\delta^{18}O$ dolomite = 25.1 to 25.8, Wheland and others, 1984) to that collected over a larger area. The sedimentary Balmat Basin, with well documented evaporitic affinities, appears to have systematically higher $\delta^{18}O$ values because evaporation removes ^{16}O preferentially. This result, if supported by further analysis, will have significance for mineral exploration and for Adirondack stratigraphy.

Other rock types in the Adirondacks, particularly meta-igneous rocks, show less variability in $\delta^{18}O$ than marbles do. Even here, the correlation with lower grade or unmetamorphosed equivalents is tenuous at best. Uncertainty (and sometimes controversy) exists as to premetamorphic isotopic composition of most Adirondack orthogneisses including metanorthosite (Taylor, 1969; Valley and O'Neil, 1979; Morrison and Valley, 1988a), metagabbros, the charnockite-mangerite suite and granitic gneisses (Shieh, 1985).

In marbles and calcsilicates with extreme values of $\delta^{18}O$ the ambiguity caused by variable premetamorphic composition can be overcome. Three categories of extreme oxygen isotopic composition: high $\delta^{18}O$ calcites ($\geq$25); low $\delta^{18}O$ wollastonites (<3.1); and sharp gradients in $\delta^{18}O$ over short distances, show that marbles are particularly well suited for studies of fluid behavior (Valley and O'Neil, 1984). These three situations all argue against the pervasive infiltration of large amounts of externally derived fluid. The rationale, simply stated, is that infiltration of high temperature fluid tends to homogenize isotopic values through fluid-rock exchange. The preservation of extreme values or gradients, which must be pre-metamorphic in origin, permits calculation of limits on the migration of metamorphic fluids.

Volatilization

Metamorphic volatilization at high temperature always tends to reduce the $\delta^{18}O$ of the rock because both CO_2 and H_2O are enriched in ^{18}O. Removal of CO_2 also reduces $\delta^{13}C$ of a rock unless fO_2 is very low (i.e. CH_4 is high), which may reverse the effect. The magnitude of ^{18}O depletion as a result of volatilization can be estimated by considering two end-member modes of fluid escape: (a) slow continuous removal and (b) pressure increase and one step removal. The natural process must lie between these end members. Continuous removal is modeled by the Rayleigh distillation law which shows that if more than 90 percent of the oxygen in a rock is outgassed, the residual rock may be severely depleted in ^{18}O (Valley, 1986). However, such large depletions are unknown because all common volatilization reactions leave better than half of the original oxygen remaining in the silicate minerals. Valley and O'Neil (1984) have modeled these volatilization processes and have shown that the average Adirondack marble has only been depleted 1 permil in $\delta^{18}O$ by volatilization, although the effect on carbon and hydrogen isotopes could be much greater. In other rock types the effect on $\delta^{18}O$ should be even less.

At Cascade Slide (Stop 22) and in the Willsboro wollastonite skarn belt (Stop 26), even if premetamorphic composition and metamorphic volatilization are conservatively estimated, it is possible, by application of these arguments, to demonstrate the absence of large amounts of pervasive fluid flow during regional metamorphism.

ECONOMIC GEOLOGY

Figure 15 is a map of economic mineral deposits in Precambrian rocks of the northeastern United States, including those discussed below, with the exception of graphite and building stone.

Northwest Lowlands

The most important mineral industry in the northwest Adirondacks is zinc mining. The Balmat-Edwards district is one of the great zinc producing districts of North America, with total production having exceeded 34 million tons of ore grading nearly 9.5% Zn and 0.5% Pb. It extends from the village of Balmat near Sylvia Lake, where several mines are located, about 16 km northeast to Edwards, N.Y. The Pierrepont Mine, some 48 km NE of Balmat, is a high grade massive sphalerite ore body in a similar geologic setting, discovered in 1979.

The local zinc industry got its real start in World War I when high prices encouraged development at the Edwards Mine. In 1926, St.

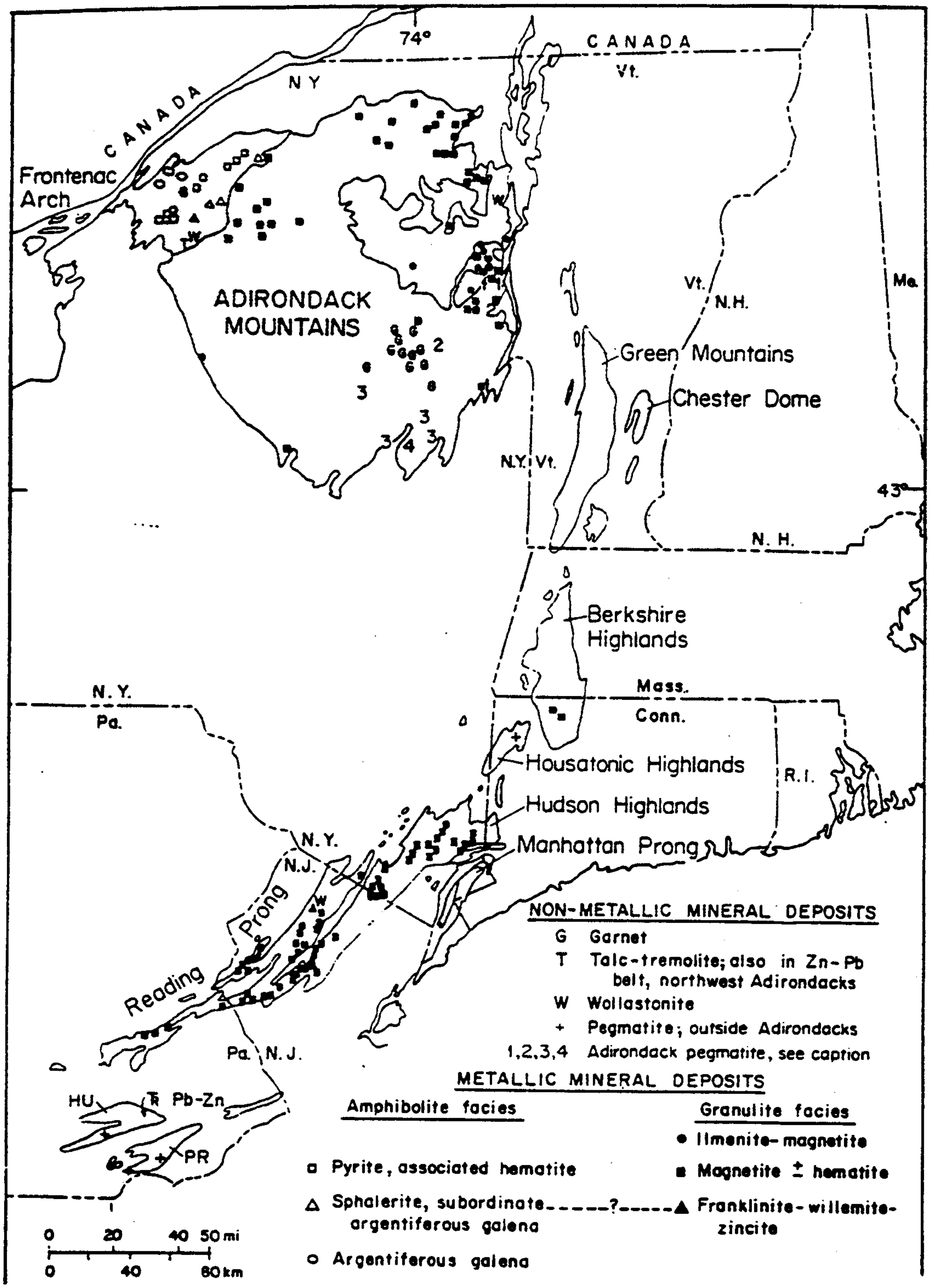


FIGURE 15 Economic mineral deposits in Precambrian rocks of the northeastern United States.

Joe Lead Company acquired the Edwards Mine from the Northern Ore Company, along with an important option on mineral rights at Balmat. The mines are now operated by Zinc Corporation of America. The Balmat #4 mine will be Stop 13a on this trip; geological details are given in the stop description.

Mining for other metals in the northwest Adirondacks has had a long and checkered history. A concise summary and references can be found in Brown (1983). Iron, first discovered in 1812, was mined as earthy and specular hematite near Antwerp and smelted into pig iron at Rossie. Mining activity continued to 1910 when competition from the Lake Superior region forced closure of the mines. The hematite ore represents the oxidized upper part of Proterozoic massive pyrite bodies that occur in gneisses within the Lower Marble Formation. Oxidation by weathering or later hydrothermal activity occurred at the unconformity between the Precambrian rocks and the Potsdam sandstone (Stop B). The unoxidized parts of other, similar, stratabound massive pyrite bodies were exploited as sources of sulfur from 1883 to 1921.

Galena was mined for lead on a small scale and intermittently from its discovery in 1829 until 1875, when the mines, near Rossie and Macomb, were finally closed. Lead and trace zinc mineralization occurs with calcite and minor fluorite in crosscutting veins which are known to extend upward into Lower Paleozoic rocks.

A variety of industrial minerals have been mined in the northwest Adirondacks. Marble quarrying for dimension stone in the Lower Marble Formation near Gouverneur from 1878 until 1941. Talc-tremolite-anthophyllite schist, known commercially as "talc", was mined as early as 1880 at Fowler and Talcville, near Balmat; talc mining continues to be an important industry today. The Wight Mine (Stop 13B), an abandoned talc operation, is situated close to an open pit talc mine operated by the Gouverneur Talc Co.. Early uses of the fibrous products included pulp and paper making while currently the main uses are in ceramics and paint as mineral fillers and pigment extenders. The talc-tremolite-anthophyllite schist is associated with evaporites in the Upper Marble Formation; the protolith of this unusually Mg-rich rock was probably a magnesite-bearing, silicious evaporite. Wollastonite is being produced from a contact metasomatic deposit a the Valentine Mine, near Harrisville (Stop 9). Although pegmatites are relatively common in the northwest lowlands, only one, the McLear pegmatite body near Richville, has ever been mined. From 1910 to 1938, perthitic feldspar was extracted for use in the glass and pottery industry.

Highlands

The history of mining in the Adirondack highlands extends over a period of about one hundred and fifty years. Important deposits include the iron ores of the Port Henry, Lyon Mountain, and Star Lake districts, the titaniferous magnetite deposit at Sanford Lake, and a variety of industrial minerals such as graphite, garnet and wollastonite. The latter two continue as important resources in the region while iron and titanium are no longer produced. Other resources that have attracted interest include dimension stone and feldspar.

Iron. Until recently, iron ore has been the major mineral resource of the region. Remarks of Emmons (1842) indicate knowledge of the iron ores of the Port Henry district as early as about 1800, but it was nearly mid-century before substantial production was realized. Brisk activity continued during the remainder of the nineteenth and into the early years of the twentieth century at numerous small mines scattered throughout the northern and eastern Adirondacks (Fig. 15). Gradually, attention was concentrated on two districts, Mineville-Port Henry and Lyon Mountain (Fig. 4). Operations at these mines by Republic Steel Corporation (later LTV Steel) continued until 1971. Recently, apatite in the tailings from Mineville has been investigated as a potential source of rare earth elements. Other districts geologically similar to Mineville and Lyon Mountain include Clintonville-Ausable Forks and Saranac (Gallagher, 1937; Postel, 1952). These low-Ti magnetite ores are associated with alaskitic and trondhjemitic gneisses, and generally occur in well-defined planar or shoot-like bodies associated with tight synclines. Postel (1952) and Gallagher (1937) considered the gneisses to be of intrusive origin and the ore to be the result of magmatic and hydrothermal processes. Whitney and Olmsted (1988) conclude, on the basis of lithologic association and geochemistry, that these gneisses are metamorphosed ash-flow tuffs that have undergone extensive diagenetic or metasomatic alteration. Many of the features of these ores and their host rocks are strikingly similar to the iron ores of southeastern Missouri (Murphy and Ohle, 1968; Emery, 1968). McLelland (1986a) has proposed that the ores originated as stratabound, exhalative deposits produced by subaqueous precipitation of iron oxides and carbonates during felsic volcanism. A small example of this type of deposit will be seen at Skiff Mountain (Stop 27).

Magnetite ores also form small deposits associated with pyroxene-rich calcsilicate skarns, such as those at the Clinton and Jayville mines in the northwestern highlands (Leonard and Buddington, 1964). A major iron deposit occurs at the Benson Mine near Star Lake in the northwestern highlands. This deposit was worked intermittently between 1889 and 1918, reopened in 1944 by Jones and Laughlin Steel, and operated continuously until 1978. The ore, averaging 26% iron oxide, consists of magnetite and hematite (martite) in a garnet-sillimanite-quartz-feldspar gneiss near the core of a tight, refolded syncline of granitic and metasedimentary gneisses. A long controversy over the origin of the ore is summarized by Palmer (1970), who interprets the deposit as a metamorphosed iron formation.

Titanium. The Sanford Lake ilmenite-magnetite deposits near the southern edge of the Marcy anorthosite massif were mined intermittently for iron for almost one hundred years. It was not until 1941 when the (then) National Lead Company acquired the mine to recover titanium oxide for use as pigment that it became a successful operation (Gross 1968). Active mining has now ceased, although magnetite tailings are being reprocessed for use in drilling mud. The ores, which consist of both ilmenite and titaniferous magnetite, occur as sheets, lenses, and crosscutting veins in metanorthosite and mafic gabbros, as well as disseminated in gabbro. An origin as late differentiates of the anorthosite suite is favored by trace element patterns in the ore (Kelly, 1979) and associated mafic rocks (Ashwal and Siefert, 1980).

Non-metallic minerals. The presence of wollastonite in the Willsboro area (Stop 26) has been known since the early nineteenth century (Buddington, 1977). Without an obvious use, the occurrence was of little interest except as a mineralogical curiosity until the early 1950's when the Cabot Corporation began mining it for use as a filler and ceramic base. Product development resulted in such uses as a tempering agent in ceramics, flux on welding rods, an alloying agent, an extender in plastics and, recently, as a substitute for short fiber asbestos. With the opening of a large open pit mine in Lewis (Stop F), ten miles SW of Willsboro, in 1980, the original Willsboro mine was closed. The wollastonite ore at these mines and in several smaller prospects occurs in a belt of metasedimentary rocks, which outcrops intermittently over a distance of almost 15 kilometers, close to or at the contact with anorthositic gneiss of the Westport Dome (Figs. 12, 21). The ore commonly consists of only three minerals: wollastonite, diopsidic clinopyroxene, and grossular-andradite garnet. The mineralogy and geologic setting favor an origin by contact metamorphism of silicious carbonate rocks at the time of anorthosite intrusion. The fact that excess calcite or quartz are ordinarily absent in this high-variance assemblage suggests that metsomatism played a major role in the ore-forming process. (Buddington, 1939; DeRudder, 1962). Anomalously low values of $\delta^{18}O$ in the ore rock result from convective circulation of heated meteoric water at the time of skarn formation (Valley and O'Neil, 1982).

Several occurrences of abrasive quality garnets are found in olivine metagabbros in the south-central Adirondacks (Fig. 15). The Gore Mountain deposit near North Creek (Stop 28) proved especially useful as an abrasive because of a unique cubic parting. This deposit for many years was the major producer of abrasive garnet in the world; it has now been abandoned in favor of the nearby Ruby Mountain deposit.

Because it is now possible to manufacture graphite more economically than to mine it, natural graphite is no longer produced in the Adirondacks. However, the common occurrence of graphite in Adirondack metasedimentary rocks prompted much exploration throughout the region prior to World War II. The most important deposits were developed in the area of Ticonderoga and Hague but occurrences are found throughout the eastern Adirondacks (Cameron and Weis, 1960). The graphite probably originated largely as fossil organic carbon (Valley and O'Neil, 1981); alternatively some may be the result of reduction of carbonate carbon during metamorphism (Lamb and Valley, 1984). We will see an exposure of one of the principal graphite-bearing horizons (the "Dixon Schist" of Alling, 1927) at Stop 32.

Building Stone. The eastern Adirondack region contains a large number of dimension stone quarries, although few are still in operation. Many nineteenth century buildings in the Adirondack region are built from locally quarried stone. Among the most notable present operations are two quarries operated by the Lake Placid Granite Company near Ausable Forks. At these quarries, production of anorthosite of two distinct colors ("green" and "blue") has continued for over twenty five years. This anorthosite finds its principal use in flooring, and many fine "outcrops" of this rock can be seen in steps and floors at the Empire State Plaza government complex in Albany. Other quarries have been operated in granite, and in Cambrian sandstone and Ordovician limestone around the edges of the Adirondack dome. The piers of the Brooklyn Bridge in New York City are constructed of Chazy limestone quarried at Ligonier Point near Willsboro (Buddington and Whitcomb, 1941).

NEOTECTONICS

The Adirondack dome is an anomalous bump on the North American craton. Several lines of evidence indicate that it is a young uplift that came into being during late Tertiary time, and is currently rising (Isachsen and others, 1983). These are: 1) Drainage basin studies elsewhere indicate that only about 18 million years are required to reduce the elevation of a mountain mass by nine tenths, even allowing for isostatic compensation. This suggests that the high-standing Adirondack dome is a late Cenozoic, probably Pliocene, uplift. 2) The drainage pattern is still largely consequent (radial), the streams being essentially unadjusted to the great variation in erosion resistance of bedrock units (Fig. 16a,b). 3) Isopach maps of Paleozoic sedimentary rocks in New York and Ontario show no evidence of domical uplift of the Adirondack region in Paleozoic time (Rickard 1973, 1975).

The evidence for contemporary uplift of the Adirondacks comes from first order releveling data (Isachsen 1975, 1976; Barnett and Isachsen 1980) which suggests current uplift rates of 3.7 mm/yr near the center of the dome and 2.2 mm/yr along the eastern margin. Uplift at these rates can only be episodic. In addition, recent controversies concerning systematic errors in geodetic measurements leave the actual uplift rates in question (Isachsen 1985). Further evidence for contemporary tectonic activity comes from recurrent, low-intensity earthquakes in many parts of the Adirondacks.

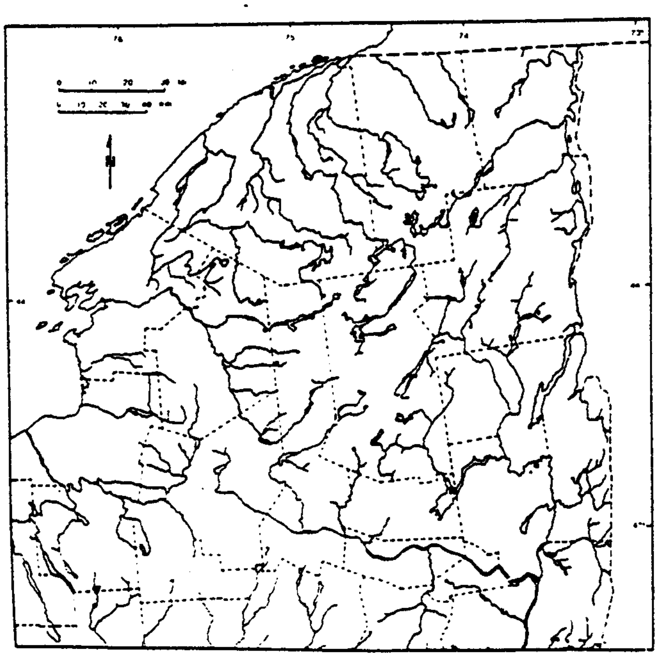
The causes of Adirondack doming are necessarily speculative, but the dome is similar to others throughout the world that are demonstrably the result of crustal expansion over thermal plumes or "hotspots". The absence of recent volcanics and lack of anomalous heat flow in the Adirondacks indicates that any thermal front which may exist must lie at least several km below the surface. However, the high electrical conductivity at 20-25 km (Connerney and others, 1980) and the subhorizontal reflectors at that depth interval discovered by the COCORP traverse (Brown and others 1983) may be the result of deep crustal intergranular melts caused by rising isotherms over a thermal high.

GEOLOGIC HISTORY

The following speculative outline is provided, not as a proposed solution to the problems of Adirondack geologic history, but as a partial definition of some of those problems, and as basis for discussion and possible future research.

Pre- 1400 Ma

Deposition of sandstones and carbonates; intrusion @ 1415 Ma by granite. Known only from the far NW Adirondacks (Stop A), these



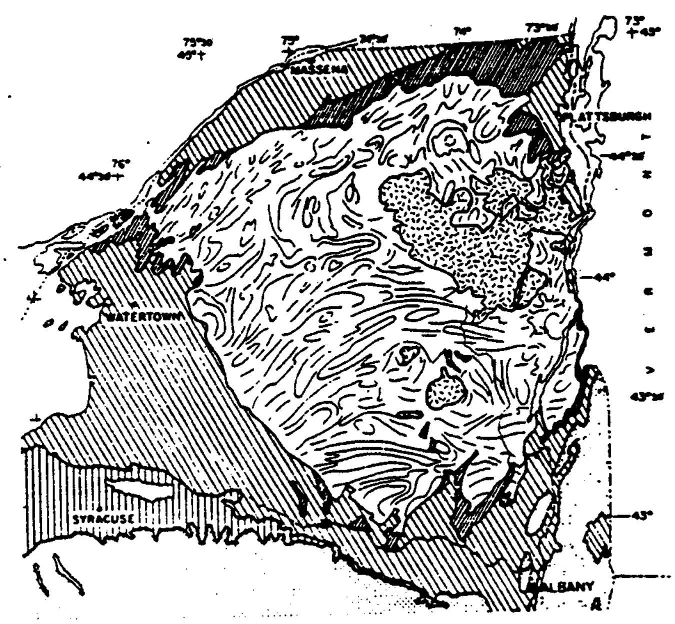


FIGURE 16 A: Radial consequent drainage pattern within the perimeter of the Adirondack dome. B: Bedrock structural trends. The streams cut almost indiscriminantly across the structural trends of metamorphic rocks of greatly varying resistance, indicating the youthfulness of the drainage pattern. Hachured and stippled areas in Fig. 16B are Paleozoic rocks.

rocks may represent an exotic terrane, or may be the basement upon which the principal metasedimentary rock sequence of the NW lowlands was deposited.

1320-1280 Ma

NW lowlands. Eruption of felsic volcanics (Hyde School Gneiss), followed by deposition of carbonate-rich sediments (Lower and Upper Marbles) with intervals of volcanism and/or clastic sedimentation (Popple Hill Gneiss). Many features of the Upper Marble, and the upper part of the Lower Marble, indicate a shallow-water, hypersaline sedimentary environment. The age of the Upper Marble is not known.

SE highlands. Intrusion of quartz diorite and tonalite into a dominantly clastic sedimentary sequence (quartzite, pelite and semipelite, subordinate marbles) of unknown age; possible deformation and metamorphism.

Central highlands. Similarities between metasedimentary rocks of the central highlands and those of both the SE highlands and NW lowlands suggest possible trans-Adirondack correlation (Wiener and others, 1984); in which case some or all of the Highlands metasedimentary suite (Lake George Group) may be of this age. However, some of these metasedimentary rocks are interstratified with apparent metavolcanic rocks that show geochemical similarities to the main highlands igneous suite (ca. 1150 Ma). Thus, there may have been more than one period of sedimentation in the highlands.

Pre- ~ 1150 Ma

Highlands. Field relations at Stop 35 demand, and those at stop 23 suggest, at least one period of deformation and high grade metamorphism prior to the major mid-1100's magmatic event. There is no direct evidence of the age of this event, but it may coincide with the earlier quartz diorite intrusive activity.

1170-1130 Ma

NW Lowlands. Various lines of evidence (Powers and Bohlen 1985; Hudson and others 1986; Rawnsley and others 1987; Mezger and others 1988) point to this time interval for the last (and only ?) major metamorphic episode, and concurrent deformation. Rocks of the Diana Complex, which straddles the Carthage-Colton Zone, were intruded at this time. Extreme ductile deformation of rocks of the Diana Complex in the CCMZ (Stops 7, 8) provides an upper limit to the age of the latest movement along this zone.

Highlands. Considerable evidence, summarized above under geochronology, indicates that the anorthosite-jotunite-ferrogabbro and mangerite-charnockite-granite igneous suites of the highlands were intruded during this time span. The anorogenic character of these suites (McLelland 1986a; Whitney 1986; McLelland and Whitney 1987) may indicate dissimilar tectonic environments in the lowlands and highlands at this time. Interstratification of apparently metavolcanic rocks of this suite and carbonate-rich, formerly evaporite-bearing metasedimentary rocks suggests some contemporaneous sedimentation. Shallow depth of intrusion of the plutonic rocks is supported by isotopic and thermodynamic arguments (Valley and O'Neil, 1982; Valley, 1985); but this has been disputed, based on the occurrence of high-pressure pyroxenes in mangeritic rocks of the High Peaks area, by Olilla and others (1984). Perhaps various depths of emplacement are represented, with deep and shallow intrusive rocks juxtaposed by later vertical tectonic movements in a doubly-thickened crust (Whitney, 1983).

1100-1000 Ma

Highlands. Most U/Pb zircon ages in low-Zr metaigneous rocks, including the anorthosite suite, fall in this range (Chiarenzelli and others, 1987), as does a metamorphic Sm/Nd isochron (1095±7 Ma) obtained from a garnetiferous anorthosite (Basu and Pettingill 1983). Recent U/Pb ages for sphenes in highlands marbles are in the 1024-1035 Ma range (Rawnsley and others 1987). These ages are consistent with isotopic resetting or growth of new metamorphic zircon and sphene under granulite facies conditions during the main Grenville orogenic event. Temperatures of 750-800^{0}C and pressures of 7-8 kbar (Bohlen and others 1985) for this event indicate a crust of double thickness, formed by stacking of slabs or slices of crust along major ductile thrusts, possibly caused by tectonic collision south and east of the presently exposed Adirondacks. Evidence of such high temperature ductile shear is widespread in the central and southeastern highlands (see, e.g., Stop 29). Kinematic indicators in these ductile shear zones are locally ambiguous, but an overall suggest a SE-over-NW sense of movement. The nearly 100 Ma span of metamorphic ages in the highlands suggests a long, complex event; the counterclockwise P-T-time path (Fig. 11) may reflect early compressional thickening and heating followed by slow uplift during cooling. Migmatization of the granulite-facies rocks is commonplace, but no unequivocally synmetamorphic intrusives have as yet been identified, leaving the heat source for metamorphism unknown. Residual heat from the major 1170-1130 Ma igneous event may have been significant, along with vertical tectonic movements of deep crustal rocks (see stages 2 and 3 of the tectonic model of Whitney, 1983). Alternatively, if the zircon ages measured by Chiarenzelli and others (1987) for the anorthosites and the felsic members of the mangerite-charnockite suite are crystallization ages, then no heat problem exists. Final closure of some isotopic systems may have been delayed until nearly 900 Ma (Ashwal and Wooden, 1983).

Post-1000 Ma

By middle Cambrian time, uplift and erosion had removed as much as 25-30 km of rock from the Adirondacks, exposing the granulite facies rocks at nearly their present level. The prominent NNE-trending fault and fracture system may also have originated by this time, as evidenced by numerous mafic dikes that parallel this trend and intrude Proterozoic rocks, but not the overlying Paleozoic rocks.

TABLE 1. U-PB ZIRCON AGES FOR META-IGNEOUS ROCKS OF THE ADIRONDACK MOUNTAINS

No.	Age (Ma)	Location	Sample No.
HIGHLANDS			
Tonalitic gneiss			
1	1321 ±60	South Bay	AM-87-12
2	1301*	Canada Lake	AM-86-12
Mangeritic and charnockitic gneiss			
3	1155 ±4†	Diana complex	
4	1147 ±10	Stark complex	AM-86-15
5	1134 ±4	Tupper Lake	AC-85-6
Older hornblende granitic gneiss			
6	1156 ±8	Rooster Hill	AM-86-17
7	1150 ±5	Piseco dome	AM-86-9
8	1146 ±5	Oswegatchie	AC-85-2
Younger hornblende granitic gneiss			
9	1100 ±12	Carry Falls	AM-86-3
10	1098 ±4	Tupper Lake	AM-86-6
11	1095 ±5	Stillwater	NoFo-1§
Alaskitic gneiss			
12	1075 ±17	Tupper Lake	AM-86-4
13	1073 ±6	Dannemora	AM-86-10
14	1057 ±10	Ausable Forks	AM-86-14
Anorthosite and metagabbro			
15	1054 ±20	Saranac Lake	AC-85-8
16	1050 ±20	Saranac Lake	AC-85-7#
17	996 ±6	Saranac Lake	AC-85-9
Xenolith-bearing olivine metagabbro			
18	1144 ±7	Dresden Station	AM-87-11
LOWLANDS			
Leucogranitic gneiss			
19	1415 ±6	Wellesley Island	AM-86-16
Alaskitic gneiss			
20	1284 ±7	Gouverneur dome	AC-85-4
Granitic gneiss			
21	1160**	Gananoque	AM-87-1
22	1137 ±1††	North Hammond	AM-87-3
HIGHLAND SAMPLES OF SILVER (1969)§§			
23	1113 ±10	Fayalite granite, Wanakena	
24	1109 ±11	Charnockite, Ticonderoga	
25	1084 ±15	Undeformed syenite dike, Jay	
26	1074 ±10	Anorthosite pegmatite, Elizabethtown	
27	1064 ±10	Metanorite, Snowy Mountain dome	
28	1054 ±20	Sheared anorthosite pegmatite, Jay	
29	1009 ±10	Magnetite-ilmenite ore, Tahawus##	

Note: Errors at two sigma.
*Minimum.
†Data from Grant et al. (1986).
§Contains 1155 Ma zircon xenocrysts.
#Baddeleyite age of >1086 ±6 Ma from this sample.
**Preliminary age, location is on Ontario Highway 2S, 8.2 km east of Gananoque.
††Monazite age.
§§Decay constants of Steiger and Jager (1977).
##Location same as Sanford Lake (SL) in Figure 1.

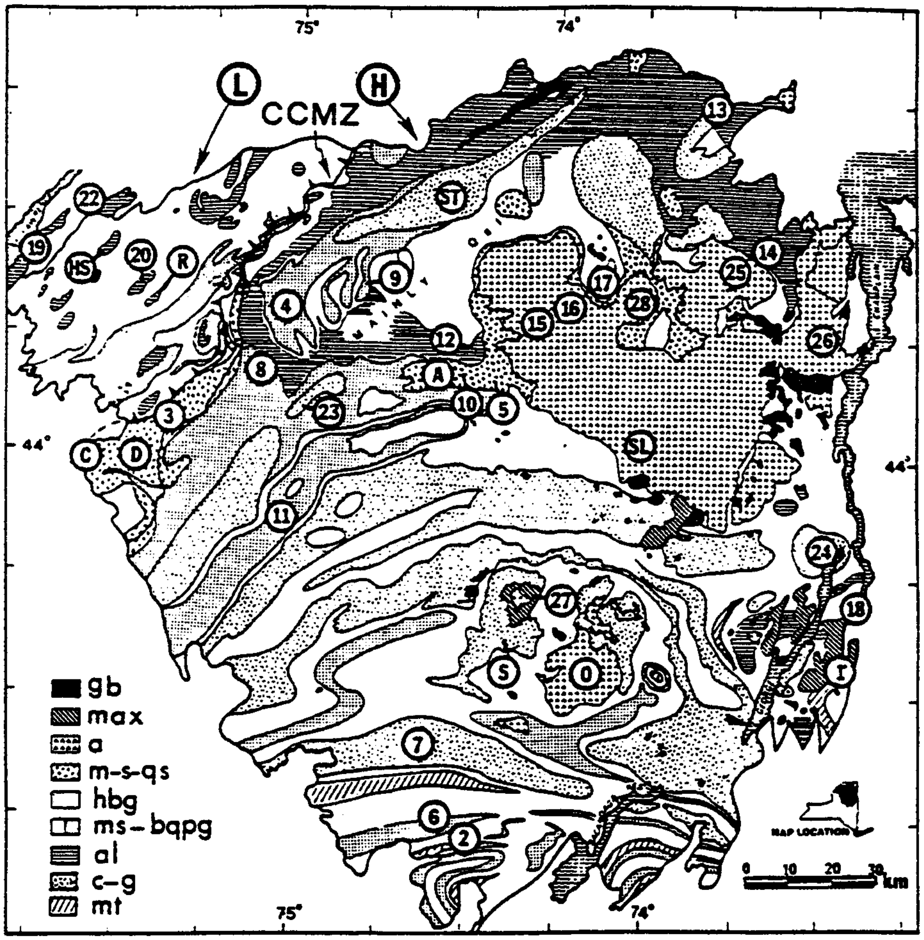


Figure 1. Generalized geologic map of Adirondacks. CCMZ = Carthage-Colton mylonite zone (heavy line with dip indicated by sawteeth) separating lowlands (L) to northwest from highlands (H) to southeast; gb = olivine metagabbro; max = mangeritic gneiss with abundant andesine xenocrysts; a = anorthosite and related gabbros; m-s-qs = mangeritic to quartz mangeritic gneiss with scattered xenocrysts of andesine; hbg = hornblende granitic gneiss; ms = metasedimentary gneisses; bqpg = biotite-quartz-plagioclase gneiss (lowlands only); al = alaskitic and leucogranitic gneiss; c-g = charnockitic to granitic gneiss with occasional andesine xenocrysts; mt = metatonalitic gneiss; A = Arab Mountain anticline; C = Carthage anorthosite; D = Diana complex; HS = Hyde School body; O = Oregon dome; R = Reservoir Hill body; S = Snowy Mountain dome; SL = Sanford Lake (Tahawus); ST = Stark complex; 1-29 = locations for sample numbers given in Table 1 (map from McLelland and Isachsen, 1986).

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McLelland - Structure

Structure and Rock Fabric Within the Central and Southern Adirondacks

James McLelland
Department of Geology, Colgate University, Hamilton, NY

INTRODUCTION

The area referred to as the southern Adirondacks is shown in Figure 1. Within this region, the Precambrian is bounded approximately by the towns of Lowville and Little Falls on the west and Saratoga Springs and Glens Falls on the east (Fig. 2).

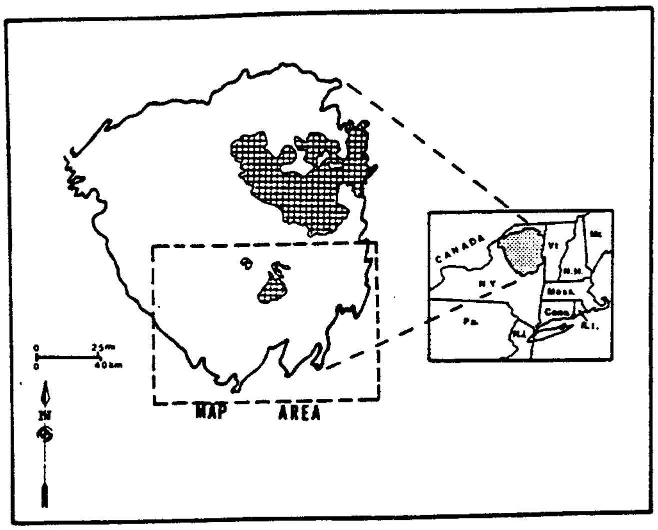

Fig. 1. Location map of the Adirondack Mts. Major anorthosite massifs are represented by the grid pattern. The central and southern Adirondacks lie within the dashed rectangle labeled "Map Area".

Mapping in the southern Adirondacks was done first by Miller (1911, 1916, 1920, 1923), Cushing and Ruedemann (1914), Krieger (1937), and Cannon (1937); more recent investigations were undertaken by Bartholome (1956), Thompson (1959), Nelson (1968), and Lettney (1969). At approximately the same time Walton (1961) began extensive field studies in the eastern portion of the area (Paradox Lake, etc.), de Waard (1962) began his studies in the west (Little Moose Mt. syncline). Subsequently de Waard was joined by Romey (de Waard and Romey, 1969).

Separately and together, Walton and de Waard (1963) demonstrated that the Adirondacks are made up of polydeformational structures, the earliest of which consist of isoclinal, recumbent folds. Their elucidation of Adirondack geology set the tone for future workers in the area. In this regard one of their most important contributions to the regional picture was that the lithologic sequence of the west-central Adirondacks is similar to that of the eastern Adirondacks.

Beginning in 1967 McLelland (1969, 1972) initiated mapping in the southernmost Adirondacks just to the west of Sacandaga Reservoir subsequently this work was extended north and east to connect with that of Walton and de Waard.

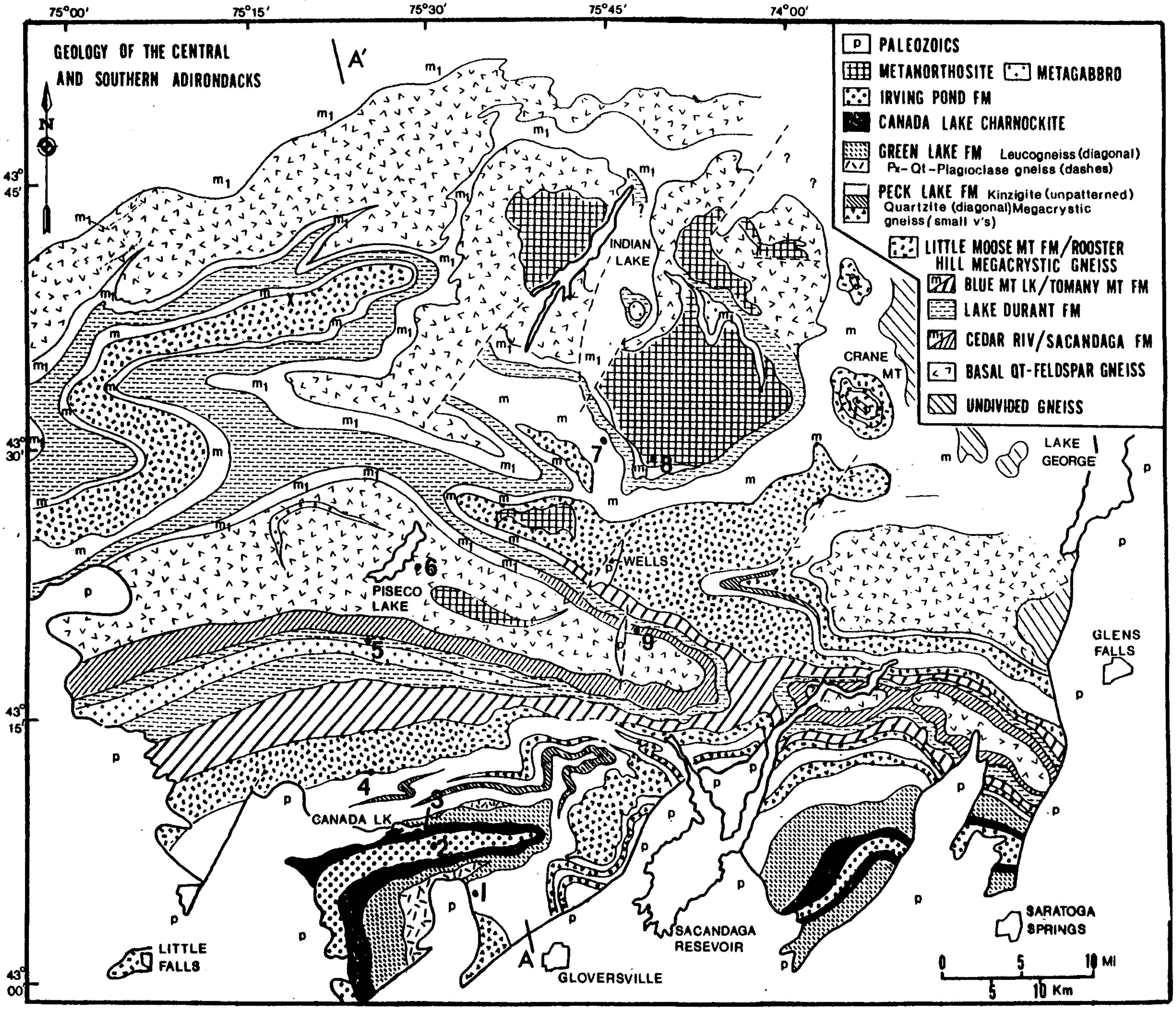

Fig. 2. Formational map and stop localities for the central and southern Adirondacks (from McLelland and Isachsen, 1980)

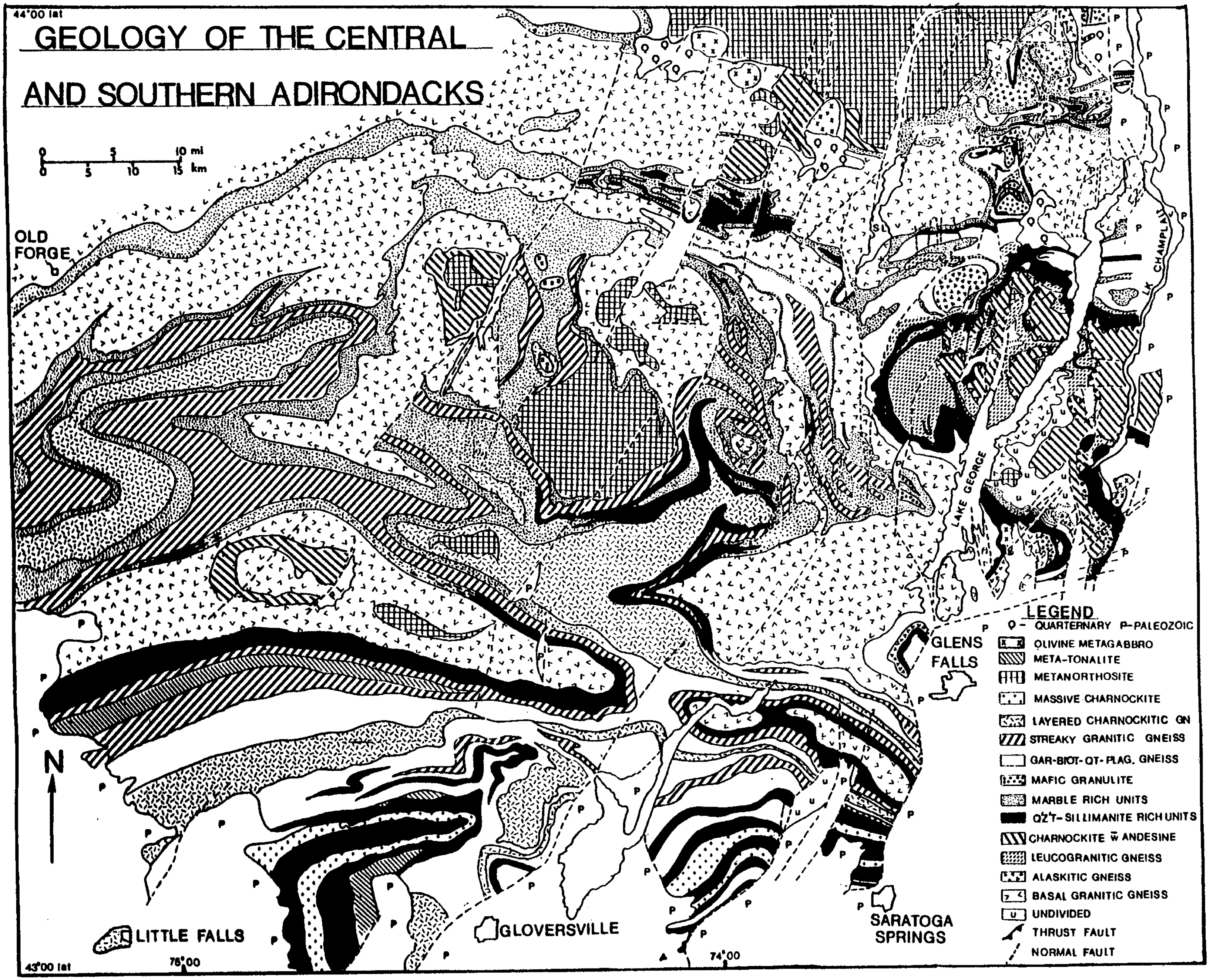

Fig. 3. Lithological map of the central and southern Adirondacks. Note that only a few high angle faults are shown.

