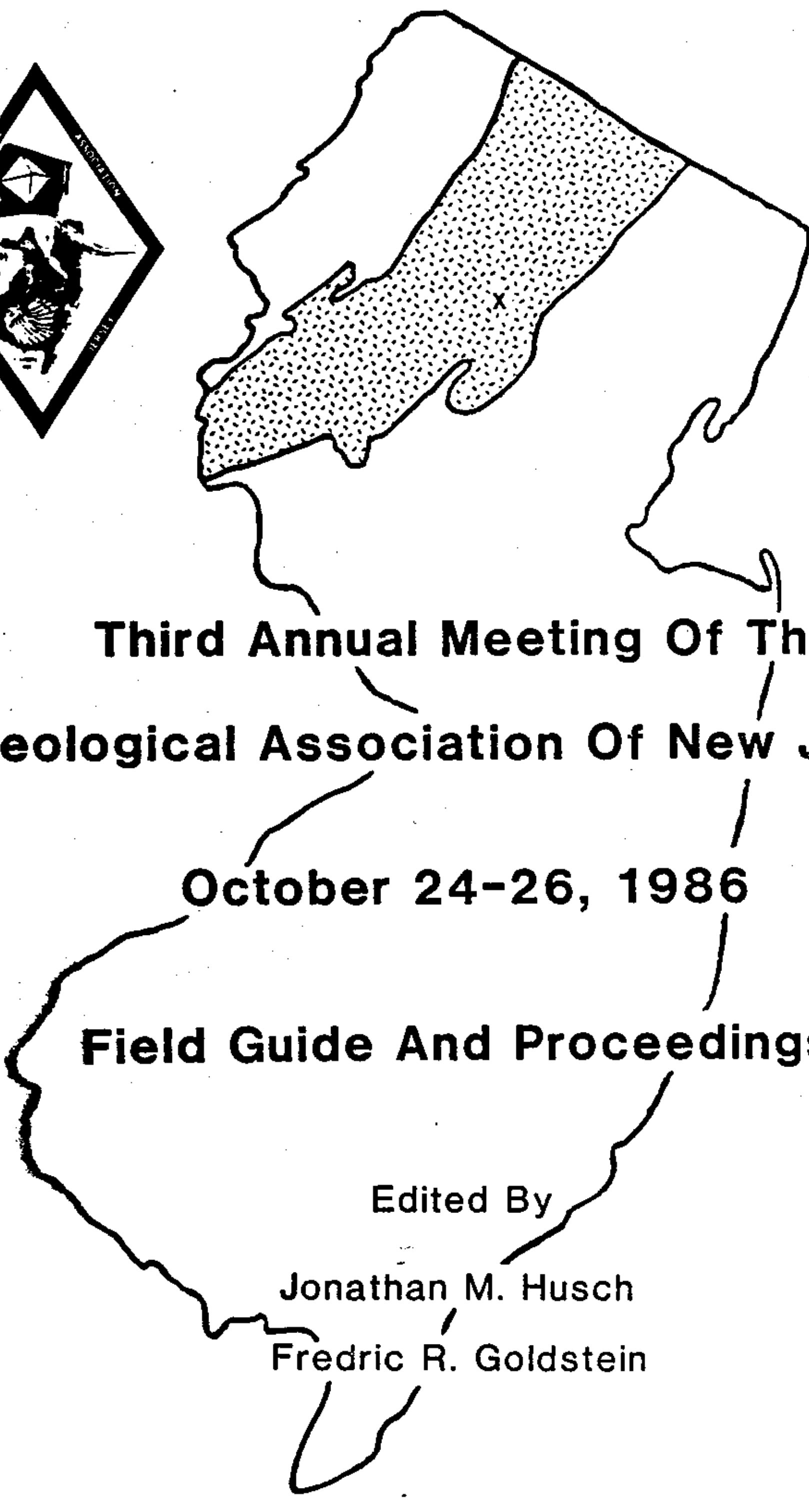
GEOLOGY OF THE NEW JERSEY HIGHLANDS

AND

RADON IN NEW JERSEY



Third Annual Meeting Of The

Geological Association Of New Jersey

October 24-26, 1986

Field Guide And Proceedings

Edited By

Jonathan M. Husch

Fredric R. Goldstein

GEOLOGY OF THE NEW JERSEY HIGHLANDS
AND
RADON IN NEW JERSEY

Edited By

Jonathan M. Husch
Department of Geosciences
Rider College
Lawrenceville, NJ 08648

Fredric R. Goldstein
Department of Geosciences
Trenton State College
Trenton, NJ 08625

Third Annual Meeting of the
Geological Association of New Jersey
October 24-26,1986

Randolph High School
Randolph, New Jersey 07869

TABLE OF CONTENTS

GANJ III PROGRAM

FRIDAY, OCTOBER 24:

7:30 PM - END: Symposium on the geology of the New Jersey Highlands. Richard Volkert (New Jersey Geolgical Survey), Avery Drake (United States Geological Survey), and Joseph Hull (Rutgers University) are participants.

SATURDAY, OCTOBER 25:

8:00 AM - 4:30 PM: All-day field trip in the New Jersey Highlands. Richard Volkert, Avery Drake, and Joseph Hull will lead.

5:30 PM - 7:00 PM: Annual Dinner and Business Meeting. GANJ executive committee elections.

7:00 PM - 8:30 PM: Radon in New Jersey Symposium. Christy Bell (New Jersey Geological Survey), Gerald Nichols (New Jersey Division of Radiation and Safety), and Jonathan Powell (Teledyne Isotopes) speakers.

8:45 PM -10:45 PM: Workshop for high school earth science teachers: Metamorphism and Metamorphic Rocks. Fredric Goldstein (Trenton State College), Raymond Talkington (Stockton State College), and Jonathan Husch (Rider College) are workshop instructors.

SUNDAY, OCTOBER 26:

9:00 AM -12:00 PM: Mineral collecting trip to the Stirling Open Cut to sample classic examples of Franklin Marble mineralization. Robert Metsger (New Jersey Zinc) is trip leader.

INTRODUCTION

The third Annual Meeting of the Geological Association of New Jersey results from a new maturity in the organization. No longer do we wonder how the Annual Meeting will be run, what functions should we have on the first or second nights, how do we publicize the meeting, etc. In a very short period of time, our traditions and goals have been established. In addition, our Association is better known (and hopefully respected), both within New Jersey and throughout the entire northeast region. In response, we have organized a meeting that is pertinent and of interest to all segments of the geological community within New Jersey, from professional researcher to amateur mineral collector.

The constitution of the Geological Association of New Jersey states the the purpose of the Association is to "encourage geologic investigations..., to foster interchange of such information..., and to disseminate knowledge of the geology of New Jersey to all interested groups and individuals." In every respect, the third Annual Meeting addresses these purposes. The symposium and field trip on the geology of the New Jersey Highlands present the results of recent professional research in the region. Furthermore, the articles in this volume by Volkert and Drake and Hull et al. provide both a general overview of Highlands geology and a detailed examination of structural elements contained within the province. The radon symposium provides a public forum for disseminating information of vital interest, not only to the professional geologist, but also to a concerned public. The earth science teachers workshop on metamorphism and metamorphic rocks adds an educational component to the association's activities. The Association can do no greater service to New Jersey than to provide a high quality continuing education program to those teaching earth sciences at the secondary level. Finally, the trip to the Stirling Open Cut allows geologists at every level a rare opportunity to sample directly the unique mineralogical assemblages located in the Franklin area.

The organization of the third Annual Meeting, as well as the production of this volume, is the result of the cooperative efforts of numerous people. In particular, the efforts of Christy Bell, Richard Bizub, Elmer Dey, Avery Drake, Betty Falkenstein, William Gallagher, Joseph Hull, Geraldine Hutner, Robert Koto, Robert Metsger, Lee Meyerson, Gerald Nichols, Jonathan Powell, Florence Sackett, Raymond Talkington, and Richard Volkert are gratefully acknowledged. We hope you find the results of their efforts to be both educational and enjoyable.

Jonathan Husch
Co-editor

Fredric Goldstein
Co-editor

Some Middle Proterozoic Rocks of the New Jersey Highlands

By

Richard A. Volkert[1] and Avery A. Drake, Jr.,[2]

[1]New Jersey Geological Survey, Trenton, New Jersey 08625

[2]United States Geological Survey, Reston, Virginia 22092

Introduction

The New Jersey Highlands constitutes the middle segment of the Reading Prong, the second largest external basement massif in the central Appalachian orogen (Drake and others, in press). The Highlands are underlain by intrusive rocks and attendant migmatites, quartzofeldspathic and calcareous metasedimentary rocks, sodium-rich rocks of probable volcanic-volcaniclastic origin, and charnockitic rocks of problematic origin, all of Middle Proterozoic age. Approximately half of this assemblage consists of plutonic intrusive and closely related rocks. These rocks are overlain by a minor proportion of distinctly younger metasedimentary and metavolcanic rocks thought to be of Late Proterozoic age. This assemblage is reasonably well understood, primarily from the work of Baker and Buddington (1970), Drake (1969, 1984), Hague and others (1956), Sims (1958), and Young (1971, 1972, 1978), although problems still exist. The purpose of this trip is to examine some of these rocks in good exposures found during field work done in connection with the preparation of a new geologic map of New Jersey. This work is a cooperative effort of the New Jersey Geological Survey and the U.S. Geological Survey and is an element of the U.S. Geological Survey's COGEOMAP program.

Losee Metamorphic Suite

The oldest Middle Proterozoic rocks in New Jersey are believed to be a highly sodic sequence that constitutes the Losee Metamorphic Suite of Drake (1984). The Losee appears to be basement to the quartzofeldspathic and calcareous metasedimentary rocks. A similar relation has been described in adjacent Orange County, New York (Offield, 1967) and farther east in the

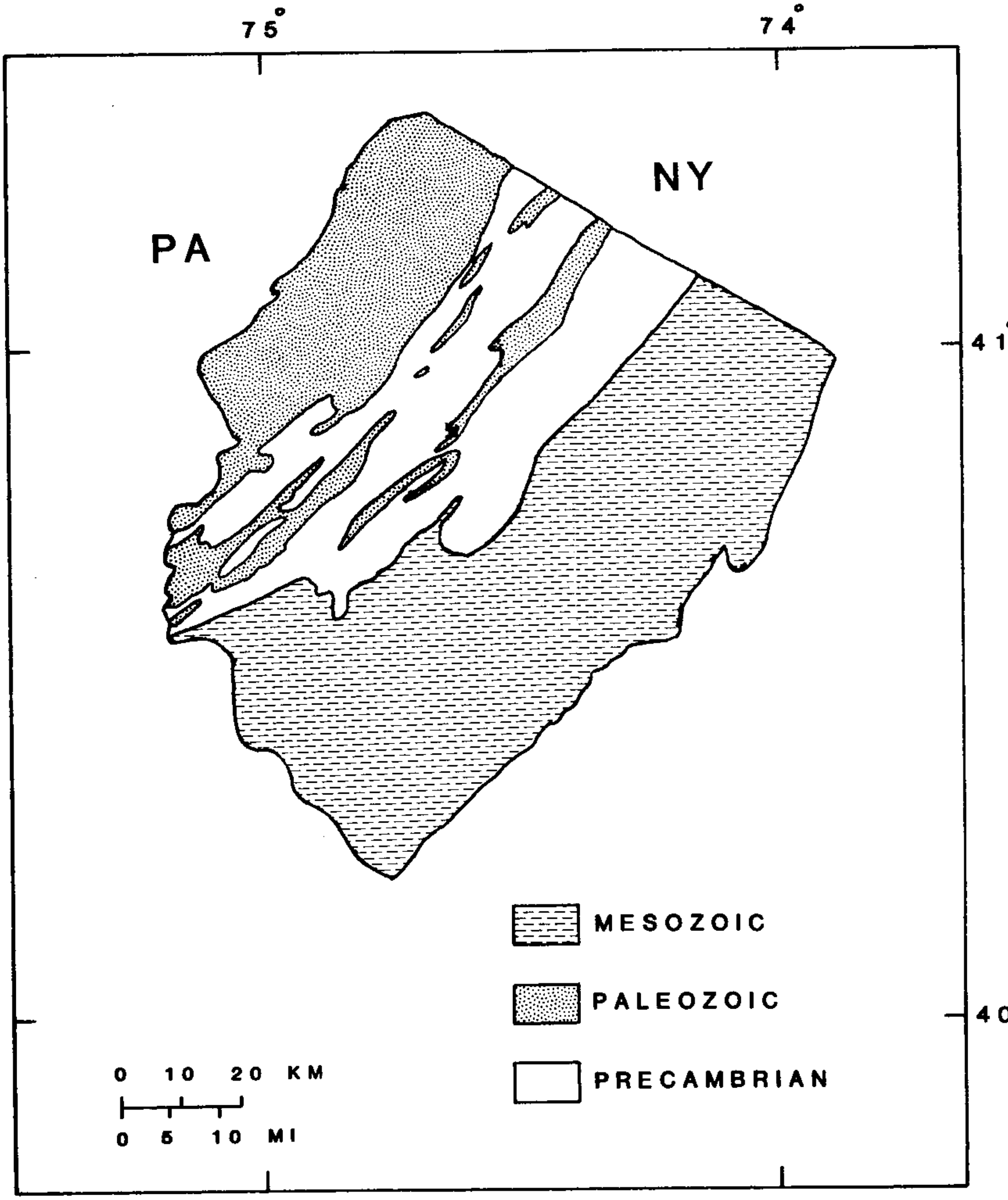


Figure 1. Location map of the Precambrian Highlands and its relation to the Triassic/Jurassic Lowlands and Valley and Ridge Provinces.

Hudson Highlands (N. M. Ratcliffe, oral communication, 1986). The only known older rocks are those of the Hexenkopf Complex (Drake, 1984) an oceanic mafic-ultramafic suite in eastern Pennsylvania.

Rocks of Losee consist largely of sodic plagioclase and quartz and rarely contain more than 5 percent of mafic minerals. Biotite is most common, but locally, the unit contains minor amounts of hornblende, magnetite, augite, or hypersthene. The mafic minerals are commonly shredded and altered to chlorite and epidote. Parts of the unit are well-layered and foliated, others are poorly foliated granofels, and still others are granitoid or pegmatitic in aspect. All these rocks have very high Na_2O/K_2O ratios (>7/1), very high normative albite (55-61 percent), and differ petrographically only in the relative amount of sodic plagioclase and quartz (Drake, 1969, 1984). The layered and granofels phases are mapped as oligoclase-quartz gneiss, the granitoid phase as albite-oligoclase granite, and the pegmatitic phase as albite pegmatite. The origin of the complex has been quite controversial, but direct observation in the field shows that the granitoid and pegmatite phases have been generated in situ by anatexis (Drake, 1969, 1984). The very high Na_2O/K_2O ratios and the very high normative albite led Drake (1984) to suggest that the protolith of the Losee was quartz keratophyre. Other authors have suggested graywacke as a protolith. The most sodic graywackes reported in the literature (Drake, 1984), however, do not approach the normative albite content of the Losee, and it seems that such rocks are an unlikely protolith. At places, the Losee contains sparse to moderate amounts of interlayered amphibolite. This amphibolite is more sodic than the other amphibolites in the Reading Prong, and therefore, is thought to be metamorphosed spilitic basalt or basaltic tuff (Drake, 1984). The Losee is probably a metamorphosed volcanic pile of quartz keratophyre and interlayered spilite that originated in an oceanic domain.

The Losee has never been satisfactorily dated in New Jersey, but similar rocks have been dated in New York. Helenek and Mose (1984) obtained a 1147 ± 43 Ma whole rock Rb-Sr isochron, Aleinikoff and others (1982) determined an U-Pb upper intercept age of 1170 Ma, and Tilton and others (1960) obtained a ^{207}Pb-^{206}Pb age of 1150 Ma, from the same belt of rocks.

The oligoclase-quartz gneiss phase of the Losee will be examined at Stop 6.

Metasedimentary Rocks

The Losee is overlain by a sequence of quartzofeldspathic and calcareous metasedimentary rocks as recognized by both Hague and others (1956) and Baker and Buddington (1970) in northern New Jersey. In immediately adjoining Orange County New York, Offield (1967) recognized a similar stratigraphic relationship as well as apparent truncation of structural trends in these rocks by the overlying metasedimentary sequence suggesting an unconformity. There is no evidence for thrust faulting. A similar relation is suggested in eastern Pennsylvania where quartzofeldspathic rocks lie both on the Hexenkopf Complex and the contact between the Hexenkopf and the Losee (Drake, 1984).