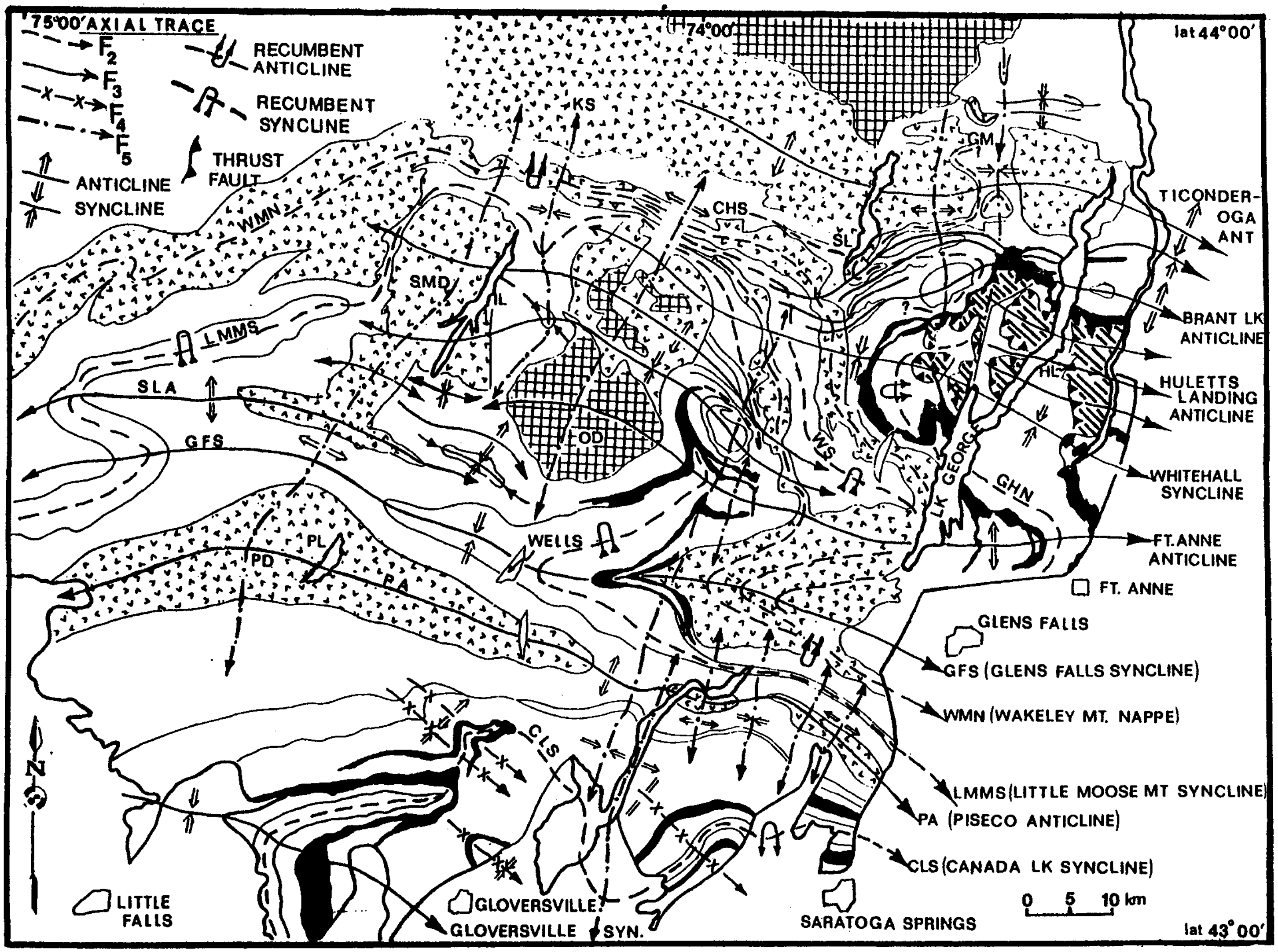

Fig. 4. Axial trace map of the central and southern Adirondacks. For lithological map symbols see Fig. 3. Abbreviations not on map: CHS - Canada Hill syncline; GHN - Green Hill nappe; GM - Glidden Marsh syncline; HL - Hullett's Landing; OD - Oregon Dome; PD - Piseco Dome; PL - Piseco Lake; SL - Schroon Lake; SLA - Spruce Lake anticline; SMD - Snowy Mt. Dome; WS - Warrensburg syncline

Geraghty (1973) and Farrar (1976) undertook detailed mapping in the eastern half of the North Creek 15' quadrangle, and tied into investigations in the Brandt Lake region by Turner (1971). Recently, Geraghty (1978) completed a detailed study of the structure and petrology in the Blue Mt. Lake area. Current investigations by McLelland and by the N.Y. Geological Survey are going forward in the general region surrounding Lake George.

The foregoing investigations have increased our knowledge of the southern Adirondacks, and this fieldtrip is designed to show as many examples of the region's structure, lithology, and petrology as time permits.

STRUCTURAL FRAMEWORK OF THE SOUTHERN ADIRONDACKS

The southern Adirondacks (Figs. 2-5) are underlain by multiply deformed rocks which have been metamorphosed to the granulite facies. The structural framework of the region consists of four unusually large fold sets, F_2 - F_5 together with an early set of isoclines represented solely by intrafolial minor folds with associated axial planar foliation (Figs. 2-4). Relative ages have been assigned to these fold sets, but no information exists concerning actual time intervals involved in any phase of the deformation. It is possible that several, or all of the fold sets, are manifestations of a single deformational continuum.

The earliest and largest of the map-scale folds are recumbent, isoclinal structures (F_2) -- for example the Little Moose Mt. syncline (de Waard, 1962) and Canada Lake nappe (McLelland, 1969) (Figs. 2 and 5). These isoclines have axes that trend approximately E-W and plunge within 20° of the horizontal. As seen in Figures 4 and 5 the axial traces of each of the F_2 folds exceeds 100 km. They are believed to extend across the entire southern Adirondacks. Subsequent useage of the terms "anticline" and "syncline," rather than "antiform" and "synform," is based on correlations with rocks in the Little Moose Mt. syncline where the stratigraphic sequence is thought to be known (de Waard, 1962).

Close examination reveals that the F_2 folds rotate an earlier foliation defined principally by platets of quartz and feldspar and axial planar to minor intrafolial isoclines. Although this foliation is suggestive of pre-F_2 folding, such an event does not seem to be reflected in the regional map patterns (Fig. 3). However, it is possible that major pre-F_1 folds exist but are of dimensions exceeding the area bounded by Figure 3. If this is the situation, their presence may be revealed by continued mapping. The existence of such folds is suggested by the work of Geraghty (1978) in the Blue Mt. area. In the vicinity of Stark Hills charnockites of the Little Moose Mt. Fm. appear to be identical to supposedly older quartzo-feldspathic gneisses (basal) which lie at the base of the lithologic sequence. Given this situation, then the Cedar River and Blue Mt. Lake Fms. are identical, and there emerges a pre-F_2 fold cored by the Lake Durant Formation. However, careful examination of the Lake Durant Formation has failed to reveal the internal symmetry implied by this pre-F_2 fold model. It is possible, of course, that the pre-F_1 foliation may not be related directly to folding (e.g. formed in response to thrusting, gravity sliding, etc.; Mattauer, 1975). Currently the origin of the pre-F_2 foliation remains unresolved. In most outcrops the pre-F_2 foliation cannot be distinguished from that associated with the F_2 folding.

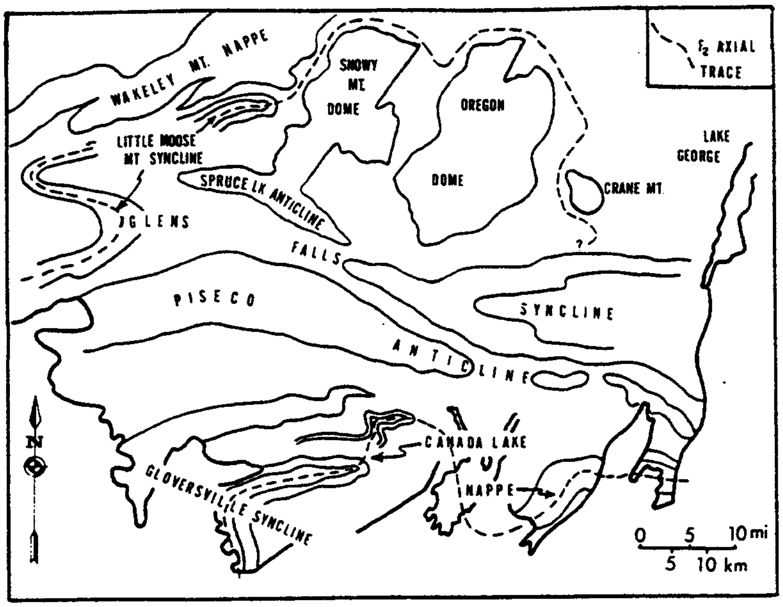

Fig. 5. Blocked out major folds of the central and southern Adirondacks (from McLelland and Isachsen, 1980).

Following the F_2 folding, there developed a relatively open and approximately upright set of F_3 folds (Figs. 2-5). These are coaxial with F_2. In general the F_3 folds are overturned slightly to the north, the exception being the Gloversville syncline with an axial plane that dips 45°N. The F_3 folds have axial traces comparable in length with those of the F_2 set. The Piseco anticline and Glens Falls syncline can be followed along their axial traces for distances exceeding 100 km until they disappear to the east and west beneath Paleozoic cover. The similarity in size and orientation of F_2 and F_3 suggests that both fold sets formed in response to the same force and field.

The fourth fold set (F_4) is open, upright, and trends NW. Within the area these folds are less prevalent than the earlier sets. However, Foose and Carl (1977) have shown that within the NW Adirondacks, northwest-trending folds are widespread and play an important role in the development of basin and dome patterns.

The fourth regional fold set (F_5) consists of large, upright NNE folds having plunges which differ depending upon the orientation of earlier fold surfaces. The F_5 folds are observed to tighten as one proceeds towards the northeast.

The regional outcrop pattern is distinctive because of the interference between members of these four fold sets (Figs. 2-5). For example, the "bent-finger" pattern of the Canada Lake nappe west of Sacandaga Reservoir is due to the superposition of the F_3 Gloversville syncline on the F_2 fold geometry (Fig. 4). East of the reservoir the reemergence of the core rocks of the Canada Lake nappe is due to the superposition on F_2 of a large F_5 anticline whose axis passes along the east arm of the reservoir (Fig. 4). The culmination-depression pattern along the Piseco anticline results from the superposition of F_3 and F_5 folds. The structure of the Piseco dome is due to the intersection of the Piseco anticline (F_3) with the Snowy Mt. anticline (F_5). Farther to the north, Crane Mt. is a classic example of a structural basin formed by the interference of F_3 and F_5 synclines (Figs. 2 and 6).

DISCUSSION AND SYNTHESIS OF STRUCTURAL RELATIONSHIPS

Over a decade ago Walton and de Waard (1963) proposed that rocks of the anorthosite-charnockite suite comprise a pre-Grenvillian basement on which a coherent "supracrustal" sequence was deposited unconformably. Rocks which would be assigned a basement status in this model are designated as basal quartzo-feldspathic gneiss in Figure 3. The basal Cedar River Fm. of the overlying "supracrustal" sequence consists of marbles, quartzites, garnet-sillimanitic gneisses, and various calc-silicates. This lowermost unit is followed upward by various quartzo-feldspathic gneisses, marbles, and other metasedimentary sequences shown in Figure 2. Although our own research agrees with the generalized lithologic sequences of de Waard and Walton, two major provisos are necessary and are given here.

(1) Anorthositic rocks intrude the so-called supracrustal sequence, and therefore the anorthosites post-date these units and cannot be part of an older basement complex (Isachsen, McLelland, and Whitney, 1976; Husch, Kleinspehn, and McLelland, 1976). Isotopic evidence (Valley and O'Neill's (1983); Ashwal and Wooden, 1984) suggests that the anorthosites intruded prior to the 1.1 Ma Grenvillian metamorphism probably during a non-compressional stage (Emslie, 1978, Whitney, 1983). Angular, rotated xenoliths within the anorthosites exhibit pre-intrusion foliation and imply an earlier orogenic event(s).

(2) Within the metastratified units of the region, there exists field evidence for primary facies changes. For example, the well-layered sillimanite-garnet-quartz-feldspar gneisses of the Sacandaga Formation grade laterally into marble-rich units of the Cedar River Fm. exposed north of the Piseco anticline (Figs. 2,3). This transition along strike can be observed just south of the town of Wells, and its recognition is critical to the interpretation of the regional structure. Thus the great thickness of kinzigites (granulite-facies metapelites) south of the Piseco anticline gives way to the north to thinner units marked by marbles, calcsilicates, and quartzites. We interpret this lithologic change as due to a transition from a locally deep basin in which pelitic rocks were accumulating to a shallow-water shelf sequence dominated by carbonates and quartz sands.

Given the foregoing information, it has been possible to map and correlate structures and lithologies on either side of the Piseco anticline. In the northwest the sequence on the northern flank proceeds without structural discontinuity into the core of the Little Moose Mt. syncline. There occurs on the southern flank a mirror image of the northwestern lithologic sequence as units are traced towards the core of the Canada Lake nappe. It follows that the Canada Lake nappe and Little Moose Mt. syncline are parts of the same fold (Fig. 6). The amplitude of this fold exceeds 70 km. and it can be followed for at least 150 km along its axial trace. The major F_2 and F_3 folds of the area are exposed through distances of similar magnitude, but their amplitudes are less than those of the F_1 isoclines. The structural framework that emerges is one dominated by exceptionally large folds.

Accepting that the Little Moose Mt. syncline and Canada Lake nappe are the same fold, and noting that the fold axis is not horizontal, it follows that the axial trace of the fold must close in space. The axial trace of the Canada Lake nappe portion of the structure can be followed from west of Gloversville to Saratoga Springs. Therefore, the axial trace of the Little Moose Mt. syncline also must traverse the Adirondacks to the north. Mapping strongly suggests that the hinge lines of this fold passes through North Creek and south of Crane Mt. (Fig. 6). From here the axial trace swings westward along the north limb of the Glens Falls syncline to a point north of Wells and thence eastward to a point south of Glens Falls. This model is depicted schematically in Figure 6 where the southern Adirondacks are shown as underlain largely by the Canada Lake-Little Moose Mt. syncline. Later folding by F_3 and F_5 events has resulted in regional doming of the F_2 axial surface and erosion has provided a window through the core of this dome. Note the western extension of the Piseco anticline beneath the Paleozoic cover. This extension is consistent with aeromagnetics of the area.

Currently attempts are underway to synthesize the structural framework of the entire Adirondacks by extending the elements of the present model to other areas. A preliminary version is shown in Figure 7 and suggests that most Adirondack structure is explicable in terms of the four regional fold sets described here. Thrust faulting has been recognized in the eastern Adirondacks (Berry, 1961) and high strain zones exist in many other areas of the Adirondacks (McLelland, 1984). Associated with these are distinctive ribbon gneisses (Fig. 8) and sheath folds (Fig. 9). These are further discussed in Stop 6, Road Log.

CONCLUDING SPECULATIONS

The ultimate origin of the structural and petrologic features of the Adirondacks remains obscure. A possible clue to the mechanisms involved is Katz's (1955) determination of 36 km as the present depth to the M-discontinuity beneath the Adirondacks. Because geothermometry-geobarometry place the peak of the Grenville metamorphism at 8-9 kb (24-36 km), a double continental thickness is suggested. Such thicknesses presently exist in two types of sites, both plate-tectonic related. The first is beneath the Andes and seems related to magmatic underplating of the South American plate (James, 1971). The second is beneath the Himalayas and Tibet and is due to thickening in response to collision (Dewey and Burke, 1973) or continental underthrusting (Powell and Conaghan, 1973). The presence of ribbon lineation, sheath folds, and subhorizontal mylonitic foliation within the region strongly suggests regional rotational strain with a dominant component of simple shear (McLelland, 1984). Rotated K-feldspar augen exhibit tails asymmetric to foliation suggesting an east side up and to the west sense of tectonic transport.

Southeastward directed subduction would be consistent with this model. The relevant plate margin presumably lies buried beneath the present day Appalachians.

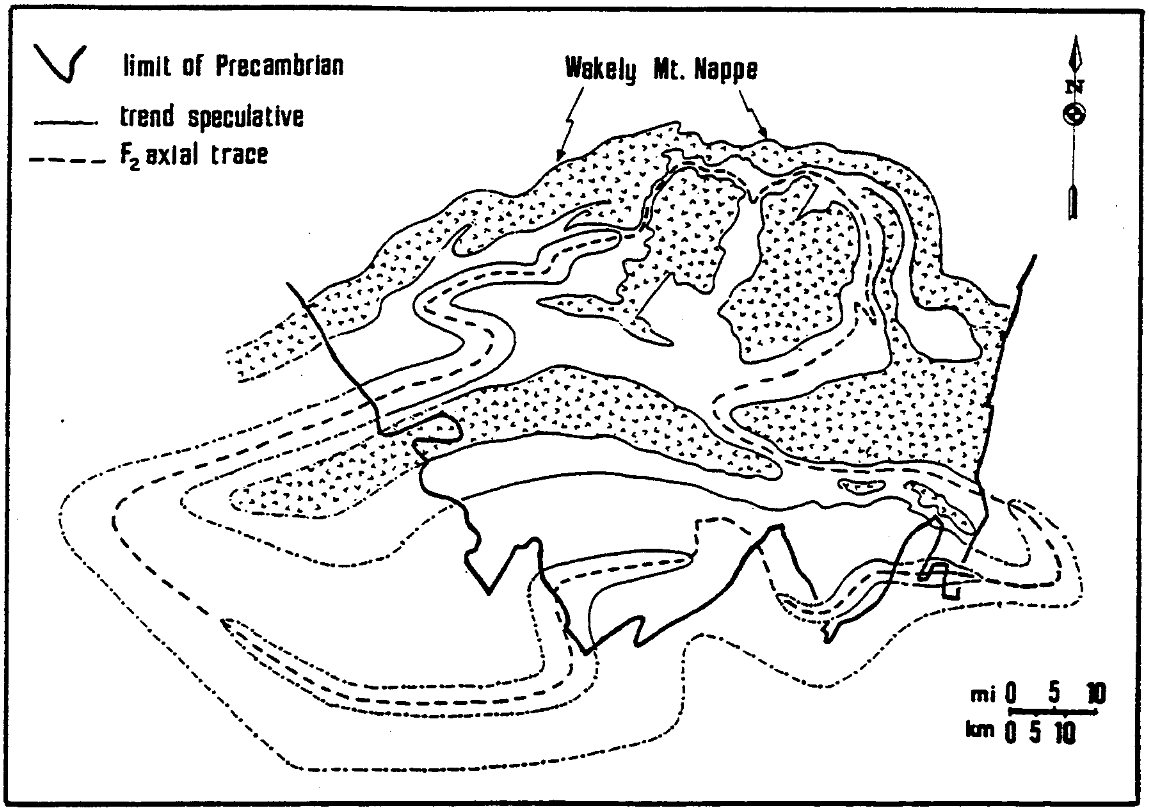

Fig. 6. Geologic sketch map showing the proposed axial trace of the F_2 Little Moose Mt. - Canada Lake syncline. The western extension of the Piseco anticline is inferred from aeromagnetic data (from McLelland Isachsen, 1980).

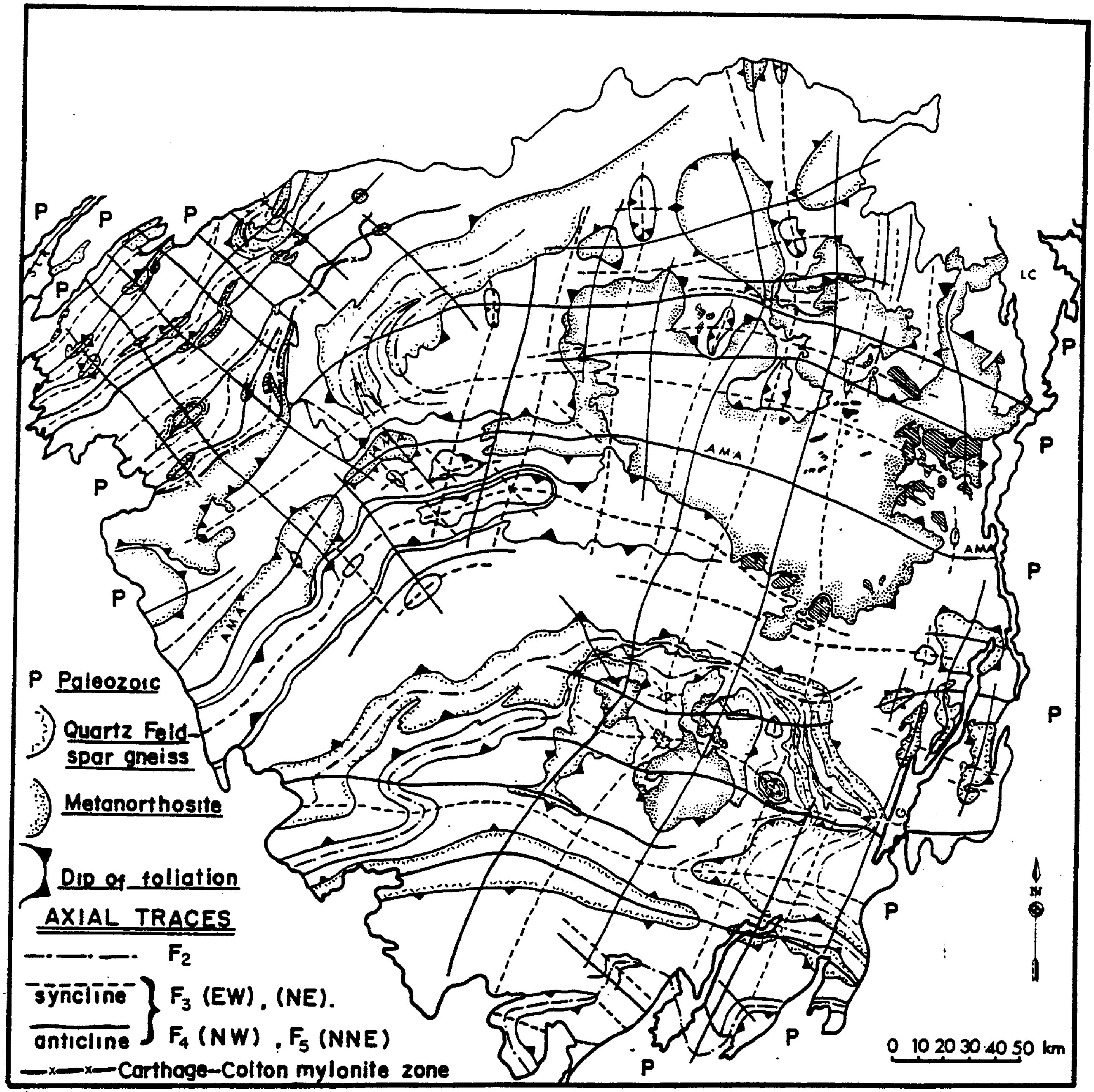

Fig. 7. Axial trace map of the Adirondack Mts. (from McLelland and Isachsen, 1980).

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Geology and Geochronology of the Southern Adirondacks

James McLelland
Department of Geology
Colgate University
Hamilton, NY 13346

PROLOGUE

Geologic investigations of the Adirondack Region began in the nineteenth century with the early surveys of Ebenezer Emmons and later with important contributions from Kemp, Smyth, and Cushing (see Buddington 1939, for complete bibliographic references of these and other early workers). in the first three decades of the twentieth century a large number of significant publications were forthcoming from Miller, Newland, Alling, and Balk, as well as those cited previously, including Buddington. The essence of these contributions is summarized in Buddington's classic Geological Society of America Memoir (1939) entitled "Adirondack Igneous Rocks and their Metamorphism". The central, dominant theme of this work is that the Adirondacks consist of a vast collection of intrusive igneous complexes ranging in composition from anorthosite to granite. This consensus view was strongly supported by a wealth of field and chemical data and may be regarded as comprising Phase I in the history of Adirondack geology. The culmination of Phase I corresponds with the publication of U.S. Geological Survey Professional Papers 376 and 377 by Buddington and Leonard (1962) and Leonard and Buddington (1964), respectively. These works were the result of government sponsored mineral exploration efforts during World War II. Both reports manifest the classical view of the Adirondacks as an igneous-plutonic domain.

Phase II begins with Engel and Engel (1958, 1964) in the lowlands and Walton and deWaard (1963) and deWaard (1964, 1965) in the highlands. Conceptually, this phase is characterized by paradigms of stratigraphy and stratigraphic correlation. Its modus operandi was the recognition and definition of rock sequences, interpreted as stratigraphic, and correlated over great distances. Depending upon the particular investigators, this stratigraphic framework was cross-bred with varying degrees of granitization and metasomatic transformations.

A late stage of Phase II is represented by the investigations of McLelland in the highlands (McLelland and Isachsen, 1986) and Carl, Foose, deLorraine, and others in the lowlands (see Carl et al. 1990 for references). In this stage, metavolcanics played a burgeoning role in the interpretation of Adirondack layered sequences (Whitney and Olmsted 1989). In addition, structural investigations documented the existence of large fold-nappe structures within most of the Adirondacks (deWaard 1964, McLelland 1984). A characteristic of this stage of Phase

II was to downgrade the importance of widespread igneous intrusion and to substitute for it metamorphic processes involving recrystallization of stratified volcanics and sediments into layered gneisses modified by local anatectic effects (Carl et al. 1990; Whitney and Olmsted 1989).

Phase III of Adirondack geology began in the mid-1970's when Eric Essene, together with his students John Valley and Steve Bohlen (see Bohlen et al. 1985 and Valley et al. 1990) established a quantitative framework for Adirondack pressure, temperature, and fluid conditions during metamorphism. This approach has been significantly augmented by the oxygen isotope studies of John Valley and Jean Morrison (see Morrison and Valley 1988 for complete reference) and the U-Pb studies of Klaus Mezger (1990), all of which have provided critical data that constrain models of Adirondack evolution. Simultaneously, McLelland and Chiarenzelli (McLelland et al. 1988, 1991a,b; McLelland and Chiarenzelli 1990, 1991) have conducted a U-Pb zircon study of the Adirondacks in order to follow up on Silver's (1969) pioneering, landmark investigations. The quantitative results of these research programs have provided unequivocal boundary conditions with which any interpretations of Adirondack geology must be consistent. These results and associated boundary conditions are presented in the text that follows. Significantly, and interestingly, the numbers demonstrate that Phase I interpretations were much closer to the truth than the elaborate stratigraphic models of Phase II. It has become increasingly clear that, in the Adirondacks, intrusive igneous rocks greatly dominate over metavolcanics, or even possible candidates for metavolcanics. Accordingly, it has become evident that layering in orthogneisses is not of primary origin but represents examples of tectonic layering of the sort described by Davidson (1984) in tectonites referred to as straight gneisses. Highly strained rocks of this sort have been described for the Piseco anticline by McLelland (1984) and are common throughout the region.

In conclusion, modern quantitative petrologic and isotopic data strongly indicates that the early, and classic, interpretations of the Adirondacks were, in the main, very nearly correct and herein lies a lesson worth pondering. These results offer additional support for the well documented thesis that granites are plutonic, intrusive rocks and that attempts to form them by circuitous, non-intrusive mechanisms are both outdated and destined to failure. This assessment applies equally well to long discredited examples of granitization and to more modern attempts to account for granites by metamorphosing acidic volcanics. Among the critical observations related to these conclusions are quantitative results from geothermometry, geobarometry, geochronology, and petrochemistry. Combined with a modern understanding of tectonic layering, these considerations can greatly constrain the interpretation of complex geologic terranes, as described below for the southern Adirondack region.

INTRODUCTION AND GEOCHRONOLOGY

The Adirondacks form a southwestern extension of the Grenville Province (fig. 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 (AB on fig. 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.

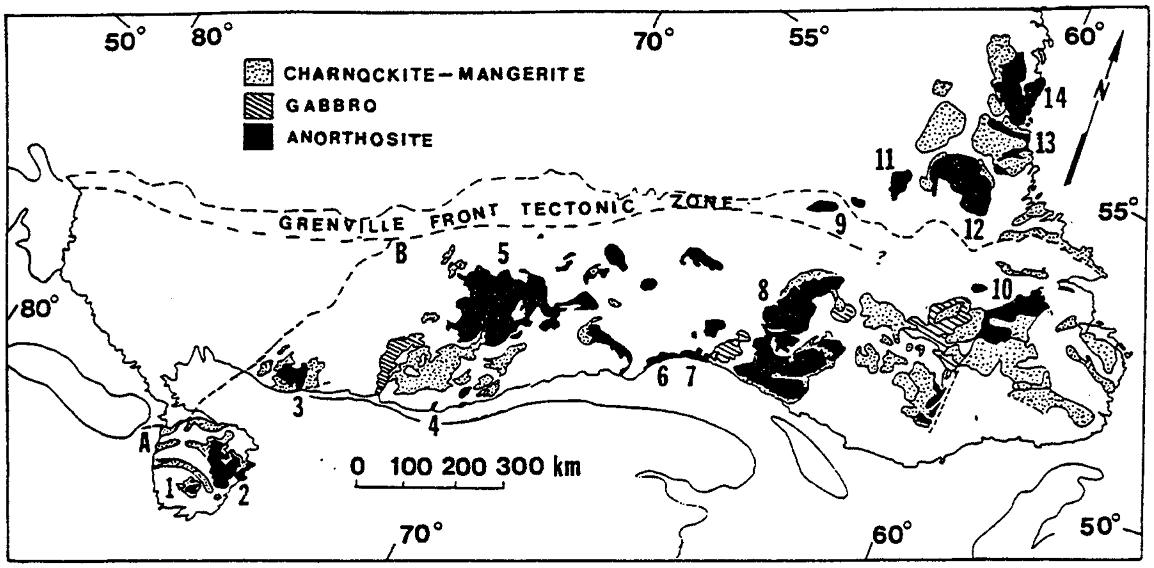

Fig. 1. Generalized map of anorthositic massifs within the Grenville Province and adjacent Labrador. The dashed line, AB, separates terranes with anorthosite massifs on the east from ones lacking them on the west and corresponds to the Carthage-Colton-Gatineau-Chibougamau Line. 1-Snowy Mt. and Oregon 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–Harvre 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 Daly and McLelland (1991), McLelland and Chiarenzelli (1991) and Marcantonio et al. (1990), 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 (fig. 4) indicate that the Adirondack-CMB region experienced wide-spread calcalkaline magmatism from ca. 1400-1230 Ma. Associated high grade (sillimanite-K-feldspar-garnet) 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 and Aleinikoff (1990), in northern Ireland by Menuge and Daly (1991), and in the Texas-Mexico belt of Grenville rocks (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 1400-1220 Ma and may have been related to the assembly of a supercontinent at this time. More locally, this magmatism along with associated metamorphism, represents the Elzevir Orogeny of the Grenville Orogenic Cycle. as defined by Moore and Thompson (1980). Within the Adirondacks Elzevirian rocks are

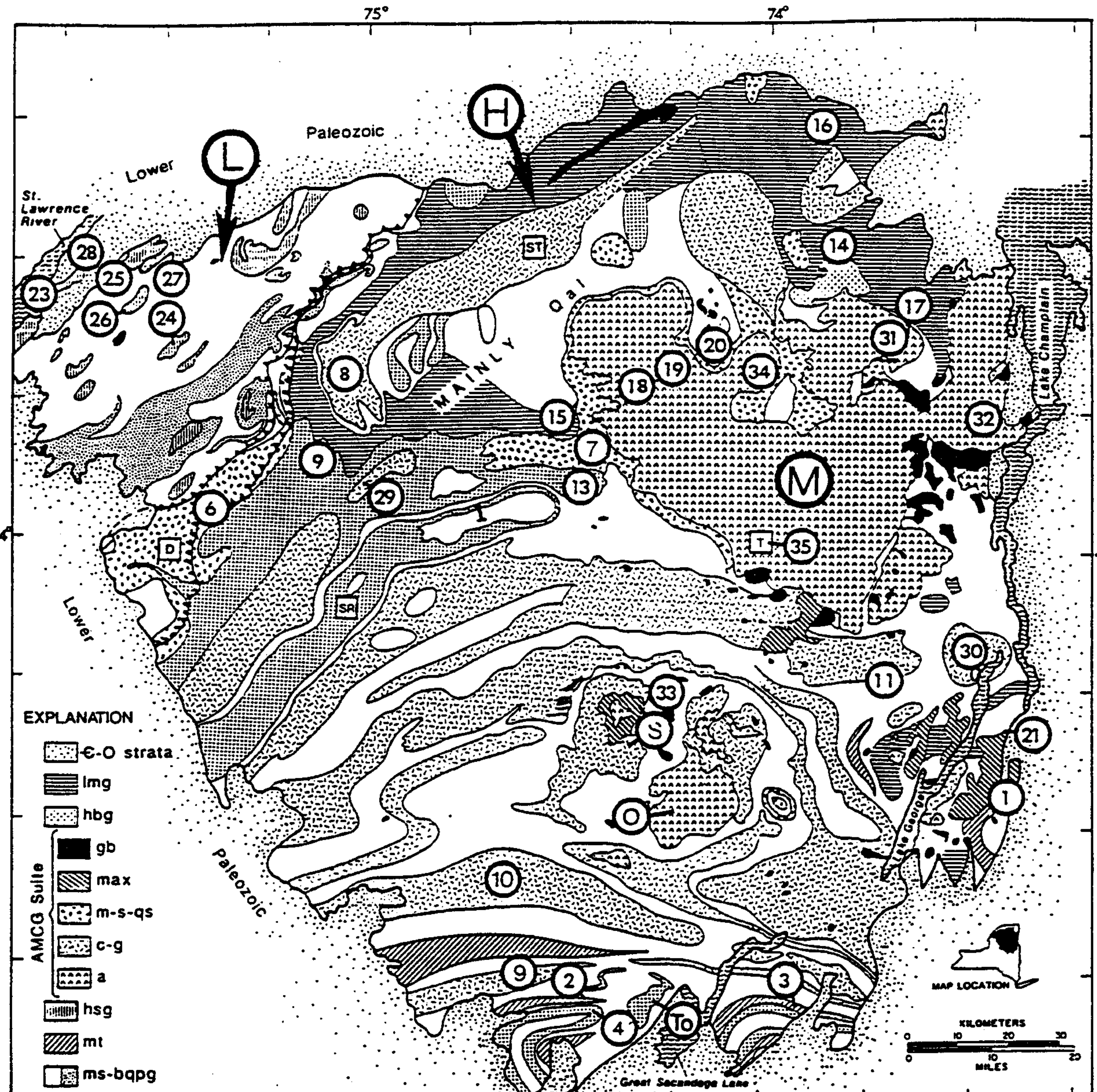

Fig. 2. Generalized geologic map of the Adirondack Highlands (H) and Lowlands (L). The Carthage-Colton Mylonite Zone (CCMZ) is shown with sawteeth indicating directions of dip. Numbers refer to samples listed in Tables 1 and 2. Map symbols: lmg=Lyon Mt. Gneiss, hbg=hornblende-biotite granitic gneiss, gb=olivine metagabbro, max=mangerite with andesine xenocrysts a=metanorthosite,m-s-qs=mangeritic-syenitic-quartz-syeniticgneiss,ms=metasediments,bqpg=biotite-quartz-plagioclasegneiss, 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).

TABLE I

U-PB ZIRCON AGES FOR META-IGNEOUS ROCKS OF THE ADIRONDACK MOUNTAINS

No.	Age (Ma)	Location	Sample No.
HIGHLANDS			
Tonalitic gneiss and older charnockite			
1	1329 ±37	South Bay	AM-87-12
2	1301*	Canada Lake	AM-86-12
3	1336*	Lake Desolation	LDT
4	1233*	Canada Lake	AM-87-13
Mangeritic and charnockitic gneiss			
5	1155 ±4	Diana complex	
6	1147 ±10	Stark complex	AM-86-15
7	1134 ±4	Tupper Lake	AC-85-6
8	1125 ±10	Schroon Lake	9-23-85-7
Older hornblende granitic gneiss			
9	1156 ±8	Rooster Hill	AM-86-17
10	1150 ±5	Piseco dome	AM-86-9
11	1146 ±5	Oswegatchie	AC-85-2
Younger hornblende granitic gneiss			
12	1100 ±12	Carry Falls	AM-86-3
13	1098 ±4	Tupper Lake	AM-86-6
14	1093 ±11	Hawkeye	AM-86-13
Alaskitic gneiss			
15	1075 ±17	Tupper Lake	AM-86-4
16	1073 ±6	Dannemora	AM-86-10
17	1057 ±10	Ausable Forks	AM-86-14
Anorthosite and metagabbro			
18	1054 ±20	Saranac Lake	AC-85-3§
19	1050 ±20	Saranac Lake	AC-85-7‡
20	996 ±6	Saranac Lake	AC-85-9
Xenolith-bearing olivine metagabbro			
21	1144 ±7	Dresden Station	AM-87-11
22	1057	North Hudson	CCAB**
LOWLANDS			
Leucogranitic gneiss			
23	1415 ±6	Wellesley Island	AM-86-16
Alaskitic gneiss			
24	1284 ±7	Gouverneur dome	AC-85-4
25	1236 ±6	Fish Creek	AM-87-4
26	1230 ±33	Hyde School	AC-85-5
Granitic and syenitic gneiss			
27	1160 ±4	Edwardsville	AM-87-5
28	1155 ±15	North Hammond	AM-87-3
HIGHLAND SAMPLES OF SILVER (1969)§§			
29	1113 ±10	Fayalite granite, Wanakena	
30	1109 ±11	Charnockite, Ticonderoga	
31	1084 ±15	Undeformed syenite dike, Jay	
32	1074 ±10	Anorthosite pegmatite, Elizabethtown	
33	1064 ±10	Metanorite, Snowy Mountain dome	
34	1054 ±20	Sheared anorthosite pegmatite, Jay	
35	1009 ±10	Magnetite-ilmenite ore, Tahawus‡‡	

Note: Errors at two sigma.
*Minimum Pb/Pb age.
Data from Grant et al (1986).
§Contains zircon cores, >1113 Ma, air abraded.
‡Baddeleyite age of >1086 ±5 Ma from this sample.
**Contains baddeleyite >1109 Ma.
Monazite age of 1137 ±1 Ma.
§§Decay constants of Steiger and Jager (1977).
‡‡Location same as Sanford Lake (SL) in Figure 1.

Table 2.: Sm-Nd data (sample numbers in Table 1)

sample	L	Zircon age[1]	t_{DM}[2]
ADIRONDACK HIGHLANDS			
Tonalites			
1 :AM87-12	t	1329 ± 36	1403
2 :AM86-12	t	1307 ± 2	1366
3 :LDT	t	>1366	1380
AMCG granitoids			
5 :DIA	s	1155 ± 4	1430
6 :AM86-15	r	1147 ± 10	1495
7 :AC85-6	m	1134 ± 4	1345
9 :AM86-17	e	1156 ± 8	1436
10 :AM86-9	g	1150 ± 5	1346
Younger granitoids			
13 :AM86-6	gd	1098 ± 4	1314
15 :AM86-4	a	1075 ± 17	1576
(repeat)			
:SK2A	tr	c.1060	1330
(repeat)			1373
Metasediment			
:JMCL-1	p	>c.1330	2075
Gabbro			
21 :Ali-1	g	1144 ± 7	1331
ADIRONDACK LOWLANDS			
Welleseley Island			
23 :AM86-16	l	1415 ± 6	1440
Fish Creek			
25 :AM87-4	a	1236 ± 6	1210
:5/90-5	t		
Hyde School			
26 :AC85-5	a	1230 ± 33	1351
:HS3	t	1230 ± 33	1397
:HS4	t	1230 ± 33	1350
Governeur			
24 :AC85-5	a	1284 ± 7	1525
ELZEVIR TERRANE			
Northbrook			
9/88-9	t	1250	1245
Elzevir			
9/88-10	t	1275	1397

1: U-Pb zircon ages in Ma from McLelland and Chiarenzelli (1990a,b) and Grant et al. (1986); 2: Sm-Nd model ages in Ma (DePaolo 1981) from Daly and McLelland (1991) for the Highlands and McLelland, Daly and Perham (1991) for the Lowlands; L: lithologies, a=alaskite, e=enderbite, g=granite, gd=granodiorite, m=mangerite, p=pelite, s=syenite, t=tonalite, tr=trondhjemite, l=leucogranite, initial digits of sample numbers refer to localities in Fig. 2.

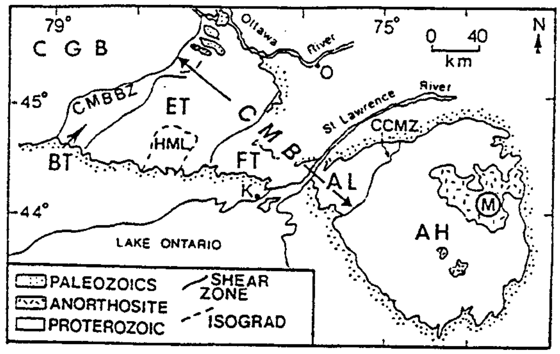

Fig. 3. Southwestern Grenville Province. CMB=Central Metasedimentary Belt, CGB=Central Gneiss Belt, BT=Bancroft Terrane, ET=Elzevir Terrane, FT=Frontenac Terrane, AL=Adirondack Lowlands, HL=Adirondack Highlands, HML=Hastings metamorphic low, K=Kingston, O=Ottawa, CCMZ=Carthage-Colton Mylonite Zone, M=Marcy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

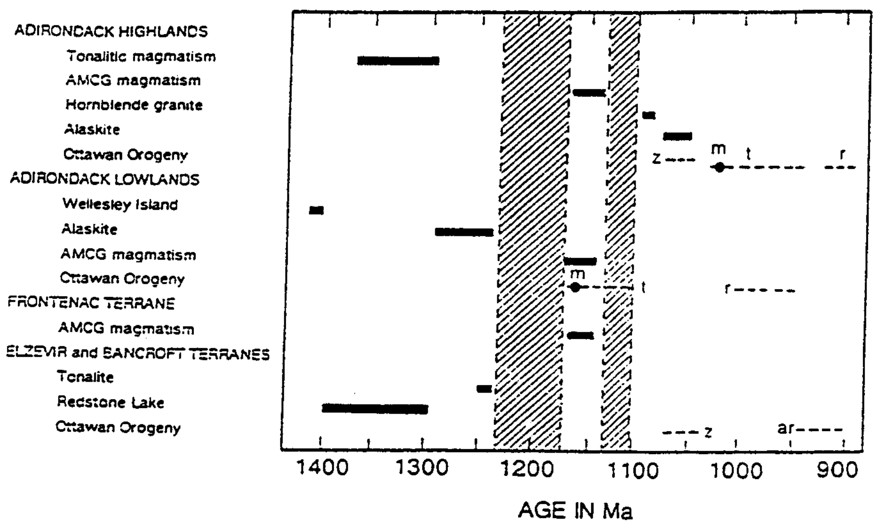

Fig. 4. Chronology of major geological events in the southwestern Grenville Province. z=zircon, t=titanite, m=monazite, r=rutile, ar=Ar/Ar. Diagonal ruling=quiescence. From McLelland and Chiarenzelli (1991).

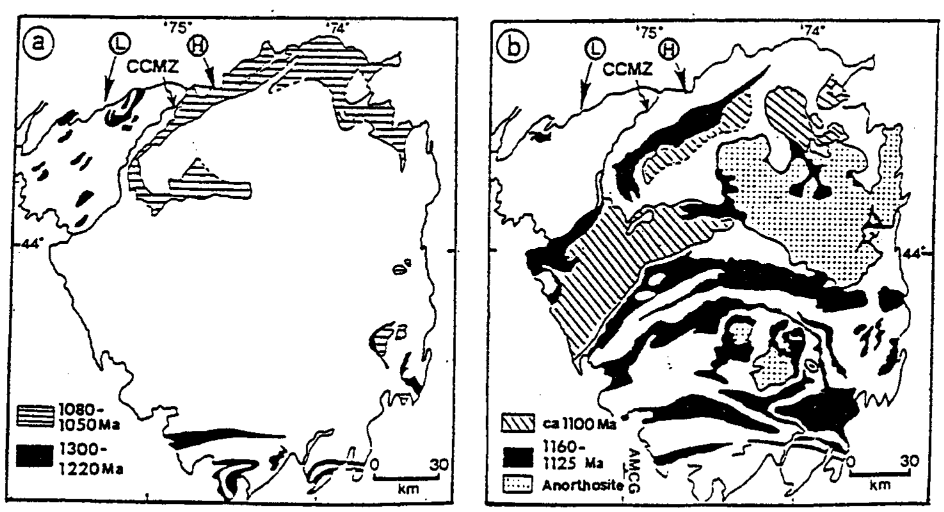

Fig. 5. Chronological designation of Adirondack units. L=Adirondack Lowlands, H=Adirondack Highlands, CCMZ=Carthage Colton Mylonite Zone. From Chiarenzelli and McLelland (1991).

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 long a NNE axis.

Shown in figure 6 are paleoisotherms determined by Bohlen et al. (1985) and zircon ages divided according to origin and degree of disturbance. Note that the locus of disturbed ages corresponds with peak paleotemperatures. This result is discussed again in the metamorphic section. Within the Frontenac-Adirondack region, the Elzevir Orogeny was followed by 40-50 Ma of quiescence terminated at 1170-1130 Ma by voluminous anorogenic (fig. 4) magmatism referred to as the 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 (fig. 5). The large Marcy anorthosite massif (fig. 2) and its associated granitoid envelope have been shown to have an emplacement age of 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 (fig. 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 (1985), McLelland and Husain (1986), and McLelland et. al. (1991a,b) have provided evidence that contact, and perhaps also regional, metamorphism accompanied emplacement of hot (~1100°C, Bohlen and Essene 1978), hypersolvus AMCG magmas. Wollastonite and monticellite occurrences related to thermal pulses from AMCG intrusions occur in proximity to AMCG intrusions (Valley and Essene 1980). In the Lowlands and the Canadian sector of the Frontenac Terrane monazite (table 1., no. 28), titanite (Rawnsley et al. 1987), and garnet ages (Mezger 1990) all indicate high temperatures (~600-800°C) at ca. 1150 Ma. Rutile ages and Rb/Sr whole rock isochron ages document temperatures not exceeding ~500 °C at ca. 1050-1000 Ma.

Following approximately 30 Ma of quiescence (Fig. 4), the Adirondacks, along with the entire Grenville Province, began to experience the onset of the Ottawan Orogeny of the Grenville Orogenic cycle (Moore and Thompson 1980). Initially the Ottawan Orogeny appears 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 Olmstead 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 single bodies of fayalite granite (ca. 1050 Ma) at Wanakena and Ausable Forks (fig. 2).

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 fig. 7a) indicating that these represent juvenile crustal additions. As seen in figure 7a, ϵ_{Nd} evolution curves for AMCG and younger granite suites pass within error of the tonalitic rocks and suggest that the tonalites, together with their own precursors (amphibolites?), 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 (table 1, fig. 7b) demonstrates that it has model neodymium ages and ϵ_{Nd} values closely similar to Highland tonalites. The 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

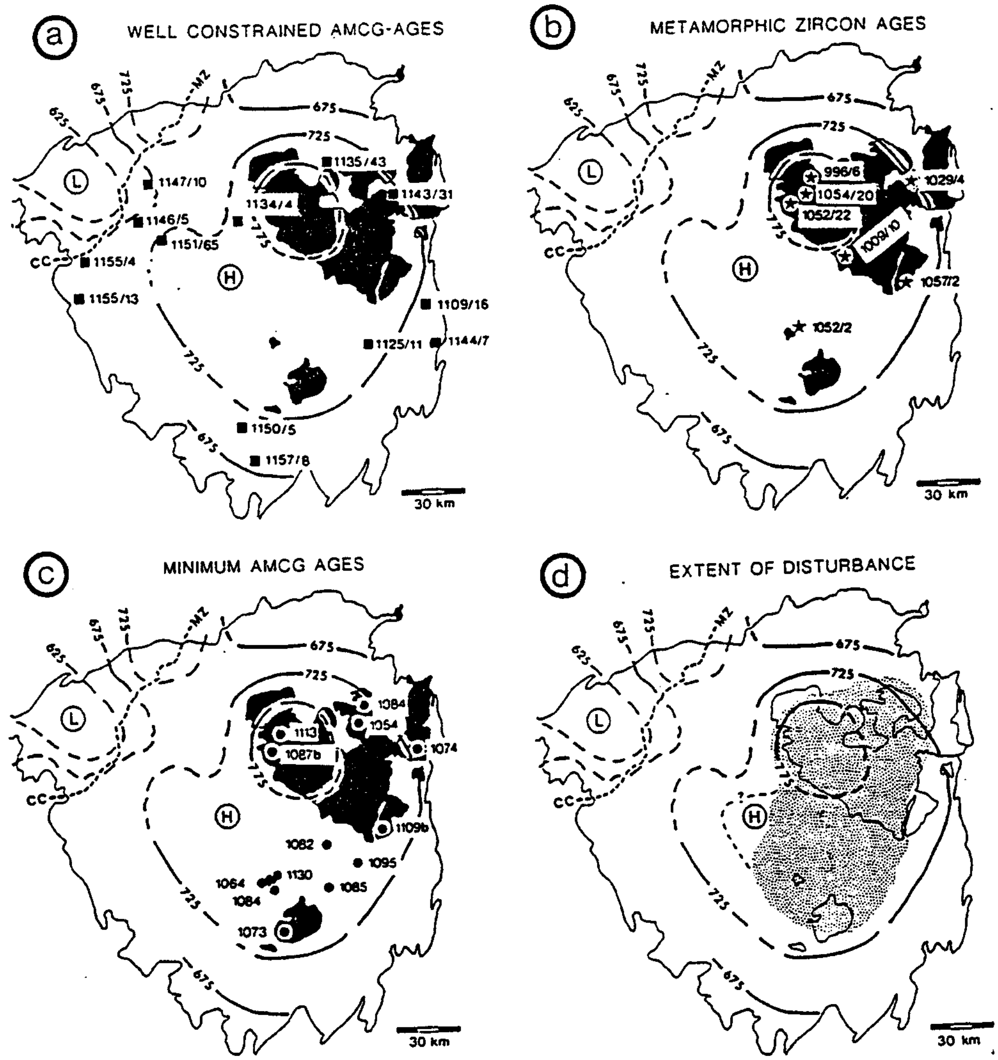

Fig. 6. Relationship between various U/Pb zircon ages and Adirondack paleoisotherms. Shaded area in (d) shows the extent of zircons whose U/Pb systematics have been disturbed. H=Highlands, L=Lowlands, CC=Carthage Colton Mylonite Zone. From Chiarenzelli and McLelland (1991).

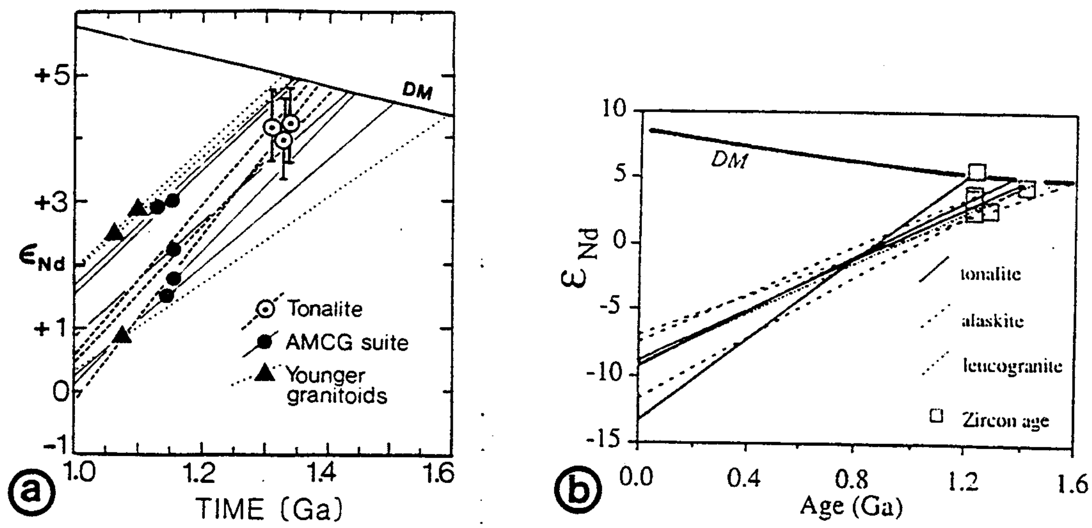

Fig. 7. ϵ_{Nd} evolution diagrams for (a) Adirondack highlands (Daly and McLelland 1991), (b) Adirondack lowlands (Hyde School Gneiss). U-Pb zircon ages are indicated by circles, triangles and squares (from table 1). DM=depleted mantle evolution curve (DePaolo 1981).

formed during the Ottawan Orogeny in response to crustal thickening by thrust stacking (Burchfiel and Royden 1985). 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.

PETROLOGY CHARACTERISTICS OF THE PRINCIPAL ROCK TYPES IN THE SOUTHERN ADIRONDACKS

The following discussion is divided into igneous and metasedimentary sections. Only rock-types occurring within the southern Adirondacks are discussed.

Igneous Rocks

A) Tonalites and related granitoids. Typical whole rock chemistries for these rocks are given in table 3. 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 9 and calc-alkali index versus silica plots in figure 10; both figures illustrate the strongly calcalkaline nature of the Highland tonalite to granitoid suite.

The tonalitic rocks, which will be visited at Stop 4, outcrop in several E-W belts within the southern 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. These are best represented by the Canada Lake Charnockite (Stop 3) and by the large Tomantown pluton (fig. 2) whose minimum emplacement age is 1184 Ma (table 1).

TABLE 3

	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	0.75	1.16	0.37	0.59	0.36	0.04	0.89	0.72	0.20
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	0.87	2.83	3.34	1.11
Fe_2O_3	6.03	6.79	1.53	1.1	0.96	0.11	1.82	1.59	1.07
MnO	0.10	0.08	0.04	0.02	0.03	0.01	0.04	0.10	0.02
MgO	0.46	1.15	0.55	0.84	0.43	0.20	0.45	0.51	0.19
CaO	2.71	2.56	1.66	2.3	1.87	0.85	2.26	2.57	0.88
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	0.10	0.51	0.10	0.24	0.09	0.01	0.23	0.19	0.04
LOI	0.39	0.74		0.40	0.27	0.17	0.19	0.39	0.17
Σ	99.87	100.96	99.42	99.63	99.65	99.61	100.59	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	60.64	61.05	60.94	62.14
TiO_2	0.88	1.55	0.65	0.51	1.14	0.78	1.39	0.36
Al_2O_3	16.40	14.87	20.32	13.90	15.27	15.98	15.91	12.35
FeO	3.96	10.25	2.81	3.1	9.28	4.60	6.51	10.32
Fe_2O_3	1.49	2.80	0.43	1.34	1.77	2.10	1.02	1.7
MnO	0.09	0.24	0.01	0.05	0.19	0.05	0.14	0.01
MgO	1.06	0.96	1.47	0.52	0.74	1.64	1.70	0.83
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	0.30	0.65	0.32	0.11	0.31	0.42	0.46	0.09
LOI	0.41	0.37	0.43	0.39	0.01	0.91	0.37	0.67
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-85-6	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	70.01	69.05	72.39
TiO_2	0.48	0.55	0.41	0.18	1.43	0.35	0.46	0.42	0.69	0.57	0.38
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	0.84	5.13	1.2	4.23	3.50	1.6
Fe_2O_3	0.93	1.18	1.12	0.61	1.16	2.1	1.11	2.8	1.5	1.42	4.13
MnO	0.06	0.07	0.05	0.01	0.02	0.02	0.14	0.03	0.03	0.01	0.03
MgO	0.49	0.45	0.11	0.01	0.67	0.33	0.08	0.56	0.01	0.17	0.29
CaO	1.99	1.81	1.43	0.45	0.75	1.55	1.25	2.35	0.94	0.35	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	0.52
P_2O_5	0.13	0.12	0.07	0.03	0.17	0.07	0.04	0.13	0.17	0.12	0.70
LOI	0.40	0.58	0.5	0.30	0.25	0.39	0.21	0.31	0.23	0.41	0.10
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

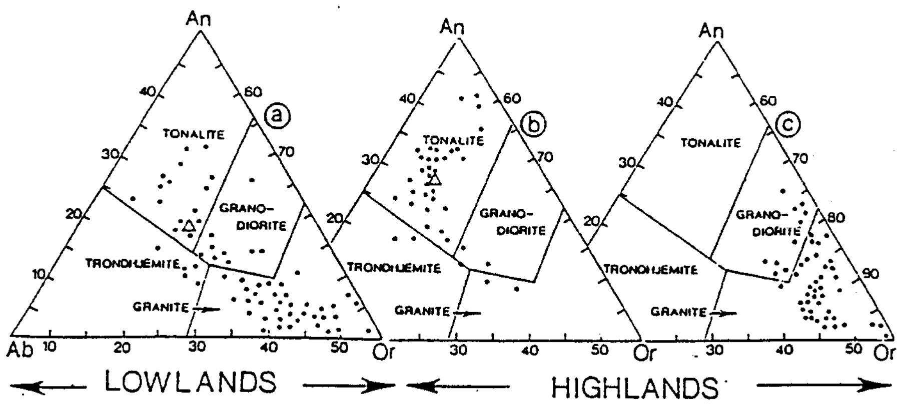

Fig. 8. Plots of normative albite (Ab)-anorthite(An)-orthoclase (Or) for (a) Hyde School Gneiss, (b) Highlands tonalites, and (c) Tomantown pluton. Open triangles give average values for tonalitic samples. Definition of fields due to Barker (1979).

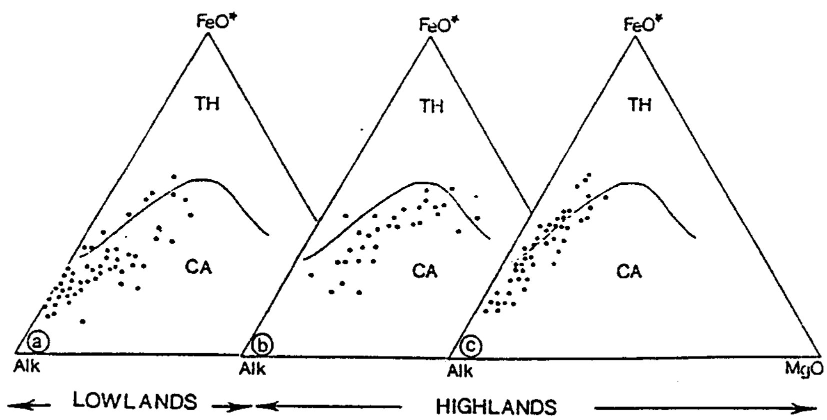

Fig. 9. AFM plots for (a) Hyde School Gneiss, (b) Highland tonalites, and (c) Tomantown pluton.

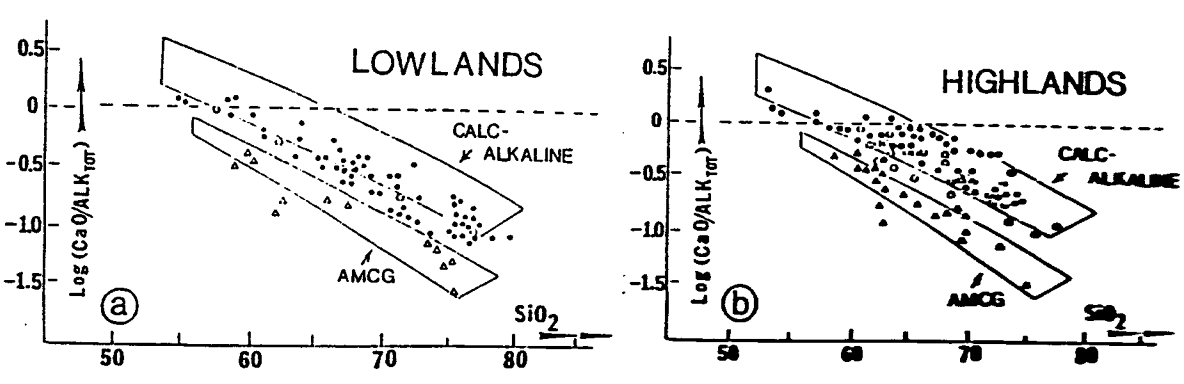

Fig. 10. Calcalkali ratio vs. wt % SiO_2 for (a) the Adirondack Lowlands and (b) the Adirondack Highlands. In (a) open circles are average values for Hyde School Gneiss, closed circles for typical Hyde School Gneiss, and open triangles for AMCG type rocks. In (b) open circles are for Tomantown pluton, closed circles for older calcalkaline rocks, and open triangles for AMCG rocks. Fields from Brown (1982).

McLelland et al. (1991b) have interpreted the early calcalkaline rocks of the Highlands as correlative with the Hyde School Gneiss of the Adirondack Lowlands (fig. 2). This interpretation is consistent with the Sm-Nd results (table 1b) discussed previously and shown in figure 7.

B) AMCG Suite. Within the southern Adirondacks AMCG rocks are widely developed and abundantly represented in the Piseco anticline (Stop 6) as well as the Oregon (Stop 8) and Snowy Mt. Domes. The chemistry of granitoid (mangeritic to charnockitic varieties of these rocks is given in table 3, especially for the older anorogenic plutonic rocks to which southern Adirondack suites belong. As shown in figure 10, the AMCG rocks have calcalkali-silica trends that are distinctly different than those shown by the tonalitic suites. McLelland (1991) and McLelland and Whitney (1991) have shown that the AMCG rocks exhibit anorogenic geochemical characteristics (figs. 11, 12, 13) and also 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, fig. 2). Bimodality is best demonstrated by noting the divergent differentiation trends of the granitoid members on the one hand and the anorthositic-gabbroic rocks on the other (Buddington 1972). This divergence is nicely exhibited by the variation of the FeO-MgO ratio with wt.% SiO_2 (fig. 12) Ga-Al_2O_3 trends (fig. 13), and by Harker variation diagrams for AMCG rocks of the Marcy massif (fig. 14) (McLelland 1989). The extreme low-SiO_2, high-iron end members (fig. 14) of the anorthosite-gabbro family will be seen at Stop 8 and are believed to represent late liquids developed under conditions of low oxygen fugacites (i.e., dry, Fenner-type trends).

C) Metasedimentary Rocks. Within the southern Adirondacks the metasedimentary sequence is dominated by quartzites and metapelites with marbles being virtually absent. The quartzites are exceptionally thick and pure and comprise an ~1000 m-thick unit referred to as the Irving Pond Quartzite (Stop 2). Of even greater extent, as well as thickness, is the Peck Lake Formation which consists of garnet-biotite-quartz-oligoclase ± sillimanite gneiss (referred to as kinzigite) together with sheets, pods, and stingers of white, minimum melt granite that commonly contains garnets (Stop 1). McLelland and Husain (1986) interpreted the kinzigites and their leucosomes as restite-anatectite pairs and attributed partial melting to heating accompanying AMCG magmatism. It is now believed that an additional period of anatexis probably preceded the 1130-1150 Ma AMCG magmatism during the 1300-1220 Elzevir Orogeny.

The occurrence of anatexis within the kinzigites is corroborated by the presence of sparse hercynitic spinel within either garnets or sillimanite-rich wisps in leucosomes. McLelland et al. (1991a) have shown that extraction of anatectic material from the least altered kinzigites can satisfactorily account for the composition of more aluminous, lower-silica kinzigites. The ultimate evolution of this process would be to produce assemblages of aluminous sillimanite-garnet-biotite gneiss together with granitic material of the sort that characterizes the Sacandaga Formation (Stop 9).

Based on the bulk chemistry of kinzigites in the southern Adirondacks, McLelland and Husain (1986) interpreted their protoliths as Proterozoic greywackes and shales. More recently, McLelland et al. (1991b) have provided evidence to support the conclusion that the Peck Lake Fm. kinzigites of the southern Adirondacks can be correlated with the markedly similar Major Paragneiss of the Adirondack Lowlands (bqpg on fig. 2). McLelland and Isachsen (1986) have also argued that the Peck Lake Fm., and associated rocks, continues eastward into the eastern Adirondacks in the vicinity of Lake George.

In contrast to the southern and eastern Adirondacks, the central Adirondacks contain only sparse kinzigite, and metasediments are principally represented by synclinal keels of marble and calcsilicate (Stop 7). It is possible that the change from carbonate to pelitic metasediments

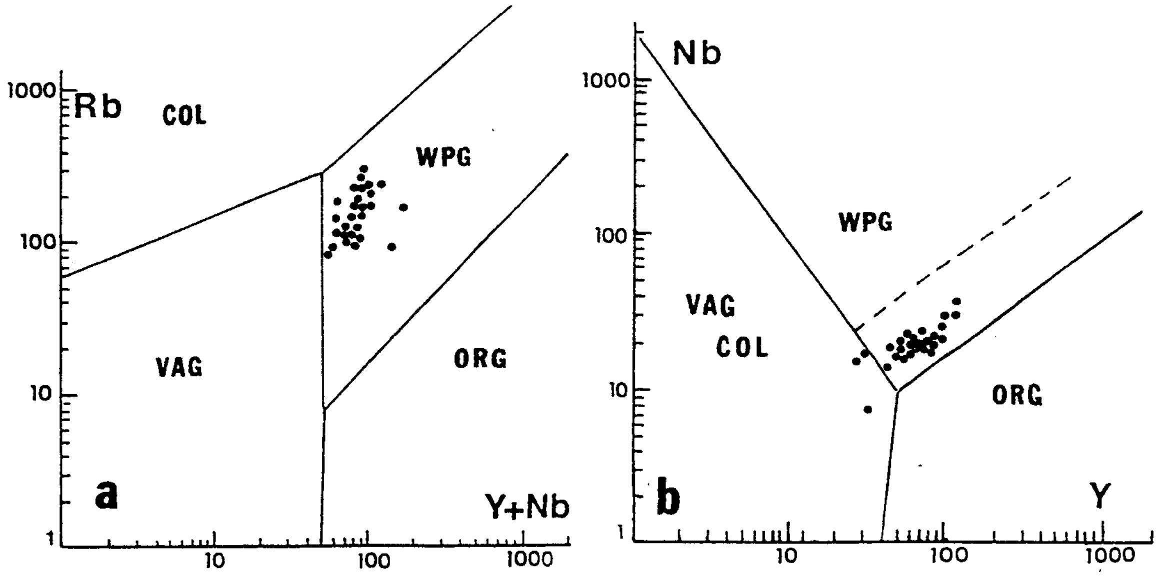

Fig. 11. Tectonic discrimination diagrams (Pearce et al. 1984) for AMCG granitoids from the Marcy massif. COL=collisional, WPG=within plate granites, ORG=-ocean ridge granites, VAG=volcanic arc granites.

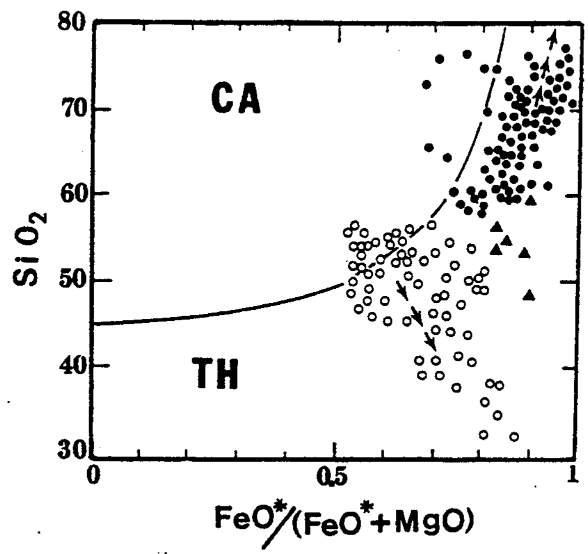

Fig. 12. FeO*-MgO vs. wt.% SiO_2 variation for anorthositic suite (open circles) and mangeritic-charnockite suite (filled circles). Triangles designate mixed rocks at contacts. Arrows indicate differentiation trends of the two suites, CA=calcalkaline, TH=tholeiitic. (after Anderson 1983)

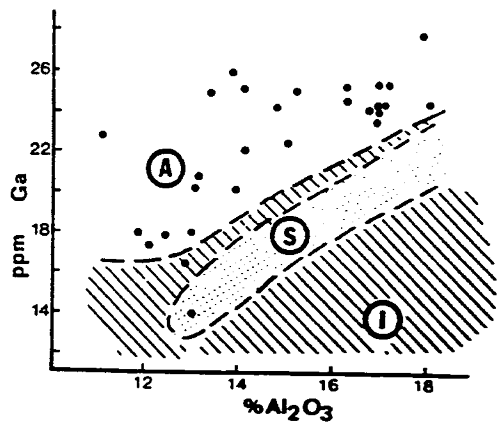

Fig. 13. Ga vs. Wt.% Al_2O_3 for AMCG suite granitoids of the Marcy massif. Fields of A-,S-, and I-type granites and shown (after White and Chappell 1983).

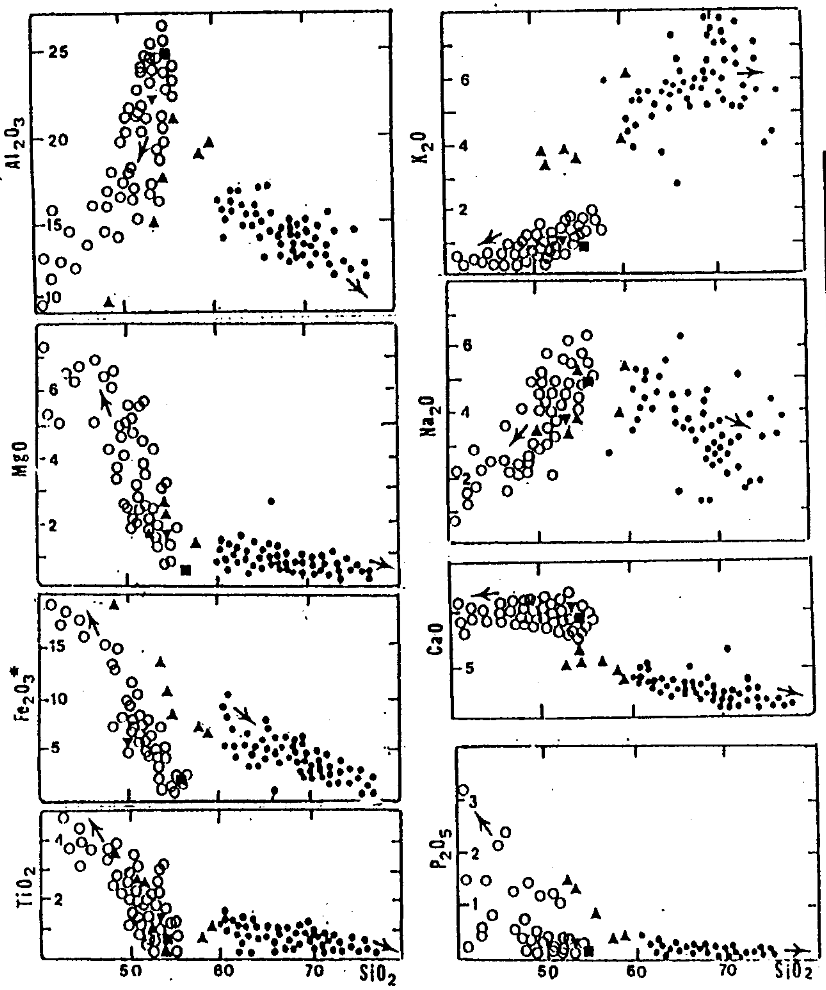

Fig. 14. Harker variation diagrams or AMCG-rocks of the Marcy massif. Open circles=anorthositic suite, filled circles=granitoid suite, upright triangles=mixed rocks, inverted triangles=Whiteface facies, square=Marcy facies. Arrows indicate differentiation trends.

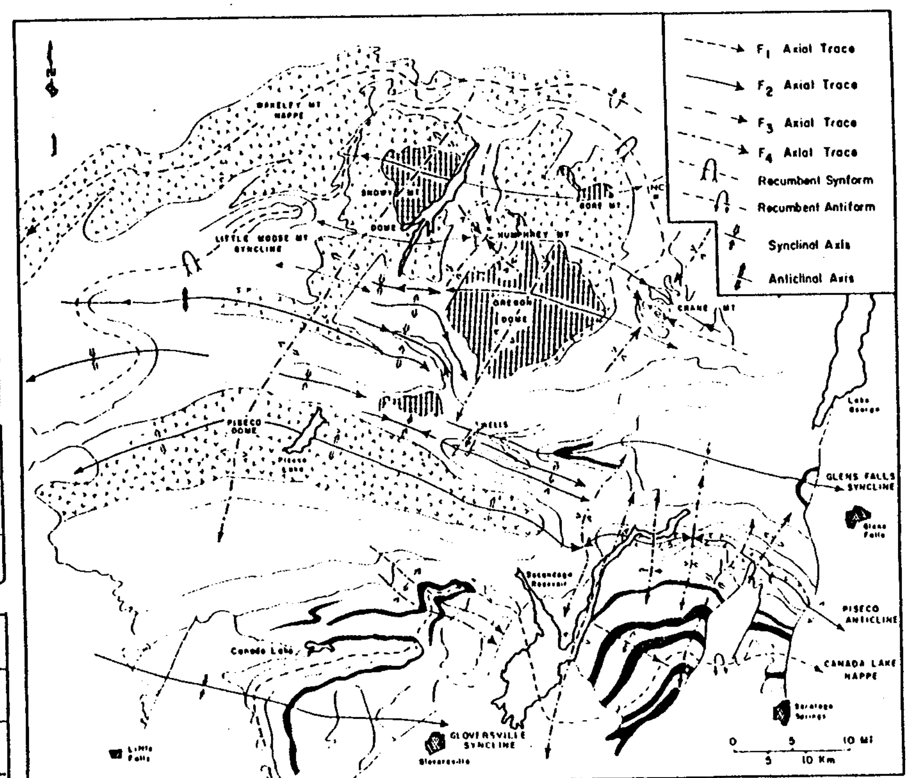

Fig. 15. Fold axes within the southern and central Adirondacks. Designation of folds as synclines and anticlines is provisional, since facing directions are not yet known.

corresponds to an original shelf to deep water transition, now largely removed by later intrusion, doming, and erosion. The Irving Pond quartzite may represent a siliceous clastic cap closing out the earlier deep water basin.

A single specimen of metapelite (no. 21, table 2) has yielded a T_{DM} of 2075 Ma. This model age approximates the time at which source rocks for the metasediment separated from the mantle. Although the age may be the result of mixing rocks >2075 Ma with younger components, the older material clearly predates any possible Adirondack sources.

STRUCTURAL GEOLOGY

The southern Adirondacks is an area of intense ductile strain, essentially all of which must postdate the ca. 1150 Ma AMCG rocks which are involved in each of the major phases of deformation, i.e., the regional strain is associated with the Ottawan Orogeny.

As shown in figures 2 and 15, 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 (fig. 16).

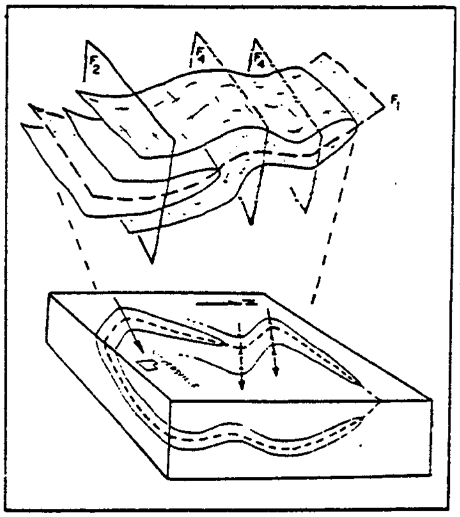

Fig. 16. Block diagram showing how interference between F_1, F_2, and F_4 fold sets produce the outcrop pattern of the F_1 Canada Lake isocline. The axial plane of the F_1 fold is stippled and its fold axis plunges 10-15° to the southeast. The city of Gloversville is shown.

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 (deWaard 1964) and the other two are antiformal, and suspected to be anticlinal, but the lack of stratigraphic facing directions precludes any certain age assignments although these are designated in figure 15 on a provisional basis. 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 (fig. 15). 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 Snowy Mt. and Oregon domes (fig. 15). 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 (fig. 15).

Kinematic indicators (mostly feldspar tails) in the area suggest that the dominant displacement 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. A movement picture consistent with this is still under investigation, although regional extension analogous to that in core complexes might resolve the situation.

METAMORPHISM

Figure 6 shows the well known pattern of paleoisotherms established by Bohlen and Essene (1977) and updated in 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 1985). The bull's eye pattern of paleoisotherms, centering on the Marcy massif, is believed to the due to late doming centered on 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).

Bohlen et al. (1985) interpret the paleotemperature pattern of figure 6 as representative of peak metamorphic temperatures in the Adirondacks, and paleopressures are interpreted similarly. Chiarenzelli and McLelland (1991) show that disturbance of U-Pb systematics in zircons corresponds with Bohlen et al.'s (1985) paleoisotherms (fig. 6), and this correlation strengthens the conclusion that the pattern is one of peak temperatures rather than a retrograde set frozen in from a terrane of uniform temperatures in the range ~750°-800°C.

The P,T conditions of the Adirondack are those of granulite facies metamorphism, and for the most part conditions correspond to the hornblende-clinopyroxene-almandine subfacies of the high-pressure portion of the granulite facies. These conditions must have been imposed during the Ottawan Orogeny in order to have affected 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 titanite and garnet U-Pb ages of ca. 1030-1000 Ma in the Highlands (Mezger 1990). 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. 1987a), and the report of some 1150 Ma U-Pb garnet ages (Mezger 1990), 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.

The present day depth to the Moho beneath the Adirondack Highlands is ~35 km (Katz 1955). Since metamorphic pressures of 7-8 kbar correspond to ~20-25 km depth of burial, it follows that during metamorphism the Adirondack region consisted of a double thickness of continental crust. Present day examples of doubly thickened continental crust are found in continent-continent collisional margins such as the Himalayas or Andean margins such as along the coast of South America. The latter model is not readily applicable to the Ottawan-age Adirondacks, because of the absence of calcalkaline magmatism of that age. On the other hand, the Himalayan-Tibetan analogue provides a strikingly consistent model, including the rather limited amount of associated magmatism. Because no suggestion of a suture exists between the Green Mts. of Vermont and the Grenville Tectonic Front, and because of the dominance of tectonic vergence to the northwest throughout the region, the Ottawan plate margin has been placed east of the Grenville inliers of the Appalachians and assigned an eastward dip. Although highly speculative, this possibility, together with other plate tectonic reconstructions are shown in figure 17.

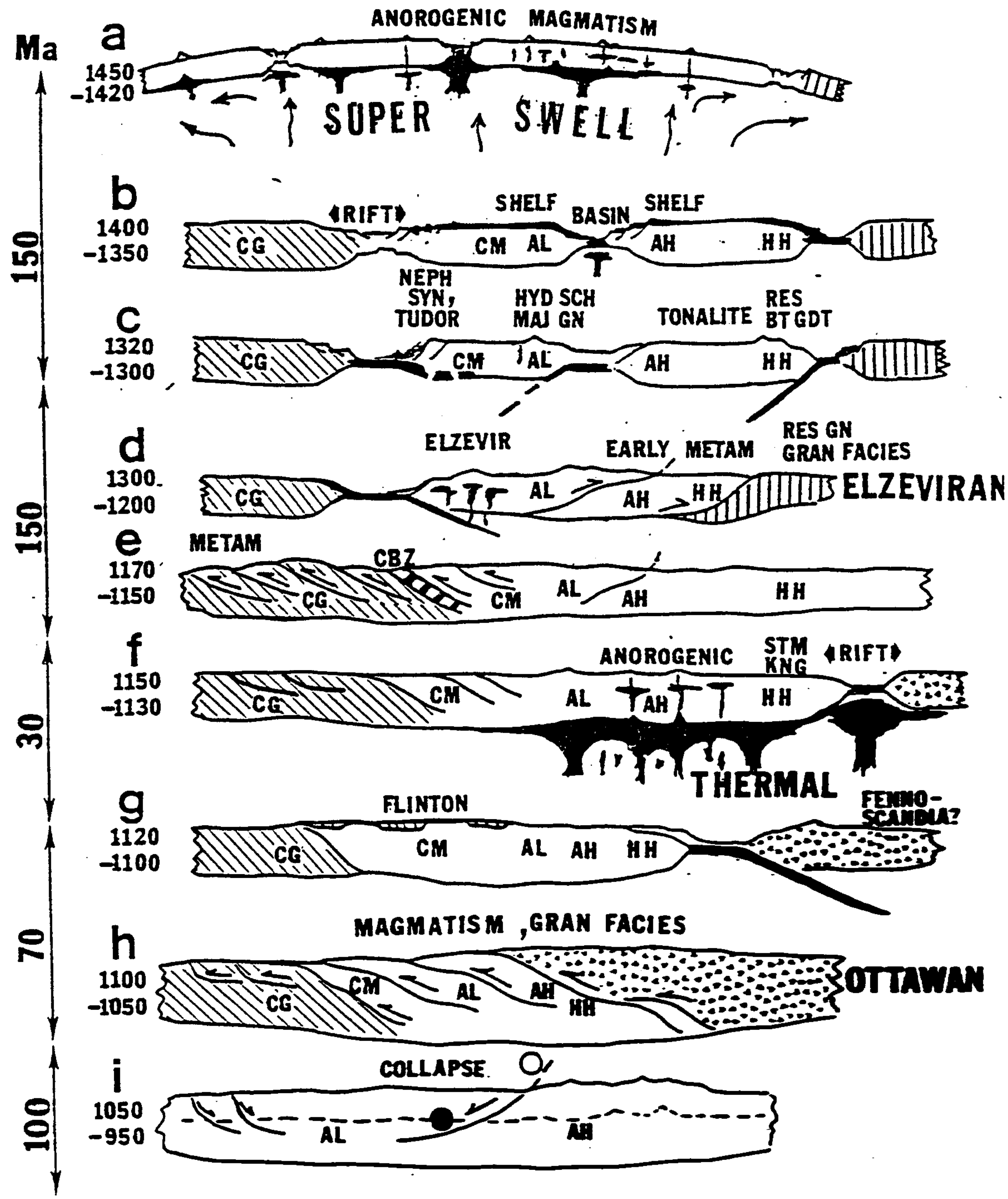

Fig. 17. Hypothetical plate tectonic scenarios for the southwestern Grenville Province.

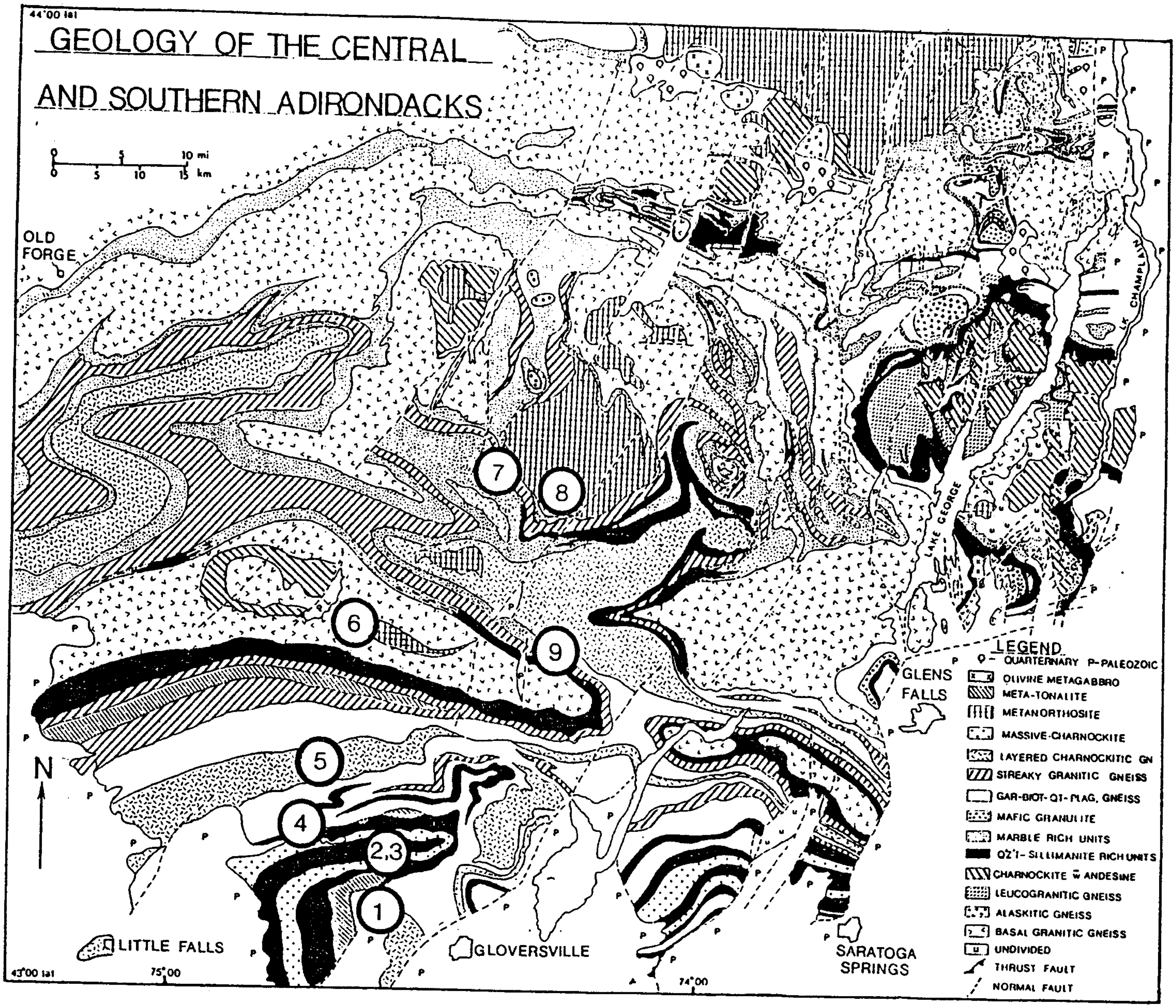

Fig. 18. Geologic map of the southern and central Adirondacks with field trip stops 1-9 indicated (McLelland and Isachsen 1986).

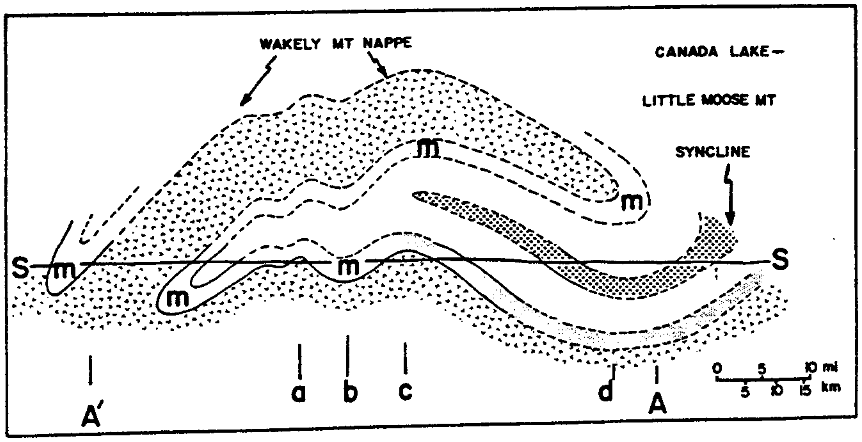

Figure 9. Cross section along A-A' of Figure 2. Several units have been omitted for sake of clarity. (a) - Spruce Lake Anticline, (b) - Glens Falls Syncline, (c) - Piseco Anticline, (d) - Gloversville Syncline. Patterned rock unit symbols as in Figure 3.

METAMORPHISM

Introduction

Metamorphism of rocks in the Adirondack highlands has been investigated extensively for the past fifteen years (see e.g. Buddington, 1963, 1965, 1966; de Waard, 1964a, 1965a, 1965b, 1967, 1969, 1971; Whitney and McLelland, 1973; McLelland and Whitney, 1977; Bohlen and Essene, 1977;

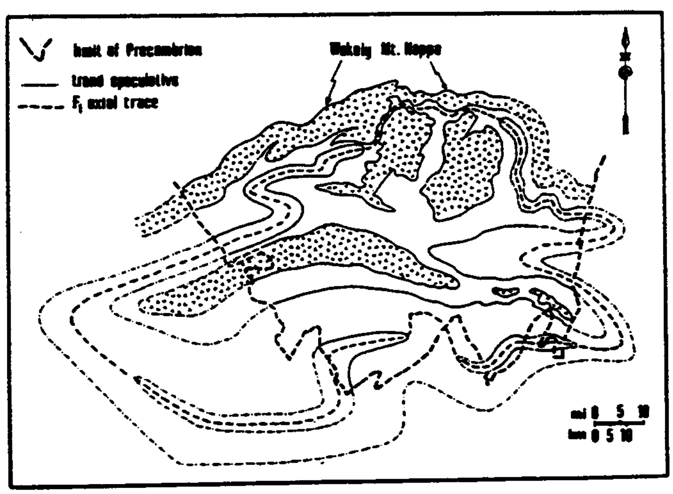

Figure 10. Generalized map showing known and projected axial trace of Canada Lake-Little Moose Mt. Syncline. Heavy dashed line marks Proterozoic-Paleozoic boundary as shown in Figures 1 and 2.