Calcareous rocks.--Calcareous metasedimentary rocks include calcite and dolomite marble, pyroxene gneiss, quartz-epidote gneiss, epidote-scapolite-quartz gneiss, and a variety of lime-silicate gneisses. Marble is most abundant along the north border of the New Jersey Highlands where two major layers can be mapped from Big Island, New York (Offield, 1967) to the Andover, New Jersey area (Hague and others, 1956). This marble is the host for the famous zinc deposits of the Franklin-Sterling, New Jersey, area. Scattered marble layers are present along strike into Pennsylvania (Drake, 1984), and only small bodies of marble have been mapped elsewhere. Calcite marble is by far the most abundant, but dolomite marble, largely altered to serpentine rock, talc rock, and tremolite rock, has been exploited in several quarries along the Delaware River (Drake, 1967). Pyroxene gneiss is not abundant but is fairly common in the Franklin area and northeast along strike, presumably reflecting the calcareous nature of much of the sediment there. Pyroxene gneiss is quite heterogeneous, having three apparent end members: clinopyroxene-plagioclase gneiss, clinopyroxene-potassic feldspar gneiss, and clinopyroxene-scapolite gneiss. Epidote-scapolite gneiss also is only common in the general Franklin area where it is a catchall unit for calcareous rocks characterized by epidote and(or) scapolite. Some phases are so quartz-rich as to be epidote- and scapolite-bearing metaquartzites. These rocks grade into magnetite-rich potassic feldspar gneiss, suggesting that they represent metamorphosed calcareous iron-formation (Drake, 1984). Another rock type, quartz-epidote gneiss, also grades into potassic feldspar gneiss and may be part of the suggested iron-formation. These rocks will not be examined on this trip.

Quartzofeldspathic Gneiss.--Quartzofeldspathic gneiss is common throughout the New Jersey Highlands. Drake (1969, 1984) originally recognized two end members: biotite-quartz-feldspar and potassic-feldspar gneiss. Our fieldwork in 1984-85, however, has shown that the microcline gneiss as mapped by Hague and others (1956) is a valid unit in north-central New Jersey. Biotite-quartz-feldspar gneiss is highly variable in both composition and texture but is characterized by conspicuous biotite and prominent compositional layering. Most specimens contain far more plagioclase than potassium-feldspar, and the more aluminous phases are sillimanitic. Other accessories include garnet, magnetite, pyrite, rutile, and graphite. At many places, the gneiss contains small lenticular bodies of amphibolite, pyroxene gneiss, or marble. Graywacke is the most likely protolith for the unit.

Potassic feldspar gneiss is characterized by its high content of potassic feldspar and quartz and its paucity of plagioclase. Much of the unit is quite heterogeneous, and some layers are so quartz-rich as to approach quartzite in composition. Sillimanite, biotite, and garnet are sporadic and other phases are rich in magnetite and resemble metamorphosed iron-formation. Another phase of the unit is a quite homogeneous granofels and is quite granitic in appearance. Both Baker and Buddington (1970) and Drake (1969, 1984) appeal to partial anatexis to develop the granitic phases within the potassic feldspar gneiss. The potassic feldspar gneiss is thought to be a metamorphosed arkosic sediment (Drake, 1984).

Microcline gneiss is an exceptionally well-layered rock composed largely of quartz and microcline, but it contains moderate amounts of plagioclase (see modes in Hague and others, 1956). The layering is on a much finer scale (1.5 to 5 cm) than that of the other potassic feldspar gneiss. Biotite and chlorite are the most common assessories, but magnetite is common and sillimanite, clinopyroxene, and hornblende are found at a few places. Granitic bodies termed microcline granite gneiss by Sims and Leonard (1952) have been generated locally within the unit by partial anatexis. The unit probably is a metamorphosed arkose. Microcline gneiss will be examined at Stop 4.

Amphibolite

Amphibolite occurs throughout this part of the Reading Prong and is associated with all other rock units. Its origin is difficult to decipher because of the lack of primary features. Much is probably sedimentary as it is interlayered with calcareous and quartzofeldspathic metasedimentary rocks. Some, however, appears to have relict pillow structure and must result from the metamorphism of basaltic flows. See discussion of Stop 8 below where this type of amphibolite will be examined. Still another type of amphibolite of the Losee Metamorphic Suite is more sodic (see above). Amphibolite clearly has more than one origin.

Intrusive Rocks

Much of the intrusive igneous rock in the New Jersey Highlands belongs to the Byram Intrusive Suite (Drake, 1984). A fair amount of rock characterized by clinopyroxene and mesoperthite or microantiperthite and ranging in composition from to syenite to monzonite crops out in north-central New Jersey and small bodies of undeformed Mount Eve Granite (Hague and others, 1956), presumably younger intrusions, occur from north-central New Jersey into Orange County, New York.

Byram Intrusive Suite.--These granitoid rocks consisting of quartz, potassic feldspar, oligoclase, and varying amounts of mafic minerals are common thoughout the New Jersey Highlands where they occur in regionally conformable sheets, pods, and refolded bodies. Although regionally conformable, these granitic bodies locally cut across structures in the older rocks. Four end members are mapped: (1) microperthite alaskite, (2) hornblende granite, (3) biotite granite, and (4) hornblende syenite which includes hornblende-quartz syenite. The hornblende granite differs from alaskite primarily in its content of mafic minerals. Biotite granite is believed to stem from the assimilation and modification of biotite-quartz-feldspar gneiss by alaskite (Dodd, 1965; Drake, 1969, 1984; Sims, 1958). The center of gravity of the available modes plots just to the left of the syenogranite-monzogranite boundary (Drake, 1984). In north-central New Jersey many of the Byram rocks become deficient in silica and, therefore, are properly hornblende-quartz syenites and hornblende syenites. The Byram rocks have a minimum melt composition and are believed to result from anatexis at depth greater than the enclosing rocks as it is clear that they were mobile in most cases. Byram alaskite forms the leucosome in the abundant migmatites in western New Jersey and Pennsylvania. Migmatites are sparse or absent farther east in New Jersey.

The Byram rocks are gneissoid, that is they have a primary flow foliation and lineation, and at many places a secondary metamorphic foliation. Only the most deformed phases are gneissic. In these rocks the microperthite has unmixed into free microcline and oligoclase.

No rocks of the Byram Intrusive Suite have been isotopically dated. These rocks do, however, have some simularity with the Storm King Granite (Lowe, 1950) of the Hudson Highlands. The Storm King has a U-Pb upper intercept age of 1118 ± 55 Ma (Aleinikoff and others, 1982).

The Byram Intrusive Suite will not be examined on this field trip.

Quartz-Poor Sequence.--Quartz-poor rocks, largely monzonite and quartz monzonite and lesser granite (Drake, 1984), but described as syenite or quartz syenite (Baker and Buddington, 1970; Young, 1971, 1972, 1978) are fairly common in north-central New Jersey. In addition to their general lack of quartz, these rocks differ from those of the Byram Suite in that their mafic mineral is ferrohedenbergite and their feldspar is microantiperthitic mesoperthite. Chemically, these rocks have much higher Fe-Mg ratios, contain more CaO, and are more femic than the Byram rocks (Drake, 1984). To date, the relation of the pyroxenic rocks to the Byram has not been established. This relation is currently one of the major research efforts in the New Jersey Highlands. These rocks have not been isotopically dated. An excellent exposure of the quartz-poor sequence will be examined at Stop 2.

Mount Eve Granite.--The Mount Eve Granite (Hague and others, 1956) forms stocks that have dike- and sill-like apophyses. It is a light-colored, coarse-grained monzogranite. In Orange County, New York it contains biotite and locally, hornblende and sparse allanite. In the Sparta, New Jersey area, clinopyroxene and magnetite are the only dark minerals. At places it has an igneous flow structure, but tectonic foliation has not been recognized. The Mount Eve should be isotopically dated because it appears to post-date the Grenville tectonism. The Mount Eve will not be examined on this field trip.

Plutonic Rocks Of Uncertain Origin

Two sequences of plutonic rocks of uncertain origin are known in this part of the Reading Prong, "granodiorite gneiss" and rocks of charnockitic aspect.

"Granodiorite Gneiss".--"Granodiorite gneiss" (Hague and others, 1956) appears to be restricted to the Newton East and Stanhope quadrangles in New Jersey. It is a crudely layered, generally well-lineated rock that contains 10 percent or less of potassic feldspar. Some phases resemble rocks of the Losee Metamorphic Suite, whereas others resemble intrusive igneous rocks. This unit will be examined at Stop 5. See description of that stop below for chemistry and more details.

Charnockitic Rocks.--Charnockitic rocks are widely distributed in the Reading Prong but appear to be much more abundant in north-central New Jersey and New York. They are granitic-appearing, but at most places have conspicuous alternating light and dark layers and contain discontinuous bodies of amphibolite and pyroxene amphibolite. Other bodies, however, are massive and appear to have igneous textures. Sims (1958) concluded that these rocks are metamorphosed intrusions. Baker and Buddington (1970) believe that they are metasedimentary. Offield (1967), Dodd (1965), and Young (1971) believe that they are metamorphosed sodic volcanic rocks, and Drake (1969, 1984) believes that they result from the high-grade metamorphism and partial anatexis of amphibolite-bearing parts of the Losee Metamorphic Suite. Perhaps the igneous-appearing bodies result from the crystallization of local total melts. An igneous-appearing charockitic body will be examined at Stop 1.

Metamorphism

Detailed metamorphic studies have not been done in New Jersey but the quartzofeldspathic rocks contain sillimanite and potassic feldspar and mafic rocks contain the critical mineral assemblage plagioclase-diopside-hyperthene-hornblende-biotite implying hornblende granulite facies metamorphism. Minimum temperatures of metamorphism of 700°C in the Lake Hopatcong area (Young, 1971) and 760°C in the Franklin-Sterling area (Carvalho and Sclar, 1979) are consistent with this facies. A retrogressive event is shown by altered hypersthene in mafic rocks and sillimanite in the quartzofeldspathic rocks.

Middle Proterozoic Structural Geology

Folds are common in this Middle Proterozoic terrane although they are difficult to map in many areas because of poor exposure. They range from a few mm to 1.7 km in wave length and to as much as 12 km in length parallel to the axis and 1.7 km in width. All mapped folds deform both foliation and layering so none are first folds. Most geologists have mapped only one phase of east-northeast-trending folds, but Hague and others (1956) mapped two phases of cross-folds in north-central and northeastern New Jersey. Drake and others (1985) have recognized three fold phases in northwestern New Jersey that deform foliation as well as lense-shaped lithologic units which are thought to be the result of regional transposition. Three fold phases also deform lens-shaped units in eastern Pennsylvania (A. A. Drake, Jr., unpublished data). It appears, therefore, that these rocks have experienced at least four phases of Middle Proterozoic folding. Outcrops displaying polyphase deformation will be examined at Stops 4 and 6.

Tectonics

A geologic history for the New Jersey and Pennsylvania parts of the Reading Prong has been suggested by Drake (1984) in which an oceanic basement of the Hexenkopf Complex as well as quartz keratophyre and spilitic basalt of the Losee Metamorphic Suite is unconformably overlain by a continental margin sequence of calcareous and quartzofeldspathic metasedimentary rocks. Perhaps part of these rocks were deposited in a rift as suggested by the potassic feldspar gneiss (potassic sandstone) and the pillowed basalt variety of amphibolite. During the intense period of deformation and metamorphism that constituted the Grenville orogeny, sheets of granitic rocks were intruded into this terrane. Subsequent to these events, the Late Proterozoic sedimentary and volcanic(?) rocks of the Chestnut Hill Formation were deposited.

REFERENCES CITED

Aaron, J.M., 1969, Petrology and origin of the Hardyston Quartzite (Lower Cambrian) in eastern Pennsylvania and western New Jersey, in Subitzky, S., ed., Geology of selected areas in New Jersey and eastern Pennsylvania and Guidebook of Excursions: New Brunswick, N.J., Rutgers University Press, p. 21-34.

Aleinikoff, J.N., Grauch, R.I., Simmons, K.R., and Nutt, C.J., 1982, Chronology of the metamorphic rocks associated with uranium occurrences Hudson Highlands, New York-New Jersey [abs.]: Geological Society of America Abstracts with Programs, v. 14, nos. 1 and 2, p. 1.

Baker, D.R., and Buddington, A.F., 1970, Geology and magnetite deposits of the Franklin quadrangle and part of the Hamburg quadrangle, New Jersey: U.S. Geological Survey Professional Paper 638, 73 p.

Carvalho, A.V., III, and Sclar, C.B., 1979, Gahnite-franklinite geothermometer at the Sterling Hill zinc deposit, Sussex County, New Jersey [abs.]: Geological Society of America Abstracts with Programs, v. 11, no. 1, p. 6.

Chapman, D.F., 1966, Petrology and structure of the Byram Cove Synform, Precambrian Highlands, New Jersey [abs.]: Geological Society of America Northeast Section Annual Meeting, p. 15-16.

Dodd, R.T., 1965, Precambrian geology of the Popolopem Lake quadrangle, southeastern New York: New York State Museum and Science Service Map and Chart series no. 6, 39 p.

Drake, A.A., Jr., 1967a, Geologic map of the Easton quadrangle, New Jersey-Pennsylvania: U.S. Geological Survey Geologic Quadrangle Map GQ-594, scale 1:24,000.

_____1967b, Geologic map of the Bloomsbury quadrangle, New Jersey: U.S. Geological Survey Geologic Quadrangle Map GQ-595, scale 1:24,000.

_____1969, Precambrian and Lower Paleozoic geology of the Delaware Valley, New Jersey-Pennsylvania: in Subitzky, S., ed., Geology of Selected areas in New Jersey and eastern Pennsylvania and Guidebook of Excursions; Rutgers Univ. Press, p. 51-131.

_____1984, The Reading Prong of New Jersey and eastern Pennsylvania: An appraisal of rock relations and chemistry of a major Proterozoic terrane in the Appalachians, in Bartholomew, M.J., ed., The Grenville event in the Appalachians and related topics, Geological Society of America Special Paper 194, p. 75-109.

Drake, A.A., Jr., Hall, L.M., and Nelson, A.E., Basement and basement-cover relation map of the Appalachian orogen in the United States: U.S. Geological Survey Miscellaneous Investigations Map I-1655, scale 1:1,000,000, in press.

Drake, A.A., Jr., Kastelic, R.L., Jr., and Lyttle, P.T., 1985, Geologic map of the eastern parts of the Belvidere and Portland quadrangles, Warren County, New Jersey: U.S. Geological Survey Miscellaneous Geologic Investigations Map I-1530, scale 1:24,000.

Hague, J.M., Baum, J.L., Hermann, L.A., and Pickering, R.J., 1956, Geology and structure of the Franklin-Sterling area, New Jersey: Geological Society of America Bulletin, v. 67, p. 435-474.

Helenek, H.C., and Mose, D.G., 1984, Geology and geochronology of Canada Hill Granite and its bearing on the timing of Grenvillian events in the Hudson Highlands, New York, in Bartholomew, M.J., ed., The Grenville event in the Appalachians and related topics: Geological Society of America Special Paper 194, p. 57-73.

Iddings, J.P. 1898, Minette from Franklin Furnace, Sussex County, New Jersey: U.S. Geological Survey Bulletin 150, p. 236-239.

Kemp, J.F., 1889, On certain porphyritic bosses in northwestern New Jersey: American Journal of Science, v. 38, p. 130-134.

Lowe, K. E., 1950, Storm King Granite at Bear Mountain, N.Y., Geological Society of America Bulletin, v. 61, p. 137-190.

Macdonald, G.A., and Katsura, T., 1964, Chemical composition of Hawaiian lavas: Journal of Petrology, v. 5, p. 82-133.

Maxey, L.R., 1971, Metamorphism and origin of Precambrian amphibolite of the New Jersey Highlands: New Brunswick, Rutgers University, Unpublished Ph.D. dissertation, 156 p.

____1976, Petrology and geochemistry of the Beemerville carbonatite-alkalic rock complex, New Jersey: Geological Society of America Bulletin, v. 87, p. 1551-1559.

Melson, W.G., and Thompson, G., 1971, Petrology of a transform fault zone and adjacent ridge segments: Philospothical Transactions Royal Societ London, Series A, v. 268, p. 423-441.

Nockolds, S.R., 1954, Average chemical composition of some igneous rocks: Geological Society of America Bulletin, v. 65, p. 1007-1032.

Offield, T.W., 1967, Bedrock geology of the Goshen-Greenwood Lake area, New York: New York State Museum and Science Service Map and Chart Series, no. 9, 78 p.

Pearce, J.A., 1975, Basalt geochemistry used to investigate past tectonic environments on Cyprus: Tectonophysics, v. 25, p. 41-67.

____1976, Statistical analysis of major element patterns in basalts: Journal of Petrology, v. 17, p. 15-43.

Pearce, J.A., and Cann, 1973, Tectonic setting of basic volcanic rocks determined using trace-element analysis: Earth and Planetary Science Letters, v. 19, p. 290-300.

Puffer, J.H., 1980, Precambrian rocks of the New Jersey Highlands, in Manspeizer, Warren, ed., Field studies of New Jersey Geology and Guide to field trips: Newark, N.J., Rutgers University, p. 42-52.

Rogers, J.J.W., Suayah, I.B., and Edwards, J.M., 1984, Trace elements in continental-margin magmatism: Part IV. Geochemical criteria for recognition of two volcanic assemblages near Auburn, western Sierra Nevada, California: Geological Society of America Bulletin, v. 95, p. 1437-1445.

Sims, P.K., 1958, Geology and magnetite deposits of the Dover district, Morris County, New Jersey: U.S. Geological Survey Professional Paper 287, 162 p.

Sims, P.K., and Leonard, B. F., 1952, Geology of the Andover mining district, Sussex County, New Jersey, New Jersey Department of Conservation and Development Bulletin 62, 46 p.

Smith, B.L., 1969, The Precambrian geology of the central and northeastern parts of the New Jersey Highlands, in Subitzky, S., ed., Geology of selected areas in New Jersey and eastern Pennsylvania and Guidebook of Excursions: New Brunswick, N.J., Rutgers University Press, p. 35-47.