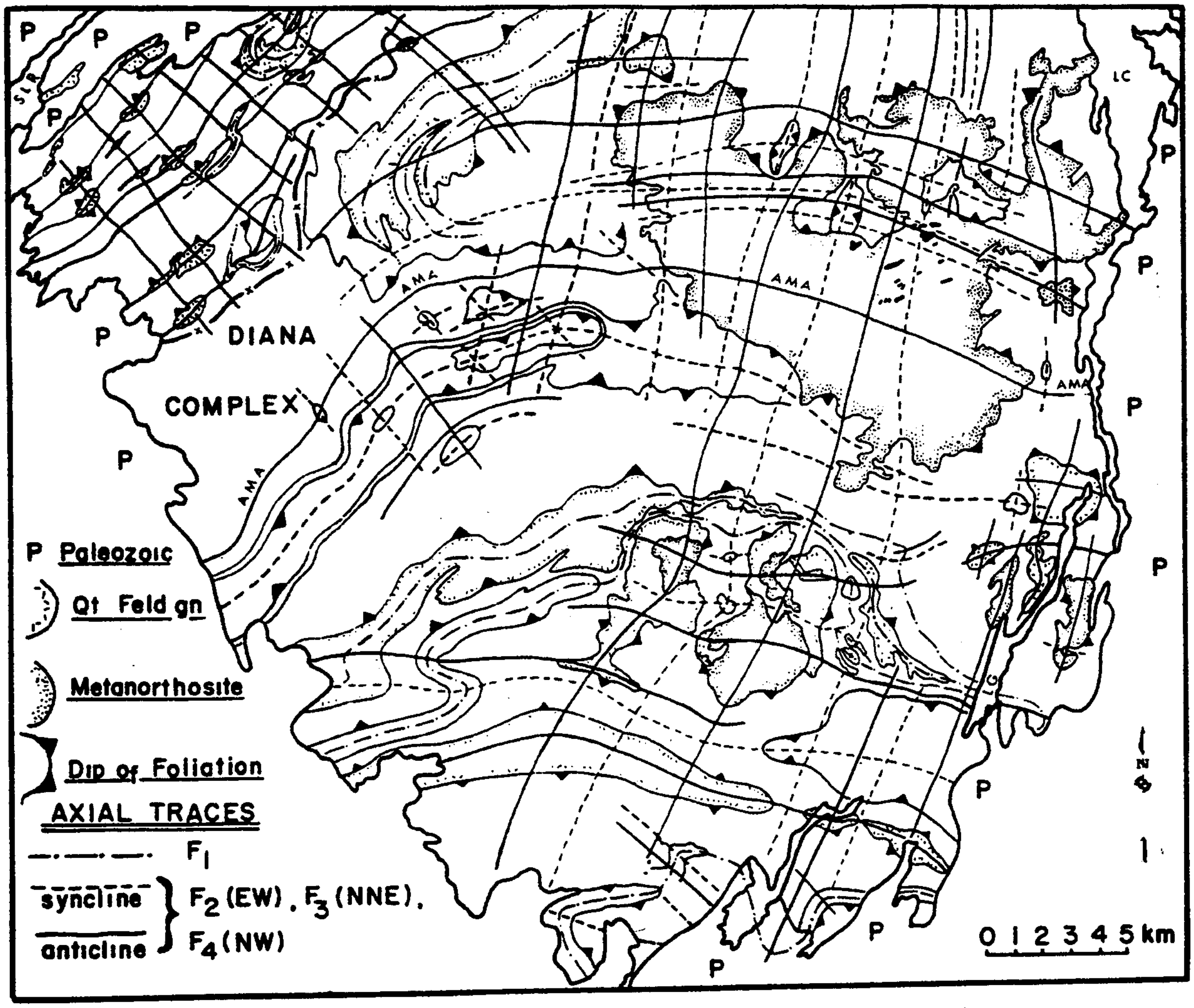

Figure 11. Hypothetical structural framework for Adirondacks. AMA-Arab Mt. Anticline.

Essene and others, 1977; Boone, 1978; Boone and others, in prep.; Valley and Essene, 1976; Jaffe and others, 1977; Stoddard, 1976). Engel and Engel (1962) made an early and fundamental contribution when they delineated in part the orthopyroxene isograd in the northwestern Adirondacks (see Fig. 12). Orthopyroxene is the diagnostic mineral of high-grade metamorphism and its regional stable occurrence with plagioclase and garnet demonstrates that metamorphic conditions of the granulite facies were attained to the east of the orthopyroxene isograd.

de Waard (1971) proposed a three-fold subdivision of the granulite facies in the Adirondack highlands (Fig. 12). The three zones, in order of progressive metamorphism, are the (1) biotite-cordierite-almandite subfacies, (2) hornblende-orthopyroxene-plagioclase subfacies, and (3) hornblende-clinopyroxene-almandite subfacies. de Waard (1971) believed the subfacies represent three stages of increasing granulite-facies metamorphism constituting an Adirondack Type of metamorphic series. All stops

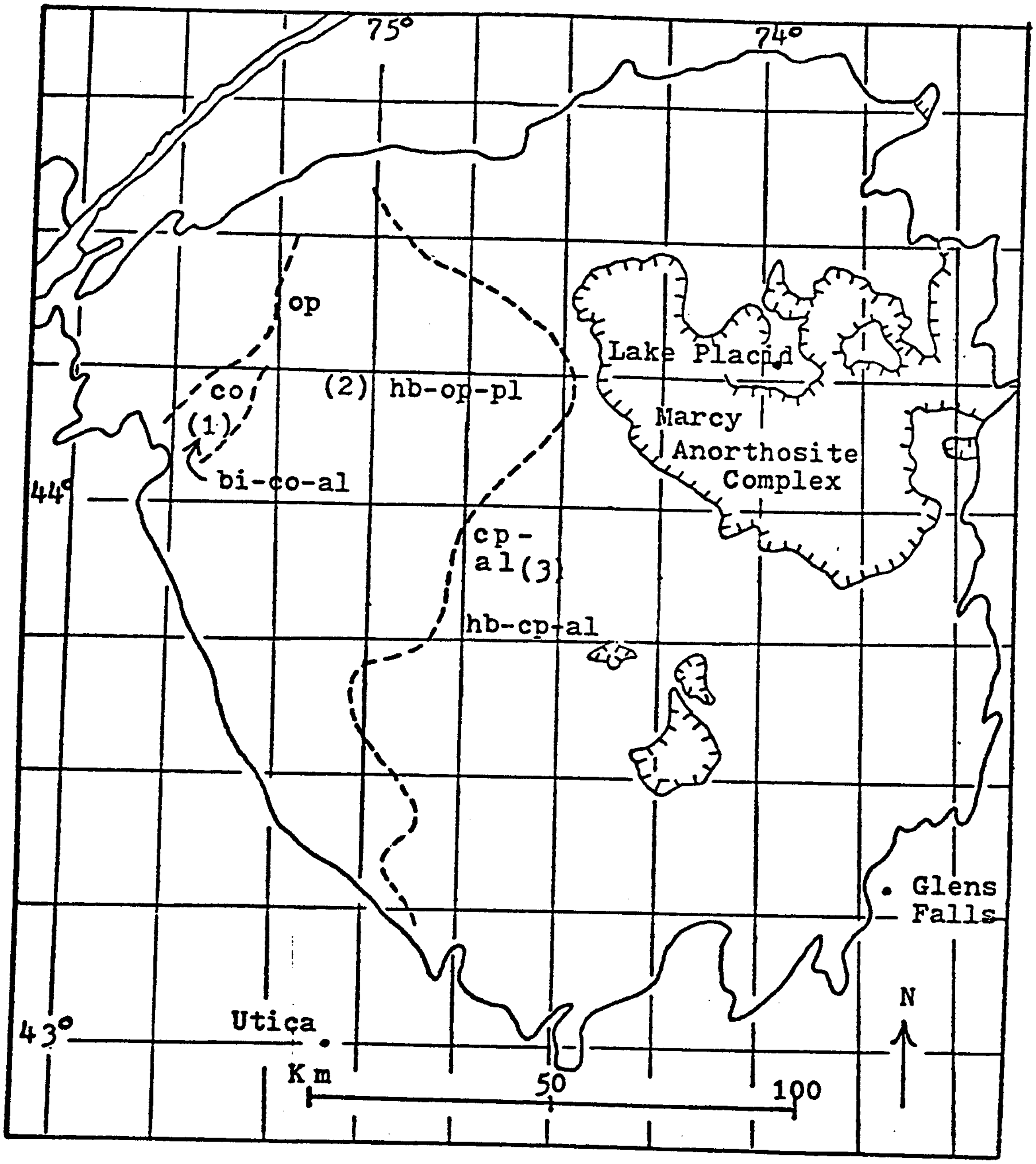

Figure 12. Outline map (modified after de Waard, 1971) of Precambrian terrane of Adirondack Mountains and Northwest Lowlands showing de Waard's proposed subdivision of granulite facies of Adirondack highlands: (1) biotite-cordierite-almandite subfacies, (2) hornblende-orthopyroxene-plagioclase subfacies, and (3) hornblende-clinopyroxene-almandite subfacies. Three isograds are shown; parts of isograds were mapped by de Waard (1971), Engel and Engel (1962), and Buddington (1963). Solid contact is trace of Precambrian-Paleozoic boundary; hatchured contacts delineate boundaries of relatively larger anorthosite complexes.

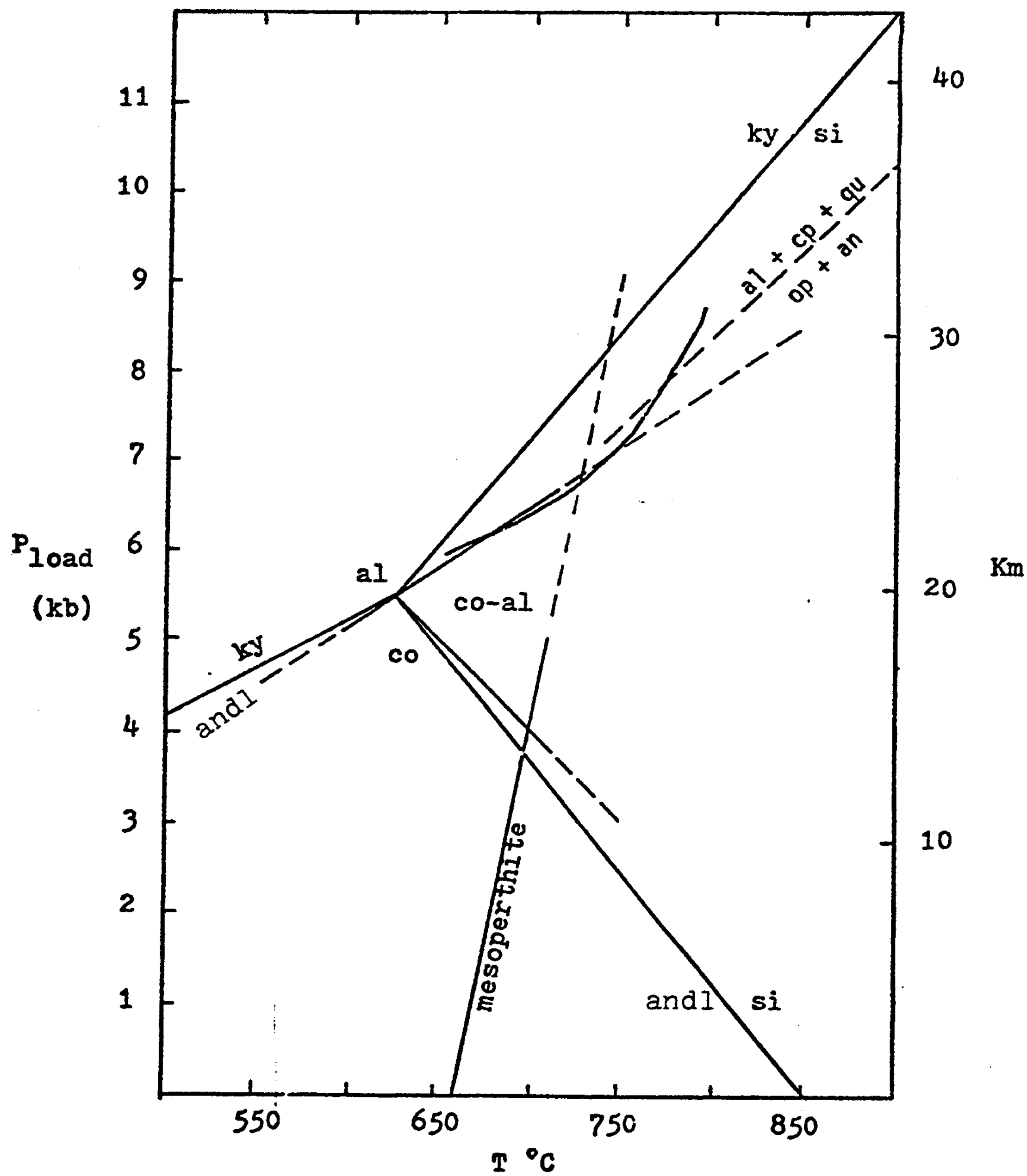

Figure 13. Petrogenetic grid (modified after de Waard, 1969, p. 129) composed of stability boundaries for solid-solid reactions involving anhydrous phases only. Curved line is geothermal gradient representing P_{load}-T conditions of metamorphism favored by de Waard. Grid is based upon following experimentally derived curves: reaction orthopyroxene + plagioclase $\rightleftharpoons$ clinopyroxene + almandite + quartz after Ringwood and Green (1966); triple point of aluminum silicates after Gilbert, Bell, and Richardson (1968) and Holdaway (1968); kyanite-sillimanite boundary after Richardson, Bell, and Gilbert (1968); cordierite and almandite stability fields after Hirschberg and Winkler (1968); and solvus temperature maximum of alkali feldspars after Orville (1963).

on this field trip are located in zone (3), to the east of the garnet-clinopyroxene isograd. This isograd, far from sharply defined, is based upon the first recognition, in the field, of garnet in quartzo-feldspathic (charnockitic) rocks (de Waard, 1971). Thus, all rocks in the field-trip area have been subjected to P-T conditions appropriate for the hornblende-clinopyroxene-almandite subfacies of the granulite facies.

In addition, the area of the biotite-cordierite-almandite subfacies has greater areal extent than exhibited in Figure 12 (P.R. Whitney, 1977, pers. comm.). Cordierite and garnet-bearing pelitic gneisses have been reported by Stoddard (1976) to occur north-northeast of zone (1). It can be inferred from these locations that zone (1) now extends north-northeast, ~parallel to the orthopyroxene isograd, almost to the Precambrian-Paleozoic boundary.

Only rock types that have yielded information about P-T conditions of metamorphism are discussed in the following section. Mineral-name abbreviations used are: al - almandite; andl - andalusite; an - anorthite; bi - biotite; ca - calcite; co - cordierite; cp - clinopyroxene; Kf - K-rich alkali feldspar, chiefly microcline, usually perthitic; ky - kyanite; ma - magnetite; op - orthopyroxene; pf - plagioclase feldspar; qu - quartz; sc - scapolite; si - sillimanite.

Charnockitic and Granitic Gneiss

Hornblende-clinopyroxene-almandite subfacies: de Waard proposed (1964a) that with increasing metamorphic conditions the typomorphic orthopyroxene-plagioclase association of the granulite facies becomes imcompatible and is replaced by the higher density almandite-clinopyroxene association. This replacement marks the start of the hornblende-clinopyroxene-almandite subfacies, and de Waard proposed (1964a) the following reaction to account for the garnet-clinopyroxene formation:

$$\text{6 orthopyroxene} + \text{2 anorthite} \rightleftharpoons \text{clinopyroxene} + \text{2 almandite} + \text{2 quartz} \qquad (1)$$

Reaction (1) has a positive P-T slope (see petrogenetic grid, Fig. 13). de Waard (1967) considered two manners in which reaction (1) proceeded to the right. In one instance, the reaction progression may indicate a gradual increase in both T and P_{load} during progressive regional metamorphism in which the central and eastern Adirondacks were subjected to hornblende-clinopyroxene-almandite-subfacies conditions. In the second instance, the reaction may have been produced by a decrease in T with little change in P_{load} during retrogressive metamorphism. However, de Waard (1967) favored the first instance of increasing P_{load} and T for the following reasons: (1) the clinopyroxene-garnet-quartz assemblage has a higher density than the orthopyroxene-anorthite assemblage and represents a reduction in molar volume of ~14 percent, thus favoring higher P_{load}, and (2) cordierite, considered indicative of relatively lower P_{load} (or higher T), is present in pelitic gneisses in the northwestern portion of the highlands, but absent in gneisses of comparable composition to the SE (Fig. 12).

Martignole and Schrijver (1971) contended that reaction (1) proceeds to the right as a consequence of de Waard's second instance: decrease in T with little change in P_{load}. They believe the formation of garnet and clinopyroxene does not represent a reaction due to progressive regional metamorphism, but, rather, represents a retrograde metamorphic reaction during slow cooling at relatively constant P_{load}. They base their interpretation on field and petrographic observations associated with their work in the Morin anorthosite complex of southern Quebec, located ~120 km north of the Adirondack highlands portion of the Grenville Province. In their field area, the garnet-quartz-clinopyroxene assemblage is restricted virtually to norites, ferrogabbros, jotunites, and mangerites that surround the anorthosite mass. Martignole and Schrijver believe this areal restruction suggests the garnet-forming reaction is genetically linked to the anorthosite complex. In addition to de Waard's reaction (1), they propose reaction

$$\text{orthopyroxene + plagioclase} \rightleftharpoons \text{Ca-Fe-Mg garnet + quartz} \quad (2)$$

and imply (1971, p. 700) that reaction

$$\text{orthopyroxene + plagioclase + Fe oxide} \rightleftharpoons \text{garnet + quartz + clinopyroxene} \quad (3)$$

involving reduction of Fe^{3+} from left to right, also was active in the formation of garnet-quartz symplectites.

Rare occurrences of cordierite (Martignole and Schrijver, 1971, p. 701) are present at the immediate contact between the anorthosite complex and supracrustal rocks. Martignole and Schrijver believe this rare occurrence of cordierite precludes using an increase of P_{load}-T near the complex to explain garnet formation as de Waard does for the Adirondack highlands. Their alternative explanation for the association of garnet-quartz symplectites and the anorthosite complex is that the anorthosite completed solidification under high load pressure and retarded regional cooling. This retarded regional cooling permitted reactions (1), (2), and (3) to proceed slowly to the right as retrograde reactions in the "dry" environment of granulite-facies metamorphism. Thus, Martignole and Schrijver contend the highest grade of metamorphism in the Adirondack highlands is preserved as the hornblende-orthopyroxene-plagioclase subfacies, zone 2 in Figure 12. Zone 3 (Fig. 12) is considered by them representative of retrograde metamorphism associated with close spatial relationship to anorthosite complexes.

McLelland and Whitney (1976, 1977) studied the origin of garnet in the anorthosite-charnockite suite of rocks in the Adirondacks. Their analysis of textural and chemical relationships suggests that the onset of the hornblende-clinopyroxene-almandite subfacies of de Waard (1964a) is marked by the following reaction:

$$2\ CaAl_2Si_2O_8 + (6-\alpha)(Fe, Mg)SiO_3 + \alpha Fe\text{-oxide} +$$

$$(\alpha-2)SiO_2 \rightleftharpoons Ca(Fe,Mg)_5Al_4Si_6O_{24} + Ca(Fe,Mg)Si_2O_6 \qquad (4)$$

where α is a function of the distribution of Fe and Mg between the several coexisting ferromagnesian phases. Reaction (4) is a general garnet-forming reaction for saturated rocks. It differs from de Waard's reaction (1) in that (a) quartz is a reactant instead of a product and (b) Fe-oxide is a reactant, as it is for reaction (3) of Martignole and Schrijver. McLelland and Whitney (1977) consider reaction (1) to be a special situation of reaction (4) where there exists, in charnockitic gneiss, a relatively high Mg/(Mg + Fe) ratio. An interesting feature of their study is that most garnet-quartz symplectites are actually garnet-plagioclase symplectites on the basis of microprobe analysis.

P-T Conditions of Metamorphism: de Waard (1969) and Bohlen and Essene (1977) estimated P-T conditions of metamorphism for the Adirondack highlands.

Figure 13 is the petrogenetic grid used by de Waard (1969) in arriving at P_{load}-T conditions of ~ 7.8 kb and 770°C at the garnet-clinopyroxene isograd (see Figure 12). de Waard estimated maximum P_{load}-T conditions to be perhaps ~ 8.3 kb and 800°C to the east of the garnet-clinopyroxene isograd (see de Waard, 1969, for a fuller discussion).

Bohlen and Essene (1977) report that pressure estimates increase from 6 kb at Balmat (northwest Adirondacks, in zone 2 of Fig. 12) to 8 kb in the central Adirondack highlands. Temperature estimates are almost 800°C in the central highlands as determined by plagioclase-orthoclase and ilmenite-magnetite thermometers (see Bohlen and Essene, 1977, for a fuller discussion).

McLelland and Whitney (1977) have estimated equilibrium temperatures for one charnockite from the Adirondack highlands assuming a P_{load} of 7.5 kb. The temperatures range from 610°C by the method of Wood (1974) to 792°C by the method of Wood and Banno (1973). The temperature methods are based on the distribution of Mg and Fe between clinopyroxene, orthopyroxene, and garnet as functions of temperature and pressure.

Metagabbro and Metadiabase

Origin of corona structures: Whitney and McLelland (1973) studied the origin of corona structures in metagabbros of the Adirondack Mountains. In the southern Adirondacks, Area I, two types of coronas are observed: (1) olivine-pyroxene-spinel coronas and (2) oxide-hornblende coronas. In the central and eastern Adirondacks, Area II, two types are also observed: (1) olivine-pyroxene-garnet coronas and (2) oxide-amphibole-garnet coronas.

Whitney and McLelland (1973) propose three partial reactions took place in the formation of olivine-cored coronas in Area I:

$$\text{olivine} \rightleftharpoons \text{orthopyroxene} + (Mg,Fe)^{++} \qquad (a)$$

plagioclase + $(Mg,Fe)^{++}$ + $Ca^{++} \rightleftharpoons$ clinopyroxene +

spinel + Na^{+} (b)

plagioclase + $(Mg,Fe)^{++}$ + $Na^{+} \rightleftharpoons$ spinel + more

sodic plagioclase + Ca^{++} (c)

Reaction (a) occurs in the inner shell of the corona structure adjacent to olivine. Reaction (b) occurs in the outer shell and reaction (c) occurs in the surrounding plagioclase, giving rise to spinel clouding in plagioclase. Summed together these partial reactions are equivalent to:

olivine + anorthite $\rightleftharpoons$ aluminous orthopyroxene +

aluminous clinopyroxene + spinel (5)

Garnet develops in olivine-cored coronas of Area II by the following partial reactions proposed by Whitney and McLelland (1973):

orthopyroxene + $Ca^{++} \rightleftharpoons$ clinopyroxene + $(Mg,Fe)^{++}$ (d)

clinopyroxene + spinel + plagioclase + $(Mg,Fe)^{++} \rightleftharpoons$

garnet + Ca^{++} + Na^{+} (e)

plagioclase + $(Mg,Fe)^{++}$ + $Na^{+} \rightleftharpoons$ spinel + more

sodic plagioclase + Ca^{++} (f)

These partial reactions [(d)-(f)] involve the products of reactions (a)-(c) and (5). Balanced, and generalized to account for aluminous pyroxenes and variable An content of plagioclase, partial reactions (d)-(f) are equivalent to:

orthopyroxene + anorthite + spinel $\rightleftharpoons$ garnet (6)

Whitney and McLelland (1973) propose the following net reaction to account for oxide-cored coronas:

olivine + anorthite + albite + ilmenite + diopside +

$H_2O \rightleftharpoons$ hornblende + hypersthene + spinel (7)

The garnet shell observed in oxide-amphibole coronas of Area II is believed (Whitney and McLelland, 1973, p. 93) to have formed by a complex reaction consuming hornblende, spinel, and plagioclase, yielding garnet, clinopyroxene (as inclusions in garnet), and a bright red, titaniferous biotite.

P-T Conditions of Corona-Structure Formation: Whitney and McLelland (1973) also have investigated the P-T conditions of corona-structure formation. Reactions (5) and (6) have been studied experimentally by Kushiro and Yoder (1966) and Green and Ringwood (1967), respectively. Figure 14 is modified after Whitney and McLelland (1973, fig. 5, p. 95). They cite several reasons for exercising caution in applying experimental results to natural systems. With those reservations, Whitney and McLelland are able to give a general estimate of P_{load} and T of corona formation. Broken lines A and B in Figure 14 illustrate two possible metamorphic histories for corona-structure formation. For garnet-bearing rocks, both paths must pass through the pyroxene spinel field prior to entering the garnet field. Path A is the prograde-metamorphic path in which gabbro and diabase intruded at shallow depths prior to maximum P-T conditions of metamorphism. Path B is the retrograde metamorphic path in which gabbro intruded at depth and cooled at constant, or increasing, pressure. A path similar to path A is favored for metagabbro of Area I (but at lower pressure

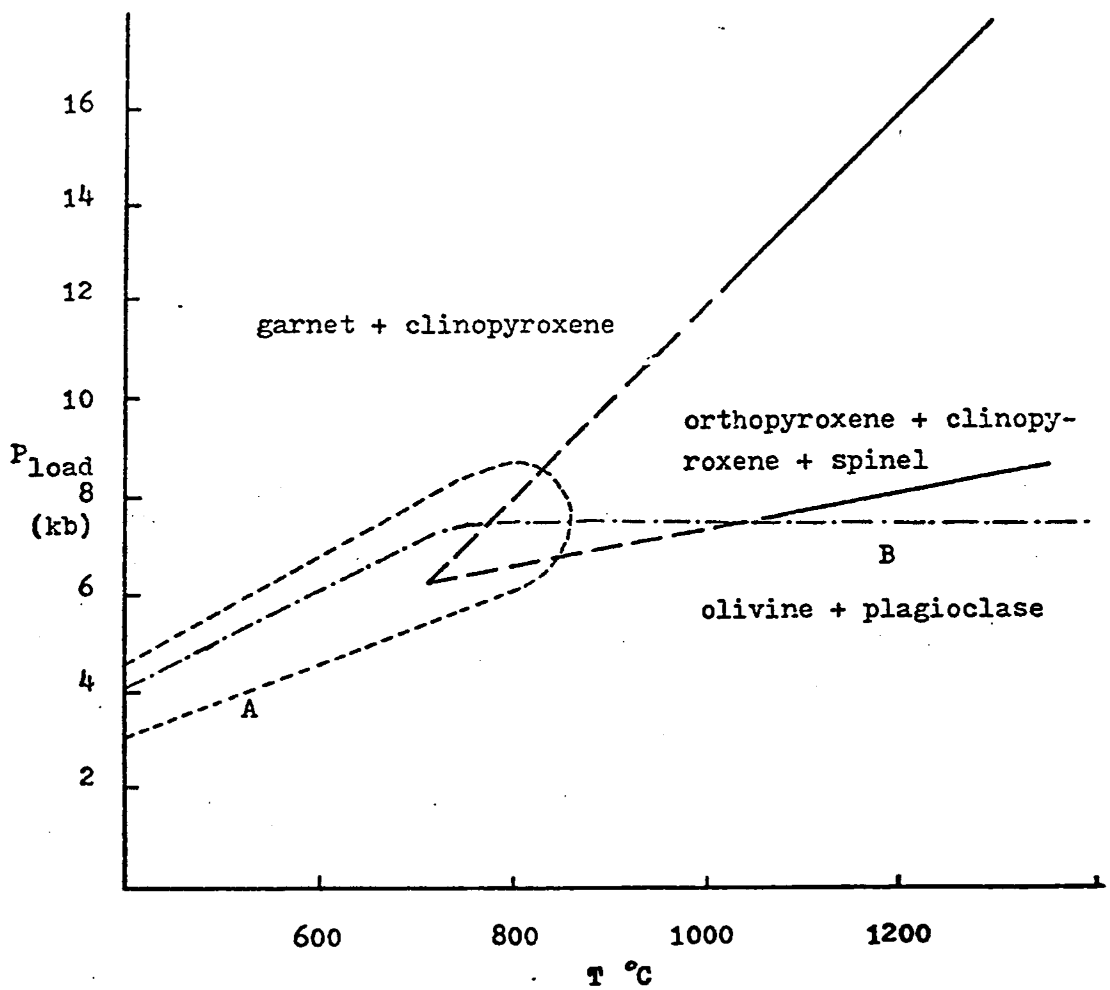

Figure 14. Stability fields of corona-structure mineral assemblages (modified after Whitney and McLelland, 1973, p. 95, fig. 5). Reaction boundaries (solid) are from Kushiro and Yoder (1966); dashed reaction boundaries are extrapolations of their work. A - path for prograde origin of garnet-bearing, olivine-cored coronas. B - path for retrograde origin of garnet-bearing, olivine-cored coronas.

because no garnet formed) whereas path B is favored for metagabbro of Area II, which is in close spatial association with the Marcy anorthosite complex. Regardless of the path followed, minimum pressure of ~8 kb and minimum temperature of ~800°C were necessary for formation of garnet-bearing coronas (see Whitney and McLelland, 1973, for a complete discussion).

Kyanite - Sillimanite-bearing Aluminous Gneiss

Description: Boone (1978) and Boone and others (in prep.) have determined mineral compositions in sillimanite-rich, quartz-feldspar gneiss at Ledge Mountain. The gneiss is situated in the core of a south-facing, recumbent antiform, and is structurally - if not also stratigraphically - the lowest unit exposed in the central Adirondack highlands (Geraghty, 1978). The gneiss consists predominantly of microcline perthite, plagioclase, quartz, sillimanite, biotite, magnetite, garnet, and minor hercynite. Lenses and alternating layers of sillimanite, magnetite, and quartz with minor garnet and hercynite, make up the remaining 20 to 30 percent of the gneiss in the central part of the mountain. Abundance of these lenses and layers decreases westward toward Route NY 28-30. Only two small patches of kyanite have been found; these occur as relatively coarse-grained, blue crystal aggregates in the feldspathic portions of the gneiss. Pegmatite lenses and discordant bodies abound.

P-T Conditions of Metamorphism: The following relationships are of interest: (1) kyanite-sillimanite; (b) biotite-magnetite-feldspar; (c) Fe/Mg distribution between biotite and garnet; and (d) Ca-contents of garnet and plagioclase. Almandine-hercynite-magnetite-quartz relationships are puzzling, and may not conform to other reaction relationships in the gneiss perhaps owing to low reaction rate. The preponderance of sillimanite effectively argues against the notion that the gneiss equilibrated on the kyanite-sillimanite univariant boundary (or divariant field in Al-Fe). Insofar as kyanite is present, however, the following enquiry was made: Taking into account the Fe^{3+}, F^- and Cl^- contents of biotite, the reaction

$$\text{Biotite } (0.4\,{}^{VI}Fe^{2+}) + \text{Quartz} \rightleftharpoons \text{K-rich feldspar} + \text{Sillimanite} + \text{Magnetite} + H_2O + H_2 + F + Cl, \quad (8)$$

was examined with reference to the redox equation of Czamanske and Wones (1973) across the temperature range of 650° - 800°C using a range of f_{O_2} compatible with the coexisting impure phases magnetite and hercynite (Turnock and Eugster, 1962). Values of calculated $P_E\,H_2O$ range from 120 bars at approximately 700°C to 600 bars at 770°C. These and volumetric data for the reaction abbreviated in equation (8) were applied to Greenwood's (1961) modification of Thompson's (1955) equation for the projected slope on P_s and T coordinates of a dehydration reaction boundary under steady-state outward diffusion conditions of H_2O with effective H_2O "pressure" less than total pressure. The resulting steep biotite dehydration boundaries are shown in Figure 15; inasmuch as they are nearly parallel to the pressure axis, the values of 695°C and 790°C may be taken as minimum and maximum for the temperature of granulite facies

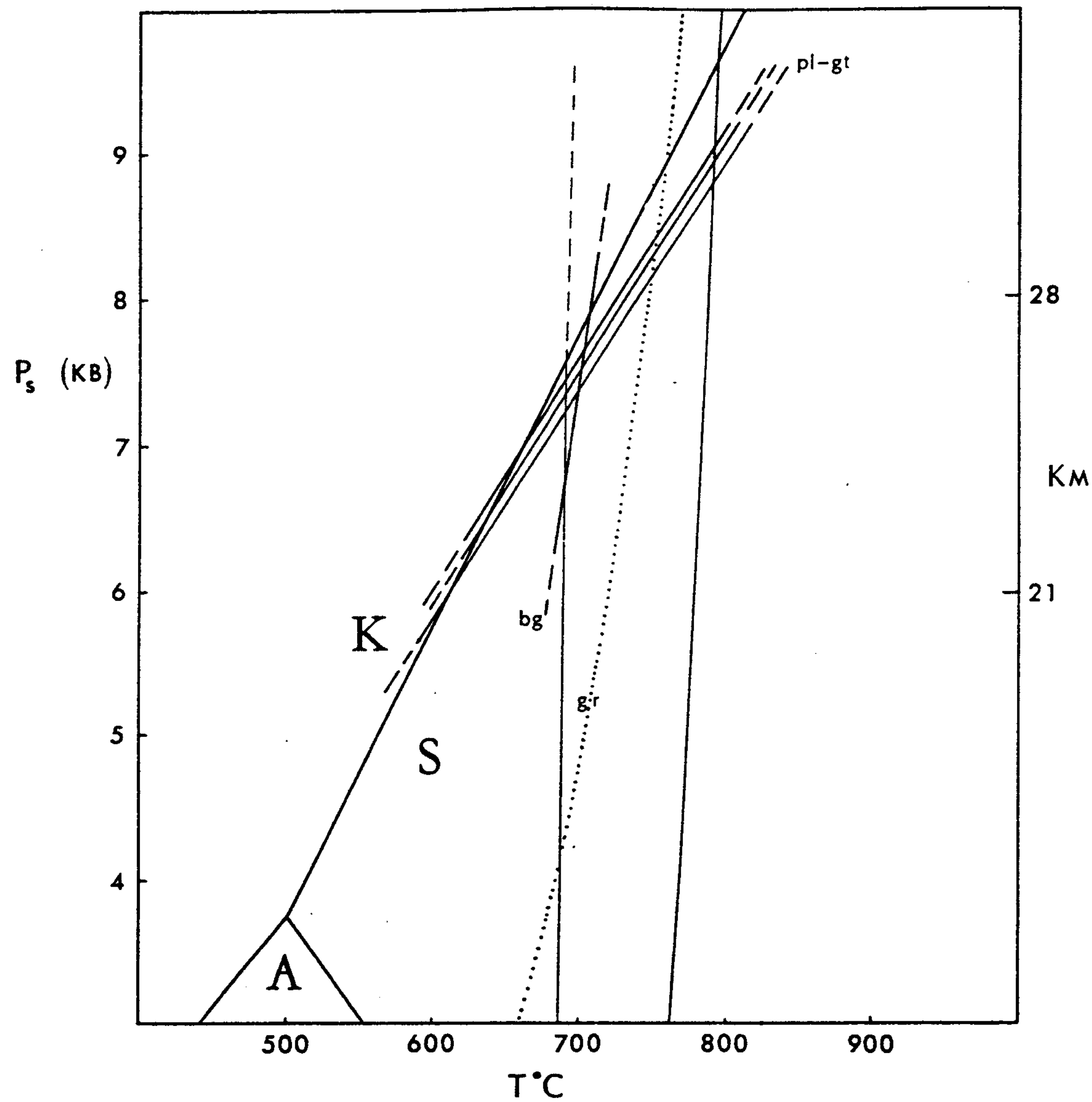

Figure 15. Petrogenic grid for Ledge Mountain aluminous gneiss showing biotite-K-rich feldspar-magnetite redox equilibria at P_EH_2O = 120 bars (left) and 600 bars (right) unlabelled boundaries. bg: biotite-garnet-Al_2SiO_5-Kf Fe/Mg equilibrium at P_EH_2O = 500 bars. pl-gt: Plagioclase-garnet equilibria. gr: muscovite granite solidus at 0.6 wgt. percent H_2O (P_EH_2O=~200 bars) from Huang and Wyllie (1973). Al_2SiO_5 phase boundaries after Holdaway (1971). Intersection of boundaries gr and pl-gt are interpreted as representing upper P_S-T limits for granulite facies metamorphism and anatexis of Ledge Mountain gneiss. Cf. text and road log (stop 2) for additional explanation.

metamorphism. Fe/Mg ratios for garnet, and biotite external to garnet porphyroblasts, when applied to Schmid and Wood's (1976) equation 11, give results shown by curve b-g (Fig. 15) for which P_{EH_2O} = 500 bars. The lack of agreement between curves b-g and P_{EH_2O} = 600 bars for equation (8) (they should be closer) probably is largely due to the lack of direct thermochemical data for the Mg end-member reaction: phlogopite + Sillimanite + quartz $\rightleftharpoons$ pyrope + Kfeldspar + H_2O.

Limiting values for total pressure were sought via the anhydrous mineral reaction relation involving plagioclase and garnet:

$$3\ \text{anorthite} \rightleftharpoons \text{Grossular} + 2\ Al_2SiO_5 + \text{Quartz}. \qquad (9)$$

Based on the estimation of mixing parameters for pairs of garnet end-members (Henson, Schmid, and Wood, 1975; Ganguly and Kennedy, 1974), grossular activity coefficients, (γ_{Gr}^{Gt}), across the above temperature range were taken between 1.23 and 1.37. Values of γ for anorthite in plagioclase were taken from Orville (1972). These and data for X_{An}^{Pl} and X_{Gr}^{Gt} were applied to the van't Hoff equation

$$-10{,}300 + 31.83T - 1.2746(P-1) = -RT\ln\left[\frac{(X_{Gr}^{Gt}\gamma)^3(0.99)^2}{(X_{An}^{Pl}\ \gamma)^3}\right] \qquad (10)$$

to obtain the reaction boundaries collectively labelled Pl-Gt summarized in Figure 15. It can be seen that within the temperature range of interest shown in Figure 15 that the plagioclase-garnet equilibria lie within the sillimanite field of stability. (One which does not is discussed in the trip log under Stop 2.) These intersect the biotite oxidation equilibrium boundaries at approximately 7.3 and 9 kb. Owing to the set of assumptions which lead to the calculation of the biotite equilibrium boundary P_{EH_2O} = 600 bars, the temperatures along this curve are thought to be too high, and therefore the value of P_s = 9 kb, also too high. Some confirmation of this view is that, with reference to the curve for the beginning of melting of aluminous granite (Huang and Wyllie, 1973), labelled gr on Figure 15, it is unlikely that temperatures much above 750°C were maintained during the metamorphism because much of the feldspathic portions of the Ledge Mountain gneiss is of granitic composition, and therefore ought to have been removed largely as anatectic granitic magma. This aspect of the problem presently is under field and analytical investigation by Ellen Metzger of Syracuse. For these reasons, the upper limit of load pressure is taken at approximately 8.2 kb (Table 1). Paths of P-T change are discussed under the heading of Stop 2.

Plagioclase-Scapolite Phase Relations

Phase relations in the systems plagioclase-calcite-halite-scapolite, high albite-halite-marialite, anorthite-calcite-meionite, and anorthite-anhydrite-sulfate meionite have been studied experimentally (Orville, 1975; Newton and Goldsmith, 1975, 1976; Goldsmith, 1976; Goldsmith and Newton,

Table 1. Summary of inferred and calculated P-T conditions of metamorphism for Adirondack highlands.

	T	P_{load}	Source
Charnockitic and granitic gneiss	>770°C<800°C	>7.8<8.3 kb	de Waard (1969)
	>700°C<750°C	~8.0 kb	Bohlen and Essene (1977)
Metagabbro	~800°C	~8.0 kb	Whitney and McLelland (1973)
Kyanite and sillimanite-bearing granulite	695°-770°C	>7.4<8.2 kb	Boone (1978)
Marble	650°-756°C		Geraghty (1978)

1977). Newton and Goldsmith (1976), in all instances, and Orville (1975), in most instances, observed that scapolite is stable in preference to plagioclase, calcite, and halite at high temperatures and pressures. This is in marked contrast to earlier discussions that gave the impression that scapolite is a metamorphic mineral resulting from retrogressive processes (see, e.g., Fyfe and Turner, 1958).

The assemblages plagioclase-calcite-scapolite and plagioclase-scapolite are observed in several thin sections of calc-silicate rock and marble from the mapped area (see Fig. 16). Compositions have been determined by microprobe. In addition, plagioclase compostions were determined optically, using the zone method of Rittman. The compositions of coexisting scapolite and plagioclase are presented graphically in Figure 17.

On the basis of analyzed compositions, it is believed the idealized reaction

$$\text{Albite + anorthite + calcite + halite} \rightleftharpoons \text{scapolite} \qquad (11)$$

took place in samples 116, 130, 215, and 289.

Direct textural evidence that reaction (11) took place is expressed in a thin section of sample 116 by the spatial association of reactants (except for halite) and product of (11). It is inferred that halite was present originally in small amount based on relatively low content of Cl in scapolite of sample 116. Textural evidence for reaction (11) is not as pronounced in other thin sections. Usually, reactants (except for halite) and product coexist in close spatial association without the development of reaction rims or corona structure. Calcite is absent in many samples, indicating that it could have been consumed in reaction (11).

Microprobe analyses were not made for all mineral phases in thin sections containing assemblages plagioclase-calcite-scapolite or plagioclase-scapolite. Thus, it is not possible to analyze in detail whether chemical equilibrium was attained in these rocks. However, it is possible to

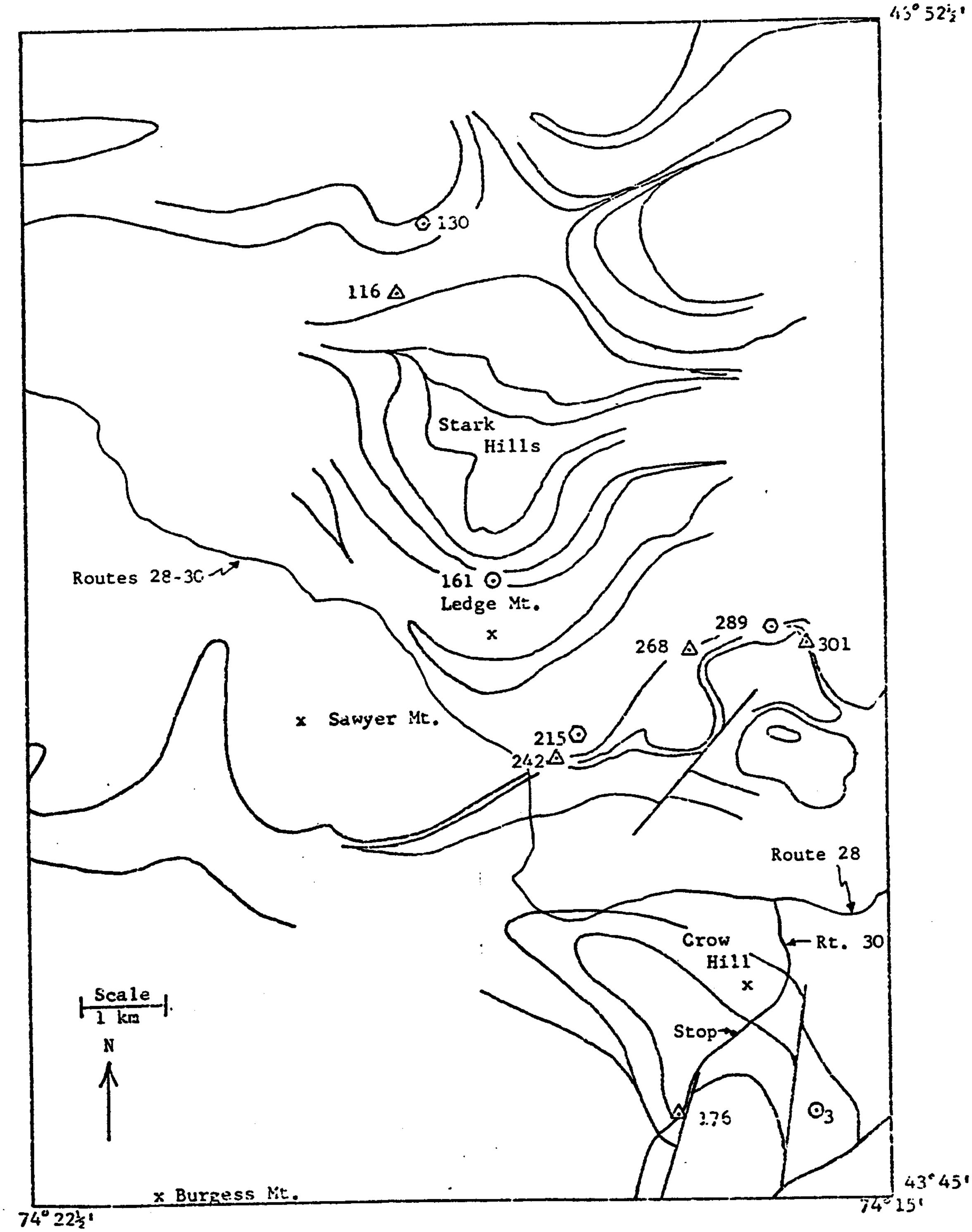

Figure 16. Location of samples used in discussion of plagioclase-scapolite phase relations (Δ - assemblage sc-ph-ca, ⬡ - assemblage (sc-pf) and in calcite-dolomite geothermometry ⊙). Map is SE ¼ of Blue Mountain 15' quadrangle; contacts between major rock units are shown for reference.