Tilton, G.R., Wetherill, G.W., Davis, G.C., and Bass, M.N., 1960, 1000-million-year old minerals from the eastern United States and Canada: Journal of Geophysical Research, v. 65, p. 4173-4179.

Volkert, R.A., 1984, A determinative study of the structural state and composition of alkali feldspars from pegmatites along Route 15, Morris and Sussex Counties, New Jersey: Upper Montclair, Montclair State College, Unpublished MA thesis, 114 p.

Volkert, R.A., Monteverde, Donald, and Drake, A.A., Jr., Geologic map of the Stanhope quadrangle, Morris and Sussex Counties, New Jersey: U.S. Geological Survey Geologic Quadrangle Map GQ- , scale 1:24,000, in press.

Young, D.A., 1969, Petrology and structure of the west-central New Jersey Highlands: Providence, Brown University, Unpublished Ph.D. dissertation, 194 p.

____1971, Precambrian rocks of the Lake Hopatcong area, New Jersey: Geological Society of America Bulletin, v. 82, p. 143-158.

____1972, A quartz syenite intrusion in the New Jersey Highlands: Journal of Petrology, v. 13, p. 511-528.

____1978, Precambrian salic intrusive rocks of the Reading Prong: Geological Society of America Bulletin, v. 89, p. 1502-1514.

Zartman, R.E., Brook, M.R., Heyl, A.V., and Thomas, H.H., 1967, K-Ar and Rb-Sr ages of some alkalic intrusive rocks from central and eastern United States: American Journal of Science, v. 265, p. 848-870.

DEFORMATION ZONES IN THE HIGHLANDS OF NEW JERSEY

JOSEPH HULL, ROBERT KOTO and RICHARD BIZUB
Department of Geological Sciences
Rutgers University
Newark, NJ 07102

Abstract

Deformation zones formed during the Grenville, Taconic (?), Alleghanian and Newark orogenies are found throughout the Reading Prong of New Jersey. Amphibolite to granulite facies zones of Grenville age are difficult to recognize, because of thorough recrystallization, but mylonite zones surrounding outcrop-sized, undeformed augen have been discovered in the Allamuchy Mountains. Retrograde deformation zones are widespread throughout the Prong, and show consistent cross-cutting relationships. Early, chlorite-rich deformation zones containing breccias to ultracataclasites cut the mylonites, but little is known about their geometry. The early chlorite zones are cut by epidote-rich faults, characterized by polished sliding surfaces (slickensides) associated with narrow zones of gouge. The epidote zones are in turn cut by late chlorite-rich zones that contain the widest variety of tectonites, including superplastic ultramylonites and deformed pseudotachylites. In the Allamuchy Mountains, the late chlorite zones show oblique extensional slip at a high angle to the regional trend. The bulk composition of the gneissic protolith can affect the retrograde mineralogy and structure. Coeval epidote and chlorite zones have been observed in adjacent felsic and mafic layers, and calc-silicate cleavage duplexes are developed in calcareous paragneisses on Jenny Jump Mountain. Two major map-scale deformation zones, the Reservoir and Wright Pond faults, are defined by broad zones of chlorite-rich quasiplastic mylonites and cataclasites. The extent of Mesozoic reactivation of these zones cannot be deduced from the tectonites alone.

INTRODUCTION

Planar zones of relatively intense deformation or strain adjacent to less-deformed protolith (country rock) are known as <u>deformation zones</u>. Deformation zones represent ideal natural laboratories for studying the micromechanisms of deformation at the grain scale, because their Effects are exaggerated by the high strains within the zones. Traverses across the deformation zone boundaries from low to high strains mimic progressive deformation within the zones themselves. Because of many studies of natural deformation zones over the last 10 years, structural geologists now have a clear idea of how various rock types deform at different grades of metamorphism. Geologists can now examine tectonites in the field and laboratory, and deduce the deformation mechanisms, the deformation history and the physical conditions of deformation.

The deformation zones themselves are no less important, especially in understanding how crystalline basement deforms in orogenic belts. The sedimentary cover usually deforms in a ductile, penetrative fashion by some combination of folding, internal deformation (producing cleavage and schistosity), and faulting. At high metamorphic grades (amphibolite to granulite facies), crystalline basement will deform ductilely as well. But at very low to moderate grades, quartzofeldspathic basement rocks are generally too strong to deform penetratively and a network of deformation zones surrounding less deformed lithons is produced instead. These zones may actually be penetrative on an outcrop or regional scale and can generate substantial strains in otherwise stiff, unresponsive basement. Unraveling the geometry and magnitude of strain in the basement produced by this population of deformation zones is a new and exciting field in structural geology.

The Reading Prong of New Jersey is composed predominantly of quartzofeldspathic rocks deformed and metamorphosed during the Middle Proterozoic Grenville Orogeny (about 1.1 Ga). These basement rocks were subsequently involved in thick-skinned deformation of the Ordovician Taconic Orogeny (about 450 Ma) and/or the Late Paleozoic Alleghanian Orogeny (about 300 Ma). The deformed sedimentary cover has been stripped off the basement throughout the Highlands, except in the Green Pond Outlier and in tiny patches adjacent to some major faults. The lack of cover rocks makes structural analysis very difficult, especially in unraveling the Paleozoic structural history. In order to understand the structural and tectonic evolution of the Reading Prong, we must focus our attention on the deformation zones in the basement.

In this paper, we will summarize some of our recent work on deformation zones and their associated tectonites in the Highlands of New Jersey. We will first introduce a new classification scheme for high strain tectonites, as well as the subject of "fault population analysis", including both the field methodology and some simple interpretative models. The different types of small, outcrop-scale deformation zones found in the basement rocks of the Reading Prong are then described, along with selected results of our fault population studies. Two examples of regional, map-scale deformation zones are also discussed. Finally we consider the tectonic history of the Reading Prong in light of the new data.

TECTONITE CLASSIFICATION

Prior to the early 1970's, structural geologists did not recognize the variety of deformation mechanisms operating in nature and ascribed almost all deformation to brittle fracture (Higgins, 1971). Bell and Etheridge (1973) were perhaps the first to point out that most mylonites are the product of plastic deformation of crystals (intracrystalline plasticity) rather than brittle deformation. Since then, a number of different classification schemes for high strain tectonites have been proposed (Zeck, 1974; Sibson, 1977; Wise et al, 1984). The most widely used classification is that of Sibson, who arranged tectonites into two series: cataclasites produced by brittle deformation, and mylonites produced by mainly plastic deformation. There is a need, however, for a more comprehensive classification that incorporates recent advances in the study of teconites.

The classification scheme should take into account the petrology (bulk composition, mineralogy, grain size) of both the protolith as well as the resulting tectonite. The scheme should incorporate the most important parameters that govern the flow of rocks in the solid state, namely temperature, pressure, grain size and stress. The classification scheme should reflect the dominant micromechanisms of deformation in crustal rocks: fracturing (brittle deformation), intracrystalline plasticity, diffusional mass transfer processes, and grain boundary sliding. Variations in strain between tectonites should also be accomodated.

We have chosen three variables for our classification scheme: bulk composition, metamorphic grade, and grain size (Figure 1a). These three parameters are relatively easy to determine in the field and laboratory, and incorporate most of the variables listed above. There are 5 categories for bulk composition, that cover most of the common lithologies: carbonate, pelitic, felsic, mafic and ultramafic (Figure 1b). Transitional compositions such as marls or andesites can be added to this scheme. Grain size is divided into 4 phi categories (Figure 1c), corresponding to clay, silt, sand and pebbles in sediments. The grid works best for rocks with uniform grain size. As pressure and temperature both increase with depth in the crust, these two variables have been combined as metamorphic grade (Figure 1d). There are 5 categories of metamorphic grade: "zero" (unmetamorphosed), very low, low, medium and high grade (Winkler, 1974).

The different tectonites are arranged within this grid, based on their occurrence in nature at different metamorphic grades for a given bulk composition. As in Sibson's (1977) classification, the tectonites are grouped by dominant deformation mechanism into four tectonite series (Figure 2a). Cataclasites produced by brittle deformation are commonly found at very low to low grades of metamorphism. Mylonites, dominated by intracrystalline plasticity, are found at medium to high grades. Quasiplastic mylonites (QP-mylonites), exhibiting both brittle and plastic deformation, are associated with low grade metamorphism. Superplastic mylonites (SP-mylonites) are fine grained teconites found at all grades of metamorphism; grain boundary sliding is the dominant deformation mechanism.

The new classification scheme is similar to a deformation mechanism map, however this tectonite map is generated from empirical observations rather than constitutive equations (Ashby and Verrall, 1978). We are presently compiling data, both from the literature and our own studies in New Jersey, that map out the boundaries of the different tectonites as a function of grade and bulk composition. Our work on carbonate tectonites (Figure 2b), for example, suggests expansion of the mylonite field to lower grades and the superplastic

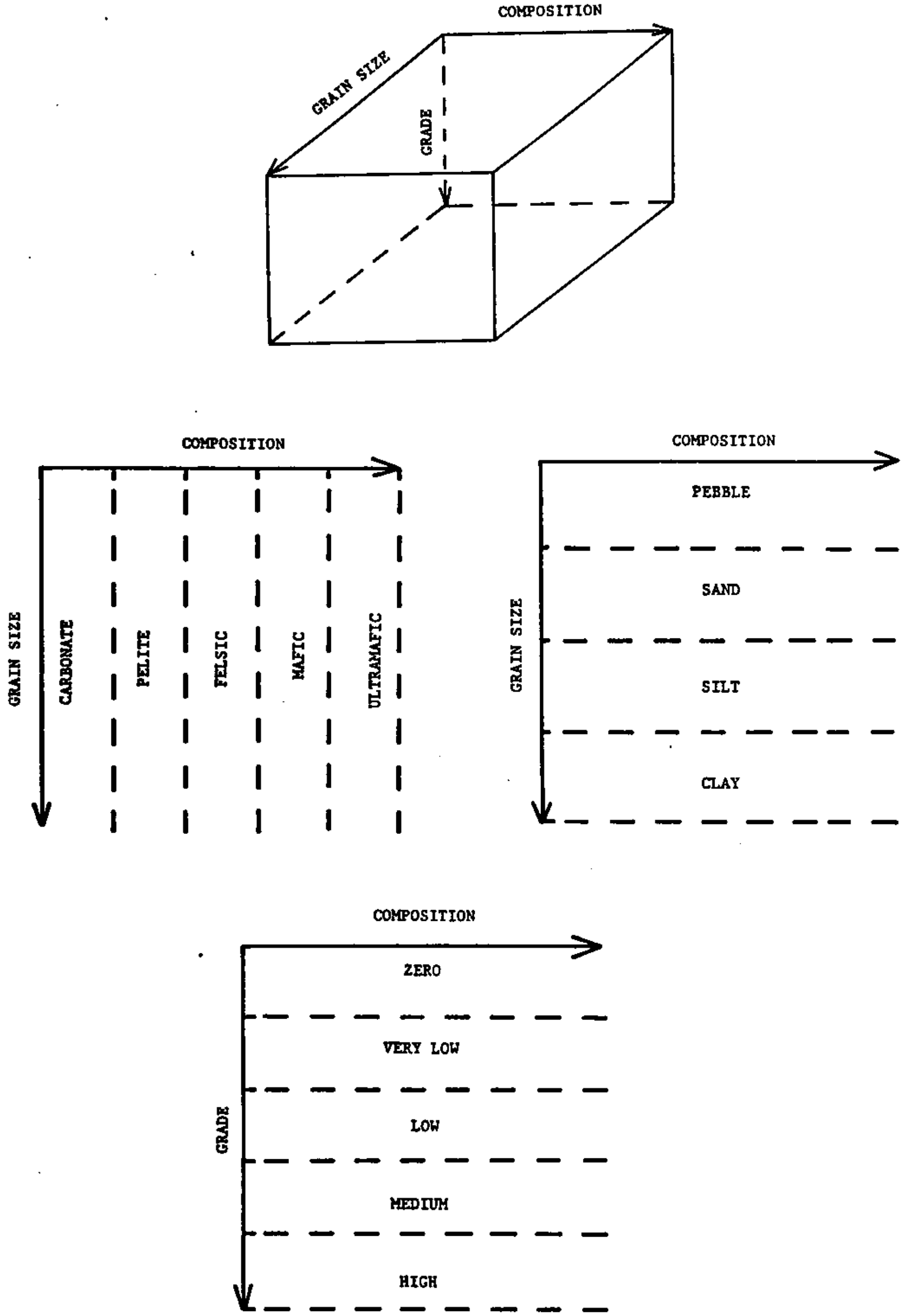


Figure 1. Tectonite grid (a) and subdivisions for composition (b), grain size (c) and metamorphic grade (d). The different tectonites are located within the grid.

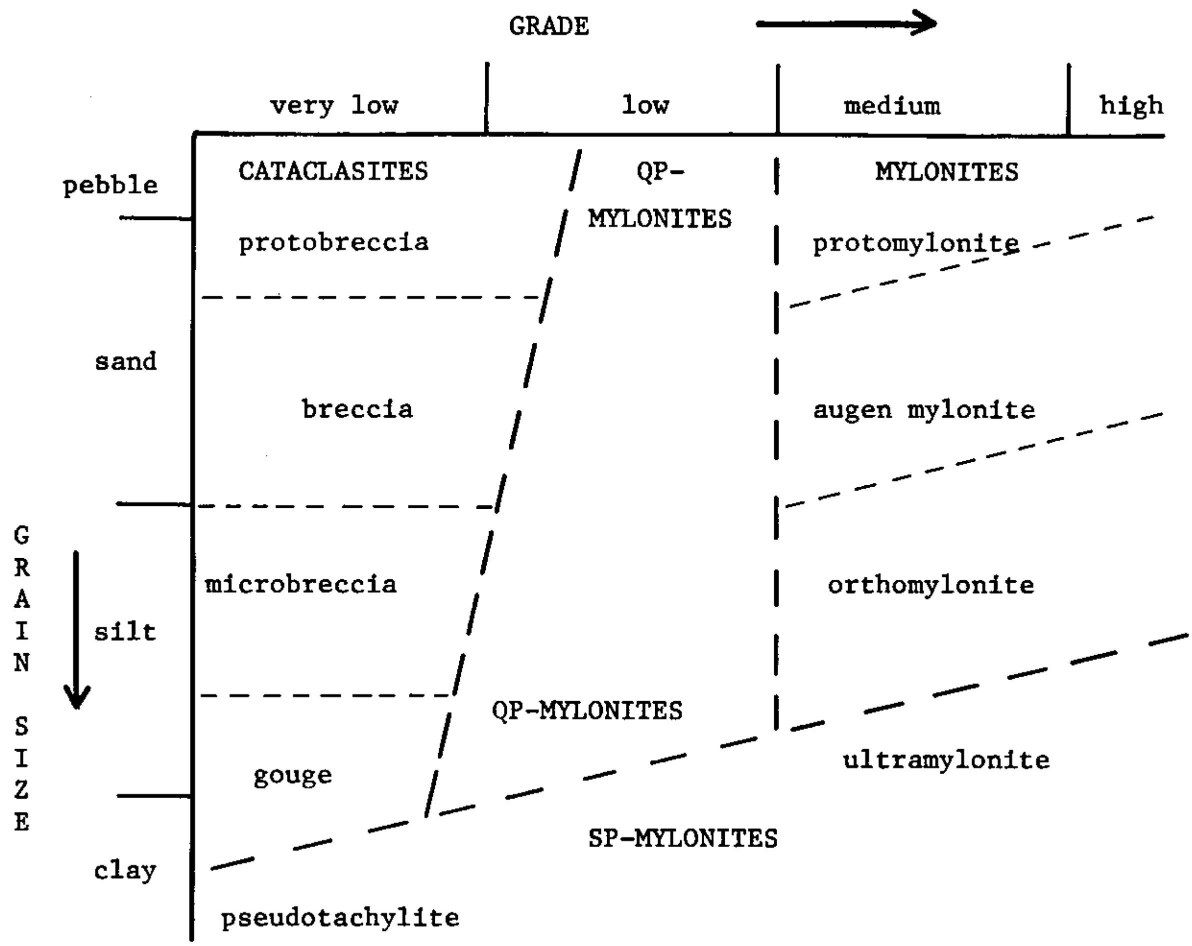


Figure 2. (a). Grade versus grain size at constant composition (felsic in this example). The divisions between the different tectonite series and their members are gradational.

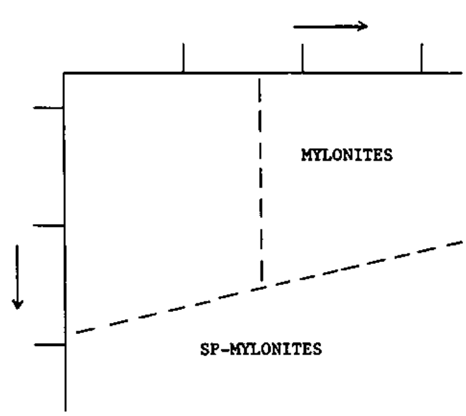


Figure 2. (b). Effect of composition on the position of tectonite series boundaries. Marls, for example, will exhibit expanded fields of plasticity and superplasticity.

field to larger grain sizes relative to quartzofeldspathic rocks. The principle deterrents to this approach are determining the P-T conditions and identifying the contribution by grain boundary sliding, which usually shows no characteristic microstructure.

There are several strengths to the new classification scheme. Much like ternary diagrams in metamorphic petrology, the scheme depicts the types of tectonites that can be expected at certain grades in different lithologies. The scheme is also based on variables which can be easily determined in the field and laboratory. These two characteristics make the classification scheme accessible to non-specialists who find the tectonite literature daunting. One of the principal drawbacks of the classification system of Wise et al (1984) is the choice of essentially unmeasurable variables (recovery rate and strain rate). For the specialist, the grid can incorporate many aspects of a structural investigation, like changes in bulk composition and progressive deformation. Finally, the classification emphasizes the protolith, which is not present in previous schemes.

The composition-grain size plane at "zero" metamorphic grade can be used for either igneous or sedimentary protoliths. For example, sand-sized felsic rocks could either be sedimentary (arkoses, wackes) or igneous (granodiorites, aplites). As the bulk composition dictates the mineralogy (to a first approximation), the differences between these protoliths are small, and the resulting tectonites will be quite similar. A wide variety of metamorphic protoliths can also be specified in this scheme, and both retrograde and prograde deformation can be shown. In addition, the system can illustrate changes in bulk composition accompanying deformation, which is common at low grades, where hydrothermal fluid circulation is most pronounced.

Grain size reduction by fracturing or recrystallization is an important phenomenon for tectonites formed from relatively coarse grained protoliths at very low to medium grades. Grain size has been used to subdivide tectonites in previous schemes (for example, Sibson, 1977): breccias, microbreccias, gouge and pseudotachylite in the cataclasite series, and protomylonites, augen mylonites, orthomylonites and ultramylonites in the mylonite series (Figure 2a). At low grades, brittle grain size is inversely proportional to strain (Draper, 1976), whereas at higher grades, recrystallized grain size is inversely proportional to stress (Weathers et al, 1979).

Grain size reduction will not accompany all tectonism, however. The importance of the initial grain size of the protolith is emphasized in the new classification. If the initial grain size is small, the full series will not be produced, and the resulting tectonite might go unnoticed. Fine grained protoliths will probably coarsen with strain, expecially at medium to high metamorphic grades, where grain boundary migration is an important phenomenon. An extreme example has been described by Hudleston (1980) for glacier ice (a metamorphic rock near its melting point), where the largest grain size is found in high strain shear zones. Tectonites in high grade terrains may show no grain size reduction or even grain growth, and may be very difficult to recognize in the field without strain markers (e.g. McClelland, 1984).

Grain size reduction also produces a change in deformation mechanism, to either mass transfer by diffusion (if the effective diffusional path length is small enough) and/or sliding along grain boundaries (if the aggregate has a equidimensional mosaic or "soap bubble" texture). Ultracataclasites and ultramylonites are therefore grouped in a separate tectonite series, the SP-mylonites, which exhibit very high strains with little or no change in their microstructure (grain size, texture, etc.). Grain boundary sliding is probably an important deformation mechanism in microbreccias and orthomylonites, showing a transition to superplastic behavior. The precise deformation micromechanisms will vary in the superplastic series. At high temperatures (ultramylonites), grain boundary sliding involves motion of grain boundary defects, whereas at low temperatures (ultracataclasites) the sliding is frictional. Fluids and/or high temperatures will promote diffusional mass transfer along the grain boundaries.

The utility of the new classification scheme can be illustrated by two examples. Mitra (1978) has described greenschist facies, Alleghanian ductile deformation zones in Grenville-age quartzofeldspathic gneisses of the Virginia Blue Ridge. Brittle cracking of sand- to pebble-sized feldspars and quartz produced a silt-sized matrix that deformed by pressure solution and plasticity. Though the Grenville assemblages are retrogressed to quartz-chlorite-muscovite-feldspar, the bulk composition has probably not changed substantially, and the retrograde deformation can be shown on a single plane of constant composition. A more extreme example has been described by Bell (1952) from Laramide retrograde zones cutting Archean orthogneisses in the Wasatch Mountains of Utah. Hydrothermal alteration is extensive along these zones, and the bulk composition has changed dramatically. The resulting tectonite is a silt- to clay-sized, superplastic phyllonite composed almost entirely of chlorite. All three variables changed during this deformation and the full grid is utilized.

DEFORMATION ZONE GEOMETRY

In studying the geometry of deformation zones, there are 4 geometric elements that must be routinely determined in the field: the deformation zone <u>orientation</u>, the <u>shear direction</u>, the <u>shear sense</u> and the <u>net slip</u>. Most low temperature, brittle deformation zones (i.e. faults) will conform to the simple shear or card deck model (Figure 3a). However at higher temperatures (more ductile deformation), bulk heterogeneous shortening (Bell, 1981) can produce deformation zones that superficially resemble simple shears (Simpson, 1981). Attempts to analyze these deformation zones in terms of the simple shear model will give spurious results. In this paper we avoid the term "shear zone," recognizing that most zones have some component of shear, but not all are simple.

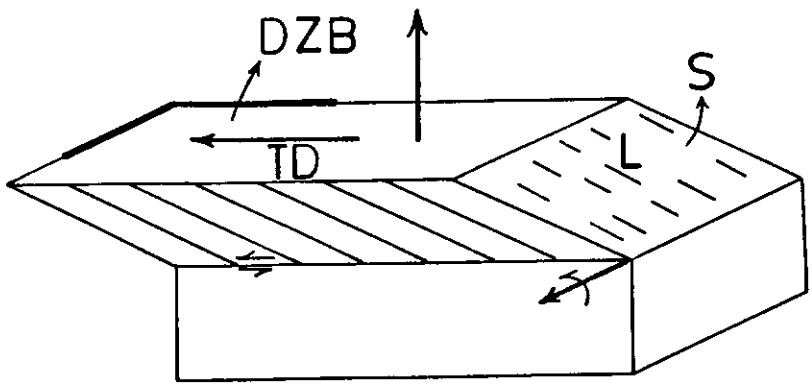


Figure 3. (a). Simple shear zone (sinistral in this view). The pole to the deformation zone boundary (DZB) and the transport direction (TD) are coplanar with the pole to the foliation (S) and the stretching lineation (L). This plane is perpendicular to the rotation axis.

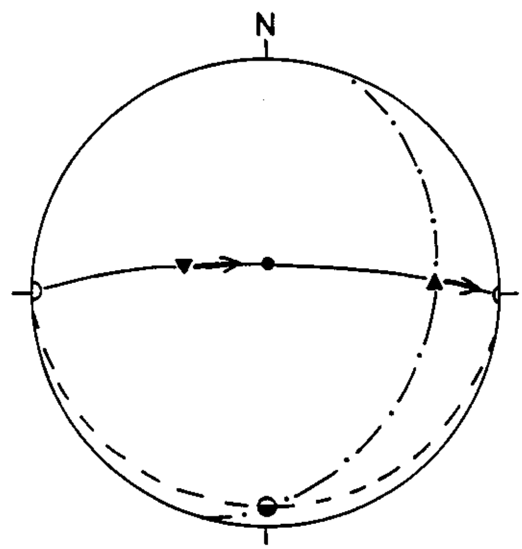


Figure 3. (b). Corresponding stereonet illustrating the simple shear geometry. The deformation zone (dashed line and solid dot), transport direction (open circle), foliation (dash-dot and triangle down), lineation (triangle up) and rotation axis (half-filled circle) are shown.

The orientation of the deformation zone is usually expressed as the pole to the deformation zone boundary. Deformation zones anastomose in three dimensions, and it is important to record this variability. Measurement of the dimensions of the zone is usually inhibited by the size of the outcrop or the quality of exposures. The terminations (either tips or branch lines) of the deformation zone and its cross-cutting relationships are important in assessing the relative age of the zone. The fabrics and structures within the deformation zone itself (foliation, lineation, subsidiary deformation zones, etc.) are also critical.

The transport or shear direction parallels the deformation zone boundary and should not be confused with the stretching lineation (Figure 3b). For continuous simple shear zones, the shear direction can be constructed stereographically from the structural elements. For discontinuous faults, the shear direction can be determined using an offset piercing point or slickenlines on the fault surface. Slickenlines may record only the last increment of movement, and to be reliable, their length should be comparable to the net slip on the zone.

There is a fast-growing literature on determining shear sense in tectonites (Simpson and Schmid, 1983; Simpson, 1986; see also a forthcoming volume of the *Journal of Structural Geology*). Rotation of markers or foliation in high temperature continuous zones are commonly used, but the zone must show simple shear. Quasiplastic tectonites often exhibit S-C fabrics: schistosity (S) rotated into small spaced deformation zones (C). Logan et al (1979) have described some of the shear sense indicators found within cataclasites. Surface features (tectoglyphs) on brittle sliding surfaces are also useful (Tjia, 1971; Angelier et al, 1985). Fibrous-type slickensides often exhibit steps, which are reliable, but steps on tool and groove type (polished) slickensides are usually absent or misleading.

The amount of displacement or net slip on discontinuous deformation zones can be determined using an offset piercing point or just offset markers if a slickenline is present. In the Highlands, these two methods yield similar results for the same fault zone. For continuous zones which satisfy the simple shear criteria, rotated markers can be used (Ramsay, 1980). The width of the deformation zone also provides a crude estimate of displacement (Hull, 1986). Width-displacement ratios vary from 1:2 for the mylonite series to 1:100 for the cataclasite series, with no fixed relationship for SP-mylonites.

FAULT POPULATION ANALYSIS

Structural geologists and geophysicists use the "right hand rule" to express the geometry of the deformation zone and its displacement (Figure 4). The upward normal or pole to the deformation zone is the index finger of the right hand, while the middle finger represents the transport or shear direction. The pole and the shear direction define

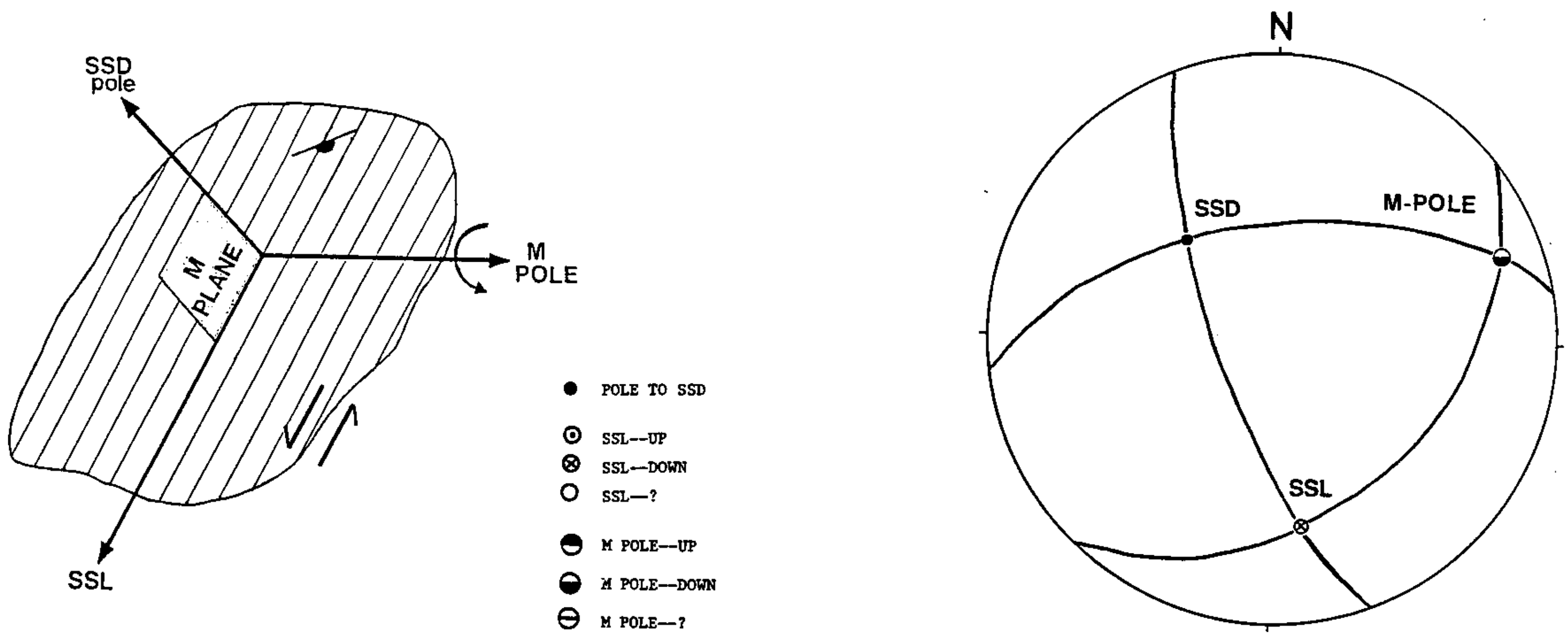


Figure 4. A slickenside (SSD)-slickenline (SSL) pair illustrating the right-hand convention. Both the slickenline and the M-pole (rotation axis or pole to the movement plane) point down in this example, as can be easily seen on the corresponding stereonet.

the movement plane (M-plane) of Arthaud (1969). The thumb is therefore parallel to the rotation axis ("M-pole"), and the fingers of the right hand curl in the direction of shear or displacement. Using the right hand rule, every deformation zone exhibits sinistral shear. The three mutually perpendicular directions are all vectors: the pole to the deformation zone always points up, whereas the transport and rotation vectors may point up or down. Transport lineations point up for reverse faults and down for normal faults. The complete geometry of a set of deformation zones, including their movement sense, can be shown easily on a stereonet using the right hand convention.

The geologist must first identify a set of deformation zones produced during a single deformation event. This task can be quite difficult, as in the Highlands of New Jersey, where basement rocks have undergone multiple deformations. Coeval deformation zones should have similar structural features and tectonites, with similar mineralogies and textures. The zones should have the same cross-cutting relationships with intrusives or other deformation zones. A coeval set of deformation zones will be mutually cross-cutting, indicating their simultaneous development. These criteria are the minimum required to establish synchroneity, but are not necessarily sufficient, as will be discussed later.

Once a deformation zone set is identified and the requisite structural data gathered, the problem of interpretation can be confronted. An excellent example of fault population analysis is presented by Wojtal (1986). So far there is no universal algorithm available that can invert the structural data to strain axes. The patterns of deformation zones in the oft-cited method of Arthaud (1969) have not been seen in the Highlands, and the Arthaud method is applicable to pure shear only (Carey, 1976). Our approach has been to examine the relationship between the strain axes and deformation zone geometries for different models of bulk deformation. Here we present some simple end-member models, illustrating both two-dimensional strain (simple and pure shear) and three-dimensional strain (pure stretching and flattening).

A population of parallel deformation zones, all with the same shear direction and shear sense, will produce bulk simple shear of the rock mass (Figure 5a). Simple shear is both plane (two-dimensional) and rotational, and there is no fixed relationship between strain and stress axes. On the stereonet (Figure 5b) the deformation zones show three point maxima, with each set of vectors showing uniform polarity. The rotation axes (m-poles) are parallel to the Y (intermediate) axis of strain, and perpendicular to the XZ strain plane. The angle between the stretching direction X and the transport direction for the deformation zones is not fixed, and decreases with increasing amounts of shear. The magnitude or amount of shear in the outcrop could be determined by summing the displacements on every deformation zone, but this fortuitous situation is unlikely in the field.

A population of conjugate deformation zones, with opposite shear

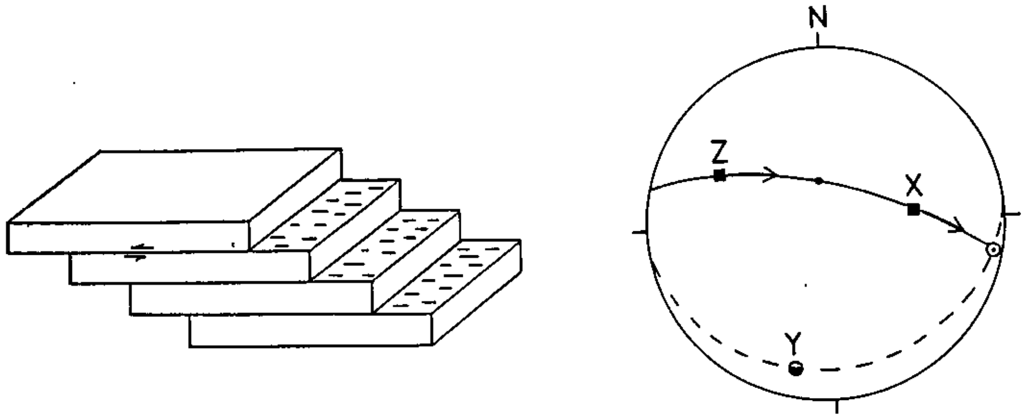


Figure 5. (a). Parallel deformation zones producing bulk simple shear. (b). Corresponding stereonet showing rotation of stretching (X) and shortening (Z) directions towards deformation zone elements.