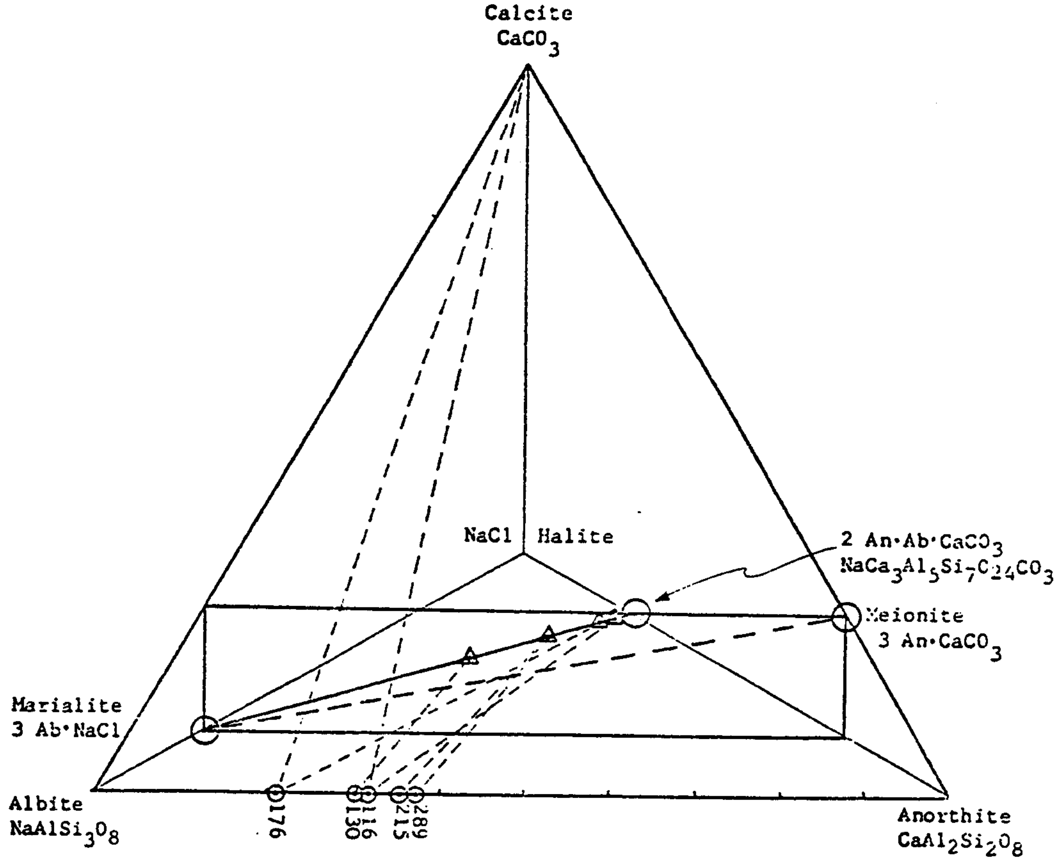

Figure 17. Composition tetrahedron albite-anorthite-halite-calcite. Scapolite composition plane 3 (albite + anorthite) : 1 (halite + calcite) is shown with solid line representing scapolite solid solution suggested by Evans and others (1969); dashed line represents previous stoichiometry. Plagioclase and scapolite compositions for analyzed samples are plotted.

Sample #	Me %	An %
116	71.6	30.9
130	45.4	29.7
176	69.0	21.0
215	60.0	35.0
189	60.0	36.7

investigate if the pairs plagioclase-scapolite were in equilibrium during metamorphism. This is attempted by examining the distribution of Na, Ca, and Al among coexisting plagioclase and scapolite from samples 116, 130, 215, and 289 (see Fig. 18). Unfortunately, only four distribution points are plotted and the clustering of points does not allow a distribution curve to be drawn. Equilibrium is suggested if the distribution curve is a straight line or smooth curve as defined by the distribution points. However, excluding data from sample 116 and using data from 130, 215, and 289, a straight line passing through or near these three samples and through the origin could be constructed for all three distribution diagrams.

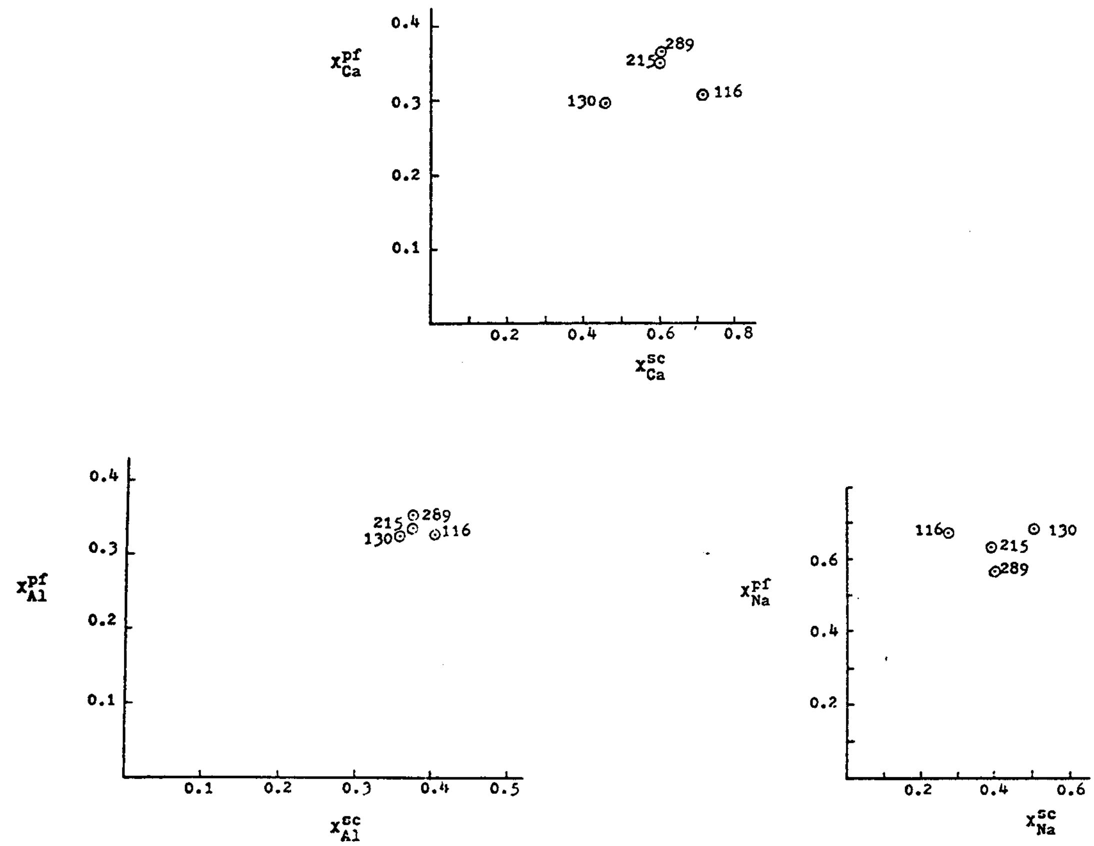

Figure 18. Distribution of calcium, sodium, and aluminum between plagioclase and scapolite.

This meager evidence argues for equilibrium between plagioclase and scapolite in these samples. The data points for sample 116, a marble, lie off the hypothetical distribution curves for data from samples 130, 215, and 289, calc-silicate rocks. The difference between sample 116 and samples 130, 215, and 289 also is expressed in Figure 17. Crossing tie lines are exhibited between two groups of samples: (1) samples 130, 215, and 289 form one group that exhibit nearly parallel tie lines between coexisting scapolite and plagioclase, (2) samples 116 and 176 exhibit tie lines between coexisting scapolite and plagioclase that cross tie lines of group (1). Group (2) samples contain calcite (see Fig. 17) and the scapolites exhibit relatively low contents of chlorine. One possible explanation for these relations between samples 116 and 176 and samples 130, 215, and 289 is that reaction (11) proceeded to the left upon falling temperature following the thermal peak of metamorphism in samples 116 and 176.

No estimate of metamorphic temperature and pressure can be made from coexisting scapolite and plagioclase, with or without calcite, of the mapped area. Newton and Goldsmith (1976) have determined experimentally the stability relations of anorthite, calcite, and meionite. However, their data can be used confidently to estimate metamorphic temperature only where the plagioclase composition is $>An_{70}$ (Goldsmith and Newton, 1977).

Based on the experimental work of Orville (1975, p. 1104), the presence of relatively sodic plagioclase (An_{21}-An_{37}) with relatively calcic scapolite (Me_{45}-Me_{72}) argues for a higher activity of $CaCO_3$ compared to NaCl in scapolite and plagioclase-bearing rocks of the mapped area.

Calcite-Dolomite Geothermometry

The amount of $MgCO_3$ in solid solution with calcite coexisting with a separate dolomite phase can be used to estimate temperature of metamorphism (Goldsmith and Newton, 1969). Graf and Goldsmith (1955, fig. 4) showed that the higher the temperature, the greater the amount of $MgCO_3$ that can be accommodated in the calcite structure. In order to use this geothermometry effectively, CO_2 pressure must have been high enough to prevent decomposition of dolomite. If noncarbonate, Mg-containing phases also are present under equilibrium conditions, they will have no effect on the Mg content of the calcite as long as dolomite is present (Goldsmith and others, 1955).

Only two of 22 thin sections of marble examined from the mapped area contain discrete grains of dolomite coexisting with calcite (see Fig. 16, samples 3 and 161). Temperature estimates for samples 3 and 161 are 650°C and 756°C, respectively. The estimated temperature recorded from sample 161 compares favorably with temperature estimates made by other methods (see Table 1).

CONCLUDING SPECULATIONS

The ultimate origin of the structural and petrologic features of the Adirondacks remains obscure. A possible clue to the mechanisms involved is Katz's (1955) determination of 36 km as the present depth to the M-discontinuity beneath the Adirondacks. Because geothermometry-geobarometry place the peak of the Grenville metamorphism at 8-9 kb (24-36 km), a double continental thickness is suggested. Such thicknesses presently exist in two types of sites, both plate-tectonic related. The first is beneath the Andes and seems related to magmatic underplating of the South American plate (James, 1971). The second is beneath the Himalayas and Tibet and is due to thickening in response to collision (Dewey and Burke, 1973) or continental underthrusting (Powell and Conaghan, 1973).

Because of the wide extent of the Grenville metamorphic belt, we prefer the Dewey-Burke model of crustal thickening in response to a continent-continent collision accompanied by reactivation of basement rocks. Mobilization of the lower crust could lead to the upward displacement of large, recumbent folds in a manner similar to some of Ramberg's (1967) scaled centrifuge experiments. This model is shown diagramatically in Figure 19.

Although it seems that the tectonic style and framework of the Adirondacks are explained satisfactorily by the Tibetan model, there are no good candidates for even a cryptic Indus-type suture in the area or within the Grenville Province itself. Dewey and Burke (1973) suggest that the collisional suture is most likely buried beneath the folded Appalachians. The

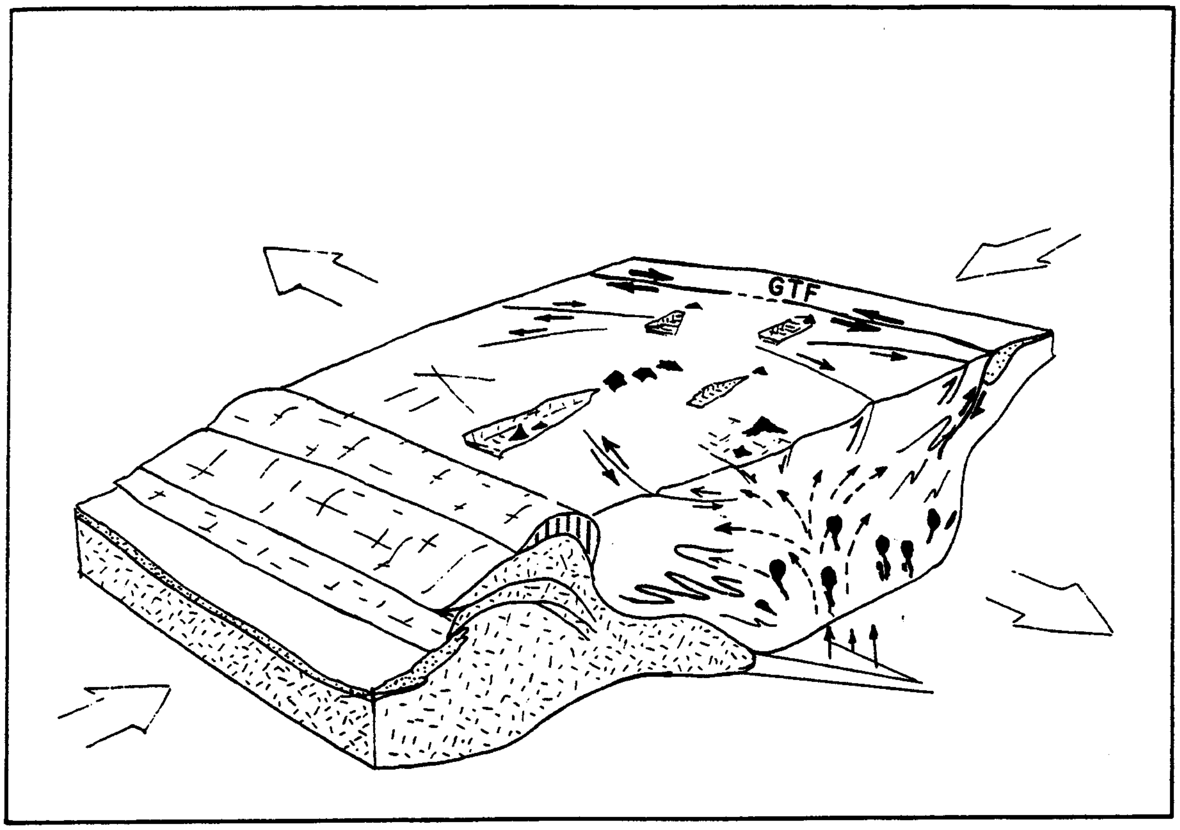

Figure 19. Himalayan type collision and associated tectonic elements.
GTF - Grenville Type Front.

Grenville Front itself cannot be a suture, and, as shown by Baer (1977), it has a large component of right lateral motion associated with it. We suggest that the Grenville Front is analogous to features such as the Altyn Tagh Fault in northern Tibet (Molnar and Tapponier, 1975), and, similar to the Altyn Tagh, accommodates the sideways displacement of large crustal blocks by strike-slip motions (Figs. 19, 20). In places the Altyn Tagh Fault lies some 1000 km distant from the Indus Suture. A similar distance measured southeast from the Grenville Front would place the corresponding suture beneath the Appalachians. Perhaps it is this buried suture that gives rise to the New York Alabama aeromagnetic lineament of Zietz and King (1977).

ACKNOWLEDGMENTS

We wish to acknowledge the editorial assistance of Daniel F. Merriam and we are grateful to Janice Potak for typing the final copy.

DAY 1

FRIDAY, MAY 5

GLOVERSVILLE - LAKE PLACID

STOPS 1 - 22

SOUTHERN AND CENTRAL ADIRONDACKS

CANADA LAKE SYNCLINE

PISECO ANTICLINE

SNOWY MOUNTAIN DOME

BARTON GARNET MINE

MARCY ANORTHOSITE

Canada Lake Area

ROAD LOG
(See fig. 18 for stop locations)

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
	0	Junction of Willie Road, Peck Hill Road, and NY Rt. 29A
1.3	1.3	Mud Lake to northeast of NY Rt. 29A
1.8	1.5	Peck Lake to Northeast of NY Rt. 29A
3.6	1.8	STOP 1. Peck Lake Fm.

1

STOP 1.

This exposure along Rt. 29A just north of Peck Lake is the type locality of the sillimanite-garnet-biotite-quartz-feldspar gneisses (kinzigites) of the Peck Lake Fm. in addition, there are exposed excellent minor folds of several generations. Note that the F_1 folds rotate an earlier foliation. The white quartzo-feldspathic layers in the kinzigites consist of quartz, two feldspars,and garnet and ar believed to be anatectic and have been folded by F_1 indicating pre-F_1 metamorphic events. Typical whole rock compositions are shown below. Spinel has been found enclosed in garnets at this outcrop. The similarity of the Peck Lake Fm. to the Major Paragneiss of the Lowlands suggests that the Adirondacks were contiguous at the time of deposition of the rocks.

Table 4.

Compositions of Representative Leucosome and Host Rock — Selected Clastics

	Leucosome		Host Rock		Average Greywacke[a] ($\Sigma = 23$)	Average PC Slate[b] ($\Sigma = 33$)	Average Slate[c] ($\Sigma = 36$)
	LL1	9-17-2A	10-29-1B	9-11-4B			
SiO_2	75.61	74.60	68.04	64.24	64.70	56.30	60.64
Al_2O_3	13.75	13.49	13.93	16.16	14.80	17.24	17.32
TiO_2	.02	.19	.86	.90	.50	.77	.73
Fe_2O_3	.51	1.47	6.08	7.44	4.10	7.22	4.81
MgO	.11	.54	1.45	1.57	2.20	2.54	2.60
CaO	.36	1.64	1.65	3.41	3.10	1.00	1.20
Na_2O	2.19	3.25	2.84	3.20	3.10	1.23	1.20
K_2O	6.82	4.69	3.27	2.92	1.90	3.79	3.69
MnO	.02	.04	.06	.09	.10	.10	...
P_2O_5	.09	.08	.18	.17	.20	.14	...
LOI	.31	.25	.66	.66	2.40	3.70	4.10
TOTAL	99.78	100.24	99.80	99.76	101.00	98.70	98.00

6.1	2.5	Junction NY Rt. 29A and NY Rt. 10
8.0	1.9	Nick Stoner's Inn on west side of NY Rt. 29A-10
8.6	.6	STOP 2. Irving Pond Fm., .5 mile north of Nick Stoner's Inn, Canada Lake. Very near hinge line of F_1 Canada Lake isocline.

2

STOP 2.

The outer portion of the Irving Pond Fm. is exposed in low cuts along the east side of Rt. 29A just prior to the crest in the road heading north.

At the southern end of the cut typical, massive quartzites of the Irving Pond are seen. Proceeding north the quartzites become "dirtier" until they develop sillimanite-garnet-biotite-feldspar (kinzigites) layers along with quartzite.

McClelland, NYSGA 1991

At the northern end of the cut, and approximately on the Irving Pond/Canada Lake Fm. contact there occurs an excellent set of F_1 minor folds. Polished slabs and thin sections demonstrate that these fold an earlier foliation defined by biotite flakes and flattened quartz grains.

At the southern end of the outcrop dark, fine grained metadiabase sheets crosscut the quartzite. Near the telephone pole erosional remnants of diabase appear to truncate approximately horizontal foliation in the quartzite suggesting that the diabase was emplaced after an early metamorphism. At the north end of the cut a diabase sheet of variable thickness is folded in the F_1 fold. The folding is interpreted as Ottawan, the diabase as AMCG in origin, and the early foliation as Elzevirian. This is consistent with the presence of quartzite xenoliths in the ca. 1300 Ma tonalites.

The Irving Pond Fm. is the uppermost unit in the lithotectonic sequence of the southern Adirondacks. Its present thickness is close to 1000 meters, and it is exposed across strike for approximately 4000 meters. Throughout this section massive quartzites dominate.

8.8	.2	STOP 3. Canada Lake Charnockite (>1233 Ma, table 1, sample AM-87-13. Now fixed at 1251±43)

3 STOP 3.

Large roadcuts expose the type section of the Canada Lake charnockite. Lithogically the charnockite consists of 20-30% quartz, 40-50% mesoperthite, 20-30% oligoclase, and 5-10% mafics. The occurrence of orthopyroxene is sporadic. These exposures exhibit the olive-drab coloration that is typical of charnockites. Note the strong foliation in the rock. Farther north along the highway there are exposed pink leucogranitic variants of this unit. The chemical composition of these is given in table 3 (ab-6). The whole rock chemistry of the charnockitic phase is similar to AM-86-17 in table 3. The lateral continuity of the Canada Lake is striking (fig. 2) but the presence of xenoliths reveals an intrusive origin.

10	1.2	STOP 4. Royal Mt. Tonalite (>1301 Ma, table 1, sample AM-86-12, now fixed at 1307±2 Ma).

4 STOP 4.

Steep roadcuts, exposed across from the Canada Lake Store, expose typical examples of the early tonalitic rocks that occur within the southern and eastern Adirondacks and that manifest the presence, throughout the region, of collisional magmatic arcs of calcalkaline chemistry that existed along the eastern margin of Laurentia from ca. 1400-1200 Ma. Amalgamation of these arcs culminated in the Elzevirian Orogeny at ca. 1250-1220 Ma.

The whole rock chemistry of the tonalitic rocks is given in table 3 and important chemical trends are portrayed in figures 8, 9, and 10. Figure 7 shows the ϵ_{Nd} characteristics of these rocks and emphasizes their petrologically juvenile character, i.e., they are not derived from any crustal rocks with long-term crustal residence but are essentially mantle derived (including derivation from melting of basaltic rock derived from the mantle at ca. 1300-1400 Ma). The ϵ_{Nd} characteristics are compared with those from Lowland tonalites and granitoids of similar age, and the similarity suggests that they are essentially the same, strongly suggesting contiguity across the entire Adirondacks at that time (~1300 Ma).

A disrupted layer of amphibolitic material runs down the outcrop to road level at the east end of the outcrop. This, and other mafic sheets in the outcrop, are interpreted as dikes an sheets coeval with the tonalite. In the eastern Adirondacks it has been possible to document mutually crosscutting relationships between these rock types. Also documented there are xenoliths of kinzigitic rock in the tonalites. Within the southern Adirondacks xenoliths of quartzite similar to the Irving Pond Fm. have been found in the tonalite.

11.8	1.8	Pine Lake, Junction NY Rt. 29A and NY Rt. 10. Proceed north on NY Rt. 10.
17.5	5.7	STOP 5. Rooster Hill megacrystic gneiss at the north end of Stoner Lake (1156±8 Ma, table 1, sample AM-86-17).

5

STOP 5.

This distinctive unit belongs to the AMCG suite and is widespread in the southern Adirondacks. Here the unit consists of a monotonous series of unlayered to poorly layered gneisses characterized by large (1-4") megacrysts of perthite and microcline perthite. For the most part these megacrysts have been flattened in the plane of foliation, however, a few megacrysts are situated at high angles to the foliation and show tails. The groundmass consists of quartz, oligoclase, biotite, hornblende, garnet, and occasional orthopyroxene. An igneous rock analogue would be monzonite to quartz-monzonite (see table 3 for chemical composition) and the presence of orthopyroxene makes the rock mangeritic to charnockitic.

The contacts of the Rooster Hill megacrystic gneiss are everywhere conformable, but the presence of xenoliths of kinzigite indicate its intrusive nature. Rocks such as the Rooster Hill are interpreted as derived from melting of ca. 1300 Ma tonalitic and lower crustal granitoid rocks with heat derived from large AMCG gabbroic intrusions that would ultimately differentiate to anorthosite. This suggestion is consistent with the ϵ_{Nd} trends of AMCG and tonalitic rocks in figure 7a and with the REE distributions shown in figure 19, where it appears that melting of tonalite so as to leave a plagioclase-rich residue can give the AMCG REE-trends.

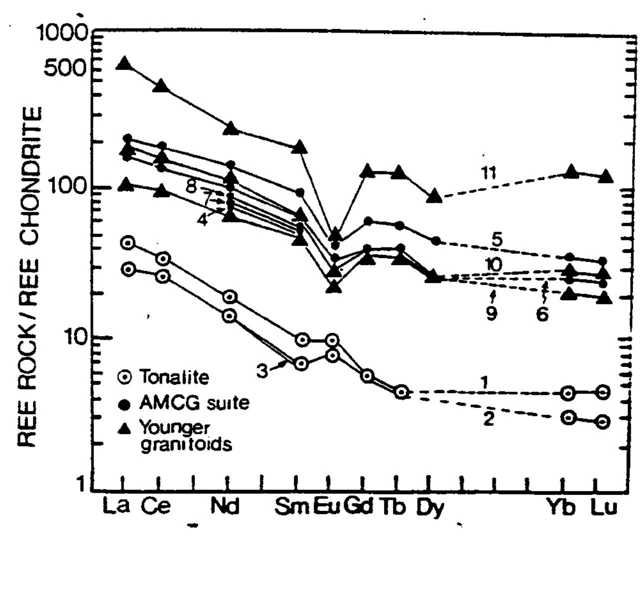

Fig. 19. Chondrite normalized REE concentrations for the Adirondack highlands. Numbers refer to samples in table 1 of Daly and McLelland (1991).

20.0	2.5	Low roadcut in kinzigites.
21.4	1.4	Avery's Hotel on west side of NY Rt. 10
22.5	1.1	Long roadcuts of pink quartzofeldspathic gneisses and metasediments of intruded metagabbro and anorthosite metagabbro. The igneous rocks are believed to belong to the AMCG suite.
23.6	1.1	Roadcut of anorthositic metagabbro and metanorite of AMCG suite.
23.9	.3	Roadcut on west side of highway shows excellent examples of anorthositic gabbros intrusive into layered pink and light green quartzofeldspathic gneisses.

24.0 5-A **Stop #5**. Lake Durant and Sacandaga Fms. intruded by anorthositic gabbros and gabbroic anorthosites.

These roadcuts are located on Rt. NY 10 just south of Shaker Place.

The northernmost roadcut consists of a variety of metasedimentary rocks. These lie directly above the Piseco anticline and are believed to be stratigraphically equivalent to the Sacandaga Formation. The outcrop displays at least two phases of folding and their related fabric elements. These are believed to be F_2 and F_3. A pre-F_2 foliation is thought to be present. Both axial plane foliations are well developed here. Several examples of folded F_2 closures are present and F_2 foliations (parallel to layering) can be seen being folded about upright F_3 axial planes.

Farther to the south, and overlooking a bend in the west branch of the Sacandaga River, there occurs a long roadcut consisting principally of pink and light green quartz-perthite gneiss belonging to the Lake Durant Fm. About half-way down this roadcut there occurs a large boudin of actinolitic and diopsidic gneiss. To the north of the boudin the quartzo-feldspathic gneisses are intruded pervasively by anorthositic gabbros, gabbroic anorthosites, and various other related igneous varieties. At the north end of the cut and prior to the meta-stratified sequences these intrusives can be seen folded by upright fold axes. They are crosscut by quartzo-feldspathic material.

Within this general region the Lake Durant Fm. and other quartzo-feldspathic gneisses seem to have undergone substantial anatexis. This is suggested by the "nebular" aspect of the rocks. Good examples of this are seen in the manner in which green and pink portions of the quartzo-feldspathic gneisses mix. Note also the clearly cross-cutting relationships between quartzofeldspathic gneiss and mafic layers at the south end of the roadcut. Here it seems that mobilized Lake Durant is cross-cutting its own internal stratigraphy. Also note that the quantity of pegmatitic material is greater than usual. This increase in anatectic phenomena correlates closely with the appearance of extensive metagabbroic and metanorthositic rocks in this area. It is believed that these provided a substantial portion of the heat that resulted in partial fusion of the quartzo-feldspathic country rock.

24.3	.3	Roadcuts of quartzites and other metasediments of the Sacandaga Fm. Mezger (1990) obtained a U-Pb garnet age of ca. 1154 from these rocks.
31.0	5.7	Red-stained AMCG quartzofeldspathic gneisses that have been faulted along NNE fractures.
31.5	.5	Junction of NY Rt. 10 and NY. Rt. 8. End Rt. 10. Turn east on NY Rt. 8.
32.0	.5	STOP 6. Core rocks of the Piseco anticline (1150±5 Ma, table 1, sample AM-86-9).

6

STOP 6.

This stop lies along the hinge line of the F_2 Piseco anticline near its domical culmination at Piseco Lake. The rocks here are typical ofthe granitic facies of quartzofeldspathic gneisses such as occur in the Piseco anticline and in other large anticlinal structures, for example Snowy Mt. dome, Oregon dome.

The pink "granitic" gneisses of the Piseco anticline do not exhibit marked lithologic variation. Locally grain size is variable and in places megacrysts seem to have been largely grain size reduced and only a few small remnants of cores are seen. The open folds at this locality are minor folds of the F_2 event. Their axes trend N70W and plunge 10-15° SE parallel to the axis of the Piseco anticline.

The most striking aspect of the gneisses in the Piseco anticline is their well-developed lineation. This is expressed by rod, or pencil-like, structures which are clearly the result of ductile extension of quartz and feldspar grains in a granitic protolith. The high temperature, grain size reduction that has occurred results in a mylonite. Where recognizable, early F_1 isoclinal fold axes parallel the lineation.

These rocks are similar in age and chemistry to other AMCG granites and are considered to be part of that suite.

Smooth outcrops of Piseco Core rocks showing exceptionally strong mylonitic ribbon lineations.

43.5	11.5	Junction NY Rt. 8 and NY Rt. 30 in Speculator. Head southeast on NY Rt. 8-30.
47	3.5	STOP 7. Northern intersection of old Rt. NY 30 and new Rt. NY 30, 3.3 miles east of Speculator, New York.

7

STOP 7.

Typical Adirondack marble is exposed in roadcuts on both sides of the highway. These exposures show examples of the extreme ductility of the carbonate-rich units. The south wall of the roadcut is particularly striking, for here relatively brittle layers of garnetiferous amphibolite have been intensely boudinaged and broken. The marbles, on the other hand, have yielded plastically and flowed extensively during the deformation. As a result, the marble-amphibolite and marble-charnockite relationships are similar to those that would be expected between magma and country rock. Numerous rotated, angular blocks of amphibolite and charnockite are scattered throughout the marble in the fashion of xenoliths in igneous intrusions. At the eastern end of the outcrop tight isoclinal folds of amphibolite and metapelitic gneisses have been broken apart and rotated. The isolated fold noses that remain "floating" in the marble have been aptly termed "tectonic fish". The early, isoclinal folds rotate on earlier foliation. The garnetiferous amphibolites have typical igneous compositions and are interpreted as flows or sills.

Near the west end of the outcrop a boudin of charnockite is well exposed. McLelland and others (1987) have presented evidence that boudin represents a local example of charnockitization by carbonic metamorphism. However, granites of similar composition outside the marble do not develop orthopyroxene, demonstrating the local nature of the process and the limited permeation of the fluid phase.

Exposed at several places in the roadcut are crosscutting veins of tourmaline and quartz displaying a symplectic type of intergrowth. Other veins include hornblende- and sphene-bearing pegmatites.

Almost certainly these marbles are of inorganic origin. No calcium carbonate secreting organisms appear to have existed during the time in which these carbonates were deposited (>1200 Ma ago). Presumably the graphite represents remains of stromatolite-like binding algae that operated in shallow water, intertidal zones. This is consistent with the presence of evaporitic minerals, such as gypsum, in Lowland marbles.

At the eastern end of the outcrop coarse diopside and tremolite are developed in almost monomineralic layers. Valley et al. (1983) showed that the breakdown of almost Mg-pure tremolite to enstatie, diopside, and quartz in these rocks requires low water activity at the regional P,T conditions. Similarly, the local presence of wollastine requires lowering of CO_2 activity, presumably by H_2O. These contrasts demonstrate the highly variable composition of the fluid phase and are consistent with a channelized fluid phase within a largely fluid-absent region.

47.5	.5	Extensive roadcuts in lower part of marble. Quartzites, kinzigites, and leucogneisses dominate. Minor marble and calcsilicate rock is present.
47.9	2.5	Large roadcuts in well-layered, pink quartzofeldspathic gneisses with subordinate amphibolite and calcsilicate rock. The layering here is believed to be tectonic in origin, and the granitic layers represent an intensely deformed granite. The calcsilicate layers may be deformed xenoliths.

49.0	1.1	STOP 8. One half mile south of southern intersection of old Rt. 30 and with new Rt. 30. Anorthositic rocks on the SW margin of the Oregon Dome.

8

STOP 8.

On the west side of the highway a small roadcut exposes typical Adirondack anorthosite and related phases.

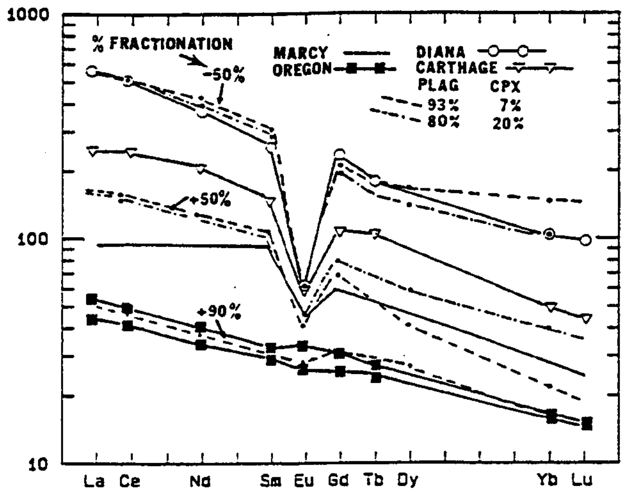

Fig. 20. Chondrite normalized REE concentrations for several Adirondack ferrogabbroic occurrences. Percentage fractionation of plagioclase and clinopyroxene are shown for a starting composition given by Carthage ferrogabbro (trianges). The Diana occurrence corresponds to sheets of breccia-bearing mafic material referred to by Buddington (1939) as shonkinite. The breccia consists of K-feldspar fragments from the host pyroxene syenite of the Diana complex.

The glacially smoothed upper surface of the roadcut reveals the presence of three major igneous phases: 1) a dark, pyroxene-rich dike that crosscuts the anorthosite, contains anorthosite xenoliths, and contains a large irregular, disrupted mass of sulfidic calcsilicate; 2) a coarse grained, Marcy-type anorthosite facies with andesine crystals 6-8" across; and 3) a fine grained anorthositic phase. Some of the coarse grained facies has been crushed and these portions bear some resemblance to the finer grained phase (note, for example, those places where fractures cross large andesine grains and produce finer grained material). However, close inspection of the finer grained material reveals the presence of ophitic texture with pyroxenes of approximately the same size as the plagioclase, and this texture and association are much better explained as igneous in origin. Therefore, the texture of the fine grained phase is interpreted as igneous in origin and may be due to chilling near the contact of the Oregon Dome massif. By contrast, large (3-4 cm) rafts of coarse grained, ophitic gabbroic anorthosite seem to be "flat" in the fine grained phase. Analyses of typical anorthositic rocks are shown in Table 5.

The pyroxene-rich ferrogabbro dike shows "soft" contacts with the anorthosite and is interpreted as essentially coeval. Zircons from it give a minimum age of 1087 Ma and, by comparison with other Adirondack anorthosites, its emplacement age is set at ca. 1135 Ma. The composition of the ferrogabbro is shown in table 5 where it is seen to be rich in TiO_2 and P_2O_5. Similar rocks occur together with other Adirondack anorthosites and are interpreted as late, Fe-enriched differentiates of a Fenner-type fractionation trend (see fig. 14). It is suggested that further differentiation within these rocks can result in liquid immiscibility and the production of magnetite-ilmenite liquids.

Table 5.

	Jotunite	Anorthositic Gabbro Green Mt.	Anorthosite Green Mt.	Anorthosite Owls Head	Mangerite Tupper Lake	Keene Gn. Hulls Falls	Marcy-type Anorthosite[a]	Whiteface Anorthosite[a]
No. of Samples	1	1	1	1	3	2	4	7
SiO_2	47.16	55.88	56.89	53.65	62.12	51.63	54.54	53.54
TiO_2	2.20	1.6	0.47	0.52	0.87	3.1	0.67	0.72
Al_2O_3	17.23	18.23	23.82	24.96	16.48	14.23	25.61	22.50
Fe_2O_3	2.75	2.4	1.21	0.41	1.49	2.1	1.00	1.26
FeO	9.24	6.57	1.3	0.70	3.96	13.5	1.26	4.14
MnO	0.15	0.09	0.02	0.02	0.09	0.16	0.02	0.07
MgO	2.71	2.08	0.65	1.45	1.06	2.63	1.03	2.21
CaO	9.04	4.87	8.19	12.21	3.27	6.5	9.92	10.12
Na_2O	6.61	4.26	5.38	3.92	4.81	2.67	4.53	3.70
K_2O	2.27	2.76	1.13	1.20	5.13	2.41	1.01	1.19
P_2O_5	0.59	0.48	0.09	0.09	0.30	0.57	0.09	0.13
H2O	0	0.09	0.42	0.04	0.32	0.07	0.55	0.12
Total	99.70	99.94	99.57	99.17	99.90	99.57	100.17	100.00

The upper, weathered surface of the outcrop affords the best vantage point for studying the textures and mineralogy of the anorthositic rocks. In several places there can be seen excellent examples of garnet coronas of the type that are common throughout Adirondack anorthosites. These coronas are characterized by garnet rims developed around iron-titanium oxides and pyroxenes. Recently McLelland and Whitney (1977) have succeeded in describing the development of these coronas according to the following generalized reaction:

Orthopyroxene + Plagioclase + Fe-bearing oxide + Quartz = Garnet + Clinopyroxene.

This reaction is similar to one proposed by de Waard (1965) but includes Fe-oxide and quartz as necessary reactant phases. The products are typomorphic of the garnet-clinopyroxene subfacies of the granulite facies (de Waard 1965). The application of various geothermometers to the phases present suggests that the P,T conditions of metamorphism were approximately 8 kb and 700±50°C respectively.

51.0	2.0	Minor marble, amphibolite, and calcsilicate rock. Predominantly very light colored sillimanite-garnet-quartz-feldspar leucogneisses interpreted as minimum-melt granitic due to anatexis of kinzigite near Oregon dome anorthosite. Enclaves of spinel- and sillimanite-bearing metapelite are present.
52.0	1.0	Junction to NY Rt. 8 and NY Rt. 30. Continue south on NY Rt. 30. To the west of the intersection are roadcuts in garnetiferous metasediment. A large NNE normal fault passes through here and fault breccias may be found in the roadcut and the woods beyond.
52.5	.5	Entering granitic-charnockitic gneiss on northern limb of the Glens Falls syncline. Note that dips of foliation are to the south.
54.8	2.3	Entering town of Wells which is situated on a downdropped block of lower Paleozoic sediments. The minimum displacement along the NNE border faults has been

		determined to be at least 1000 meters.
58.3	3.5	Silver bells ski area to the east. The slopes of the ski hill are underlain by coarse anorthositic gabbro that continues to the west and forms the large sheet just south of Speculator.
60.3	2.0	Entrance to Sacandaga public campsite. On the north side of NY Rt. 30 are quartzo-feldspathic gneisses and calcsilicates. An F_1 recumbent fold trends sub-parallel to the outcrop and along its hinge line dips become vertical.
60.8	.5	Gabbro and anorthositic gabbro.
62.0	1.2	STOP 9. Pumpkin Hollow.

9

STOP 9.

Large roadcuts on the east side of Rt. 30 expose excellent samples of the Sacandaga Fm. At the northern end of the outcrop typical two pyroxene-plagioclase granulites can be seen. The central part of the outcrop contains good light-colored garnet-microcline-quartz gneisses (leucogneisses). Although the weathered surfaces of these rocks are often dark due to staining, fresh samples display the typical white vs. grey color of the Sacandaga Fm. The characteristic and excellent layering of the Sacandaga Fm. is clearly developed. Note the strong flattening parallel to layering. Towards the southern end of the outcrop calc-silicates and marbles make their entrance into the section. At one fresh surface a thin layer of diopsidic marble is exposed.

At the far southern end of the roadcut there exists an exposure of the contact between the quartzo-feldspathic gneisses of the Piseco anticline and the overlying Sacandaga Fm. The hills to the south are composed of homogenous quartzo-feldspathic gneisses coring the Piseco anticline (note how ruggedly this massive unit weathers). The Sacandaga Fm. here has a northerly dip off the northern flank of the Piseco anticline and begins its descent into the southern limb of the Glens Falls syncline.

The pronounced flaggy layering in the Sacandaga Fm. is not of primary sedimentary or volcanic origin. Instead it is tectonic layering within a "straight" gneiss. Hand specimen and microscopic inspection of the light layers, particularly, reveals the existence of extreme grain size reduction and ductile flow. Long quartz rods consist of rectangular compartments of recovered quart and annealed feldspar grains occur throughout. The rock is clearly a mylonite with its mylonitic fabric parallel to compositional layering.

The chemistry of the light colored layers in the Sacandaga Fm. indicates that they are minimum melt granites. As one proceeds away from the core of the Piseco anticline, these granitic layers can be traced into less deformed sheets and veins of coarse granite and pegmatite. In the most illustrative cases the granitic material forms anastamozing sheets that get grain size reduced and drawn into parallelism as high strain zones are approached. The Sacandaga Fm. is interpreted as an end result of this process and represents a mylonitized migmatite envelope developed in metapelites where they were intruded by AMCG granites at ca. 1150 Ma and then intensely strained during the Ottawan Orogeny at ca. 1050 Ma. This interpretation is consistent with field relationships, the presence of spinel and sillimanite restites in the leucosomes, and with the fact that similar metapelitic rocks are crosscut by ca. 1300 Ma tonalites. The latter observation makes the Sacandaga Fm. protoliths older than the ca. 1150 Ma granitic rocks in the Piseco anticline and makes an intrusive relationship obligatory despite the conformable contact at the south end of the roadcut.

62.5-67.0	.5-4.5	All exposures are within the basal quartzo-feldspathic gneisses at the core of the Piseco anticline.
67.0	4.5	Re-enter the Sacandaga Fm. Dips are now southerly.
68.0	1.0	In long roadcuts of southerly dipping pink, quartzo-feldspathic gneisses with tectonic layering. The coarse grain size of the gneissic precursors can be seen in many layers.
70,4	2.4	Cross bridge over Sacandaga River.
74.4	4.0	Bridge crossing east corner of Sacandaga Reservoir into Northville, N.Y.

END LOG

Speculator - Blue Mtn Lake - 88

from McLelland et al, NYSGA 1978

29.6-30.2 Passing through thick charnockite layer in the Upper Marble.

30.2 Passing through units of the Upper Marble. Generally low dips have resulted in broad exposure of this unit. Note horizontal foliation in some roadcuts. At 32.9 cross contact with Blue Mt. Fm. which cores a local F_2 syncline.

34.3 Contact of Blue Mt. Fm. with Upper Marble. Passing into the southern limb of the F_2 syncline.

34.7 Long roadcuts of garnetiferous amphibolite in Upper Marble. Some garnets attain diameters of 5-6". A large pegmatite is also present. Note that this outcrop sites astride the hinge line of an F_2 anticline.

follow up ↑

34.8 Contact of Upper Marble amphibolites and Blue Mt. Fm.

35.4 Junction of Routes NY30 and 8 in center of Speculator.

-- Side trip, no cumulative mileage --

0 Head southeast on Rt. NY30

1.5-2.8 Charnockites of Blue Mt. Fm. At 2.8 cross into Upper Marble.

10 = 3.4 Stop 5: Northern intersection of old Rt. NY30 and new Rt. NY30, 3.3 miles east of Speculator, New York.

7

The Upper Marble Fm. is exposed in roadcuts on both sides of the highway. These exposures show typical examples of the extreme ductility of the carbonate-rich units. The south wall of the roadcut is particularly striking, for here relatively brittle layers of garnetiferous amphibolite have been intensely boudinaged and broken. The marbles, on the other hand, have yielded plastically and flowed with ease during the deformation. As a result the marble-amphibolite relationships are similar to those that would be expected between magma and country rock. Numerous rotated, angular blocks of amphibolite are scattered throughout the marble in the fashion of xenoliths in igneous intrusions. At the eastern end of the outcrop tight isoclinal folds of amphibolite and metapelitic gneisses have been broken apart and rotated. The isolated fold noses that remain "floating" in the marble have been aptly termed "tectonic fish." The early, isoclinal folds rotate on earlier foliation.

Features such as those seen within this roadcut have led this writer to question the appropriateness of assigning an unconformity to the base of the Lower Marble Fm. Tectonic phenomena in rocks of high viscosity contrast can account for the fact that the marbles are able to come into contact with a variety of lithologies.