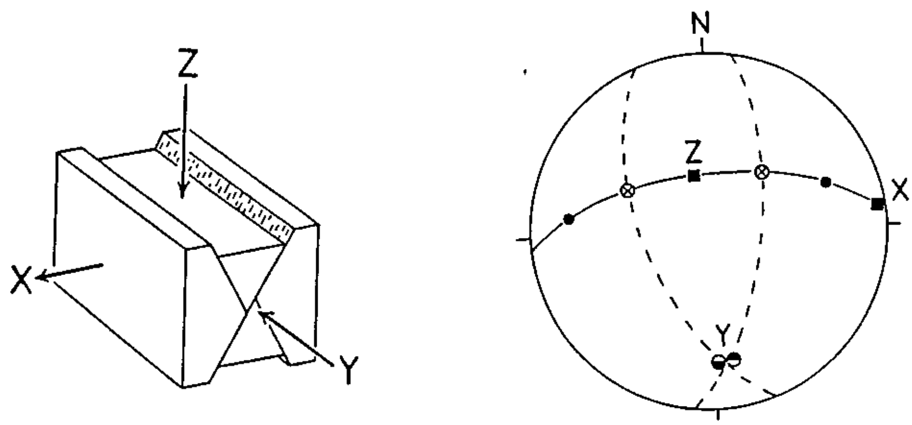


Figure 5. (c). Anderson-type conjugate deformation zones in the shallow crust (dihedral angle = 60°). (d). The stretching direction (X) bisects the acute angle between the poles to the deformation zones.

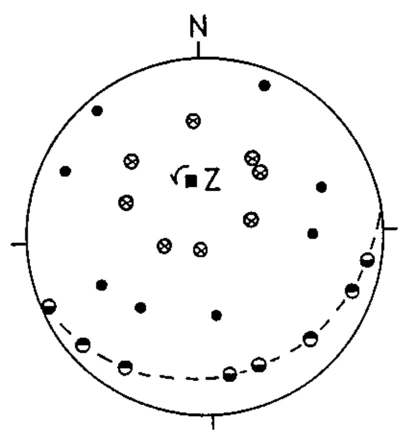


Figure 5. (e). Pure flattening illustrating conical distribution of poles to deformation zones and transport directions, and single girdle of rotation axes of mixed polarity.

senses (a "double couple"), will produce bulk pure shear of the rock mass (Figure 5c). This Anderson model of strain is also plane but irrotational; the axes of stress and strain are fixed. Rotation axes define a single point maximum but of dual polarity, distinct from the simple shear model (Figure 5d). Poles to deformation zones and transport directions define either double maxima, or with the expected scatter, partial girdles. The dihedral angle between the conjugate deformation zones determines the orientation of the other two principle strain axes. For shallow deformation zones (faults), the Z axis (shortening direction) bisects the acute angle between the zones, whereas the X axis (stretching direction) bisects the acute angle for zones formed at higher metamorphic grades (Kligfield et al, 1984). For fault zones, the shortening direction Z is thus parallel to the partial girdle of transport lineations, a somewhat contraintuitive result.

Pure flattening and pure stretching are both three-dimensional deformations, producing pancake- and cigar-shaped strain ellipsoids, respectively. Unfortunately the geometries of the deformation zones will probably be identical for these two strain states. A pure flattening deformation (for example) can be produced by revolving the Anderson model around the shortening direction Z (Figure 5e). Rotation axes will define a single great circle girdle whose normal (pole) is the shortening direction. For a pure stretching deformation, the pole to the great circle is the extension direction X. Poles to deformation zones and transport lineations will define small circle girdles (conical distributions).

The number of faults produced in a general, three-dimensional deformation can be predicted by analogy with intracrystalline plasticity (Mitra, 1979). Von Mise's criterion states that a minimum of 5 independent slip or glide systems are required for an aggregate of crystals to deform plastically, without producing gaps or overlaps. Mitra has observed 6 sets of deformation zones in basement rocks of the Virginia Blue Ridge (see also Aydin and Reches, 1982). The geometry of these zones and the principal directions of strain are difficult to predict, however.

The principal strain axes determined in our studies by fault population analysis compare well with minor structures such as veins and folds, however the methodology is not without problems. A pre-existing fabric could influence the orientation of deformation zones, though for retrograde zones in stiff basement this problem seems to be minimal. The analysis can only be applied to the last tectonic event, as a second deformation will redistribute an earlier fault population. Once formed, deformation zones may rotate as material elements towards the stretching direction with progressive deformation. Rotation of deformation zones is usually not a problem at low temperatures, where lithons are stiff and undeformable. Near major faults, the basement is often highly retrogessed, and folding of deformation zones can be seen.

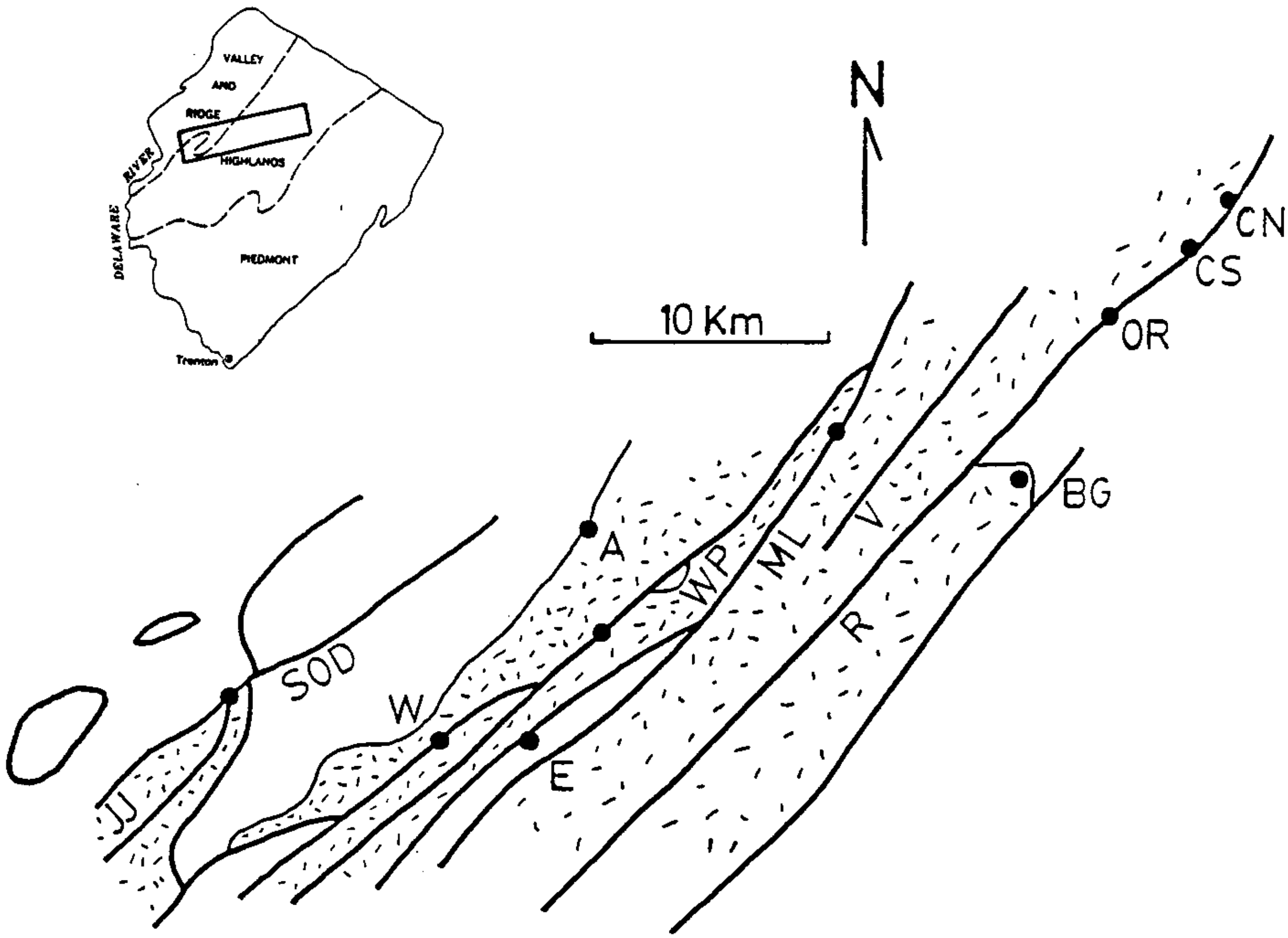


Figure 6. Location map showing distribution of major faults (heavy lines), basement (ticked), the basal unconformity (light lines) and localities referred to in the text. Deformation zones (from west to east) include the Jenny Jump (JJ), Shades of Death (SOD), Wright Pond (WP), Morris Lake (ML), Vernon (V) and Reservoir (R) faults. Localities (from west to east) include the west (W) and east (E) flanks of Allamuchy Mountain, the quarry near Andover (A), Bowling Green Mountain (BG), and three exposures of the Reservoir fault: Oak Ridge Reservoir (OR), southern Clinton Reservoir (CS) and northern Clinton Reservoir (CN). Map data compiled from Baker and Buddington (1970), Barnett (1966), Drake and Lyttle (1986), Drake et al (1978), and the present work.

GRENVILLE DEFORMATION ZONES

Grenville tectonism in the Reading Prong is characterized by penetrative deformation at amphibolite to granulite facies (Drake, 1969; Drake, 1970; Drake, 1984). The ages of the varied protoliths are not known, but the deformation is dated at 1.1-1.2 Ga (Dallmeyer et al, 1975). The Grenville metamorphic rocks exhibit a schistosity and/or stretching lineation that is parallel to gneissic banding of unknown and probably varied origin. Most rocks are thoroughly recrystallized, obscuring primary textures and microstructures, and occasionally the type of protolith as well (orthogneiss versus paragneiss).

The relationship between protolith and gneiss is preserved in at least one locality, in the Allamuchy Mountains along I-80 (Figure 6). This exposure is described in Stop 3 of the accompanying field guide. The roadcuts expose quartzofeldspathic and amphibolite gneisses that are both cut by undeformed pegmatites. The contact between the two gneisses is irregular and generally subparallel (though locally oblique) to foliation (both schistosity and gneissosity). Near the contact, tabular xenoliths of amphibolite within the felsic gneisses suggest an intrusive origin.

Essentially undeformed (though metamorphosed) protolith is preserved in outcrop- to football-sized lithons surrounded by gneiss (Figure 7a). The larger lithons are quite rounded at their terminations and show little internal deformation (except for a few thin zones), whereas smaller augen are thin and tapered, containing small deformation zones and a weak fabric. Their boundaries may be sharp or gradational into the adjacent gneiss. In places the foliation makes a small angle to the boundary of the augen (Figure 7b), though in general foliation is subparallel to the boundaries. The gneiss exhibits a mineral and shape lineation (S > L tectonite) at a low angle to the rotation axis for simple shear deformation (Figure 8a). A similar fabric geometry has been observed by Simpson (1981) in amphibolite facies orthogneisses of the Maggia Nappe, Switzerland. The concordant foliation, flattening strain and fabric geometry all suggest deviation from the simple shear model. The deformation zones were most likely produced by heterogeneous bulk flattening, rather than simple shear (see Bell, 1981).

Several authors (e. g. Puffer, 1980) have argued that the presence of layering (gneissosity) implies a sedimentary precursor. The field relationships at Stop 3 illustrate how even a relatively homogeneous igneous protolith can be transformed to gneiss simply by deformation. Unfortunately the undeformed protolith is rarely preserved in the Grenville of New Jersey. For example, the quartzofeldspathic gneiss to the north and east has the same fabric and structure as the mafic gneiss but macroscopic augen (lithons) have not been observed. The difference in mineralogy (quartz, alkali feldspar and hornblende versus plagioclase, hornblende and diopside) probably accounts for a small difference in strength at this grade of metamorphism.

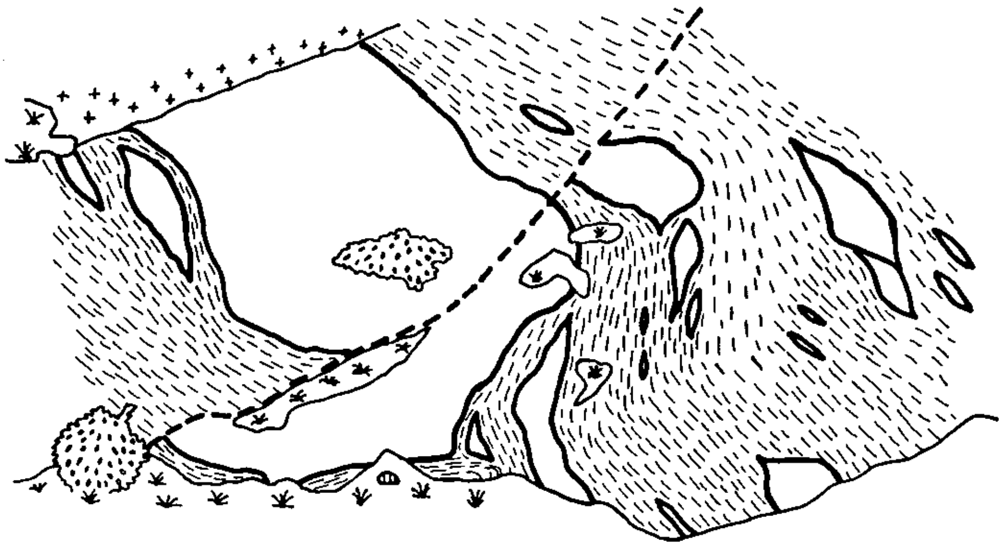


Figure 7a. Outcrop sketch illustrating undeformed lithons or augen of mafic metaporphyry surrounded by gneiss. Both are cut by pegmatite (at top) and epidote-rich fault (dashed line). Scale is 1:200. Stop 3.

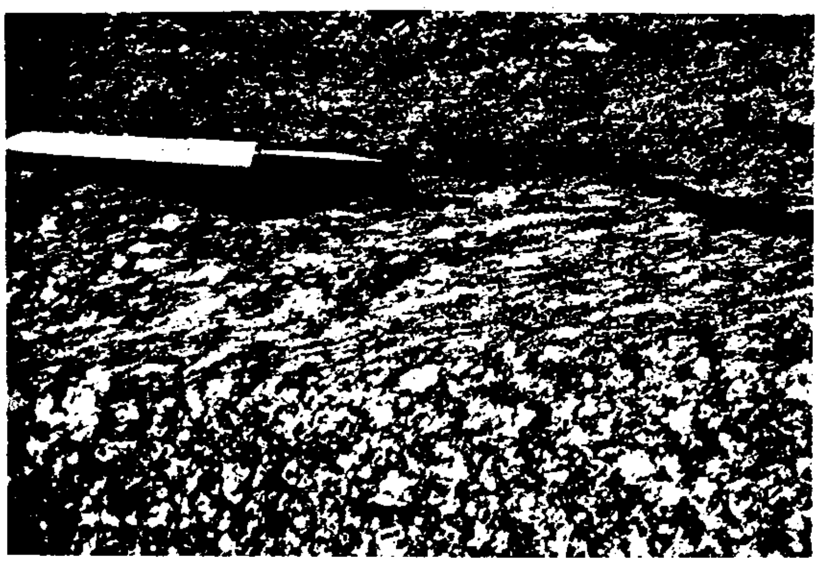


Figure 7b. Undeformed plagioclase metaporphyry in sharp contact with laminated gneiss containing ellipitical plagioclase augen. Schistosity makes a small angle to the deformation zone boundary. Pen for scale.

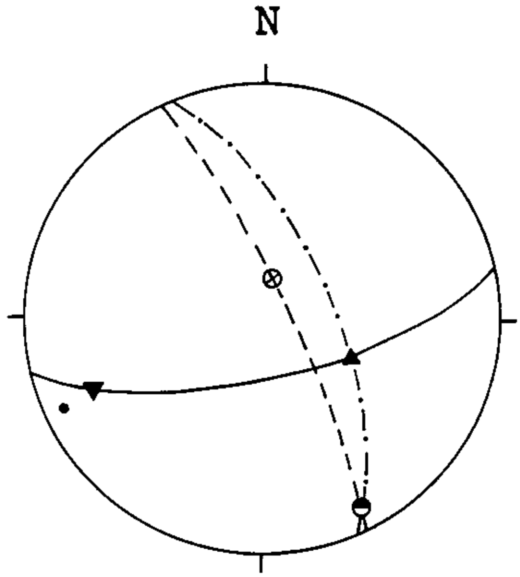


Figure 8a. Structural elements for a lithon in Figure 7a. The fabric and deformation zone elements are not coplanar, suggesting deformation other than simple shear. Lower hemisphere, equal area projection.

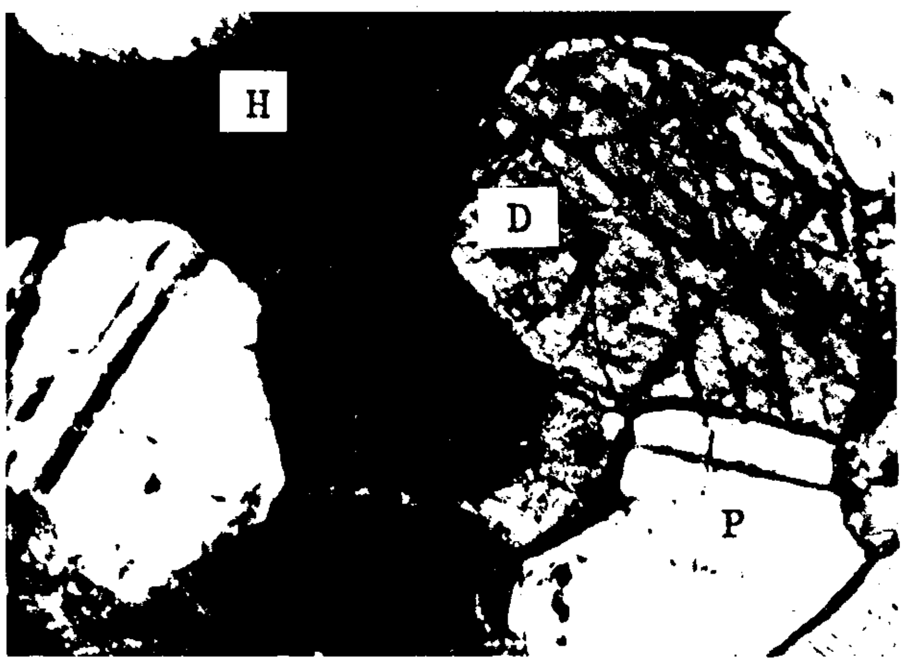


Figure 8b. Photomicrograph of amphibolite gneiss composed of plagioclase (P), hornblende (H) and diopside (D). Note equant mosaic, 120° triple junctions and curved grain boundaries indicating high temperature, syntectonic recrystallization. Scale 20X.