Snowy Mountain Dome

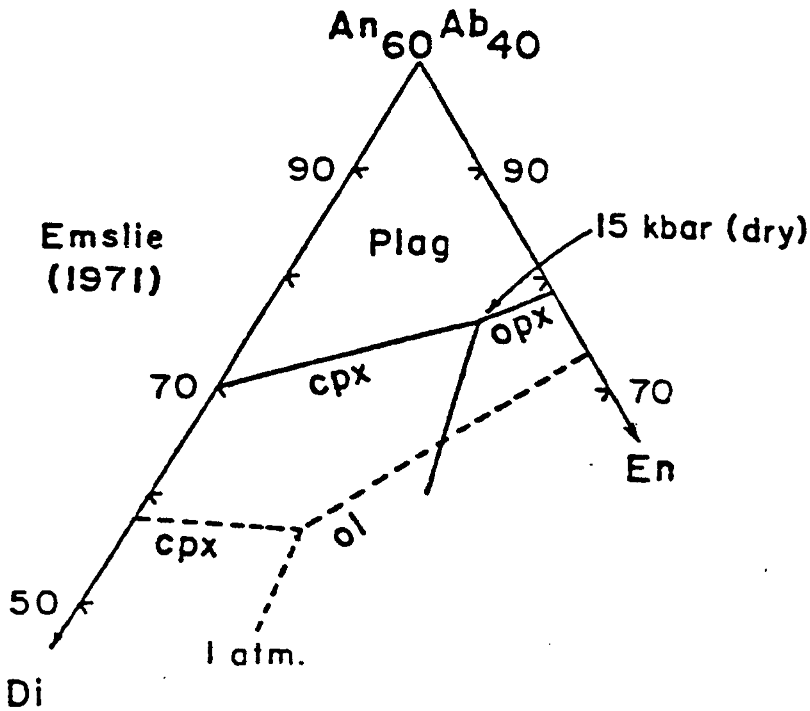

Figure 23. Liquidus phase relations in system $An_{60}Ab_{40}$-Di-En at 15 Kb load pressure, from Emslie (1971).

21.2 Charnockites of the Snowy Mt. Dome.

These type "a" quartzo-feldspathic gneisses tend to be more massive than charnockites higher in the sequence. They also contain xernocrysts of andesine. de Waard and Romey (1969) believed that these charnockites were comagmatic with the anorthosites of the Snowy Mt. Dome. Buddington (1963) has argued that they are later than the anorthosites. Isachsen, McLelland, and Whitney (1975) suggested that these, and similar, charnockites are the products of melting accompanying intrusion of the anorthosites into quartzo-feldspathic country rocks. Rb/Sr whole rock ages obtained by Hills and Isachsen (1975) yield results of ~1.2 by and do not suggest that these charnockites are part of an "older" basement complex.

23.2 State Campsite.

27.4 Mason Lake Parking Area. The lower Lake Durant Fm. and the Lower Marble are exposed in this general vicinity.

27.9 Contact of the Lower Marble with the Lake Durant Fm.

28.7 Contact of the Lake Durant Fm. with the Upper Marble.

change was prograde along a geothermal gradient which penetrated the kyanite field, followed by partial melting and decrease of lithostatic pressure into the sillimanite field and retrograde cooling within that field, with residual kyanite being trapped within feldspar-rich (solidus?) portions of the gneiss.

11.3 Intersection of Routes NY 28 and 30 in hamlet of Indian Lake. Head south of Rt. NY30.

12 12.5 <u>Stop 3</u>: Scenic overlook on east side of Route NY30.

Mountainous area to the southeast is part of the Thirteenth Lake complex, cored by anorthosite and charnockite. Overlying gneisses dip to the north (left) and west (towards us) off the complex. To the south metasedimentary rocks dip to the northeast off of Snowy Mountain dome (not visible), which also is cored by anorthosite and charnockite.

Walk south on highway NY30 to roadcut on west side of road. This roadcut is composed of a distinctive "diopside-clot" gneiss and is situated close to the axial core of the Crow Hill synform. The rock is a zircon-apatite±plagioclase±calcite±garnet-sphene-scapolite-clinopyroxene-quartz-microcline granulite. It is part of the distinctive basal portion of the Lake Durant Formation.

Chemical compositions of scapolite and clinopyroxene from this unit at another location were determined by electron-probe microanalysis to be Me_{69}, $(Na_{1.17},Ca_{2.63})_{3.80}(Al_{4.67},Si_{7.33})_{12.00}O_{24}((CO_3)_{0.97},Cl_{0.03})_{1.00}$ and salite, $(Ca_{0.93},Na_{0.06})_{0.99}(Mg_{0.62},Fe_{0.39},Mn_{0.01},Al_{0.02})_{1.04}(Si_{1.96}Al_{0.04})_2O_6$, respectively. Plagioclase composition is oligoclase, An_{21}, based on petrographic determinations using the zone method of Rittman.

11 17.2 <u>Stop 4</u>: Described by de Waard (1964) as follows: "Large roadcut on the hill 0.4 miles southwest of the intersection of highway 30 with the lake shore road through Sabael. Anorthosite at the lower end of the outcrop is overlain by metanorite (unfoliated andesine-pyroxene-hornblende gneiss) which is in turn overlain by streak andesine-pyroxene-hornblende augen gneiss. Both "Marcy-" and "Whiteface-" type anorthosites are present. The grain size of metanorites ranges from coarse to fine, and the original texture of the rock is preserved to various degrees in different parts of the exposure. Several small amphibolite (metadolerite) lenses may be observed in the streak gneiss. Foliation is nearly horizontal. Walk up the steep hillside above the road to see massive ledges of anorthosite, metanorite, and a rock which is texturally and compositionally intermediate between these two types."

The origin of, and relationships within, the anorthosite-charnockite suite of rocks has been debated for decades. Those favoring a comagmatic association have tended to postulate a dioritic parent magma which yields plagioclase (anorthositic) cumulates and charnockitic residua (de Waard and Romey, 1969). Those who do not accept a comagmatic relationship between these rocks, have generally postulated a parent of gabbroic anorthosite composition (Buddington, 1972). A variant of the gabbroic anorthosite parent is the high-alumina basalt of Morse (1975).

The snowy Mt. Dome is the type area of de Waard and Romey's (1969) comagmatic differentiation process. By detailed mapping beginning at the core of the dome, they showed that there exists an outward gradation from central anorthosite through metanorite, to noritic augen gneiss, to charnockitic gneisses (see Fig. 22). This they interpreted as reflecting a differentiation sequence and variation diagrams were constructed to portray these trends.

A critical aspect of the compositional variation within this suite is that grains (xenocrysts) of andesine occurs within the charnockitic rocks. These xenocrysts increase in abundance as the anorthositic core rocks are approached. Concomitantly the amount of K-feldspar and quartz decrease. Although these changes do result in a gradation of rock types, the transition seems to be mechanical rather than chemical. This is suggested by the constancy of xenocryst composition and the widespread presence of cross-cutting relationship between end-member rock types.

Based upon field and chemical data Buddington (1939, 1972), suggested that the charnockitic rocks are distinctly later than, and unrelated to, the anorthositic rocks. He presented variation diagrams of major oxides demonstrating that the anorthositic and charnockitic rocks follow separate differentiation trends and that discontinuities exist between their paths. Simmons (1976) and Goldberg (1977) have studied trace element and REE patterns in Adirondack anorthosite-charnockite lithologies and concur with Buddington that the two are unrelated. They also show that a gabbroic anorthosite parent is consistent with their trace-element studies. Simmons suggests that such a parent can be produced from dry melting at high load pressure of a gabbroic source rock. Figure 23 shows Emslie's (1971) results for such a system at P_L = 15 Kb and at 1 atm. The minimum melt generated at 45-50 km is essentially a gabbroic anorthosite. As it rises the field boundaries move so as to enlarge the domain of plagioclase crystallization. In this manner anorthosites may result from reasonable petrogenetic processes.

The origin of the charnockitic rocks in the suite remains largely unresolved. Buddington (1972) suggests that they represent an independent magma series in which contamination of granitic magma by garnetiferous amphibolite has been important. Husch, Kleinspehn, and McLelland (1975), as well as Isachsen, McLelland, and Whitney (1975), have suggested that the charnockite-mangerite envelope results from fusion of quartzo-feldspathic country rocks of the intruding anorthositic magma (crystal mush ?). Early in the process the anorthositic rocks attain complete crystallization and are subsequently intruded by the lower melting temperature quartzo-feldspathic lithologies. Wiebe (1975) has suggested a similar mechanism for adamellites near Zoar (Nain), Labrador. All fusion models of this sort depend critically on the initial temperature of the charnockitic rocks and the heat budget within the system. Although the lack of data on heat capacities, heats of fusion, etc., preclude detailed calculations, it does seem possible that at 8-10 Kb anorthositic intrusives with temperatures of 1200-1300°C can melt substantial quantities of quartzo-feldspathic gneisses initially at 800°C. Whether or not this mechanism actually operates is a question deserving of extensive research. It is certainly consistent with field evidence suggesting that stratigraphically continuous units undergo increasing anatexis as anorthositic rocks are approached. Some examples of this anatexis will be seen at Stop 7.

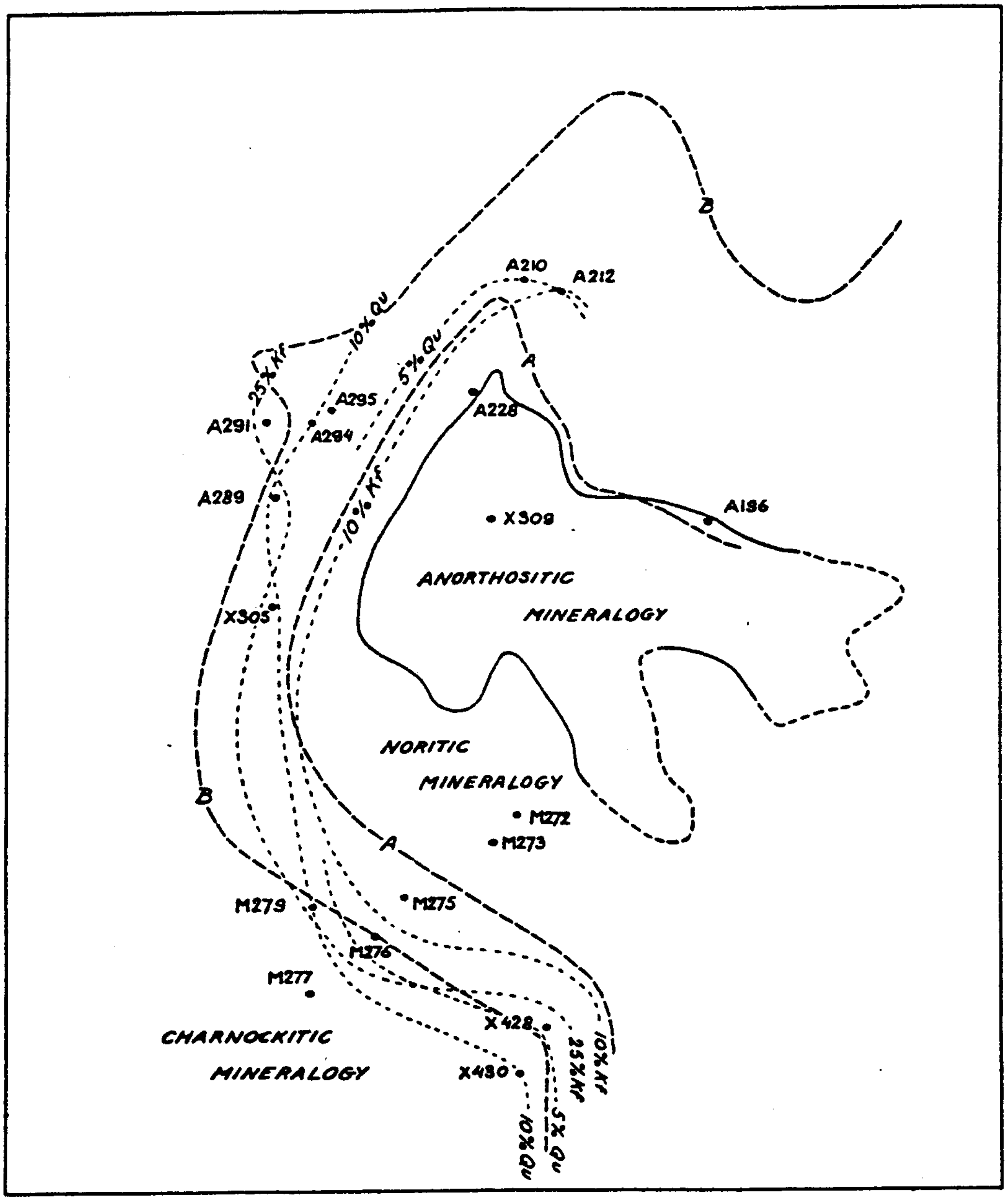

Figure 22. Textural and compositional boundaries and gradations in central part of Snowy Mountain dome. Arbitrary textural lines from center outward: (A) indicates the approximate location of transition from deformed blastonoritic texture to augen-gneiss texture; (B) indicates approximate zone in which number of andesine augen decreases to less than one per square meter. Compositional boundaries and isopleths: solid line indicates boundary zone between anorthosite and metanorite; dashed lines are isopleths of 10 and 25 percent modal K feldspar, and 5 and 10 percent modal quartz. Intersection of structural line A with anorthosite-metanorite boundary zone reflects occurrence of finer grained and foliated Whiteface-type anorthosite developed along this part of the boundary (from de Waard and Romey, 1969 and de Waard, 1964b).

Barton Garnet Mine

parking area on R. Get Permission at office.

13 STOP 28. Barton Garnet Mine This garnet deposit was mined intermittently between 1882 and 1924, and continuously from 1925 to 1983. In 1984, the Barton Mines Corporation transferred its operations to the nearby Ruby Mountain deposit; at the time of its closing the Barton Mine was the oldest continuously operated, family owned mine in the United States. The mine produced high quality abrasive garnet, which owes its exceptional abrasive properties to a well-developed cubic parting in the garnet crystals.

The elongate open pit, oriented roughly ENE-WSW, is located in a small olivine metagabbro body, which is in contact with gabbroic anorthosite gneiss on the north, with a fault contact against quartz mangerite on the south (Fig. 23). Along the north wall of the pit, typical olivine metagabbro with well-preserved igneous textures is exposed. Faint igneous layering is locally visible, and a xenolith of foliated anorthosite was reported by Luther (1976). The igneous mineralogy of this rock was plagioclase-olivine-clinopyroxene-ilmenite. During metamorphism, coronas of two pyroxenes and garnet have formed between olivine and plagioclase, and coronas of biotite, hornblende and garnet have formed between plagioclase and ilmenite (Whitney and McLelland, 1973, 1983). As a part of the corona-forming process, the plagioclase has become clouded with a host of minute (1-10 micron) grains of green spinel (Whitney, 1972).

Going S across the pit, the gabbro undergoes a nearly isochemical transition (columns P and Q, Table 1) into garnet amphibolite, with garnet porphyroblasts commonly up to 0.3 m and rarely up to 1 m in diameter. It is this garnet amphibolite that constitutes the ore; interestingly, the modal garnet in the ore is approximately the same (roughly 15-20%) as in the coronitic gabbros. The composition of the garnet (Levin, 1950) is approximately 43% pyrope, 40% almandite, 14% grossular, 2% andradite and 1% spessartite; zoning, where present, is very weak and variable (Luther, 1976). Toward the W end of the pit, garnet hornblendite with little or no plagioclase is locally present, probably representing ultramafic layers or pods in the original gabbro. In the more mafic zones of the ore body, garnet porphyroblasts are rimmed with up to several cm of hornblende; where the host is less mafic, the garnets have plagioclase ($\pm$ hypersthene) rims. The details of the ore-forming process are not well understood, however, the ore body probably represents a zone within the gabbro where fH_2O was locally higher during metamorphism, facilitating diffusion and the growth of the large garnets and favoring the extensive development of hornblende at the expense of pyroxenes and plagioclase. The S wall of the pit is marked by a thin zone of more leucocratic, garnet bearing rock ("light ore"), which is in fault contact with the adjacent quartz mangerite. This leucocratic gabbro locally shows ductile deformation and development of foliation parallel to the fault.

Return to parking area, turn around and return to Rte. 28 via Barton Mines Road.

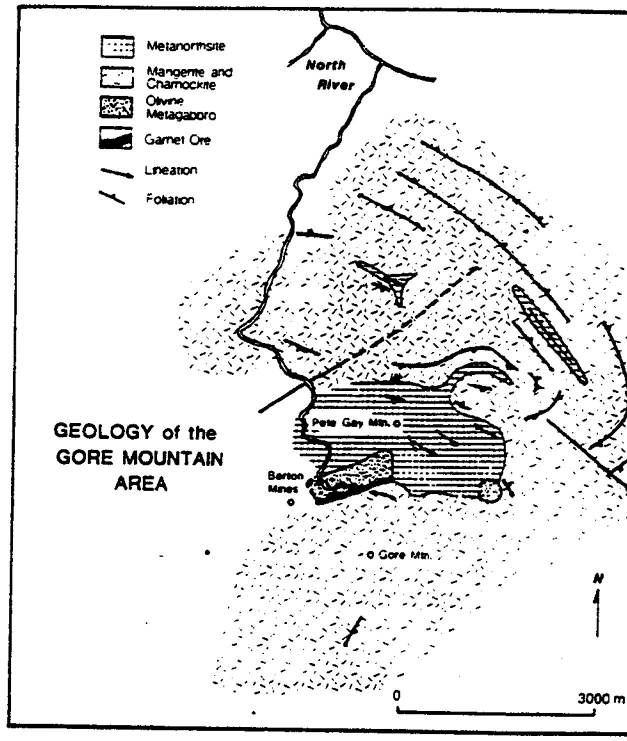

FIGURE 23 Geologic map of the vicinity of the Barton garnet mine. (after Bartholeme, 1956)

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Road Log

Mileage

0 Intersection of Routes NY28N-30 and NY28-30 in Blue Mt. Lake. Head south on Rt. 28-30.

1.6 - ASSEMBLY POINT IN ROADSIDE PARKING AREA, E, END, NORTH SHORE OF LAKE DURANT. ASSEMBLY TIME 7:30 AM, SATURDAY, 23 SEPTEMBER.

15 Stop 1: Large roadcut on north shore of Lake Durant. This location is the type section of Lake Durant Formation (D. de Waard, 1970, pers. comm.).

This stop has been described by de Waard (1964b) as follows: "The section of diverse, layered metamorphic rocks includes pink and greenish leucocratic gneisses with thin metabasic layers, marble, and calc-silicate rocks. The section forms part of the supracrustal sequence which overlies the leptites of the Wakely nappe exposed in the hills visible towards the south across the lake, and which underlies the Blue Mountain charnockite sequence towards the north. Lineations on foliation planes indicate a 30° NE plunging fold axis. The intrusive nature of marble into boudinaged layered gneiss is shown on the west end of the north side of the road cut."

Outcrop mapping to the east and south has revealed that the Lake Durant Formation contains large amounts of hornblende granitic gneiss and biotite-hornblende granitic gneiss both above and below the well-layered sequence exposed in the type section at Lake Durant. In addition, a distinctive rock sequence of biotite granitic gneiss (bottom), calc-silicate rock, and platey-quartz gneiss (top) makes up the basal portion of the Lake Durant Formation in areas to the south.

8.0 Trail-side parking area (south) on Route NY28-30. (This is about 1.25 mi north of intersection of Rt. NY28-30 and the Cedar River Road.)

14 Stop 2: Ledge Mt. Hike 1 mi east through open woods, to well-exposed south-facing cliffs. This is on the southward culmination of the recumbent Ledge Mountain antiform. Quartz-sillimanite lenses increase in size and relative amount from west to east, until they assume the proportions of major layering in the gneiss. Kyanite occurs here in two feldspar-rich portions of the gneiss. We have sought more, without success. If you should discover additional kyanite, PLEASE OBSERVE PETROLOGIC ETIQUETTE OF PHASE PRESERVATION! NOTIFY TRIP LEADER, WHO WILL OFFER SUITABLE REWARD. Note different proportions of magnetite, garnet, and biotite in feldspathic portions of gneiss, as well as in pegmatite. The structural relationships of pegmatite to host gneiss also differ. Note in Figure 21A that biotite compositions within garnet porphyroblasts are Mg-richer and Al-poorer than "Free" matrix biotite. Also, of the four plagioclase - garnet equilibria shown in Figure 21B (representing five pairs), all represent 'probed rims of grain pairs each of which is in mutual contact. The highest-P boundary is that calculated for a relatively large plagioclase grain within a garnet porphyroblast; the others are of small plagioclases within garnet porphyroblasts, and of "free" plagioclases against garnet rims.

It is deduced from these relationships, and from ubiquitous but small-scale late corona structures of albite on magnetite, that the path of P-T

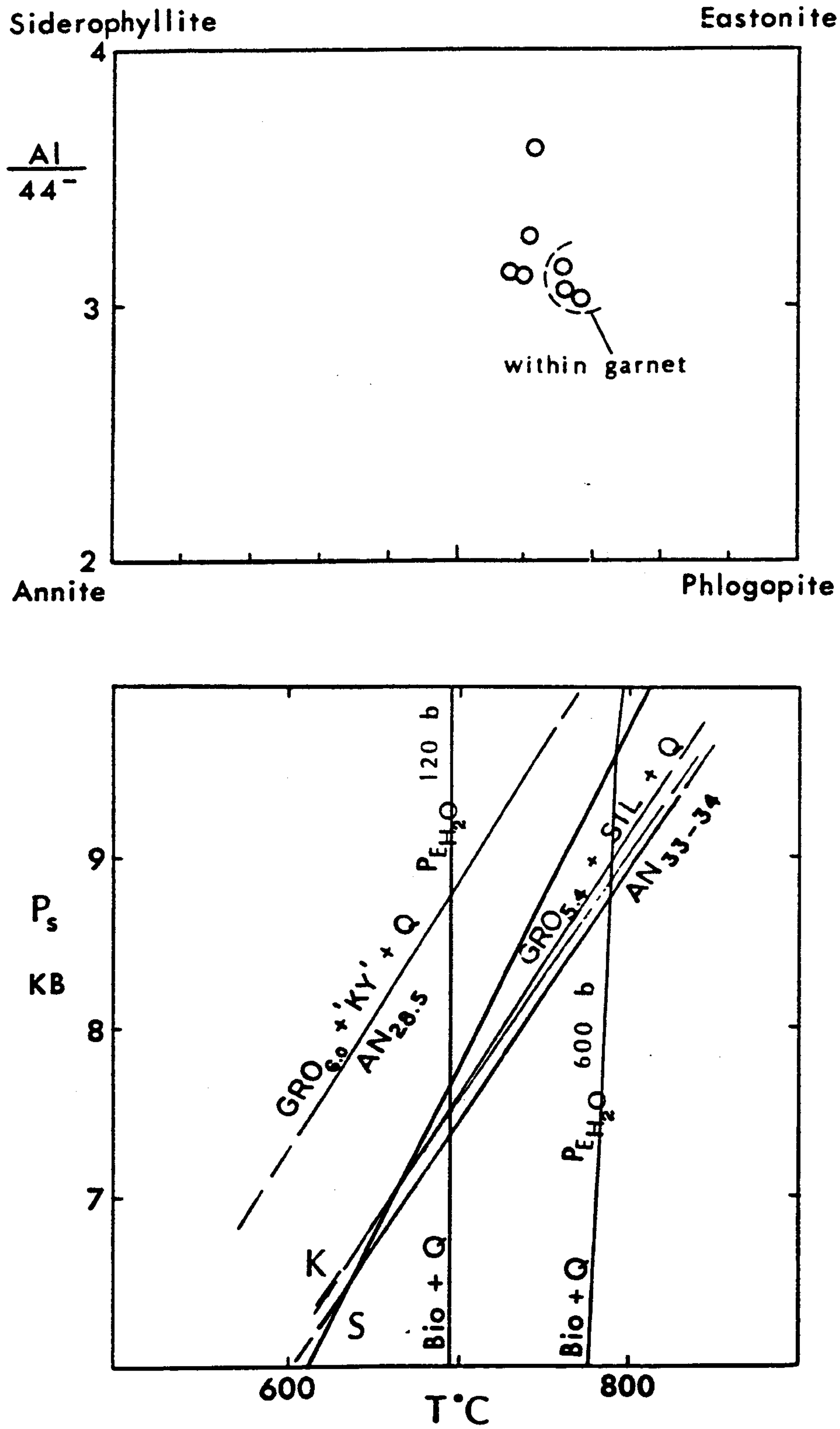

Figure 21. Biotite-garnet-plagioclase relationships. Ordinate in A is number of Al atoms per total of 44 anionic charges. Grossular 6.0% - $An_{28.5}$ (moles) in B refers to large plagioclase inclusion within garnet porphyroblast.

Long Lake Area

Fallon, NYSCA, 1984

(16) 2.7 Buck Mountain Fold Complex

This stop offers the opportunity to see the diopside augen quartz gneiss and other quartz rich gneisses and well exposed examples of second and third phase folds.

Road Cuts: Glassy quartzites and diopside augen quartz gneiss with small sparse diopside augen are exposed on the western side of Route 30. The small outcrop on the eastern side of the highway contains a few large dark green diopside porphyroblasts.

1-A.) Buck Mountain Syncline. Walk east from the highway 600 feet down the logging road. Bear left at a small cabin onto an overgrown skidder trail. Walk 1000 feet approximately N45E along a contour. The outcrop on the steep slope to the southeast exposes the hinge of the Buck Mountain Syncline, a large third phase fold (see figs. 8 and 9). This is a tight fold which plunges moderately to the southeast on a northwest trending, steeply south dipping axial surface.

1-B.) Proceed west from the highway on an overgrown trail from the south end of the western road cut for about 1000 feet. Turn north and climb the hill to a large pavement outcrop.

The rock is typical diopside augen quartz gneiss. The augen which have weathered out, giving the rock this distinctive texture, are large single crystals or aggregates of smaller grains of green to white diopside. Scapolite is common. Minor K-feldspar, and traces of calcite, sphene, and zoisite are present.

A second phase isoclinal fold, outlined by a thin amphibolite, is refolded by a tight third phase fold, forming a hook interference pattern. Note the lack of an axial plane foliation in the hinge of the second phase fold. A more prominent axial plane feature is present in the hinges of third phase folds. Other folds, and a nice view to the south, are visible higher on the hill.

Blue Mtn Lake –
Lake Placid

DAY #4

0.0 Jct. of Rtes. 30 and 28 in Blue Mtn. Lake Village. Proceed N on 30.

0.6 Entrance to Meadowbrook Conference Center on L.

1.0 Entrance to Adirondack Museum on L; park in Museum parking lot, upper level, for Stop #16.

1.1 17 STOP 16. Hornblende Granitic Gneiss Walk to cuts on the W side of Rte. 30. This rock is typical of the hornblende granitic gneisses of the Adirondack highlands. The principal feldspar is mesoperthite and the chief mafic mineral is a hastingsitic hornblende. Geochemically, these rocks appear to be near the felsic end of a continuum which also includes charnockite (Stop 17) and possibly mangerite (Stop 18) (Whitney, 1986; McLelland and Whitney, 1987). An analysis of the rock at this outcrop is shown in Table 1 column H. Both this rock and the charnockite at Stop 17 are migmatites; the texture can be best observed on the glaciated upper surface of the cut. The coarse to locally pegmatitic leucosomes are generally concordant with the foliation but locally crosscut it, suggesting syntectonic partial melting. Almost all granitoids in the Adirondack highlands are migmatites, a fact easily overlooked on fresh surfaces due to the (usually) small contrast in color index between leucosome and host.

Return to Rte. 30 and continue N.

3.4 18 STOP 17. Charnockite Park off road to the R and cross to a long, dark outcrop of charnockite (orthopyroxene-clinopyroxene-hornblende-plagioclase-quartz-mesoperthite gneiss). Compare the analysis for this rock (Table 1, Col. I) with those for the hornblende granitic gneiss of Stop 16 (Col. H) and mangerite of Stop 18 (Col. J). Also present in this outcrop are several layers of amphibolite. Foliation in the amphibolite is roughly parallel to that in the charnockite, but not always to the contacts, suggesting that the amphibolites may be pre-tectonic mafic dikes. At first glance the charnockite looks relatively homogeneous, but the glaciated surface at the top shows the rock to be a migmatite.

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10.7 Jct. of Rtes. 30 and 28N in Long Lake Village; continue N on 30.

11.4 Crossing Long Lake. This lake is the westernmost of several large Adirondack lakes that occupy prominent NE-trending lineaments (see discussion of "Brittle Structure").

17.8 Road to Sabattis on L. Continue on Rte. 30.

29.4 19 STOP 18. Mangerite Exposed in a long roadcut on the southeast side of the highway are typical, olive-gray mangeritic rocks of the Tupper-Saranac Complex. The mangerites consist principally of mesoperthite, clinopyroxene, orthopyroxene, and minor Fe-Ti oxides, quartz, apatite and zircon. Igneous textures have been partially preserved and suggest a cumulus origin for at least some of the mesoperthite. Some hornblende is present and may be secondary. Xenocrysts of plagioclase zoned from An_{50} to An_{20} occur and have cores clouded with oriented inclusions of Fe-Ti oxides. The xenocrysts are thought to be derived from anorthositic rocks which contain similarly clouded andesine. Several late dikes cut the outcrop. One of these is a pyroxenite similar to the orthopyroxenite dikes associated with anorthosite at Roaring Brook (Stop 23). Samples of mangerite from this roadcut yield a tightly constrained U/Pb zircon age of 1134 ± 4 Ma.

33.0 Jct. of Rtes. 30 and 3 in Tupper Lake Village; proceed E on 3.

34.4 Entering series of roadcuts in mixed mangerite and anorthosite.

37.0 20 STOP 19. Anorthosite-Mangerite Contact Relations Park off road to R and cross to a small gravel road on L. Walk up this road to a small abandoned quarry. Roadcuts along Route 30 near the quarry entrance, as well as exposures in the quarry, show light colored mangerite crosscutting anorthosite, yielding anorthosite xenoliths as well as abundant andesine xenocrysts in the mangerite. Staining of the outcrop surface in the quarry with sodium cobaltinitrite demonstrates that the mangerite permeates the anorthosite, filling interstices between the plagioclase grains. This suggests that the anorthosite was not totally solidified when intruded by the mangerite, and that the two rocks are essentially contemporaneous. The whole rock chemistry of the mangerite is similar to that of the rock at Stop 18, but SiO_2 is lower and FeO and CaO are higher at the quarry site, as might be expected if some mixing had taken place between mangerite and a late differentiate of the anorthosite.

DAY 2

SATURDAY, MAY 6

LAKE PLACID - TICONDEROGA

STOPS 23 - 41

EASTERN AND NORTHEASTERN ADIRONDACKS

FACIES OF THE MARCY ANORTHOSITE

BRANT LAKE ANTIFORM

WHITEHALL SYNCLINE

38.55 Rte. 30 branches N; continue E on 3.

21

40.76 STOP 20. Metanorthosite of the Marcy Massif Park in sandy area on L side of road and walk E to first outcrops. This coarse grained andesine rock is typical, in both composition and relict igneous
40 texture, of the most voluminous member of the anorthosite series. This exposure contrasts 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 (commonly An_{40}-An_{50} 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 only occasionally seen 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 (locally present as coronas around hypersthene, oxides, and -rarely- apatite, and as discrete porphyroblasts where the anorthosite has been extensively deformed and recystallized); secondary clinopyroxene, amphibole, and, less commonly, biotite, clinozoisite, and scapolite.

Zircons extracted from the anorthosite at this location are small and clear, and yield an age of 1040±43 Ma (Chiarenzelli and others, 1987). This suggests either that the anorthosite was emplaced at this time and is thus approximately synmetamorphic, or that the zircon itself is a metamorphic mineral. Unequivocal evidence of the crystallization age of the anorthosite is still lacking.

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 metanorthosite in the NW lobe of the Marcy Massif are extremely homogeneous ($\delta^{18}O = 9.3 \pm 0.2$), which in conjunction with the preservation of magmatic features (Davis, 1969), indicates that the oxygen isotopic composition reflects magmatic values rather than exchange with metamorphic fluids.

22

45.6 STOP 21. Cumulus-textured metanorthosite Park in small parking area on L at bottom of hill; walk back uphill to outcrops on N side. In contrast with the previous stop, the plagioclase in the anorthosite at this stop shows a strong preferred orientation, with lath-shaped crystals having their long axes in the horizontal plane. We interpret this as an adcumulus texture. Note the faint, nearly horizontal, igneous layering visible on the weathered surface at the west end of this outcrop. Preferred orientation of igneous plagioclase is rarely this pronounced in Adirondack anorthosites, however a weaker preferred orientation is commonly observed, and has usually been attributed to flow differentiation (Buddington, 1969). Zircons from the anorthosite at this exposure resemble those at the previous stop, and give an essentially identical date (1054±20 Ma; Chiarenzelli and others, 1987). Continue E on 3.

49.8 Cross Saranac River.

53.8 Jct. of Rtes. 3 and 86 in Saranac Lake Village. Continue straight ahead, now on 86.

60.5 Jct. of Old Military Road; continue on Route 3.

62.8 Sharp right turn at bottom of hill in Lake Placid Village.

63.4

63.8

65.8

67.0

70.2

72.2

100

Eastern Marcy Massif

NYSGA, 1988

GRENVILLE CALC-SILICATE, ANORTHOSITE, GABBRO, AND IRON-RICH SYENITIC ROCKS FROM THE NORTHEASTERN ADIRONDACKS

By

ELIZABETH B. JAFFE and HOWARD W. JAFFE
Department of Geology and Geography
University of Massachusetts, Amherst

INTRODUCTION

The Marcy anorthosite massif is delineated by a major NW-SE-trending lobe and a smaller N-S-trending lobe which coalesce to the S to form a heart-shaped outcrop pattern covering 5000km^2. (Fig. 1). In section , the major NW-trending lobe approximates a piano bench or slab 3-4.5 km thick with two legs or feeder pipes extending at least 10 km down according to the geophysical model of Simmons (1964), whereas Buddington (1969) favored an asymmetrical domical shape based on extensive field mapping and other considerations. The massif consists of a coarsely crystalline core of apparently undeformed felsic andesine anorthosite thrust over a multiply deformed roof facies consisting of gabbroic-noritic anorthosite, gabbroic anorthosite gneiss, and quartz-bearing ferrosyenite-ferromonzonite facies (Pitchoff Gneiss). Remnants of a siliceous carbonate- and quartzite-rich metasedimentary sequence and associated garnet-pyroxene-microperthite granulites form discontinuous screens and xenoliths in the roof facies. Xenoliths of any kind are very rare inside the felsic anorthosite of the core, but abound in the gabbroic anorthosite of the roof facies.

We will visit ten outcrops which include all of the major rock types (Fig. 1) of the massif, and that lie principally in its multiply deformed roof facies.

That regional metamorphism took place at high pressure, in the range of 8-10 kbar at about 800^{o}, is indicated by the occurrence of orthoferrosilite, Fs95 + quartz, and the absence of any vestiges of fayalite in the ferrosyenite facies of the Pitchoff Gneiss (Jaffe et al., 1978). Because recent Sm147-Nd143 age dating yields 1288 M.Y. for the age of magmatic crystallization of the Marcy anorthosite (Ashwal and Wooden, 1983), and older Pb-U age dating by Silver (1969) yields about 1130 M.Y. for crystallization and about 1100-1020 for metamorphism, the Grenville orogeny may have spanned as much as 200 M.Y. and it is difficult to fix the peak of metamorphism with a specific thermal or tectonic event within this time span.

23

Stop 1. FAYALITE-FERROHEDENBERGITE GRANITE, AUSABLE FORKS AREA, 15' AUSABLE FORKS QUADRANGLE, LOCALITY AF-1-A

Iron-rich granitic-syenitic-quartz monzonitic rocks, ascribed to a charnockitic gneiss series, are abundant in the northeastern and central Adirondacks, where they occur in close association with anorthositic and metasedimentary calc-silicate rocks in the Marcy Massif. Most of these contain iron-rich orthopyroxene (eulite or orthoferrosilite) with quartz, an assemblage stabilized at the high operable regional metamorphic pressure of about 8 kbar, with T = 700-770°C (Jaffe, Robinson, and Tracy, 1978: Bohlen and Boettcher, 1981). Other members of this alkali-feldspar-rich series contain fayalite and quartz in place of orthoferrosilite and quartz. From tables 1, 2, 3, 4, and Fig. 9 and Table 7 from Jaffe, Robinson and Tracy (1978), (Appendix I) it is reasonable to assume that both of these rock types were recrystallized under similar metamorphic conditions. Work by Bohlen and Essene (1978) and by Ollilla, Jaffe, and Jaffe (1988) indicate that these rocks had igneous precursors that crystallized above 900°.

JAFFE ET AL.: IRON-RICH PYROXENES (1978)

Table 1 Modes of pyroxene-microperthite gneiss

	Po-13	SC-6	Po-17	FFG-6	AF-1A	Po-2
Quartz	4.1	9.1	13.8	18.4	69.1	1.5
Microperthite	37.2*	46.5*	52.6*	69.2**	20.5	84.5
Plagioclase***	32.7	25.1	18.2	3.5	Trace	3.6
Augite	6.2	8.7	6.6	3.7	5.5	4.1
Orthopyroxene	3.9	2.4	1.9	None		2.4
Olivine	None	None	None	0.6	3.0	None
Hornblende	3.8	2.1	4.1	3.9	Trace	1.8
Garnet	7.1	3.6	1.7	None	None	0.8
Magnetite + Ilmenite	3.0	1.9	0.8	0.5 Trace	1.1	1.0
Zircon	0.2	0.3	0.1	0.1	0.3	Trace
Apatite	1.8	0.3	0.1	0.1	0.3	0.3
Fluorite					0.2	–
	100.0	100.0	99.9	100.0	100.0	100.0
Color Index	26	19	15	9	10	15
Mole % An*** Plagioclase	23	16	[illegible]	11	12	21
Mole % Fs Orthopyroxene	85-88† 83-85††	97 89	95 91			83-9
Mole % Fa† Olivine				99	99	

*Contains abundant exsolution lamellae of albite and is partially inverted to microcline microperthite.

**A mesoperthite containing an oligoclase, An_{11} component.

***An values for plagioclase host grains determined by measurement of the α index of refraction in oils. For probe analyses see Table 8.

†100(Fe + Mn)/(Fe + Mn + Mg).

††100 Fe/(Ca + Fe + Mn + Mg) as used by Smith (1971a).

Table 2. Optical properties of iron-rich pyroxenes

	Po-13	SC-6	Po-17	$FeSiO_3$*
Orthopyroxenes				
γ	1.7763	1.785	1.786 Z=pale blue-green	1.789
β	1.7765	1.774	1.774 Y=pink-yellow	1.780
α	1.7560	1.764	1.765 X=pale pink	1.772
γ−α	0.0203	0.021	0.020	0.017
$2V_{calc.}$	88°,(−)	88°,(+)	79°,(+)	86°,(+)
Disp.	r<v,strong	r>v,strong	r>v,strong	
$\frac{Fe+Mn}{Fe+Mn+Mg}$	.88	.92	.95	1.00

	Po-13	SC-6	Po-17	FFG-6	Fe Augite**
Augites					
γ	1.7505	1.759	1.760	1.7645	(1.7646)
$\frac{Fe+Mn}{Fe+Mn+Mg}$	.83	.88	.93	.95	1.00

*Indices as reported by Lindsley, Davis, and MacGregor, 1964. They report 2V = 58°(+) which is inconsistent with their indices.

**The value of γ for pure Fe augite is from the equation of Jaffe et al., 1975.

Table 3. Lattice parameters of natural and synthetic orthoferrosilite

	Po-17	Synthetic $FeSiO_3$ Burnham (1965)	Synthetic $FeSiO_3$ Sueno et al. (1973)
a (Å)	18.42	18.431	18.418
b (Å)	9.050	9.080	9.078
c (Å)	5.241	5.238	5.237
V (Å^3)	873.68	876.60	875.14

Table 4 Representative electron probe analyses of coexisting pyroxenes and olivine, and compositions of theoretical end members

	Po-13A	Po-13B	SC-6	Po-17	FFG-6	$CaFeSi_2O_6$
Augite						
SiO_2	47.81	47.84	48.51	48.69	47.66	48.44
Al_2O_3	1.31	1.17	1.00	1.00	.65	
TiO_2	.12	.10	.09	.14	.06	
Cr_2O_3	.05	.05	.00	.06	.03	
MgO	3.78	3.08	2.07	1.17	.91	
ZnO	.09	.07	.14	.19	.01	
FeO	25.73	26.75	27.68	29.18	29.14	28.96
MnO	.27	.38	.37	.43	.58	
CaO	19.54	19.92	19.43	19.07	20.32	22.60
BaO				.00		
Na_2O	.73	.72	.78	.72	.73	
K_2O	.00	.00	.00	.00		
Total	99.43	100.08	100.07	100.65	100.09	100.00
Augite						
Si	1.926	1.925	1.965	1.971	1.949	2.000
Al	.062	.056	.035	.023	.031	
Fe^{3+}	.012	.019			.020	
Total	2.000	2.000	2.000	2.000	2.000	2.000
Al			.013	.025		
Ti	.004	.003	.003	.004	.001	
Cr	.002	.002		.002	.000	
Fe^{3+}	.122	.124	.078	.044	.106	
Mg	.227	.185	.125	.071	.054	
Zn	.003	.002	.004	.006	.000	
Fe^{2+}	.733	.757	.860	.947	.871	1.000
Mn	.009	.013	.013	.015	.020	
Ca	.844	.859	.843	.830	.891	1.000
Na	.057	.056	.061	.057	.057	
Total	2.001	2.001	2.000	2.001	2.000	2.000

	Po-13A	Po-13B	SC-6	Po-17	FFG-6	$FeSi_2O_6$
Orthopyroxene					Fayalite	$Fe_2Si_2O_6$
SiO_2	46.15	45.90	45.65	45.50	28.62	45.54
Al_2O_3	.30	.31	.27	.36	.08	
TiO_2	.06	.08	.13	.11	.00	
Cr_2O_3	.06	.02	.05	.06	.03	
MgO	4.62	3.52	2.48	1.36	.39	
ZnO	.28	.26	.38	.37	.31	
FeO	46.36	47.60	49.28	49.81	69.51	54.46
MnO	.51	.86	.72	1.10	1.57	
CaO	.80	.75	.79	.84	.05	
BaO				.06		
Na_2O	.00	.00	.00	.03		
K_2O	.00	.01	.00	.00		
Total	99.14	99.31	99.75	99.60	100.56	100.00
Orthopyroxene					Fayalite	$Fe_2Si_2O_6$
Si	1.969	1.971	1.969	1.980	.964	2.000
Al	.015	.016	.014	.019	.003	
Fe^{3+}	.016	.013	.017	.001	.033	
Total	2.000	2.000	2.000	2.000	1.000	2.000
Ti	.002	.003	.004	.004		
Cr	.002	.001	.002	.002	0	
Fe^{3+}	.025	.023	.022	.014	.036	
Mg	.294	.225	.159	.088	.019	
Zn	.009	.008	.012	.012	.007	
Fe^{2+}	1.613	1.674	1.739	1.798	1.892	2.000
Mn	.018	.031	.026	.041	.045	
Ca	.037	.035	.037	.039	.001	
Ba				.001		
Na				.003		
K		.001				
Total	2.000	2.001	2.001	2.000	2.000	2.000

Here, at Stop 1, we will visit outcrops of the fayalite-ferrohedenbergite granite and later, at Stop 4, we will visit a eulite-ferrohedenbergite syenite gneiss on Pitchoff Mt. in the 15' Mt. Marcy quadrangle.

A circular outcrop area, about 1 km in radius from the center of Ausable Forks village, was mapped by Kemp and Alling (1925) as an olivine-bearing quartz nordmarkite. They located several quarries within this outcrop area. On a fresh break, the rock shows the greenish cast typical of the ferrosyenites and charnockitic rocks of the northeastern and central Adirondack region. It is a medium-grained (1-5mm) hypersolvus granite, in places gneissic, with a color index of 5-15. Similar fayalite-bearing granitic rocks were described by Buddington and Leonard (1962) from the Cranberry Lake quadrangle, near Wanakena, from the St. Lawrence Co. magnetite district of the central Adirondacks. At Wanakena, and very likely at Ausable Forks, eulite- or orthoferrosilite-ferrohedenbergite granitic-quartz-monzonitic gneiss is closely associated with the fayalite-ferrohedenbergite granitic rock. Fayalite and ferrosilite, together with quartz, have not thus far been found in the same specimen; if they were it would provide a precise geobarometric value for the pressure of regional metamorphism. From Fig. 9 and Table 7, Jaffe, Robinson and Tracy, (1978) (see Appendix I) it will be seen that the assemblage orthoferrosilite + quartz gives a minimum P, whereas fayalite + quartz gives a maximum P. A range of 7-9 kbar at 600^{O} or 10-12 kbar at 800^{O} outlines the extremes of the metamorphic P-T conditions. Recent work by Ollila, Jaffe and Jaffe (1988) indicates that the orthoferrosilite in Pitchoff Mt. syenite gneiss is actually an inverted pigeonite crystallized from a magma above 9 kbar and 900^{O}c, conditions in excess of those accepted for the regional metamorphic peak.

In the Ausable Forks area, fayalite-ferrohedenbergite granite contains only trace amounts of hornblende: in outcrops where hornblende becomes abundant, fayalite is pseudomorphously altered to a brown fibrous serpentine or talc. The granitic rocks are cut by dikes of hornblende granite pegmatite and diabase.

The fayalite-ferrohedenbergite granite differs from the orthoferrosilite-ferrohedenbergite granitic-syenitic gneisses in several important aspects;

1) the fayalite granite is massive to poorly foliated, while the ferrosilite granitic gneiss is well foliated.