The protolith is a coarse-grained metaporphyry composed of plagioclase phenocrysts set in a hornblende-plagioclase-diopside groundmass. The cm-sized phenocrysts are compositionally zoned and contain various twin sets (mostly albite law). The metaporphyry is deformed to produce augen gneiss and finely laminated "pin stripe" gneiss. Individual phenocrysts are flattened into millimeter thick plagioclase laminae, illustrating the magnitude of the strain in the orthogneiss. Grain size reduction by high temperature recrystallization accompanies the deformation (Figure 8b). The resultant equigranular, subequant mosaic is characterized by abundant 120° triple junctions and curviplanar grain boundaries, indicating grain boundary mobility.

The gneisses at Stop 3 are intruded by a network of unmetamorphosed and essentially undeformed pegmatites, many of which are rich in magnetite. These pegmatites are in turn cut by the retrograde deformation zones described in the next section. The pegmatites thus provide upper and lower bounds (respectively) to the ages of the two principal deformations in the Highlands, however the pegmatites in New Jersey have not been dated. The pegmatites might be correlative with or slightly postdate the late orogenic Canada Hill intrusive suite in the northernmost Reading Prong, which has been dated at 900 Ma (Helenek and Mose, 1984).

A consistent sequence of retrograde deformation zones cutting the Proterozoic rocks and structures can be found throughout the Highlands (Figure 9a). Early chlorite-rich deformation zones are offset by epidote-rich zones (Figure 9b), which are in turn cut by late chlorite zones. The ages of these retrograde zones are not well known.

EARLY CHLORITE-RICH ZONES

The early chlorite faults are characterized by broad, irregular zones of retrograde deformation. Both the boundaries and geometries of these early faults are difficult to define, due to later tectonism. Folding of these faults is suggested by ubiquitous minor folds of the retrograde foliation. Almost all of the tectonites associated with the early deformation belong to the cataclasite or SP-mylonite suites.

The protolith adjacent to the chlorite faults shows extensive retrogression of pyriboles to chlorite and zircon, and lesser alteration of feldspars to muscovite and a little epidote. Unstable (transgranular) shear fractures and veins filled with fibrous chlorite are usually found at low angles to the chlorite zones (Figure 10a). Quartz exhibits undulatory extinction and irregular deformation lamellae at high angles to long narrow subgrains (polygons) but little recrystallization is observed. Feldspars are not plastically deformed.

In small faults, feldspars are quite altered and tartan twins, myrmekite and perthitic exsolution lamellae are abundant. Recrystallization of quartz along grain boundaries has produced a core and mantle microstructure that superficially resembles grain boundary suturing.

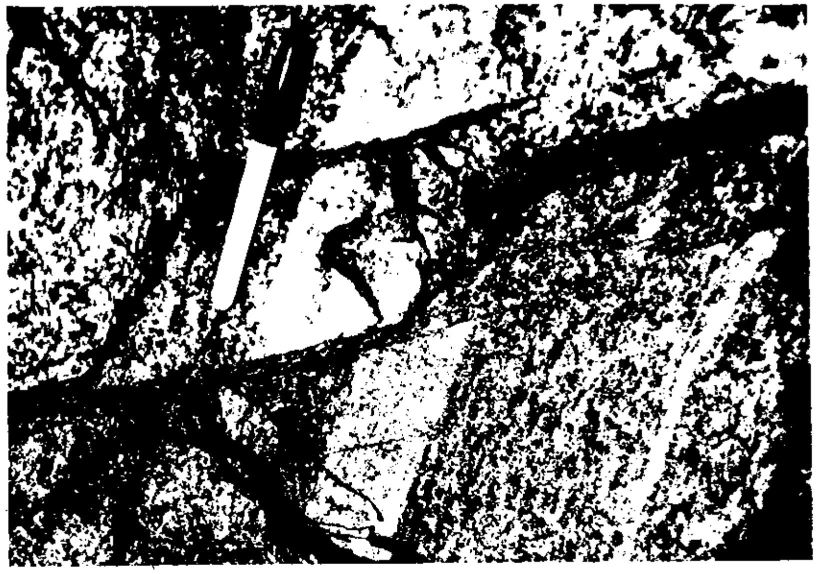


Figure 9a. Chlorite-rich brittle deformation zone (BDZ) offsetting foliated aplite sills or dikes in felsic gneiss. Transport direction lies parallel to page. Pen for scale. Allamuchy Mountains.

Figure 9b. Cross-cutting relationships among retrograde BDZ. Chlorite zone (C) cut by vertical, dip-slip epidote zone, in turn cut by horizontal epidote zone (pen cap parallel to slickenline). Allamuchy Mtns.

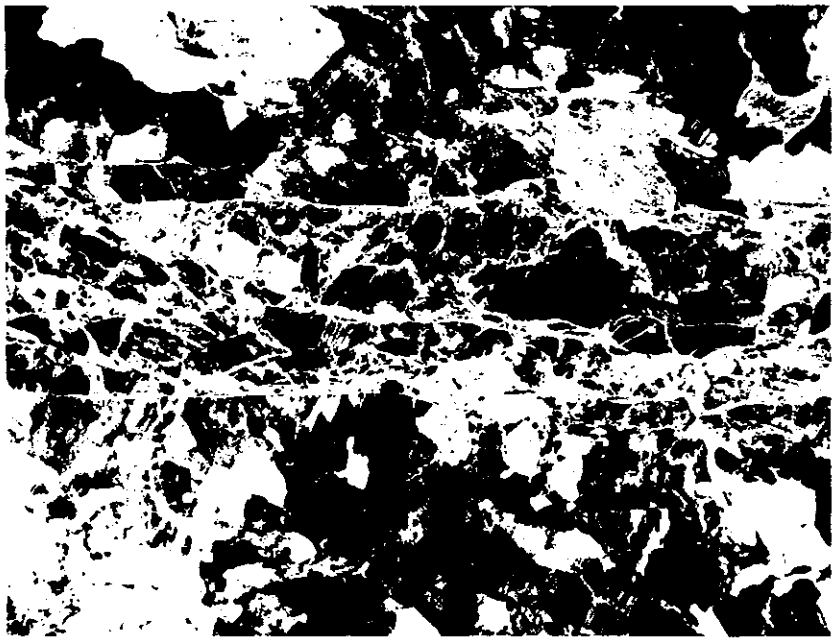


Figure 10a. Negative image of chlorite-rich protobreccia. The negative image is produced by placing the thin section directly in the enlarger with polaroid sheets. Several anastomosing sets of small BDZ contain microbreccia. Scale 12X.

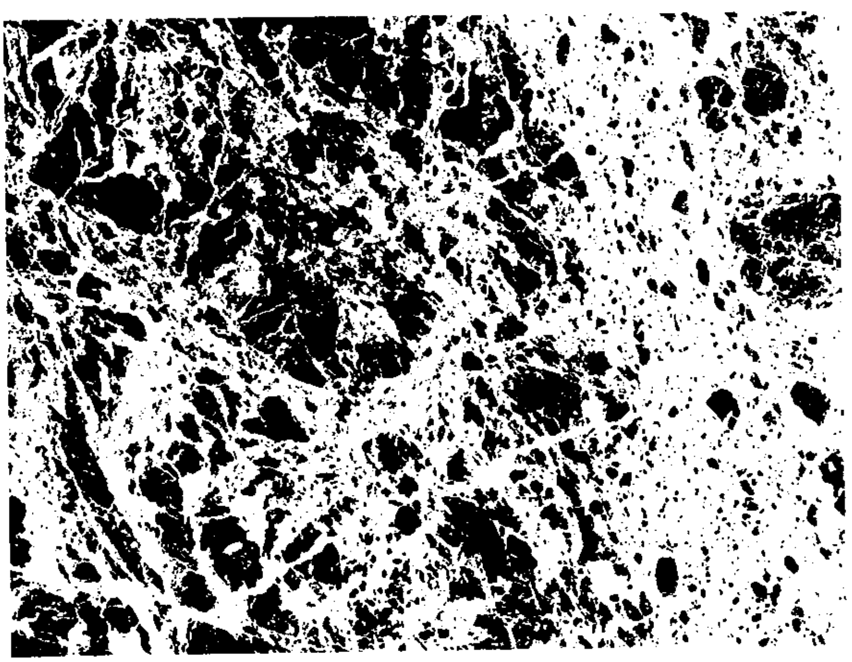


Figure 10b. Negative image of highly retrogressed protobreccia cut by sets of small BDZ and strongly foliated breccia, microbreccia and gouge. The foliated cataclasites are rich in chlorite and therefore appear opaque (white) under partially crossed polars. Scale 5X.

Fracturing and plasticity are often associated in these tectonites. Narrow zones of finely recrystallized quartz are seen along shear fractures, and feldspar twin lamellae are kinked at fracture tips. Fracturing is the dominant mechanism in reducing grain size as recrystallization is still limited.

In the largest zones, the breccias and microbreccias are initially composed of variably sized, angular fragments of quartz and feldspar with a small proportion of matrix (Figure 10b). With further deformation, both unstable and stable (intragranular) fracturing produces a matrix composed of chlorite, zircon, opaque, and small monocrystalline fragments of quartz and feldspar. The clasts are rounded to angular, and rotate by propagating extension fractures and deformation of the ductile matrix. The clasts are dominated by feldspar that shows a limited amount of recrystallization. The matrix of some dense, hard, almost flinty ultracataclasites appears to be more recrystallized.

Determining the geometry and strain axes of the early chlorite suite is difficult, because they are usually overprinted by the epidote zones, which redistribute the chlorite population. We have found several locales where epidote zones are essentially absent, but it then becomes difficult to identify the relative age (early or late) of the chlorite-rich faults. A rather simple population of chlorite zones can be seen in an abandoned gravel quarry near Andover (Figure 6), which will be visited at Stop 4.

The quarry exposes quartzofeldspathic paragneisses (?) intruded by highly deformed pegmatites. The paragneisses are composed of quartz, microcline, plagioclase and varying proportions of biotite, sillimanite and hornblende. Layers rich in biotite, hornblende, spinel and calc-silicates form distinctive marker horizons. Foliation dips shallowly to the east while the lineation plunges to the northeast (S>L tectonite). Scar folds associated with boudinage of layering are common and sheath folds have been observed in one locality, but overall the foliation is not folded. Highly deformed felsic pegmatites show both chocolate-tablet boudinage (parallel to layering) and ptygmatic folds (perpendicular to gneissosity). A few pegmatites postdate the deformation.

The zones contain coarse breccias and microbreccias (Figure 11a), with both tool and groove and fibrous type slickenlines on the fault zone boundaries. There is some evidence that suggests the different slickenlines formed sequentially. Most of the chlorite zones cut the Grenville foliation at a high angle and offsets are easily determined (Figure 11b). The zones seem anomalously wide given the small displacements typically found.

The chlorite zones form steep, north- and south-dipping conjugates with the south-dipping set predominant (Figure 12a). These oblique normal faults extend the subhorizontal gneissosity. Lineations plunge to the west but show some girdling in an east-striking vertical plane. A single maximum of m-poles with dual polarity plunges moderately to the east. The fault population thus suggests a small amount

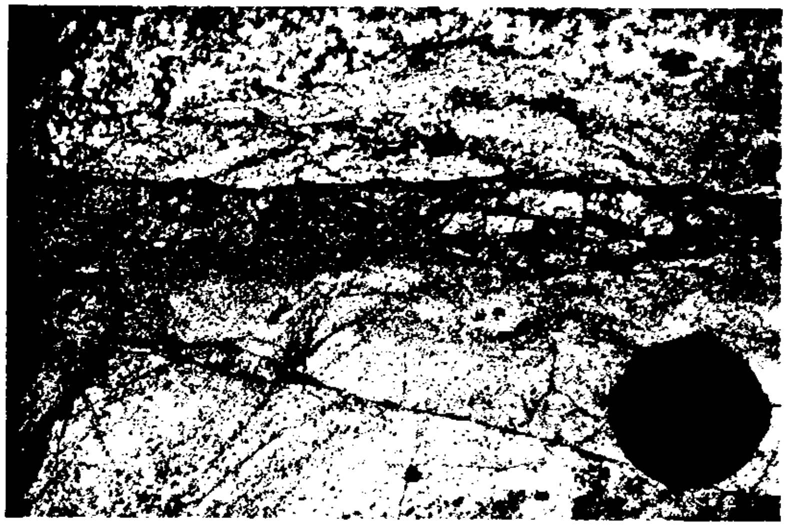


Figure 11a. Chlorite breccia containing tabular clasts of paragneiss, exposed on foliation surface (lineation trends northeast). Fracturing parallel to the BDZ boundary can be seen. Stop 4.

Figure 11b. Conjugate chlorite BDZ cutting paragneiss. Normal, oblique displacement of the marker M (towards the viewer) suggests subhorizontal north-south extension. Stop 4. Scale 1:10.

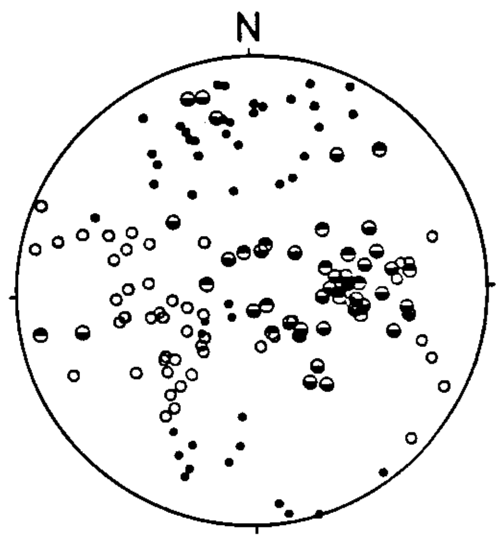


Figure 12a. Chlorite BDZ from Stop 4 near Andover. The two sets of BDZ (south-dipping set dominant) have parallel rotation axes with opposite polarities. Slickenline polarity not shown (mostly normal). Lower hemisphere, equal angle.

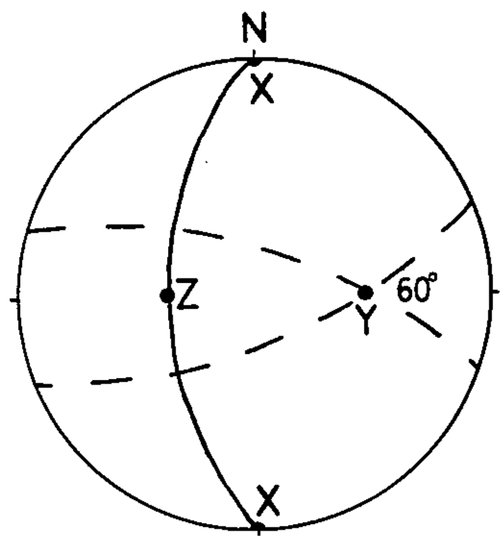


Figure 12b. Corresponding plane strain, pure shear geometry with subhorizontal, NS extension (X) and west-plunging shortening (Z). Anderson conjugates form a 60° angle. Lower hemisphere, equal angle.

of pure shear (Figure 12b), with north-south, horizontal extension and moderately west-plunging shortening. A component of simple shear (down to the east) is indicated by the dominance of the south-dipping set. In addition, a few (though substantial) subhorizontal west-directed thrust faults can be found here.

The quarry is located near the north-striking contact between Grenville basement and Cambro-Ordovician sedimentary rocks to the west (Drake et al, 1978). The contact has been mapped as the basal unconformity, but our work to the south in the Tranquility quadrangle favors a thrust (though the contact is not exposed). The structural setting is not really known. One possibility is longitudinal extension along a frontal ramp; minor extension parallel to fold axes of ramp anticlines in the cover is common in fold and thrust belts. The extension could also be related to emplacement of the Beemerville-Cortlandt suite, for example.

EPIDOTE-RICH ZONES

Some of the most striking retrograde deformation zones in the Highlands are rich in apple green epidote. These deformation zones can be quite extensive (Figure 13a), as they are usually not offset by later faults. The zones vary from millimeters to decimeters in thickness, with sharp boundaries against the country rock. Like all deformation zones, the epidote-rich faults anastomose or curve in three dimensions, with one of two geometries. Most zones have an axis of curvature that parallels the transport direction; slip will not produce rotation. But a few zones have axes of curvature that are perpendicular to the slickenline, producing large rotations during slip. These latter zones are not folded, but probably form to accomodate space problems near deformation zone intersections.