2) the fayalite granite is hypersolvus, carrying only a "strip" or "striped" microperthite that is slightly unmixed to sodic plagioclase and orthoclase, whereas the ferrosilite granitic gneiss is subsolvus, containing blebby and patchy microperthite more unmixed to sodic plagioclase and partly inverted to microcline, and this microcline microperthite coexists with an intermediate plagioclase,

3) the fayalite granite does not contain garnet because of the absence of intermediate plagioclase, whereas the ferrosilite granitic gneiss always carries garnet.

All of this suggests that the fayalite granite might be younger than the ferrosilite granitic gneiss. We concur with Buddington and Leonard (1962) who suggested that the fayalite granite could have originated from the fractional remelting at depth of the pyroxene granitic gneisses, with its intrusion occurring during the waning stages of deformation.

24

Stop 2. ANORTHOSITE NEAR COVERED BRIDGE AT JAY, 15' AUSABLE FORKS QUADRANGLE

Outcrops just beyond a covered bridge about 0.32 km E of Ausable Forks center show the characteristic textures of anorthositic rocks along the margins of the Marcy Massif. In the roadcut, anorthositic block structure shows up to 2m blocks of coarse, cumulate-textured andesine anorthosite, and coarse hypersthene (to 25 x 15cm) enclosed in a gabbroic anorthosite. Dark layers, up to 2.5cm thick, occur within the country rock. Outcrops in the East Branch of the Ausable River are principally of coarse andesine anorthosite with a megacryst index of about 20-30 and a color index of only 1. Numerous shear veinlets crisscross the anorthosite, trending N40E and N10W for the most part.

25

Stop 3. GRENVILLE-ANORTHOSITE HYBRID GNEISS 3.4 KM S OF UPPER JAY ON RTE 9N IN THE 15' LAKE PLACID QUADRANGLE

We are located in a 6.4 x 1.6 km north-trending section of Grenville strata consisting largely of calc-silicate-amphibolite-marble assemblages. Graphitic marbles occur across the Ausable River to the W of this roadcut. All these have been intruded and pervaded by sills of gabbroic anorthositic composition, resulting in the formation of Grenville-anorthosite-hybrid gneisses. Subsequently these were intruded by a sill of gabbro that forms the center of the E side of the large road cut on Rte. 9N (Fig. 2).

The section may be divided into three parts:

1) a lower unit consists principally of a mottled granular black and white sphene-augite-andesine calc-silicate gneiss discontinuously interlayered with black hornblendite-amphibolite lenses presumably of volcanic origin. The granular mottled host rock consists of white andesine, An32-39, black augite, and 5-10% of red-brown to yellow-brown sphene. The black amphibolitic layers contain mostly brown hornblende along with white plagioclase now altered to prehnite and calcite.

2) a central unit is a biotite-hornblende-hypersthene-augite-garnet-plagioclase metagabbro sill. The abundance of garnet, 20%, suggests that the gabbro sill may have been olivine-rich. The sill shows sharp contacts above and below with the anorthosite-calc-silicate-hybrid gneiss.

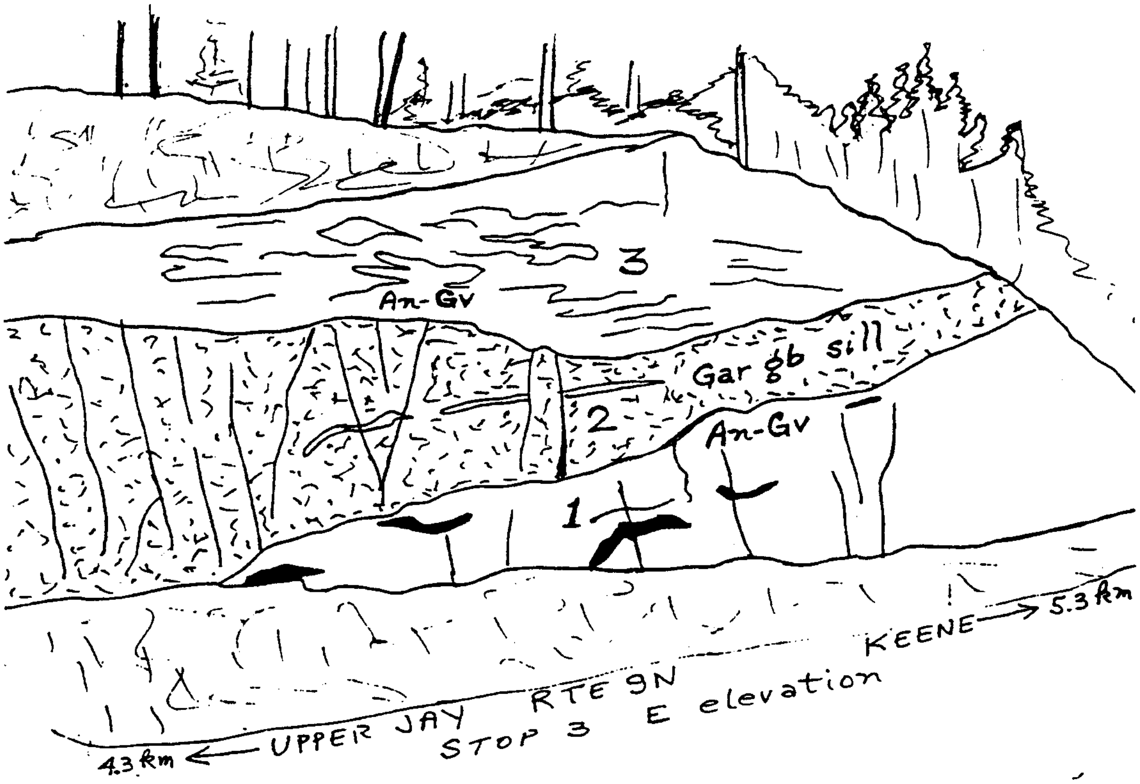

Figure 2. Grenville calc-silicate anorthosite hybrid.

3) above the upper contact of the gabbro sill, the rock is augite-andesine An39-40 gabbroic anorthosite gneiss intercalated with dark amphibolite and calc-silicate layers.

Sporadic large garnets up to 6 cm in diameter occur along contacts of anorthositic and mafic layers.

The outcrop on the W side of the road shows a well-developed high strain pencil lineation, oriented N25E.

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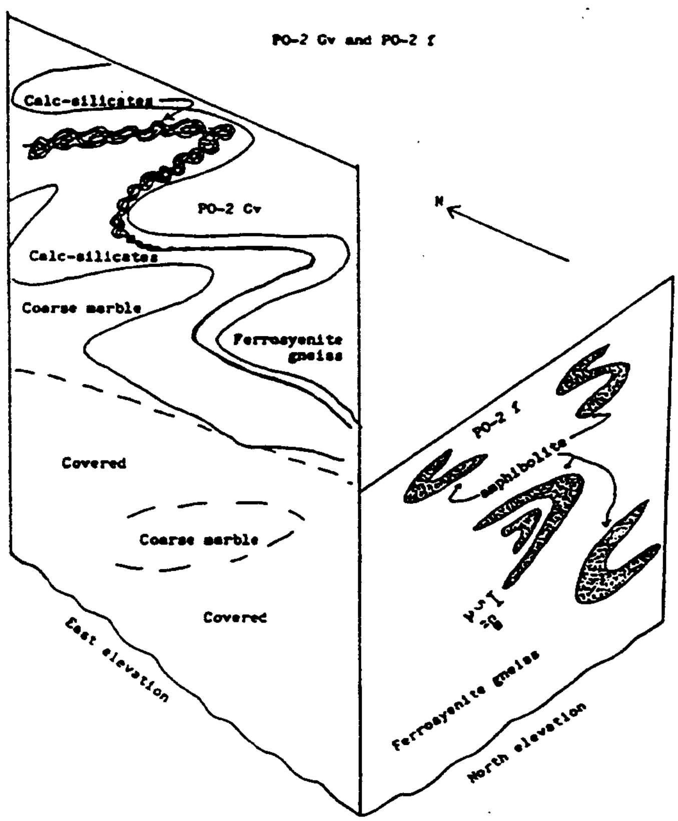
Stop 4. THE PITCHOFF GNEISS - FERROSYENITE FACIES, PO-2

Turn right (north) on Route 73, and go through Keene Village, where Route 73 turns west. Follow it about 4.5 miles (7.2 km) to the foot of Lower Cascade Lake. Park at the lakeside in the second parking area on the left, just past a sign FALLEN ROCK 1 1/2 MILE. Be very careful cutting across traffic: this is a busy high-speed highway. Recross the road on foot, again with great caution. Walk back toward the FALLEN ROCK sign, to a rough trail up the talus just short of the sign. The talus and cliff are both steep and full of loose rocks: be considerate of those below and behind.

The prominent, southeast-facing cliff we are climbing to is a ferrosyenite gneiss that crystallized from a melt prior to its metamorphism. It is one of a group of quartz-poor, alkali-feldspar- and ferroan-pyroxene-rich igneous rocks that acquired their gneissic fabric during an episode of isoclinal and recumbent folding associated with the Grenville orogeny at about 1100 M.Y. The persistent proximity and intimate intercalation of these gneisses with "Grenville" supracrustal rocks suggests that they may have initially been iron-rich felsic volcanics, or perhaps sills, interlayered with siliceous dolomites, calcareous quartzites, marls and basaltic flows comprising a Proterozoic series of rocks deposited about 1350 M.Y. This Grenville age is estimated from Ashwal's (1983) recent Sm^{147}-Nd^{143} date of 1288 M.Y. believed to represent the age of crystallization of the Marcy anorthosite massif-core facies. Following the model of McLelland and Isachsen (1980) for the southern Adirondacks, we suggest that, in the High Peaks Region of the northeastern Adirondacks, a "typical" Grenville supracrustal sequence correlative with rocks of the Central Metasedimentary Belt (Wynne-Edwards, 1972) was buried in a plate-tectonic event or events to a depth of about 70 km (42 mi), that of a doubly thickened crust. Following Emslie (1977), we envisage the birth of an anorthositic magma from the fractionation of copious amounts of orthopyroxene from an already-fractionated Al-rich gabbroic magma. Under these deep-seated conditions, the high pressures and temperatures plus the availability of Grenville-strata-derived CO_2-rich fluids initiated the formation of potassium- and iron-rich, relatively quartz-poor, melts of syenitic to monzonitic composition (Wendlandt, 1981). Ascent, intrusion, and emplacement of the syenitic melt at levels on the order of 25-35 km (15-21 mi) and temperatures of 800-900^{o} induced deep contact metamorphism of appropriate Grenville rock types. Here, at the easternmost part of the PO-2 outcrop, designated PO-2Gv (Fig. 3), a calc-silicate sequence infolded with the ferrosyenite contains the assemblage: wollastonite-diopside-grossular-quartz. Because anorthosite is absent, the contact-metamorphic origin of the wollastonite must be attributed to the intrusion of syenitic melt. Further, because the ferrosyenite contains relict inverted pigeonite, which now consists of host orthoferrosilite, 100Fe/(Fe+Mg)=85-92, the melt must have crystallized at temperatures of 850-900^{0} (Lindsley, 1983 and Ollila, Jaffe and Jaffe, 1988). Alternatively, shallow emplacement with

Figure 3. PO-2gv. Folded amphibolite, marble, and wollastonite-bearing calc-silicate rock in ferrosyenite gneiss. Pitchoff Mt. cliff, above north end of Lower Cascade Lake. Mt. Marcy quadrangle.



crystallization of fayalite and quartz, later deeply buried and converted to orthoferrosilite and quartz, is conceivable, yet unlikely, because no relict olivine, whatsoever, has been observed by Jaffe or by Ollila in quartz-bearing syenitic rocks of the Mt. Marcy and the Santanoni quadrangles. Fayalite (Fa_{95}) plus quartz, but with orthopyroxene absent, has been described from quartz-syenitic rocks in the Cranberry Lake quadrangle to the west (Buddington and Leonard, 1962, Jaffe et al. 1978) and in the Ausable Forks quadrangle to the northeast (Kemp and Alling, 1925 and Jaffe et al, 1978). A deep emplacement with high pressure crystallization is consistent with field observations and experimental data for all of these rocks.

At the PO-2 outcrop, we will split into several smaller groups: the footing can be a bit tricky. Remember not to step back for a better look at the outcrops. The first or westernmost cliff consists of strongly foliated ferrosyenite gneiss, N45E30W, with a large inclusion of shonkinite granulite (Fig. 4, Table 5). The foliation continues through the inclusion. The southwest end of the inclusion is sharply cut off by the gneiss but the northeast end fingers out. The inclusion is cut by a discordant vertical tongue of gneiss which becomes a subhorizontal pegmatite vein. At the northeast end of this cliff, the ferrosyenite gneiss is cut by an unfoliated aplite dike. Small amphibolite inclusions can be seen in the gneiss.

We will proceed cautiously about 300' (91.5 m) along the base of the cliff to the northeast, across a stream and a gully. Here we see several larger folded amphibolite inclusions in the ferrosyenite gneiss (Fig. 3). The axial planes of these folds are approximately parallel to the pervasive foliation. We will now crawl a few feet up the gully: in its east wall are exposed marbles and calc-silicates intimately infolded in the ferrosyenite gneiss. Wollastonite occurs in these calc-silicate beds (Fig. 3). If we make allowance for the plasticity of the marble, these folds are also approximately parallel to the pervasive foliation. There is a cave in the marble a little higher up this gully. On the opposite side of Lower Cascade Lake, in the anorthosite, another cave can be found a few hundred feet higher up. Caves are common in New York State, but these two must be among the few in Precambrian rocks.

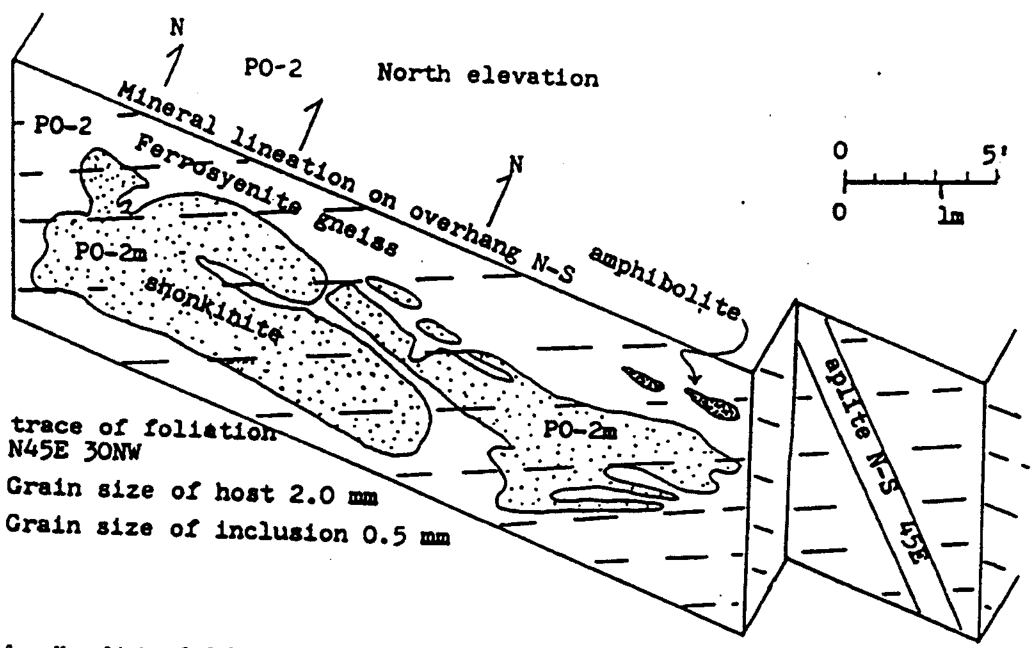


Figure 4. Xenolith of folded shonkinite layer PO-2m and amphibolite remnant in quartz ferrosyenite gneiss PO-2. Aplite dike cuts syenite gneiss. Pitchoff Mt. cliff above north end of Lower Cascade Lake, Mt. Marcy quadrangle.

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Stop 5. GRENVILLE MARBLE-SYENITE-ANORTHOSITE SECTION SOUTH OF KEENE

Drive five miles (8 km) northeast on Route 73 to the first right turn after a Gulf gasoline station and almost into Keene village center. Turn right (south) on the westernmost of two small roads that parallel both sides of the East Branch of the Ausable River. Drive about 0.75 (1.2 km) miles south on this western side of Hulls Falls Road and park judiciously along the edge of this little travelled road. Descend about 25' (7.6 m) to the bank of the river watching out to avoid standing or sitting in Rhus toxicodendron which commonly grows in Grenville marble terrain. A fine river outcrop of folded Grenville marble consists of calcite (white), diopside (green), fluopargasite (black), and minor glistening flakes of phlogopite (brown) along with less abundant graphite. Pink quartz leucosyenite and black-streaked gray-white gabbroic anorthosite gneiss have been dragged into highly contorted syntectonic folds enhanced by the plasticity of the marble and the probable molten state of the quartz syenite and gabbroic anorthosite (Fig. 5). Occasional tongues of gabbroic anorthosite cross-cut the syenitic rocks. A major vertical fracture zone, the Keene Fault Zone

runs parallel to the river in a N-S direction, and is well exposed about one-half mile (0.8 km) south in a granulated anorthosite outcrop. The Keene Fault has dragged the preexisting, gently north dipping, isoclinally folded strata into fairly steeply plunging folds at this locality. A late, brittle stage of movement on the same fault has granulated and retrograded all of the brittle rock types. Feldspar in syenite is sericitized, intermediate plagioclase in gabbroic anorthosite has been albitized and veined by calcite, grossular-diopside calc-silicate rocks have been prehnitized and chloritized, but marble merely goes along for the glide.

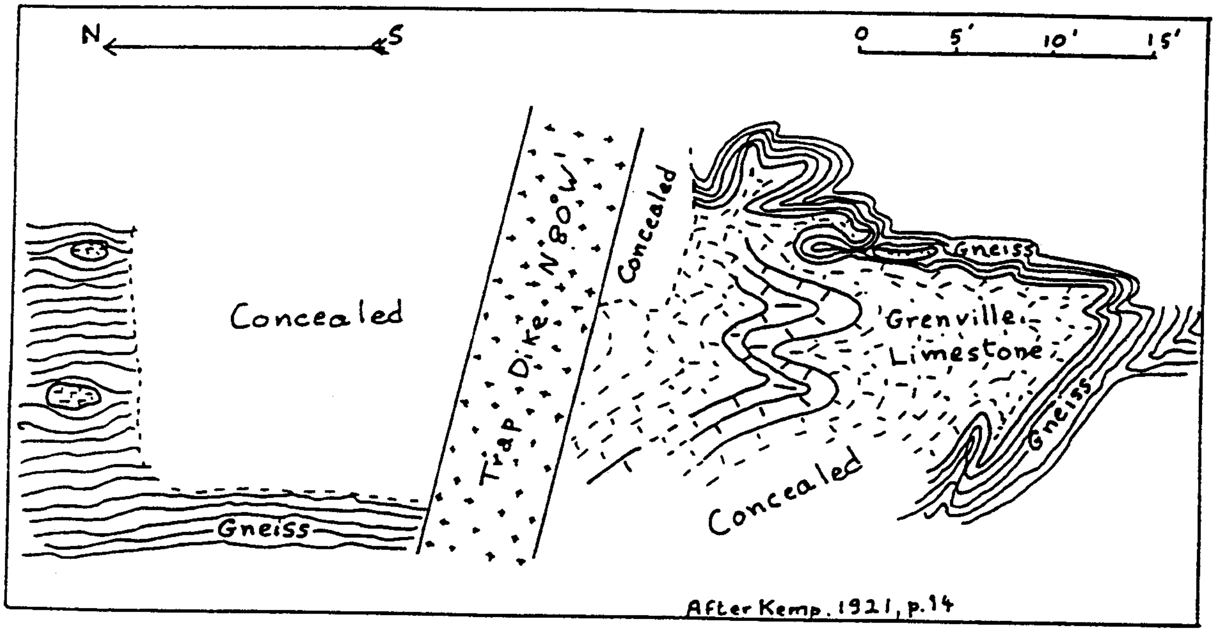

Figure 5. Contorted folding in diopside-calcite-pargasite-calcite marble, quartz leucosyenite gneiss, and anorthosite gneiss in the West Branch of the Ausable River south of Keene, N.Y. A camptonite dike cuts the marble-gneiss section. After Kemp, 1921. Mt. Marcy quadrangle.

At the northernmost end of the outcrop, the diopsidic marble and the quartz syenite are transected by a 4' (1.2 m) wide N80W90 trending lamprophyre dike (Fig. 5) which displays good chilled margins. It is a classic lamprophyre: a dark, dense, porphyritic dike rock in which the ferromagnesian minerals occur in two generations and in which only the dark minerals form the phenocrysts. It consists of 1-5 mm diameter phenocrysts of partially serpentinized magnesian olivine, and zoned clinopyroxene with augite cores and titanaugite rims, which display spectacular zoning, intense anomalous interference colors and dispersion, and hourglass structure. The groundmass contains a second generation of

microphenocrysts of titanaugite, kaersutite, titanian biotite and abundant very thin needles of apatite in a quasi-isotropic base that has too high an index of refraction to be analcime or leucite; it has a mean index of refraction = 1.525 and is either untwinned anorthoclase or a zeolite. The dike may be classified as either a camptonite or a monchiquite, but exactly conforms to neither.

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Stop 6 THE 1063 MYLONITE

On the west side of Route 73, 1.3 miles (2.1 km) south of the village of Keene Valley at BM 1063 and opposite the Beer Bridge across the Ausable River, the anorthosite is cut by a well-developed mylonitic zone about 2' (0.6 m) wide which trends N55E35NW. Park on the west shoulder of the road south of the outcrop and walk back. The anorthosite here appears to be the normal gabbroic type; however, the mafic minerals occur in clots and aggregates of augite+apatite, many of which are bent into mini- and microfolds. These clots and the sparse megacrysts of labradorite form a foliation which trends N70W70SW. The coarse grained contaminated felsic anorthosite and the mafic clots of partially assimilated augite+apatite represent either remnants of a Grenville phosphate-rich calc-silicate rock, or a mafic cumulate segregated from a gabbroic anorthosite melt. The iron content of the augite, 100Fe/(Fe+Mg) = 47, while too high for felsic anorthosite, is representative for anorthositic gabbro. Further, the profusion of "100" and "001" metamorphic pigeonite exsolution lamellae in the host augite (Jaffe et al., 1975) suggest that the clots may derive from anorthositic gabbro, where such are common, rather than from a Grenville calc-silicate lithology, where they are rare. Labradorite megacrysts in this rock are nevertheless higher in anorthite than felsic anorthosites of the Marcy region, and show $An_{50.5-54.5}$ rather than the typical An_{46-48}, suggesting a probable assimilation of calcium from the augite-apatite-rich clots of xenoliths. For this reason we classify such rocks as a percalcic subfacies of the gabbroic anorthosite facies.

Just north of a small waterfall, the outcrop changes dramatically: the rough foliation gives way to fine layering along which the dark minerals occur as streaks and schlieren, though occasional megacrysts have escaped granulation and appear as flaser. The mylonite is focused in a 2' (0.6 m) zone which dies out gradually to the north after about 20' (6 m) giving way again to percalcic anorthosite. Just beyond a covered interval, the north end of the outcrop contains a mafic rock, in rudely vertical attitude, but somewhat bent about a sub-horizontal axis, perhaps earlier than the mylonitization. It has been named "aproxite" by one of the authors, in allusion to its bimineralic apatite + pyroxene composition, which is identical with that comprising the mafic clots in anorthosite host rock at the south end of the outcrop. The mineralogy thus suggests that the "aproxite" is a folded layer in anorthosite rather than a mafic dike.

The mylonitic zone does not retain any of its primary magmatic or high grade metamorphic mineralogy but is totally retrograded to a fine-grained mixture of albite, prehnite, sericite, quartz, chlorite, pumpellyite, epidote, and calcite. Labradorite is altered by the following probable retrograde reaction:

$$4[Na_{0.5}Ca_{0.5}Al_{1.5}Si_{2.5}O_8] + KAlSi_3O_8 + 2H_2O$$

4 Lab + Or + Water

$$= 2[NaAlSi_3O_8] + Ca_2Al[AlSi_3O_{10}](OH)_2 + KAl_2[AlSi_3O_{10}](OH)_2 + SiO_2$$

= 2 Ab + Prehnite + Sericite + Qz

Augites have been drawn out into elongate lenses, spindles, and schlieren, and totally retrograded to a mixture of fibrous, isotropic chlorite and pumpellyite with a little calcite. Apatite, alone, remains unaltered, appearing as microflaser in the mylonitized base (Fig. 6).

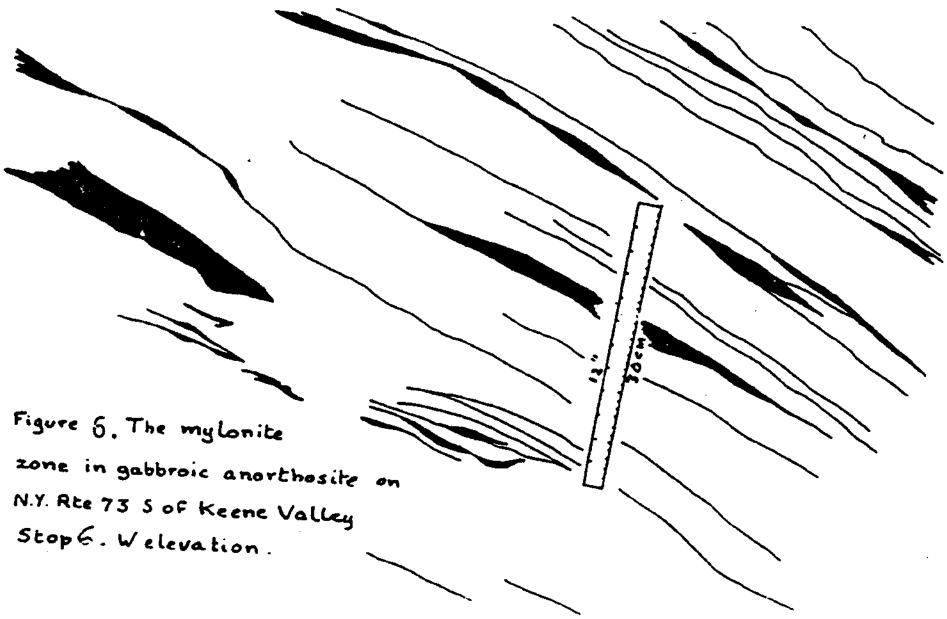


Figure 6. The mylonite zone in gabbroic anorthosite on N.Y. Rte 73 S of Keene Valley Stop 6. W elevation.

This wet assemblage is inconsistent with deep, ductile shear and suggests that the mylonitized zone originated by brittle shear or cataclasis in a wet, relatively shallow crustal setting.

That the shear zone was initially a deep, ductile mylonite, later retrograded, remains a possibility.

Stop 7. CHAPEL POND ANORTHOSITE, EAST CENTRAL MARGIN OF 15' MT. MARCY QUADRANGLE

Outcrops on the E side of Rte. 73 consist of andesine anorthosite containing shear zones of scapolite gneiss and a garnetiferous aplite dike (Fig. 7). For a detailed description of these outcrops, see Kelly (1974).

Most of the rock here is Marcy facies andesine anorthosite.

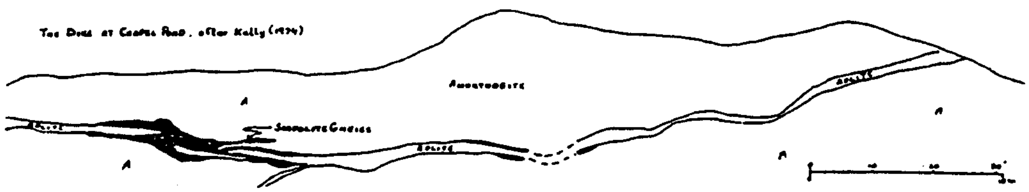


Figure 7. The dike at Chapel Pond, after Kelly (1974).

It consists of 30-40% dark blue-gray megacrysts of calcic andesine, An45-49, set in a matrix of white andesine, hypersthene, augite, and a little hornblende. Plagioclase makes up about 90% of the rock. A fine-grained aplite dike, up to 1 m thick, trends roughly parallel to the road where it may be seen to crosscut the foliation marked by aligned megacrysts of plagioclase in the anorthosite. The aplite contains about 60% microperthite, 25% quartz, 10% altered plagioclase, 3% magnetite, and 1% each of garnet and altered ferromagnesian silicates.

Along anorthosite-dike contacts, fracture zones in the anorthosite contain clear andesine, An44-47, abundant scapolite, Me36-47, dark olive hornblende, and a little hypersthene. Fluids carrying Cl and CO_2 apparently migrated into fractures in the anorthosite to convert plagioclase into scapolite.

The abundance of the plagioclase megacrysts here and the overall texture of the rock is typical of the anorthosite that forms the core of the Adirondack high peaks to the west.

Stop 8. GABBROIC ANORTHOSITE PROTOMYLONITIC GNEISS, SOUTH CENTRAL ELIZABETHTOWN QUADRANGLE

The prominent outcrop on the W side of Rte. 9 is a gabbroic anorthosite protomylonite or straight gneiss located in one of the prominent northeast-trending fault zones that abound in the NE Adirondacks. Note that the size reduction and mylonitization of this gabbroic anorthosite are not so intense as that seen at Stop 6, the 1063 mylonite.

The rock consists of very fine crenulations and streaks of hornblende, garnet, ilmenite, and augite in a fine matrix of white andesine, An32, and anorthoclase. Recrystallization took place under dry conditions, and all minerals are fresh.

The high strain nature of the gabbroic anorthosite gneiss is evident, and is illustrated by the total granulation and virtual absence of plagioclase megacrysts. An occasional block or xenolith of felsic anorthosite is present. Punky grey veins of finely altered rock occur in fracture zones.

31

Stop 9. MULTIPLY DEFORMED LAYERED GABBROIC ANORTHOSITE GNEISS, RTE. 9, 2 KM S OF ELIZABETHTOWN

The prominent cliff on the W side of Rte. 9 is a hydrothermally altered layered gabbroic anorthosite or leucogabbro located in one of the prominent northeast-trending fault zones that abound in the NE Adirondacks.

Figure 8. Refolded layered gabbroic anorthosite neiss.

The rock contains kaolinitized andesine, titanian brown hornblende, augite, and garnet, with a little ilmenite and apatite. Compositional layering is marked, with more mafic layers rich in augite and garnet, and more felsic layers richer in altered andesine and hornblende and without garnet. Many thin veinlets of white prehnite, calcite, and chlorite crosscut the gneiss. The layered gneiss was subsequently intruded by one or more aplite dikes, composed of altered albitic plagioclase, potassium feldspar, and quartz, along with garnet, and small amounts of apatite. Garnet is altered to chlorite. The rock is sliced on a mm scale, and slickensided, fracture surfaces often being coated with bright green pistacite.

The gneiss was recumbently folded under a high strain rate and then subjected to open folding (Fig. 8). Do the conspicuous open folds delineated by the pink aplite represent:

1) refolding of the recumbently folded gneiss?

2) a sheath fold squeezed or squirted up inside the gneiss?

3) are there one or two aplite dikes, and does the aplite layer close into a fold hinge beneath the road?

Just across Rte. 9 on the E side, the rock has a completely different texture. About half the rock consists of 0.5-5cm green, euhedral kaolinite pseudormorphs after plagioclase phenocrysts lying in a foliated matrix of brown hornblende, quartz, anorthoclase, augite, magnetite, and apatite. The fine black matrix is unaltered, in contrast to the large plagioclase phenocrysts. The green kaolinite pseudomorphs show good albite and Carlsbad twinning, but under the microscope only trace amounts of unaltered plagioclase remain. Compositionally, the rock is a quartz gabbro or quartz diorite. However, it may represent a porphyritic gabbro with its groundmass recrystallized to a metamorphic assemblage.

32

Stop 10. THE WOOLEN MILL GABBRO AND ANORTHOSITE, RTE. 9N, 15' ELIZABETHTOWN QUADRANGLE

On Rte. 9N about 1.6 km W of the intersection of Rtes. 9 and 9N at the golf course on the southern edge of Elizabethtown are outcrops of gabbro and anorthosite at the site of an old, long disused broken dam. Recent reconstruction of the dam and installation of a penstock along the Branch River just N of the road may limit our examination of a fine exposure of anorthosite block structure along the river.

On the S side of the road, a prominent cliff exposes the very irregular contact of a garnet- and magnetite-rich gabbro with a white gabbroic anorthosite in which almost all of the plagioclase megacrysts have been granulated. A rude foliation may be observed in both rocks. Modes of the gabbro and optically determined compositions of plagioclase, augite and hypersthene from the gabbro and anorthosite are given in Table 6. The contrast in Fs content of coexisting clino- and orthopyroxene in gabbro and anorthosite is marked. Note that the iron-rich gabbro has abundant garnet, while the relatively iron-poor anorthosite has none. The presence in the gabbro of isolated blue-gray, well-twinned (Carlsbad and albite) labradorite xenocrysts and occasional xenoliths, as well as contact relations, suggests that the gabbro has intruded the anorthosite before regional deformation. Toward the center of the roadcut, a pink aplite dike has intruded both the anorthosite and the gabbro.

The complexity of the contact relations here has led different geologists to different interpretations. Kemp and Ruedemann (1910) and Kemp (1921) reported that river outcrops showed "anorthosite tonguing in to the dark supposed gabbro" and suggested that the gabbros might represent "surviving inclusions of Grenville sedimentary gneisses impregnated with matter from the anorthosites." They suggested that the Woolen Mill gabbro was thus an old Grenville rock hybridized by later intrusion of anorthosite. Commenting on this interpretation, Miller (1919) assigned these plagioclase-megacryst-bearing gabbros to the Keene gneiss, which he believed to be a syenitic magma containing assimilated anorthosite! Buddington (1962) described the Woolen Mill gabbro as a typical olivine gabbro intrusive into anorthosite, in which olivine was extensively converted to garnet during regional metamorphic recrystallization. No evidence of relict olivine could be found under the microscope.

How would you interpret the outcrop relations?

Table 6. MODES AND MINERAL COMPOSITION OF GABBRO AND ANORTHOSITE FROM THE WOOLEN MILL LOCALITY, RTE 9N, 1.6 KM W OF ELIZABETHTOWN

<u>Mineral</u>	<u>metamorphosed gabbro</u> <u>% by volume</u>		<u>anorthosite</u>
orthoclase	0.5--------0.5		tr
andesine	43.0-------50.1	$An_{33,5}$ 1/	andesine $An_{44.5}$ 2/
augite	12.9-------16.4	Fs_{48} 3/	augite Fs_{29} 4/
hypersthene	4.5--------4.1	Fs_{60} 5/	hypersthene $Fs_{40.5}$ 6/
almandine	9.9-------10.2		none
ilmenite	8.2--------2.7		tr
magnetite	15.0-------10.0		none
apatite	6.0--------6.0		none
	100.0 100.0		

<u>1/</u> mol % An from optical meas., $\alpha = 1.546$

<u>2/</u> " " " " " " $\alpha = 1.5520$

<u>3/</u> mol % Fs, Fe/(Fe+Mg) from meas. of $\gamma = 1.730$

<u>4/</u> " " " " " " " " $\gamma = 1.717$

<u>5/</u> " " " " " " " " $\gamma = 1.740$

<u>6/</u> " " " " " " " " $\gamma = 1.715$

A plagioclase xenocryst in gabbro has $\alpha = 1.555$, $An_{50.5}$

For optical composition curves, see Jaffe, Robinson, Tracy and Ross (1975). Amer. Mineral. 60, 9-28.

Brant Lake Area

Turner, NYSGA, 1979

MILE - FIRST LEG

0.0 Field trip starts at end of ramp of northbound Exit 24, Interstate 87. Turn right toward Bolton Landing. Within 50 meters turn right again and head south on the River Road.

33 0.5 STOP #1 - Outcrops on right (west) side of road are largely alaskitic granites with some strongly contorted layers of paragneiss. Although the outcrop is just outside the leader's field area, it was mapped at his suggestion by McConnell in 1964 and is deemed to be "basement". A two-meter exposure of an isoclinally refolded isoclinal fold is one of the main features of this stop. The first-formed fold has axial plane foliation and is believed to be an F-1 fold. The second-formed fold deforms foliation and is believed to be an F-3 fold. The other important feature of this exposure is the strong suggestion that the granite is an anatectic product of the paragneisses.

Turn cars around, head back to I87 and continue north to Exit 25. RESET mileage for the second leg of the field trip which begins at the end of the ramp for northbound Exit 25.

MILE - SECOND LEG

0.00 End of ramp, Exit 25. Turn right onto N.Y. Route 8 and proceed eastward through the hamlet of Brant Lake and thence along the lake of the same name (Bolton Landing 15' quadrangle).

7.3 Turn left onto Palisades Road. It is difficult to see this turn until you are practically on top of it. Go 1.3 miles to the first stop of this leg.

34 8.7 STOP #1 - At the T-intersection with the Beaver Pond Road, an outcrop of Older Paragneiss (informal stratigraphic name) is located on the south side of the road. The outcrop contains a heterogeneous group of gneisses, quartzites and amphibolites. Of particular interest are the rotated (penetrative) clots of sillimanite and minor folds whose axes plunge about 20° in an azimuthal direction of 60°-65°, which is the local axis of F-3 folding. This unit is believed to lie unconformably below the base of the supracrustal metasedimentary rocks, and in the opposite stratigraphic direction grades into granitic gneisses. A whole-rock, Rb-Sr age of 1210± 45 m.y. has been obtained from this outcrop.

Turn around and head back to N.Y. Route 8; turn left on Rt. 8.

35 10.3 STOP #2 - Brant Lake Gneiss (informal name) is exposed on the south side of the road. This rock is found throughout a structural dome to the south of this exposure. This granitic gneiss has a very uniform modal composition of approximately 35% quartz, 25% microcline, 30% sodic plagioclase, 3% mesoperthite, 5-7% biotite and 1% opaques. The granitic rocks of this dome yield a whole-rock, Rb-Sr age of 1119±39 m.y. It has been proposed that this granitic gneiss is an anatectic product of the Older Paragneiss (Bickford and Turner, 1971). This is a very brief stop.

36 10.7 OPTIONAL STOP - (1979 NEIGC-NYSGA will not stop here) - Older paragneiss in the road cut on the north side of Rt. 8 consists of quartzofeldspathic gneisses with accessory biotite and garnet. Most, if not all, minor isoclinal folds in this exposure show axial planar foliation. This paragneiss mantles the structural dome whose granitic core rock was observed at Stop #2, and may be traced almost continuously around the dome (see Plate 2 of accompanying description).

37 14.7 STOP #3 - In the road cut on the east side of the road is perhaps the simplest set of folds in the Swede Mountain structural complex. A large isoclinal synform is outlined by the contrast between quartzite and sillimanitic schists. Sillimanite needles just above the quartzite closure plunge about 5° in an azimuthal direction of 80°. Closer to the road, a minor fold crenulation plunges 20° in an azimuthal direction of 80°. The

synform is thought to be an F-3 fold. The isoclinal synform is clearly refolded. A stereographic beta diagram of attitudes of compositional layers shows an intersection which plunges 22° in an azimuthal direction of 116°, which comports with the axis of F-4 folding. About 15 meters uphill along the road cut, a pair of refolded isoclinal folds about 3 meters long may be observed. Measurements of hinges and crenulations in these produce the same pair of data as for the large refolded synform.

38 15.4 STOP #4 - Pull into turn-out area on the right and park. Walk back across Rt. 8 and continue east for about 50 meters. On the south side of the highway is a low road cut about 70 meters long. An isoclinal fold, whose limbs may be traced 55 meters to its closure, is exposed. Crenulations in the hinge plunge about 10° in the azimuthal direction of 103°. At least one minor isoclinal fold within the larger isocline displays axial plane foliation. This is probably an F-1 fold. In the eastern part of the road cut, the limbs of the isocline have been refolded by an F-4 fold. A beta diagram of the refold shows an axis plunging 30° in the azimuthal direction of 117°.

Returning to the western end of the road cut, and at about right angles to the cut, the outcrop portion of the exposure contains evidence of three episodes of folding (see Figures 1,2 and 3 of accompanying description). A complex F-1 isocline has been isoclinally refolded into an F-? synform, and several smaller F-4 folds have been superimposed on the refolded mass. Axial plane foliation in the F-1 fold has been rotated by F-?, and a weak F-4 foliation with strong quartz rodding penetrates the outcrop. A beta diagram of compositional layers in the nose of the synform shows an intersection plunging about 5° in an azimuthal direction of 110°. This does not correspond with an F-3 axis, and may represent an F-2 refold. A beta diagram of the limbs of the refold shows an intersection plunging 15-20° in an azimuthal direction of about 120°, which corresponds with F-4 fold axes. Although the beta maxima are only 10° apart in azimuth, the difference is believed to be real because measured hinge axes correspond with the 110° beta direction and measured quartz rods with the 120° beta direction.

Return to vehicles, continue around turn-out loop back to Rt. 8, turn left and proceed back in a westerly direction.

16.0 Pull into turn-out area on left. At this point the group may wish to split into two parties. Assuming that time permits, the leader will take a group of physically able participants on a 4-kilometer hike (total distance) to examine evidence

OUTCROP - STOP #4 - SOUTH SIDE OF N.Y. ROUTE 8, NEAR NORTH POND, BOLTON LANDING 15' QUADRANGLE - THREE FOLD SETS.

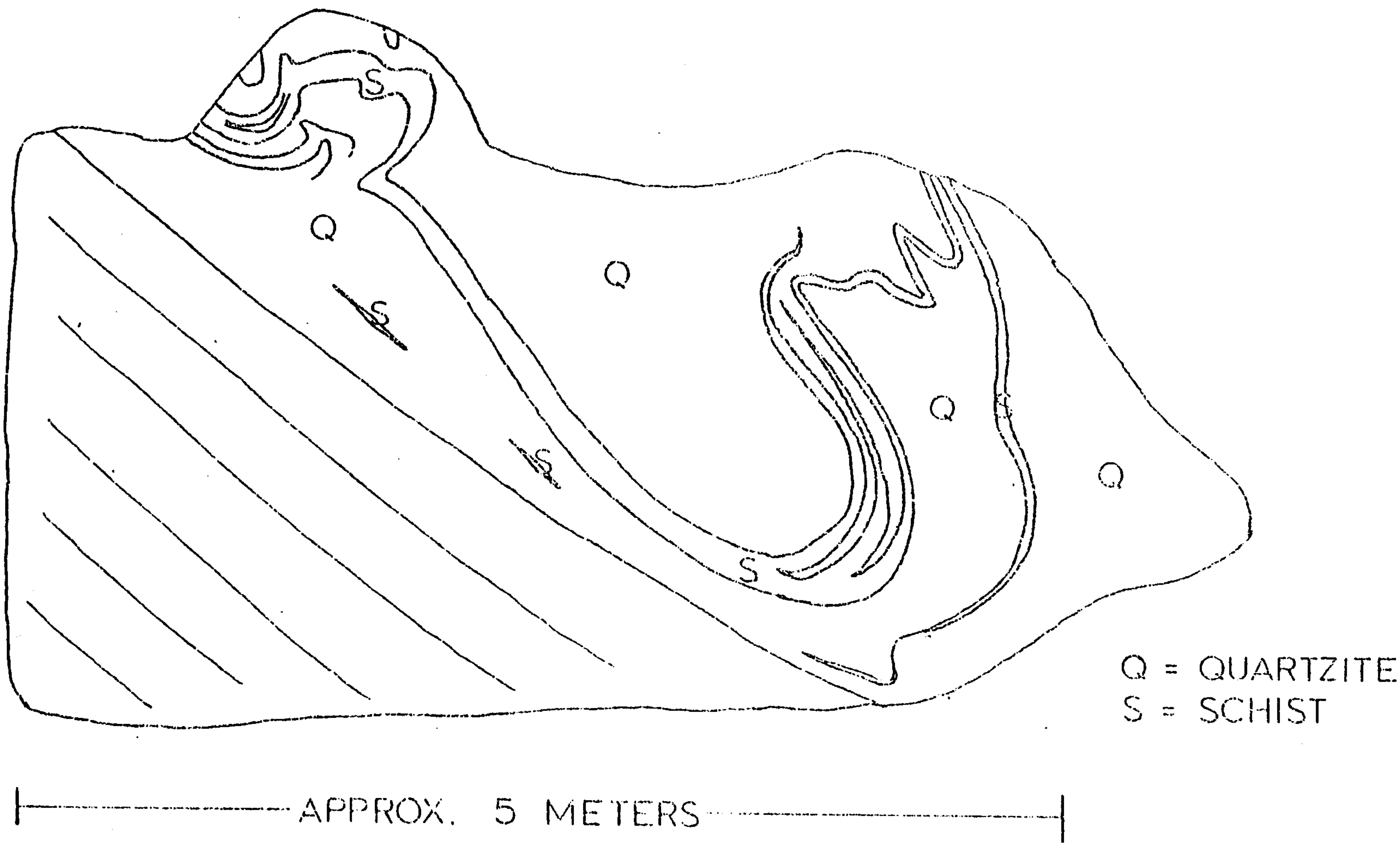
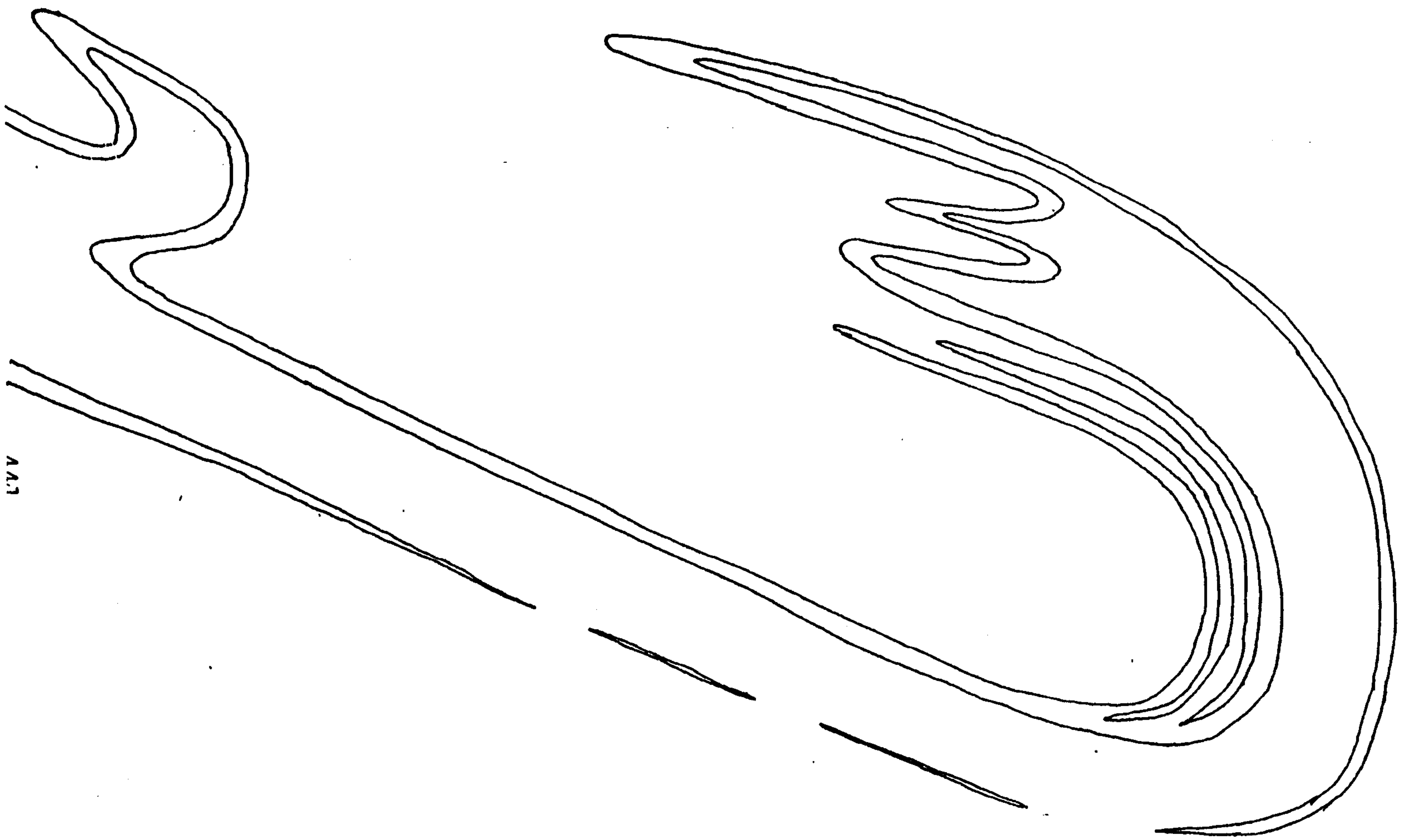
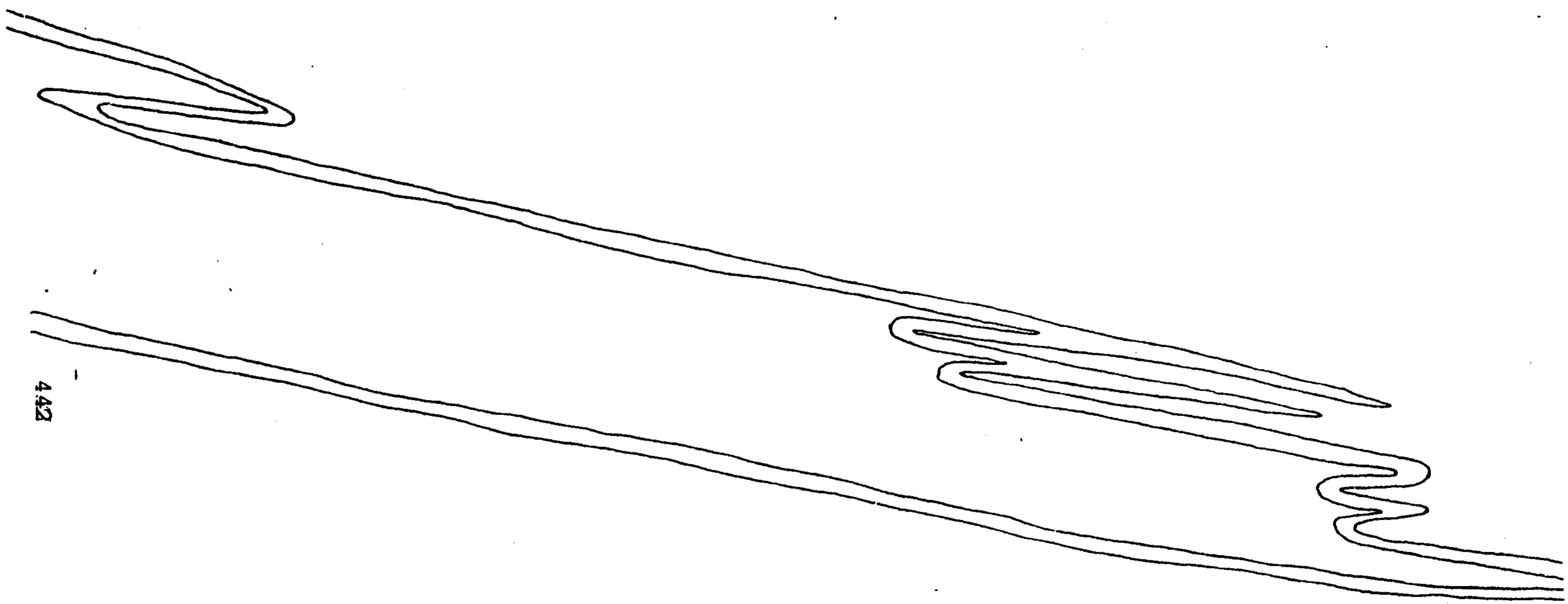


FIGURE 1



PROJECTED CONFIGURATION OF FIG. 1 IF LAST FOLDS (F-4) WERE REMOVED. F-2 (?) SYNFORM IMPOSED ON F-1 ISOCLINE.

FIGURE 2



PROJECTED CONFIGURATION OF FIGURE 1 IF BOTH REFOLD SETS (F-4 and F-2 (?)) WERE REMOVED.

F-1 ISOCLINE REMAINS.

FIGURE 3

of large nappe structure in Swede Mountain. Those persons not wishing to take the hike are invited to examine the 1.0 kilometer of nearly continuous exposure of Swede Mountain beside Rt. 8, which includes the structures seen at Stop #3.

After returning from the hike, if time still permits, an attempt will be made to round up all participants for a final but optional stop. Proceed westerly on Rt. 8 toward I87.

28.3 Within seeing distance of I87, turn right onto the Starbuckville Road, continue until a long one-lane bridge is crossed and look for an intersection.

29.0 Turn left onto the River Road and go 0.25 miles.

38a 29.25 STOP #6 (optional) - You are in the northeast quadrant of the North Creek 15' quadrangle, mapped by Geraghty (1973). The outcrops on the east side of the road are predominantly quartzofeldspathic schists and gneisses with abundant biotite, sillimanite and garnet (kinzigite) and quartzite. The author believes the kinzigite is an iron-rich facies of the basal khondalite seen in Swede Mountain. Some minor folds display axial plane foliation and may be F-1 folds. Numerous axial traces of minor folds strike 60°-70° in azimuth, a biotite-sillimanite crenulation shows plunge of about 42° in an azimuthal direction of 59°, a beta diagram of one fold shows an intersection at plunge 26° in an azimuthal direction of 53°, and one minor fold hinge plunges 15° in an azimuthal direction of 70°. The foregoing reflect F-3 folding. Several F-4 cross folds are present and measured lineations and hinges have a range of plunges of 45° to 55° in azimuthal directions of 115° to 120°. At least one F-5 fold is apparent in the cutcrop, and its axial trace is about 10° in azimuthal direction.

GOOD LUCK ON YOUR TRIP HOME!

DAY 3

SUNDAY, MAY 8

TICONDEROGA - SARATOGA SPRINGS - NEWARK

STOPS 42 - 84

SOUTHEASTERN ADIRONDACKS

LAURENTIAN PLATFORM SEDIMENTARY ROCKS

PASSIVE MARGIN CARBONATES AND STROMATOLITES

TACONIC ACCRETIONARY PRISM AND THRUST FAULTS

Brant Lake – Ticonderoga

174.9 Jct. of Rtes 30 and 8; turn L (E) on Route 8.
198.6 Jct. of Rtes. 8 and 28 in Weverton; Continue straight ahead on 8.
202.5 Crossing Hudson River at Riparius.
209.7 Cross under Northway (US 87) and proceed E on Rte. 8 through Brant Lake Village and along S shore of Brant Lake.
221.1 Parking area on R opposite large cut in migmatitic biotite-garnet-quartz-plagioclase gneiss.

222.5 Parking area on R near Swede Mtn. Pond. Park here for Stop 32.

39 STOP 32. Graphitic "Dixon" schist
Cautiously cross road and walk a short distance downhill to roadcuts on N side. Alling (1917) mapped the graphite deposits of the eastern Adirondacks, proposed stratigraphic sequences, and assigned unofficial formational names. The unit here exposed is his Dixon Schist which was, in 1917, the greatest source of flake graphite in the Adirondacks, and "possibly in the United States" (Alling, 1917, p. 43). The formation is a quartz schist that contains about 5 to 7 percent graphite, and small amounts of biotite and pyrite. Its thickness ranges from about 1 to 7 m. Immediately overlying the Dixon Schist is the Faxon "limestone" (= marble), and above that the Swede Pond quartzite, atop Swede Mountain to the south of the road. If time permits, we will walk the short trail to the top of Swede Mountain to examine this unit.

Carbon isotopic analysis demonstrates an organic, syngenetic origin for much of the Adirondack graphite. One graphite, in silicious schist at Faxon Pond, is clearly organic ($\delta^{13}C = -27.3$) while other graphites, 2-4 m away across strike, are in exchange equilibrium with adjacent marbles ($\delta^{13}C(Gr) = -4.2$ to -5.2, $\delta^{13}C(cc) = -0.7$, Weis and others, 1981). The results are consistent with other data showing that in Adirondack marbles, graphite and calcite attain carbon isotopic equilibrium (Valley and O'Neil, 1981; Valley, 1986). Such exchange raises $\delta^{13}C$ of graphite and decreases that of calcite so that low values (<-20 permil), indicative of organic origin, are preserved only in carbonate-poor lithologies. The sharp gradient in $\delta^{13}C$ at Faxon Pond (25.1 permil/2m) demonstrates that carbon exchange is limited to minerals in close proximity, suggesting fluid-absent diffusion as a mechanism. Clearly, carbon mobility across strike of compositional layering has been very limited.

5.7 The low area to the L is a small graben floored by Cambrian Potsdam sandstone.

7.1 40 STOP G. Charnockitic "pencil gneiss"
The roadcut on the R and the smaller outcrop near barn on L are a coarse-grained charnockitic gneiss. This rock yields a U/Pb zircon age of 1113±10 Ma (Silver, 1969). The rock has a strong lineation defined by quartz rods ("pencils"); these are oriented N10°-20°E, which is the dominant lineation direction in the northeastern Adirondacks and is nearly perpendicular to the ESE lineations typical of much of the southern Adirondacks (e.g. Stop 29).

9.1 Traffic circle around monument in Village of Ticonderoga; continue N on 9N.
9.8 Jct. of Rtes. 9N and 22; turn R(S) on Route 22.
12.7 Lake Champlain on L.
14.7 Entering Warren Co.

Whitney et al., 28th IGC, 1989

18.3 41 Stop 33 Paleozoic/Proterozoic Unconformity This roadcut exposes approximately 3 meters of the basal Ausable Member (medial Cambrian?) of the Potsdam Sandstone, resting unconformably on Proterozoic gneisses and calcsilicates. The Ausable Member here consists of green-gray, matrix-rich arkosic conglomerates and sandstones. Thickness patterns and paleocurrent data suggest that deposition of the sandstones took place in localized basins bounded by NE-trending faults. Petrographically, the rock consists of angular quartz and microcline grains set in a matrix of green chlorite. The chlorite is in part secondary after detrital hornblende and biotite. Additional chlorite occurs as a cement, intergrown with authigenic quartz. Detrital accessory minerals include rutile, zircon, monazite, xenotime, and chlorite, all derived from nearby pegmatites in the Proterozoic basement. Intense diagenetic/hydrothermal alteration of the detrital suite has occurred, as indicated by the chloritization of the ferromagnesian minerals, and the alteration of magnetite and ilmenite to anatase and pyrite. Authigenic overgrowths of xenotime on detrital zircon and monazite on detrital monazite also occur.
Continue S on Rte. 22.

19.2 Parking area on L; park here for Stop 34.

NY 22, Whitehall area 126

(42) STOP 34. Isoclinal folding in metasedimentary rocks and granitic gneisses . Cautiously cross road to cuts

19.2 on W side. Here are numerous isoclinal folds with steeply dipping axial planes and approximately horizontal, E-W oriented axes. Steep axial planes are uncommon in isoclinal folds in the Adirondacks; this outcrop affords a rare look at such folds in cross section. Note that the folds appear to have been flattened. This outcrop is discussed in some detail by Granath and Barstow (1980).

25.4 (43) STOP 35. Thrust fault; metagabbro; xenolith. This long roadcut, on both sides of Rte. 22 near Dresden Station, exhibits several different rock types including (from south to north) garnetiferous ferrogabbro, assorted metasedimentary rocks, olivine metagabbro and garnet-sillimanite-quartz-feldspar ± biotite gneiss ("khondalite") interlayered with marble. All rocks dip steeply N. Approaching its contact with the metasedimentary rocks the ferrogabbro becomes first grain-size reduced and then mylonitic. This contact has been interpreted as the sole of a large, folded thrust sheet that places charnockite and minor ferrogabbro over metasedimentary rocks. The rocks beneath the thrust display isoclinal folding which may be pre- or syn-thrusting.

Most of the metasedimentary rocks in the outcrop consist of impure marbles and khondalite; near the northern end of the cut on the E side the latter contains a rusty, sulfidic zone resembling the Dixon Schist (Alling, 1917; see Stop 32). At the southeastern end of the roadcut an olivine metagabbro is exposed in contact with the khondalite. The metagabbro, which is locally lineated and contains granulite facies garnet coronas, can be seen boudinaged (or isoclinally folded ?) on the W side of the road. On top of the outcrop, at the SE end, another contact is exposed that shows metagabbro truncating a strong foliation in the khondalite, which here appears to be a large, incompletely exposed xenolith in the metagabbro. The foliation in the khondalite is defined by flattened lenses containing sillimanite; sillimanite is also present as inclusions in large (up to 1 cm) rolled garnets, some of which are truncated by the contact. The foliation and metamorphic minerals thus clearly predate the intrusion of the gabbro. The significance of these relationships is that at least two episodes of deformation, and two metamorphisms, have affected the rocks at this locality. Further details are given by McLelland and others (1988).

The metagabbro at this stop has yielded a U/Pb zircon age of 1144±7 Ma. This rock is indistinguishable from other olivine metagabbros in the eastern highlands and the high peaks region, where they intrude anorthosite. If these metagabbros are all of the same age, then the 1144 Ma date is a minimum emplacement age for the anorthosite.

Continue S on Rte. 22.

29.1 Road to Clemmons on L, continue on 22.

29.5 Road to Hullett's Landing on R; continue on 22.

32.7 (44) Stop 36 Metatonalite. The roadcut on the L (E), directly opposite the junction with old Rte. 22, exposes a gneiss consisting of andesine, quartz, orthopyroxene and hornblende, whose whole-rock chemistry (Col.R, Table 1) corresponds to tonalite. Texturally the rock is homogeneous and massive except for amphibolite interlayers interpreted as disrupted mafic dikes. There is little evidence for metamorphic flow of the host rock between the amphibolite boudins, suggesting that intrusion and disruption of the dikes occurred either prior to complete solidification of the tonalite, or that the tonalite became partially melted during later metamorphism and deformation.

Rocks of tonalitic composition are found in several locations in the southern and southeastern Adirondacks, but are absent elsewhere in the highlands. Samples from this outcrop yield a U/Pb zircon age of 1320±60 Ma, which is close to the ~1300 Ma age of a similar metatonalite from near Canada Lake (see discussion of geochronology). In any case, these rocks are significantly older than the anorthosite and mangerite-charnockite suites. They may represent early orogenic magmatism within the proto-Adirondacks, or arc magmatism in a separate terrane later accreted to the proto-Adirondacks.

34.4 Bridge across South Bay of Lake Champlain.

35.1 Turn L into abandoned quarry near town highway garage. Get permission from landowner.

NY 22: WHITEHALL TO FORT ANN
ROAD LOG

Modified from Whitney, NYSGA, 1985

Interval
mileage

0.0 South Bay on Lake Champlain. The rocks underlying the valley to the north are Paleozoic strata downdropped along a major normal fault which here forms the western side of the Pinnacle Range. To the south, this fault intersects the Welch Hollow fault at an oblique angle. The fault follows the shore of South Bay south of the bridge, then strikes inland and follows the line of cliffs visible to the north. Estimated vertical displacement on this fault, based on offset of Paleozoic cover rocks, is in the vicinity of 300 m. West of the bay, outcrops of gently E-dipping, highly deformed Precambrian rocks resume.

0.2 Farmhouse of owner of property at Stop ; ask permission.

0.2 **STOP 45**. Entrance to abandoned quarry and Washington County Highway Department garage. This is posted **private property**; ask permission at the large brick farmhouse 0.2 miles back up the road on L. **Be extremely careful climbing and hammering here - there is much loose rock and the quartzite is very splintery when hammered.**

This thick unit of quartzite is interlayered with lesser amounts of a greenish rock that forms bands and streaks from a fraction of a millimeter to several tens of centimeters thick, with knife-sharp contacts against the quartzite.

The quartzite, which is visibly foliated in hand sample. comprises over 95% strongly flattened quartz, flattened and elongated grains of K-feldspar and sericitized plagioclase, lensoid garnets, green biotite, and chlorite. The interlayered green rock ranges from plagioclase-quartz-garnet-biotite-hypersthene gneiss to a retrograded epidote-chlorite-quartz-plagioclase rock with some muscovite and at least two minerals not yet identified. Some well-crystallized chlorite is present as flakes parallel to the foliation, but chlorite also occurs locally as an alteration product of garnet. Distribution of the retrograde assemblage within the outcrop has not been determined. If it is related to fracture patterns, it may be low-temperature alteration along the E-W brittle fault that parallels the road. If not, it may be evidence for localized retrogression associated with renewed, layer-parallel shearing during the latest Proterozoic or during the Taconic event. Supporting the latter hypothesis are slickensides on foliation surfaces at an acute angle to the lineation.

All rocks at this site show extreme foliation and a well-developed lineation, here close to E-W, with a 0-20° plunge. Numerous minor folds are present. These are of two distinct types, both of which are recumbent with axes parallel to the lineation. One type consists of intrafolial, highly asymmetric, isoclinal folds defined by thin micaceous layers in the quartzite. Among folds of this type is an apparent sheath fold strongly flattened in the plane of foliation. The other type is not quite isoclinal and more symmetric, and it visibly folds the foliation in the quartzite. The minor folds and petrofabrics at this outcrop have been described by Granath and Barstow (1980), who attribute the deformation primarily to severe flattening strain.

0.5

South end of series of cuts in highly fractured metasedimentary rocks.

0.7

STOP 46. Pull off the road close to the smaller outcrop. The rocks here are typical metapelites, consisting of quartz, K-feldspar, sillimanite, lavender garnet and varying amounts of biotite and graphite.

0.8

Junction NY 22 and US 4 in Whitehall; continue S on NY 22

0.6

Whitehall village line on south side of town

1.9

Flat outcrops on slopes to the R are a dipslope on foliation in highly strained gneisses. A short distance S, on West Mtn., a mylonite zone close to 300 m thick is exposed. The hills across the valley to the L, and on Skene Mtn. straight ahead, are Cambro-Ordovician carbonates of the Whitehall Formation, resting on Potsdam sandstone.

1.4

STOP 47 Park as far off the road to the L as possible. Examine briefly the outcrops on the R. These are typical Adirondack olivine metagabbros, with well preserved igneous textures as well as coronitic reaction rims around olivine and ilmenite (see introductory section). The interiors of the olivine coronas here have been retrograded to chlorite and carbonate; otherwise the rocks are quite fresh.

1.5

Whitehall town line

0.2

STOP 48 North end of long roadcut. Pull off road on right, and begin examination at N end of the outcrop and work south. Probably the best way to approach these rocks is to move rather rapidly to the S end, scanning the rocks on both sides as you go for major lithologic changes; and then work your way back N more slowly, looking at the rocks in detail.

The sequence of rock types going S on the W side is as follows:

A interlayered (or interleaved ?) marbles and paragneisses

B garnetiferous, quartzo-feldspathic gneisses intruded by unmetamorphosed mafic dikes

\- gap -

C charnockitic gneiss

D thin marble with numerous exotic blocks

E mafic gneiss

F calc-silicate and marble

G interlayered (interleaved ?) paragneiss, charnockite, marble and calc-silicate with amphibolite boudins. An unmetamorphosed mafic dike forms the face of much of this section of the cut.

DETAILS WALKING NORTH:

G Note the wide variety of rock types, including charnockitic gneisses, amphibolites, marble (carefully examine the marble-amphibolite contact), lineated sillimanite-bearing metapelites, and calc-silicates. Note the local slickensides along foliation surfaces, as well as on vertical fractures. Work out the sense or shear using the steps on the slickensides. **Has there been late movement parallel to the foliation? When do you think that movement could have occurred? What could it have been related to?**

F This thin calcareous unit consists of marble near the base and a complex calc-silicate zone adjacent to the contact with the overlying mafic gneiss. Major minerals in the calc-silicate zone are grossularite, diopside, quartz, calcite and K-feldspar, with lesser amounts of plagioclase and chlorite as well as several minor phases yet to be identified. The calc-silicates probably originated by contact metamorphism at the time of intrusion of the igneous precursor of the overlying mafic gneiss. This contact is irregular and appears to have been folded. The thickness of the calc-silicate layer varies widely, both in this outcrop and on the opposite side of the road.

E This mafic gneiss contains plagioclase, clinopyroxene, hornblende, biotite, garnet and minor quartz and K-feldspar. The composition is probably similar to a monzodiorite. Similar rocks elsewhere in the Adirondacks have been called jotunite, and are associated with the anorthosite suite of rocks. The rock is well foliated throughout, but becomes more so toward the sharp upper contact. On the east side of the road, a detached sliver of the mafic gneiss is found in the overlying marble, and contains carbonate-filled fractures.

D The next unit upward is a thin (usually < 1 m) band of marble with numerous rotated fragments of other rocks. No calc-silicates are developed near the sharp contact with the mafic gneiss beneath, and the foliations in both mafic gneiss and overlying charnockitic gneiss are strongly developed and parallel to the marble band. In the outcrop on the E aside of the road, foliation in the charnockite is locally truncated by the marble. This marble is a good example of a possible detachment zone between the mafic gneiss and the charnockite, with relative movement of uncertain direction and magnitude. The absence of a contact-metamorphic zone of calc-silicates by at the mafic gneiss contact may result from its having sheared off during displacement. Alternatively, this marble may be a tectonically emplaced younger rock (see discussion under **A** below). Considerable displacement may have taken place along most or all of the marble layers in this outcrop.

C Above the marble is a thick unit of charnockitic gneiss. This rock, close to granite in composition, consists of quartz, microcline, plagioclase, hornblende, garnet, clinopyroxene, and orthopyroxene. The orthopyroxene is extensively chloritized, which is characteristic of many Adirondack charnockites. The typical green color is well developed toward the center of the unit. Near the northern end of the outcrop both green and white varieties are present, with diffuse color boundaries which crosscut foliation. Immediately beyond the charnockite unit is a gap in the outcrop, possibly indicating the presence of a fault or thick marble layer.

B Following the gap is a short section of well foliated, garnetiferous quartzofeldspathic gneisses similar to the charnockite but with green color less well developed. Note the unmetamorphosed mafic dike just back from the face of the outcrop, and roughly parallel to it. A few meters farther N is a complex vertical fault with a zone of carbonate-cemented breccia.

A The last section of the outcrop, roughly 100 m long, consists of interlayered (interleaved?) paragneiss and marble, with minor amphibolite and thin calc-silicate bands in the paragneiss. Contact surfaces are frequently slickensided and/or coated with graphite. At least two types of marble are present; one is dark, relatively fine-grained, brown-weathering dolomite marble, which has a slightly fetid odor when struck with a hammer; the other is coarser-grained, has a somewhat lighter color and considerable calcite as well as dolomite. Both marbles contain abundant rounded to angular silicate rock and mineral fragments, including quartz, feldspar and serpentine, and larger rotated blocks of various rock types including amphibolites, serpentinite, paragneiss and calc-silicate granulite. Quartz, dolomite and serpentine coexist in these rocks with no evidence of mutual reaction, indicating that the rock as presently constituted has never undergone high temperature metamorphism. Temperatures must have been sufficiently low to prevent reaction of quartz with either dolomite or serpentine. It is probable that these marble zones, as well as those of units G, F, and D, are tectonic breccias formed along thrust faults or low-angle normal faults under conditions that permitted the carbonates to recrystallize and deform in ductile fashion, while silicates behaved in a more brittle manner. The interleaved paragneisses, by contrast, are similar to the gray gneisses seen in previous stops, have a high-T metamorphic assemblage and show little evidence of retrogression.

The age of the tectonic interleaving of the gneisses and marbles may be either late Proterozoic or Taconic. The marbles themselves may be Proterozoic with retrograde serpentine after forsterite and entrained fragments of quartz and feldspar, or they may be Paleozoic mafic rocks. This question is now under study and will be discussed on the outcrop.

0.2

South end of the Stop 7 outcrop.

0.4

STOP 49. (optional) Pull onto shoulder, **cautiously** cross the roadway, and briefly examine the outcrops.

The rock here is a pale gray, biotite-quartz-two feldspar-garnet-sillimanite-graphite paragneiss with thin layers and lenses of calc-silicate. Compared to previously examined "kinzigites", this rock is finer-grained, more aluminous, and has distinctive lavender garnets. The abundant white layers look like leucosomes in a migmatite, but they contain significant amounts of sillimanite and are thus probably more aluminous than minimum-melt granite. Note the flattening of quartz in these layers. Look carefully for lineations defined by sillimanite and quartz.

0.3

STOP 50 N end of next major road cut. Walk S along the W side. At the S end are more gray gneisses, here with a distinct reddish tinge caused by an abundance of garnet. The gray gneisses here are nearly devoid of K feldspar. They become more strongly foliated toward the contact with overlying pink granitic gneisses. The contact itself is extremely sharp (but note the late spherical, undeformed garnets, some of which are situated directly on the contact). The pink granitic gneisses, which contain biotite, chlorite and garnet, are strongly foliated, approaching mylonitic texture in places, and display prominent quartz ribbon lineation. The less deformed parts of these gneisses contain K-feldspar megacrysts (phenocrysts? porphyroblasts?) in various stages of deformation and recrystallization.

Continuing N, pass a large gabbro pod, broken at the base and injected with granitic material, in part pegmatitic. Look S across the road; the similarly shaped body of gabbroic rock in the pink gneisses is probably the same pod. Then re-enter gray gneisses, here with somewhat more K feldspar, which is concentrated in the leucosomes. Notice the prominent discontinuity in the foliation which is visible for some distance along the cut. Even though little textural evidence (e.g. grain size reduction) for shear displacement exists along the discontinuity, other explanations for this feature are even more difficult to defend. Toward the N end of the cut is another body of gabbroic rock, which also appears to continue on the opposite side of the road. These mafic rocks, which intrude both the gray and pink gneisses, are generally fine grained and massive with a distinct relict igneous texture. Much garnet is present in the form of indistinct coronas. These rocks are the equivalent of the coronitic olivine metagabbros, a more typical example of which will be seen at Stop #8. These gabbroic bodies (several are present here) are lensoid to sigmoidal in cross section, but apparently elongated in a roughly N-S direction. Their crudely sigmoidal shape yields opposite estimates of shear sense depending on whether they are pre- or syntectonic in origin.

Cross the road to the E side, and note the prominent minor folds in the migmatitic gray gneisses near the N end of the cut. Note also the open, upright folds, which warp the foliation of these rocks, then compare the orientation of these with the recumbent, isoclinal minor folds and with the lineation. Then walk S along the E side and return to the starting point. The petrology of the rocks at this outcrop has been studied in detail by William Glassley and students at Middlebury College. Dr. Glassley (per. comm., 1985) reports the following:

"Garnet-clinopyroxene and garnet-biotite temperatures were computed from microprobe data. Average temperatures from eight samples ranged from 770 C to 850 C, with a strong mode at 810 C. Pressures, calculated from the assemblages garnet-plagioclase-clinopyroxene-quartz and garnet-plagioclase-orthopyroxene-quartz using the method of Newton and Haselton, average 7.5 kb + 0.5 kb.

Two unusual assemblages can be found along the contact between the two gneiss units. Within 50 cm. of the contact occur 1-3 cm. long augen which contain the assemblages clinopyroxene-garnet-rutile and biotite-sillimanite-hercynite-kspar-garnet. The former assemblage is a typical eclogite assemblage. Garnets from these eclogitic lenses are similar to those reported from basal gneiss eclogites in Western Norway. The clinopyroxenes, however, are poor in jadeite component, with only 5% of this component present. The sillimanite-spinel-bearing assemblage is clearly consumed and biotite and sillimanite are being generated. The significance of this assemblage for P-T conditions remains obscure, in that we do not yet have compositional data for all of the minerals in the assemblages nor do we have water fugacity values that would allow calculation of the equilibrium conditions."

0.4

Junction NY 22 and US 4; NY 22 southbound crosses canal just E of here and goes past the state prison at Comstock. Continue S on 22 and 4.

0.4

STOP 51 Pull off on R side as close to the guardrail as possible. The rocks immediately to the R are strongly foliated biotite-quartz-2 feldspar-garnet gneisses. This version of the gray gneiss is commonly referred to as "kinzigite". Present in this outcrop are thin quartzo-feldspathic pegmatites in various stages of tectonic disintegration and reorientation. The large K-feldspars survive the tearing-apart process better than quartz, and remain visible as large porphyroclasts, either in strings or as isolated individuals. Be alert for evidence of tectonic rotation of these feldspars, which can be a useful indicator of the sense of shear.

Also observe the variable shape and appearance of the garnets: some are rounded and others elliptical; some are nearly inclusion-free while others are "spongy". Careful study of this variation might, if combined with probe analysis of garnet compositions, yield information on the interrelation of metamorphism and deformation of these rocks.

From here, walk northward along the road past a gap in the outcrop, then enter the S end of a long cut. The first rocks are strongly foliated and lineated gray gneisses with lenses and pods of calc-silicates. Roughly 30 m. northward and uphill, these overlie amphibolitic rocks, which comprise most of the remainder of the cut. The bulk of these rocks are strongly foliated garnet amphibolites and mafic gneisses. Numerous lenses and pods of calc-silicates, garnet hornblendite, and ultramafic rocks are present. (Students: the coarse grained ultramafic pods are a fine opportunity to test your mineral recognition skills). About 90 m. northward along the cut a large pod of calc-silicate granulite (grossular-diopside-quartz) is visible in the mafic gneisses on the opposite side of the road. Near the N end of the cut, still on the R (E) side, two large pods or megaboudins of massive, relatively fine grained, garnet-rich metagabbro are surrounded by strongly foliated amphibolites. The transition between foliated and unfoliated rock is very abrupt. Patches of tourmaline-bearing pegmatite are present at the broken (?) end of one of the megaboudins.

0.1

Road crosses small pond.

0.1

Outcrops on R are extensively fractured granitic gneisses close to as N-S brittle fault.

1.3

Outcrops at edge of woods on L are fine-grained white arkosic sandstones.

0.2

STOP 52 Turn off main road and park on dead end road which leads downhill toward the Champlain Canal.

The outcrop on the R side of Rte. 4 just beyond the intersection exposes the unconformity between Proterozoic and Paleozoic rocks (missing: roughly 500 million years of the geologic record, and 20-25 km of Proterozoic rock). The Paleozoic rocks here are coarse arkosic sandstones and quartz-pebble conglomerates of the Cambrian Potsdam Formation, locally with carbonate cement. Measure the strike and dip of the unconformity surface, and compare this with the 10-15 degree easterly slope of the fault block as observed driving N out of Fort Ann.

Observe the lack of evidence for deep weathering of the Precambrian rocks beneath the contact, and the absence of a paleosol layer. This suggests deep erosion and scouring (by wave? ice?) shortly before deposition of the Potsdam.

Walk a few meters along the S face of the outcrop, towards the canal. Note the complex fracturing of the gneisses, and the filling of the fractures with dark, fine-grained dolomitic rock. The significance of this feature is unclear, and it will be discussed in more detail on the trip. Note the deeply weathered zone where these rocks are exposed at the surface.

After examining the unconformity, cautiously cross the road to the cut in complexly deformed gray gneisses on the opposite side. Measure several lineations here and compare with what you saw at Stop #2. Note not only differences in orientation, but also in the nature of the lineation.

0.2

STOP 53 Roadcut on R (SE) side of road. Strongly foliated quartz- 2 feldspar-pyroxene-hornblende-garnet gneisses, +/- biotite. Leucocratic bands contain numerous pyroxene megacrysts, both clino and ortho, the latter showing characteristic rusty weathering color. These rocks are close to the charnockitic end of the migmatitic gray gneiss spectrum. Note the presence of at least two types of amphibolite. One is relatively coarse grained, boudinaged and injected with leucocratic veinlets. The foliation within the boudins is locally truncated by that in the enclosing gneisses. The other amphibolite is dark, fine-grained, biotite-rich, and lacks the leucocratic veining and prominent foliation of the coarser amphibolite. The fine-grained, massive amphibolite forms a megaboudin or recumbent fold (which is it?) near the center of the cut. Do these amphibolites represent one, two, or more generations of mafic intrusives?

Numerous complex minor folds are present within the gneisses; also observe the warping of the foliation by larger, open folds. Measure and record lineations and attempt to relate them to the fold axes of both types. Is more than one lineation present in these rocks?

Thin, folded dark bands near the N end of the cut are a peculiar, fine grained carbonate-rich rock with poorly oriented biotite.

0.1

Flat Rock Road, on L.

0.5

Outcrops in woods to R of road are coarse marble with numerous detached and rotated blocks of amphibolite and gneiss, around which the foliation of the marble is wrapped.

0.2

STOP 5Y Turn into parking area on R (SE) side of road and cautiously cross road to outcrops on opposite side. A plaque on the face of the outcrop commemorates the Battle of Fort Ann (July 8, 1777). The rocks here are intensely foliated and fractured representatives of the anorthosite suite. Although the characteristic andesine megacrysts found in anorthosites elsewhere in the Adirondacks are absent here, they can be found sporadically in other outcrops along the West side of the Pinnacle Range. Minerals in the anorthosite at this stop consists of recrystallized and sericitized plagioclase, hornblende, clinopyroxene and garnet. Large garnets (please do not sample) are surrounded by leucocratic haloes which locally obliterate the foliation, which suggests that the garnet grew at the expanse of mafic minerals which define the foliation, and that it postdates at least the first deformation.

At the eastern end of the outcrop is a large mass of gabbroic rocks (plagioclase-clinopyroxene-garnet-ilmenite) which displays little or no foliation, and around which the foliation in the anorthositic gneiss is deflected. Also present in the outcrop are a breccia zone and numerous closely spaced fractures with a general northeasterly trend; a major high-angle fault may exist roughly parallel to the road.

0.8

Jct. of NY 149 and US 4 in Fort Ann Village. Take R (south or west) on NY 149.

As you leave the village, notice the range of hills directly behind you. The eastern slope is a dip slope on a fault block of Precambrian rocks (known as the Pinnacle Range) bounded on the West by the Welch Hollow Fault (Hills, 1965). The Paleozoic rocks previously noted are on the downthrown (W) side. This is the easternmost of several such fault blocks, all showing a gentle, regional eastward dip of 10-15 degrees on the Precambrian rocks and overlying Paleozoics. The eastern slope of this block is a dip surface close to or at the unconformity, which will be seen in outcrop at Stop 3. Is the tilt of this surface a result of a rotation of the fault blocks, or a reflection of the Tertiary-Recent doming of the Adirondacks (Isachsen, 1975)?

1.8

Outcrops of Cambro-Ordovician Beekmantown Group carbonates.

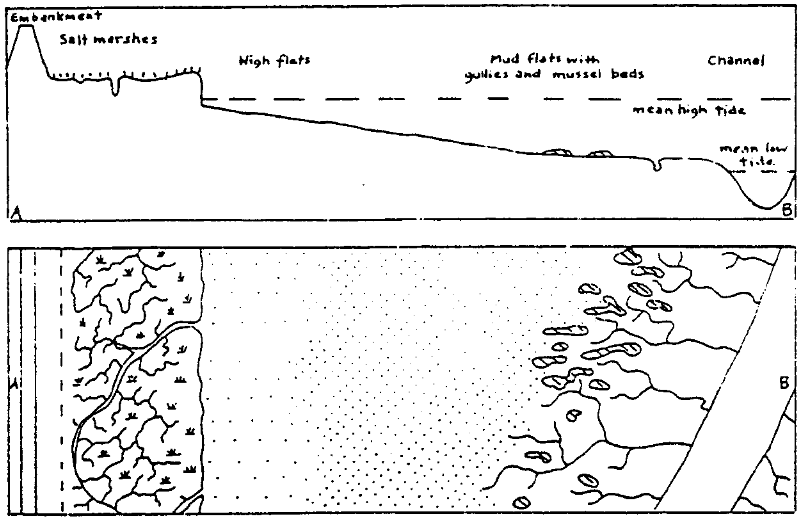
10.0

Junction NY 149 and US 9.   Bear L (east) on NY 149.

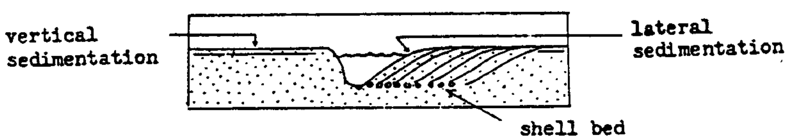
0.6

Entrance to exit 20 on I-87.

Saratoga Springs Stromatolites



Wadden Sea Tidal Flat Environments



Vertical and Lateral Sedimentary Processes

Figure 3. Sedimentary environments of the Wadden Sea intertidal zone (after Van Straaten, 1954, p. 27; Johnson and Friedman, 1969, p. 472).

tion was occurring was a primary structural control. This structure formed a barrier to terrigenous material that was moving westward from the source area into the marine basin, making it possible for carbonate sediment to accumulate. The clastic material accumulated in a basin-margin trough or depression which subsided intermittently as deposition continued. During the transgressive phase landward migration of the strandline caused river mouth drowning and resulted in more widespread estuarine (tidal) conditions as the Tully interval was accumulating.

ITINERARY

Figure 4 is the road log.

Cambro-Ordovician Shoaling and Tidal Deposits

Depart from the parking lot of the Performing Arts Center and turn north on NY 50.

MILES FROM LAST POINT	CUMULATIVE MILEAGE	ROUTE DESCRIPTION
0.6	0.6	Bear left following sign to NY 29.

1.2	1.8	Drive to traffic light and turn left (west) on NY 29.
2.1	3.9	Turn right (north) on Petrified Gardens Road; drive past "Petrified Gardens" to Lester Park.
1.2	5.1	Alight at Lester Park.

84 STOP 1. PRODUCTS OF INTERTIDAL ENVIRONMENT: DOMED ALGAL MATS (CABBAGE HEADS)

This locality is the site of one of the finest domed algal mats to be seen anywhere preserved in ancient rocks. On the east side of the road in Lester Park a glaciated surface exposes horizontal sections of the cabbage-shaped heads composed of vertically stacked, hemispherical stromatolites (Figure 5). These structures, known as Cryptozoons, have been classically described by James Hall (1847, 1883), Cushing and Ruedemann (1914), and Goldring (1938); an even earlier study drew attention to the presence of ooids as the first reported ooid occurrence in North America (Steele, 1825). Interest in these rocks has been revived as they are useful environmental indicators (Logan, 1961, Fisher, 1965; Halley, 1971). The algal heads are composed of discrete club-shaped or columnar structures built of hemispheroidal stromatolites expanding upward from a base, although continued expansion may result in the fusion of neighboring colonies into a Collenia-type structure (Logan, Rezak, Ginsburg, 1964). The stromatolites are part of the Hoyt Limestone of Late Cambrian (Trempealeauan) age. An intertidal origin has been inferred for these stromatolites (Fig. 6, 7 and 8).

Figure 5. Top view of algal stromatolites showing domed laminae known as cabbage-head structures, Hoyt Limestone (Upper Cambrian), Lester Park, New York (Stop 1).

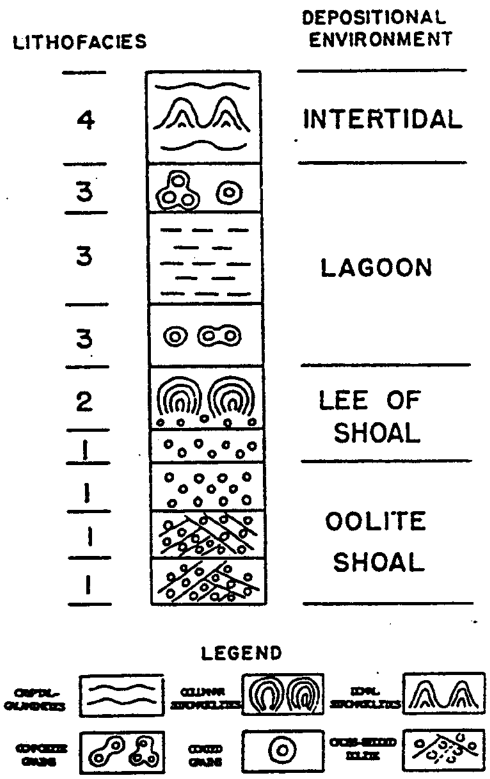


Figure 6. Vertical sequence, lower Lester Park section. The vertical sequence shown by this section reflects a vertically continuous progradational sequence. The upward increase in lithofacies number suggests progressivly shoreward deposition. (R. W. Owen and G. M. Friedman, 1984, Fig. 8, p. 230.)

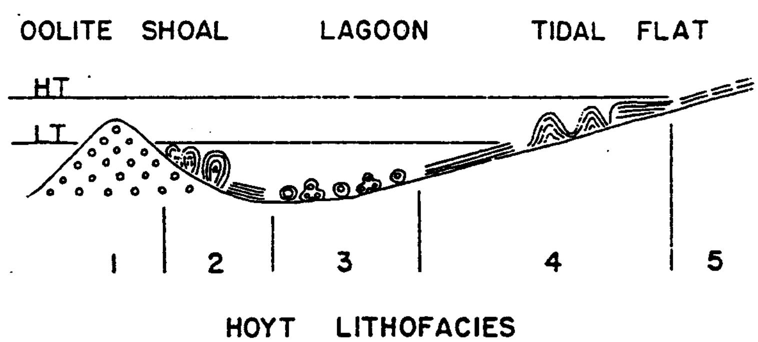


Figure 7. Hypothesized depositional model, cross-section view. Note the similarity in horizontal sequence of lithofacies and vertical sequence of lower Lester Park section (figure 8). Vertical scale greatly exaggerated. (R. W. Owen and G. M. Friedman, 1984, Fig. 13, p. 233.)