As an epidote-rich brittle deformation zone is approached, the wall rock shows increased static alteration (to epidote, chlorite, muscovite, etc.) and abundant, essentially monomineralic veins (Figure 13b). Epidote veins often cut chlorite patches (after pyriboles). Undulatory extinction, deformation lamellae and elongate subgrains in quartz record a small amount of intracrystalline plasticity, but recrystallization is very restricted. Plastic deformation of feldspars is minor (undulatory extinction and bent twins) but exsolution (myrmekite and perthite) is common. Transgranular (unstable) cracks are associated with kinked and offset plagioclase twins.

A network of shear fractures subparallel to the deformation zone boundary produces elongate to discoidal, polymineralic fragments that define a foliation in the resulting breccia. Progressive fracturing and alteration of clasts and wall rock produces a fine grained matrix rich in epidote, quartz, chlorite, zircon and opaques. Some of these shear fractures propagated along the epidote veins, as in the Sierras (Segall and Simpson, 1986). Relatively large, euhedral epidote crystals along BDZ boundaries and broken epidote crystals and clasts of coarse epidosite in the breccia represent the early vein fill. Most of

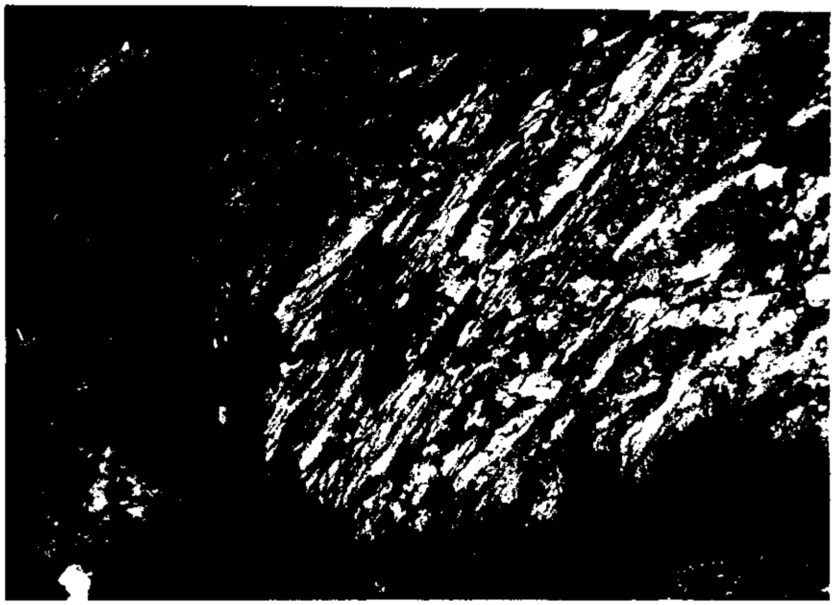


Figure 13a. Large, dip slip epidote zone cutting felsic gneisses in the Allamuchy Mountains. The zone is over 10 cm thick and contains several polished sliding surfaces, including the one exposed. Bob Koto gives scale.

Figure 13b. Epidote zone in thin section (negative image). Offset of feldpar twins in the protolith (lower) can be seen. The BDZ contains breccia and microbreccia. The zone is bounded on both sides by sliding surfaces associated with ultracataclasites. Scale 8X.

the clasts are cut by intragranular (stable) cracks that die out in the matrix at the clast boundaries. The small clasts in the microbreccias are almost always single crystals of quartz or alkali feldspar, and may be slightly elongate to round. The role of frictional wear in reducing grain size and shaping clasts is not known.

The deformation micromechanisms of the matrix are not well understood. Grain boundary sliding is probably quite important but does not produce diagnostic microstructures. The actual sliding mechanism (frictional sliding, pressure solution slip, or high temperature gbs) is not known. Occasional strain hardening in the matrix is shown by "breccia in breccia" texture (clasts of breccia in the cataclasite) and quartz veins at a high angle to the slickenline. Among other late adjustments of the cataclasites, coarser-grained, opaque-poor epidote veins and replaces (?) the epidote breccia.

Sliding surfaces (slickensides) are usually found along one or both of the BDZ boundaries, as well as within the zone itself (Figure 13b). The slickensides are tool-and-groove type surfaces, varying from pasty to highly polished, but never fibrous. The surfaces may be knife sharp and often cut across tectonite banding, juxtaposing different cataclasites. The grooves vary from small, tapered scratches to corrugations with wavelengths over a meter. The sliding surfaces are associated with micron- to submicron-sized, highly indurated (recrystallized?) gouge. Flow structures characteristic of pseudotachylite have not been observed. It is not clear if the slickensides are the natural result of progressive grain size reduction or if the sliding surfaces develop in pre-existing ultracataclasite. Microstructural evidence for both scenarios can be found in a single thin section.

Aydin (1978) has shown that most of the displacement on brittle deformation zones is associated with the sliding surfaces. The large displacements on narrow zones indicate very high angular shear strains (displacement/width), consistent with the fine grain size of the ultracataclasites. Epidote-rich zones in the Highlands have angular shear strains of approximately 100 (Hull, 1986), but the displacement due to the slickensides alone cannot be factored out. In addition, the sliding surfaces may exhibit a different sense and/or orientation of slip than the breccia zone itself. On the few zones where both displacement methods are available, the slip determined using markers and slickenlines is similar to that determined using an offset piercing point.

In addition to grooves, a variety of other tectoglyphs are present on the sliding surfaces (Tjia, 1971; Angelier et al, 1985), including steps, cornices and scoops. Steps are unreliable indicators of shear sense on tool-and-groove type slickensides, and are usually produced by other shear fractures intersecting the slickenside surface. Cornices of microbreccia and gouge dragged over open fractures are the most reliable tectoglyph on the epidote zones. We are currently studying the geometry of tectoglyphs on epidote zones whose shear sense is independently known.

The variable geometry of epidote-rich brittle deformation zones is illustrated by fault populations from two localities on either side of the Allamuchy Mountains (Figure 6). On the west flank of the range along I-80, epidote zones are well developed to the west of a broad chlorite-rich zone. Many large polished surfaces are exposed here (Figure 13a), however displacement sense is very hard to determine. Bulk simple shear is suggested by the uniform orientation and shear sense of the largest zones (Figure 14a), though at least three other smaller sets are present. This relatively simple geometry contrasts with that on the east side of the range (Figure 14b). The dominant set is rotated about 60° in strike, however many other sets are well developed, perhaps representing a general deformation mimicking Von Mises' criterion (5 or more sets of zones).

LATE CHLORITE-RICH ZONES

Late chlorite-rich deformation zones are found throughout the Allamuchy Mountains (Figure 6). The zones are narrow, usually less than a half meter wide and are recognized by their relatively sharp, planar boundaries that are rarely offset. Most of these chlorite zones can be traced across the entire outcrop. The epidote zones terminate against the late chlorite zones (as in Figure 13a), establishing their relative age.

The orientation of the late chlorite zones is consistent across the Allamuchy Mountains, with oblique normal slip directed towards the southeast on NE-dipping zones (Figure 15). Many small scale structures are associated with the main zone, including networks of small faults in the wall rocks up to a meter on either side of the boundary, subsidiary deformation zones in the interior (shear bands and antithetic and synthetic microfaults) and late sliding surfaces. Most of these structures are subparallel to the main zone of chlorite-rich tectonites.

A wide variety of tectonites can be found in the late chlorite zones, many of which are similar to those of the early chlorite suite. Here we describe just two distinct tectonites of the SP-mylonite series, a transitional ultramylonite and an atypical pseudotachylite. Ultramylonites can be found near Stop 3 in the eastern Allamuchy Mountains (Figure 6) in narrow zones a few decimeters thick. The zones are planar and not cut by later faults, but are restricted in extent. The displacement sense can be determined in the field by rotation of mylonitic foliation into ultramylonite zones (S-C fabric).

The ultramylonites (Figure 16a) are composed of thin bands of microaugen mylonite and more uniform ultramylonite (grain size banding) that are both felsic and mafic (compositional banding). Feldspar exhibits both brittle and plastic deformation (undulatory extinction, subgrains and neocrysts), but quartz shows only plastic features. The ultramylonites are considered transitional as grain size reduction is produced by both fracturing and recrystallization. Feldspars dominate

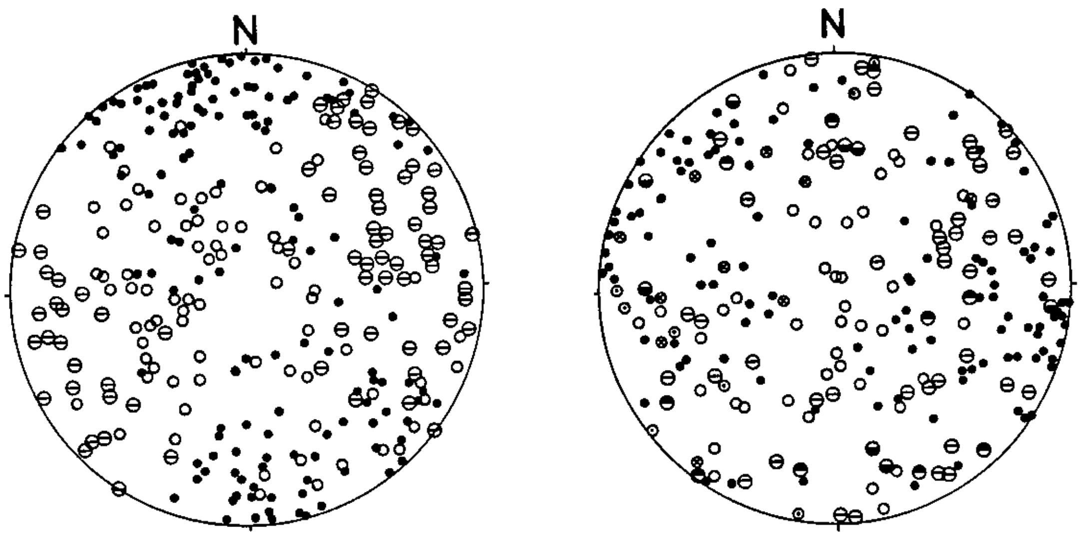


Figure 14a (left). Epidote population from western Allamuchy Mountains (see Figure 13a). ENE-striking, subvertical dip slip sets dominate. The few shear sense determinations (not shown) are all reverse. Bulk simple shear is thus suggested. Lower hemisphere, equal area. Figure 14b (right). Epidote population from eastern Allamuchy Mountains (Stop 3). A NNE-striking set dominates but the orientations are highly scattered. This pattern may reflect a general deformation (5 or more sets). Lower hemisphere, equal area.

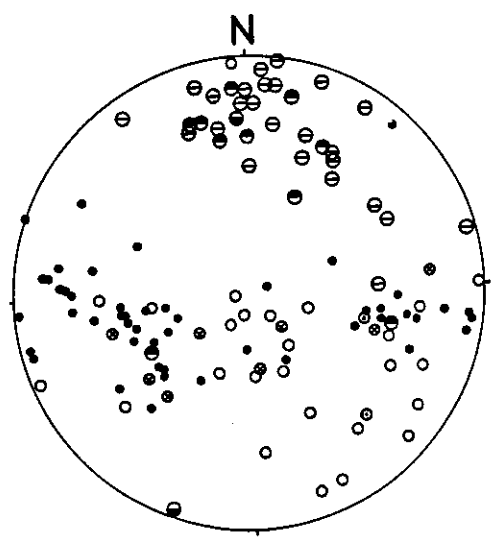


Figure 15. Structural elements of late chlorite BDZ from eastern Allamuchy Mountains (Stop 3), including both footwall and hangingwall zones, the main zone itself and late sliding surfaces. The main zone dips moderately northeast, with oblique, southeast-directed normal (?) slip. Lower hemisphere, equal area.

the augen (the large ones have biotite beards), and both sigma and delta types are present (Simpson, 1986). Shear sense can also be determined from synthetic and antithetic shear fractures cutting feldspars, microfolds of layering and faint shear bands (extension crenulation cleavage).

The ultramylonite is very fine grained (10 micron range) and shows a strong crystallographic preferred orientation (texture) with an oblique subfabric. Quartz is throughly recrystallized, and along with opaques, forms monomineralic ribbons that help define the ultramylonitic layering. Chlorite-, biotite- and even amphibole-rich layers are also present. The mineralogy and deformation mechanisms are consistent with higher temperatures than the cataclasites previously described. The similar geometry to other late chlorite zones may thus be fortuitous, but no definitive cross-cutting relationships have been seen.

Breccias and microbreccias in the chlorite zone are further reduced in grain size, producing "breccia in breccia" structures (Figure 16b). The end product of extreme grain size reduction is thin, discontinuous, anastomosing bands of black, glassy pseudotachylite. These ultracataclasites are brown to opaque in thin section with a submicron grain size, and are atypical in that contorted flow banding characterisitic of most pseudotachylites is absent. Most of the anomalous features of these ultracataclasites are related to later deformation of the chlorite zones. For example, sliding surfaces within the zones cut across and juxtapose the different cataclasites and SP-mylonites.

COEVAL CHLORITE AND EPIDOTE ZONES

A consistent temporal sequence of early chlorite-epidote-late chlorite is seen throughout the Highlands. However locally there is good evidence that chlorite and epidote zones formed simultaneously, and that the bulk composition of the protolith exerted an influence on the mineralogy of the retrograde deformation zones. An excellent example of this lithologic control can be seen adjacent to the Morris Lake fault along Route 15 in the Stanhope quadrangle (Figure 6).

The Morris Lake fault separates rocks of the Losee metamorphic suite (Drake, 1984) from hornblende syenite to the west (Baker and Buddington, 1970). Rocks on both sides of the Morris Lake fault along Route 15 contain a network of retrograde deformation zones, though the main fault zone itself is not exposed here. The east block (hanging-wall?) exposes quartz-oligoclase gneiss (Losee sensu stricto) and large bodies of amphibolite gneiss near the fault (not shown on Baker and Buddington). An igneous protolith for the Losee is suggested by quartz-oligoclase dikes (now folded and foliated) that intrude the amphibolite. Both gneisses are cut by folded and foliated felsic pegmatites.

The mineralogy of the retrograde deformation zones is strongly dependent on the gneissic protolith. Deformation zones rich in epidote

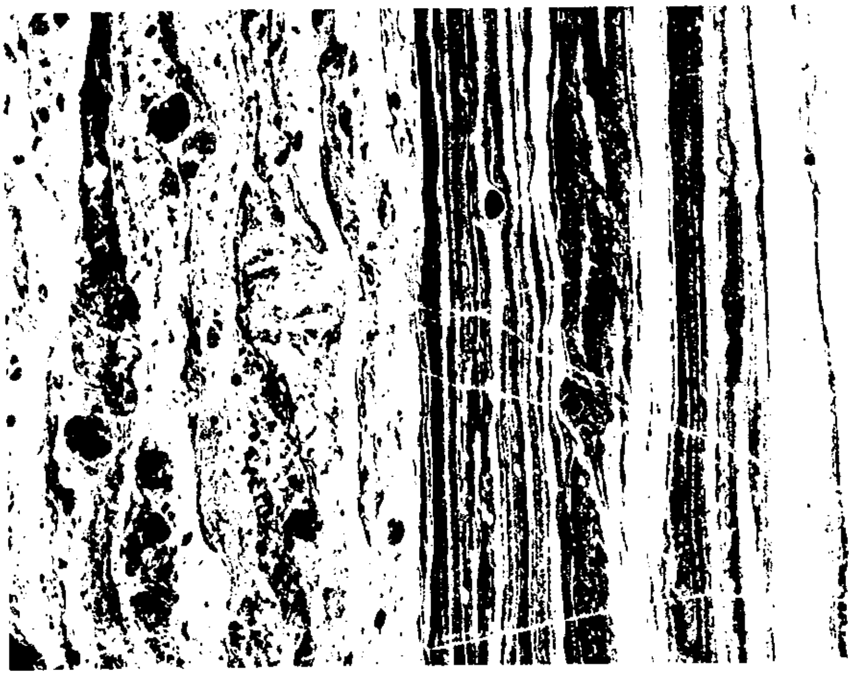


Figure 16a. Negative image of microaugen mylonite and ultramylonite (fine grained light and dark bands) from late chlorite zone near Stop 3. Both delta- and sigma-type feldspar augen are present. Scale 12X.

Figure 16b. Thin section of late chlorite zone at Stop 3. Several varieties of microbreccia are themselves brecciated and cut by razor sharp sliding surfaces. Extremely fine-grained ultracataclasites (pseudotachylite) are almost totally opaque, and appear white in this negative image. Scale 13X.

are found within the quartz-plagioclase protolith, while chlorite zones are associated with the amphibolite (hornblende-plagioclase-diopside). The mineralogy of a single zone changes across the contact between the two gneiss types, while the geometry remains constant (Figure 17a). This change in mineralogy also affects the small scale structures associated with the deformation zones; chlorite zones are more likely to develop fibers and steps, for example. Both of the protoliths show brittle and plastic microstructures that increase in intensity towards the retrograde zones. The resulting tectonites are similar to those previously described.

The retrograde deformation zones form a conjugate set of reverse faults (Figure 17b), one dipping moderately to shallow northeast (with oblique to dip slip), and the other dipping moderately southeast (with oblique to strike slip). Though reverse, both these faults extend the layering, as gneissosity is near vertical in these exposures. Poles to movement planes (rotation axes) form a broad point maximum in the southeast quadrant. The Anderson-type conjugates indicate subhorizontal, northeast-directed shortening and steeply west-plunging extension. To a first approximation, the deformation is plane strain, with the intermediate axis (Y) plunging shallowly towards the southeast (parallel to the M-poles). The Morris Lake fault strikes 035 (the dip and dip direction are unknown), suggesting shortening parallel to the fault zone. Further work on both the Morris Lake and the outcrop-scale faults is needed to resolve this peculiar geometry.

A single, foot-wide basaltic dike (unmetamorphosed) intrudes the amphibolites subparallel to gneissosity (060,80SE) at this locality. The dike is not offset by any retrograde zones, though there is some slip along the dike margins that is post-emplacement. As discussed in a later section, this and other unmetamorphosed basaltic dikes are probably Mesozoic in age, rather than Eocambrian (proto-Atlantic rifting) or Ordovician (Cortlandt-Beemerville trend). This dike gives an upper bound to the age of the retrograde deformation.

The deformation zones in the Losee gneisses illustrate the role of the protolith in controlling the retrograde mineral assemblage. The abrupt change in mineralogy along a zone suggests that fluid flow was restricted in a partially closed system. The resultant high variance mineral assemblages both formed at the same grade. Therefore the temporal sequence of early chlorite-late epidote found throughout the Highlands may be related to changes in fluid composition rather than changes in physical conditions.

CALC-SILICATE RETROGRADE ZONES

The influence of lithology on the types of deformation zones formed and their associated tectonites can also be seen at Jenny Jump Mountain in the Blairstown quadrangle (Westgate, 1896; Drake and Lyttle, 1985). Jenny Jump Mountain is a large horse bounded by branch lines between the Jenny Jump thrust (placing Grenville on Paleozoics) and the Shades of Death thrust. At the northern end of the mountain, a

Figure 17a. Offset of foliated pegmatite dike in amphibolite gneiss suggests vertical extension and northeast shortening in the hangingwall (?) of the Morris Lake fault. Individual deformation zones change mineralogy as they cross lithologic boundaries. Scale 1:25.

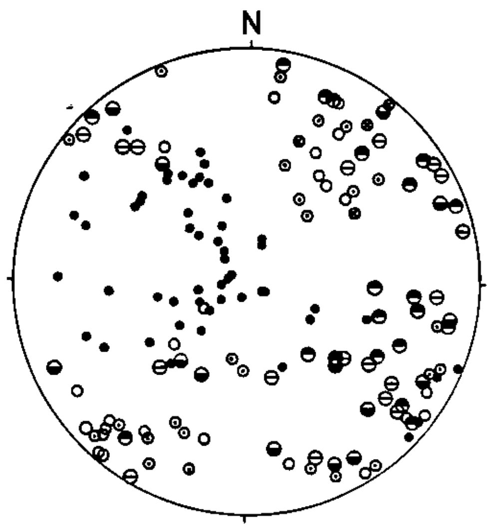


Figure 17b. Geometry of retrograde deformation zones. Conjugate sets of slickensides with shallowly plunging slickenlines of mixed polarity indicate pure shear deformation. Lower hemisphere, equal area.

smaller horse of calcareous gneisses in the hangingwall (?) of the Shades of Death thrust is exposed along the I-80 roadcuts (Figure 6).

The smaller horse is composed of calc-silicate paragneisses, amphibolites, metapelites, metasandstones and other lithologies. The sequence is strongly banded, with unusual and distinctive marker horizons; much of this layering may be bedding. In the roadcuts gneissosity dips moderately to the east-southeast with little variation, but ptygmatically folded boudinage can be seen to the south near the branch line with the Shades of Death thrust. The sequence may be correlative or transitional with the Franklin marbles to the northeast and/or the Chestnut Hill Group to the southwest (Drake, 1984). The gneisses are intruded by vertical, east-striking basaltic dikes with glassy chill margins and diabasic interiors. In the diabase, phenocrysts of plagioclase and xenocrysts of quartz are set in a groundmass of plagioclase, augite and opaque.

The most unusual deformation zones seen in the I-80 roadcuts are calc-silicate cleavage duplexes (Figures 18a and 18b). The cleavage duplexes are less than a meter wide and anastomose subparallel to gneissosity. The cleavage or schistosity is asymptotic to both the upper and lower deformation zone boundaries (sigmoidal cleavage), consistent with reverse oblique slip on the deformation zone. The foliation surrounds less deformed lithons in the interior of the zone, and lineations and steps indicate reverse oblique slip on the cleavage itself. The poles to foliation and the deformation zone boundary and the lineation and transport direction are all coplanar, consistent with simple shear deformation (Figure 19a). Tight folds of cleavage plunge shallowly to the southwest at a high angle to the deformation zone boundaries, suggesting some later modification of the cleavage geometry. Sliding surfaces on both the upper and lower deformation zone boundaries are polished to fibrous and indicate a similar displacement direction to that of the cleavage duplex itself.

Overall the geometry is very similar to Alleghanian cleavage duplexes developed in very low grade shales of the Pennsylvania foreland (Nickelsen, 1986). The duplex geometry in the Pennsylvania examples is produced by both rotation of an early steep cleavage (formed at the thrust tip) and by imbrication from the floor to the roof thrusts (slip on the cleavage itself) to form miniature horses. The duplex represents the greatest amount of shortening in the smallest volume (Boyer and Elliott, 1982), and displacements on these zones should be quite large. One diabase dike, about 10-20 m wide, is offset along the cleavage duplex (Figure 18a). Using the south contact of the dike and the slickenline on the sliding surface, approximately 9 m of reverse oblique slip can be calculated. The small amount of slip is inconsistent with the duplex geometry, suggesting reactivation of the sliding surfaces following intrusion of the dike.

The medium to high grade assemblages in the country rock are retrogressed to calcite, talc, chlorite and amphibole. Calcite is strongly deformed by intracrystalline plasticity (Figure 18b), with

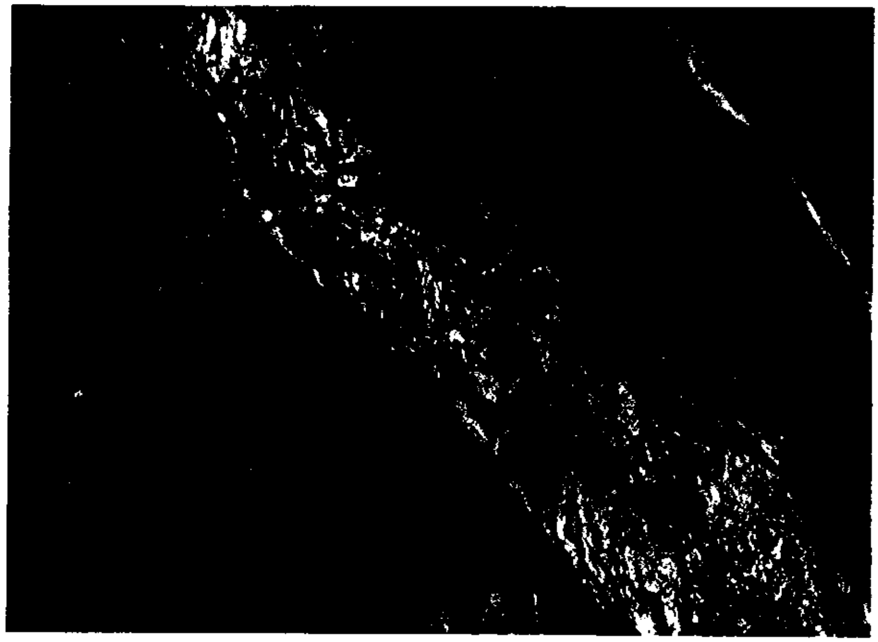


Figure 18a. Calc-silicate cleavage duplex subparallel to gneissosity at Jenny Jump Mountain. Amphibolite in hangingwall, diabase dike intruding calcareous gneiss in footwall. Cleavage asymptotic to both sliding surfaces, indicating sinistral shear. Late folding of cleavage can also be seen (dead center). Bob Koto gives scale.

Figure 18b. Thin section from cleavage duplex, showing plastic deformation of coarse calcite (bent twins, undulatory extinction) and a small, foliated retrograde zone rich in talc and amphibole. Fibrous calcite in tension vein at upper right. Negative image. Scale 17X.

deformation twins (subsequently bent and kinked) and narrow zones of recrystallization that are crystallographically controlled. Fracturing seems to play a lesser role in the development of the retrograde tectonites. Grain size reduction by recrystallization and syntectonic retrogression is enhanced in ductile deformation zones. The resulting low variance tectonite is strongly foliated, with a sigmoidal habit that suggests thin section scale (fifth order) duplexing. Syntectonic and post-tectonic extension veins filled with fibrous calcite are common.

Further evidence of post-intrusive deformation comes from sets of chlorite deformation zones that cut the basaltic dike (Figure 19b). Chlorite slickensides rotate into and terminate in shear veins with fibrous chlorite, quartz, calcite and biotite. Some early actinolite needles on these slickensides have been crenulated and overprinted by the short chlorite fibers. The slickensides are rotated about a constant orientation of slickenline (shallow to the south) and show both senses of shear. If the amount of shear was equally distributed on all slickensides, no bulk deformation would result, but net slips have not been measured.

A number of different deformation zones shorten the Grenville section, including wedge faults, insertion thrusts and the cleavage duplex. All of these contraction faults are probably Paleozoic (Alleghanian?) in age, given their low grade mineral assemblages and thrust nature. The dike is most likely Mesozoic (Maxey, 1973); Eocambrian dikes are metamorphosed to greenstones, and Ordovician dikes are highly alkalic. The dike must therefore intrude the cleavage duplex, though this contact has not been seen. Mesozoic reactivation of the cleavage duplex bounding surfaces and development of chlorite slickensides are the latest events. The field relationships are more consistent with an Ordovician age for the dike, though alkalic basalts have not been described in the Cortlandt-Beemerville suite (Maxey, 1976; Ratcliffe, 1981).

The small, outcrop-scale deformation zones described here are relatively easy to study, as the zones can be completely exposed, displacements can be measured, and all of the resulting tectonites can be seen, sometimes in a single thin section. The information gathered on the small faults can be compared with map-scale, regional deformation zones, where the displacement history is usually unknown and complete exposures are almost always lacking. There is growing evidence that deformation zones on widely different scales are approximately self-similar in aspects such as surface roughness and displacement/width ratios. Thus there is some justification in comparing the two, and viewing map-scale zones as just bigger versions of those on the outcrop scale, with perhaps a more complex displacement history. This viewpoint is supported by the preliminary results of our study of two regional deformation zones, the Wright Pond and Reservoir faults.

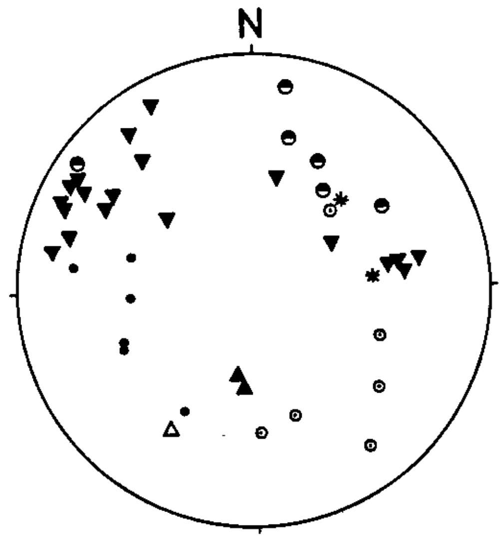


Figure 19a. Structural elements for the cleavage duplex. Poles to foliation (solid triangles down), two lineations (solid triangles up), poles to sliding surfaces and slickenlines are all coplanar. Axes of curvature of individual horses (asterisks) and a hinge line for late folds of cleavage (open triangle up) are also shown. Lower hemisphere, equal area.

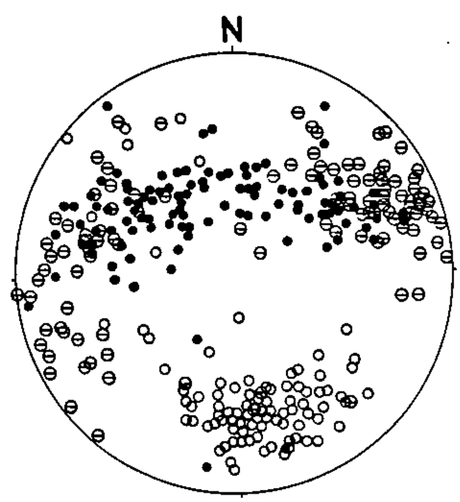


Figure 19b. Geometry of chlorite zones cutting diabase dike. Polarities of slickenlines and rotation axes are mixed (not shown). Lower hemisphere, equal area.

WRIGHT POND FAULT

The Wright Pond fault strikes northeast across the Stanhope quadrangle, which has been recently mapped by Rich Volkert, and is one of several Highland faults that carry small patches of Cambrian sedimentary rocks in their east walls. The fault is nicely exposed northeast of Route 206 in some uranium prospect pits (Figure 6). The hangingwall (east block) at this locality is composed of hornblende gneiss with a shallow east-dipping fabric that may steepen towards the fault. The footwall rocks are pink metagranites with little foliation but a good stretching lineation that plunges very shallow to the northeast.

The retrograde deformation is concentrated in a variable zone 10's of meters wide, but (as usual) complete exposures are absent. Both the subparallel fault network and the cataclastic foliation dip steeply southeast and a weak lineation (defined by elongate clasts) rakes moderately northeast. The shear sense has not been determined but Cambrian rocks in the hangingwall suggests at least some normal slip in the displacement history.

The Wright Pond tectonites are typical of the early chlorite suite of cataclasites and ultracataclasites (Figure 20a). The retrograde assemblage is unremarkable and quartz and feldspar exhibit familiar brittle and plastic microstructures. To a first approximation, the Wright Pond fault appears to be simply a larger version of the small zones described previously, growing in size as displacement increased. There are some subtle differences. The Wright Pond fault contains a distinctive breccia with very large, rounded to subangular clasts, that could be mistaken for a retrogressed pebbly sandstone. The grain size distribution in these breccias is very heterogeneous, with clasts ranging from large pebbles to silt. The coarse breccias are themselves cut by a network of small faults that contain finer grained and more foliated cataclasites (Figure 20b). The geometry of these faults is not known, as they are usually visible only in thin section, but larger cross-cutting zones dip to the southwest.

RESERVOIR FAULT

The Reservoir fault marks the western boundary of the Green Pond Outlier, an enigmatic structure that best illustrates the Alleghanian deformation style in the Highlands (Bizub and Hull, 1986). There are many partial exposures of the Reservoir fault along its over 40 km trace, but complete cross-sections have not been seen. The lack of total exposure is crucial, as Ratcliffe (1980) has proposed Mesozoic reactivation of the Reservoir fault. The Mesozoic zones are probably narrow and could easily be missed without perfect exposures.

The Reservoir fault contains at least 10 m of tectonites, though non-penetrative retrograde deformation extends up to 100 m to the west in the footwall gneisses. The fault does not have sharp boundaries (where exposed), and so its orientation must be inferred from

Figure 20a. Small brittle deformation zone containing foliated, chlorite-rich cataclasites from the Wright Pond fault. Plagioclase shows kinked and offset twins, while quartz exhibits deformation lamellae parallel to the BDZ and subgrains rotating into the BDZ. Positive image. Scale 120X.

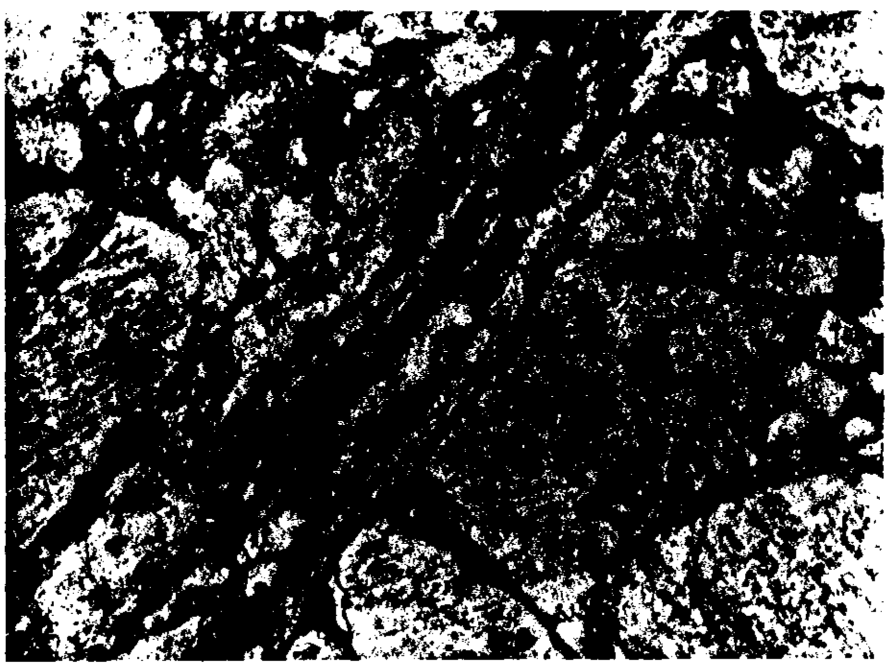


Figure 20b. Weakly foliated chlorite breccia from the Wright Pond fault, cut by small foliated BDZ. Positive image, plane light. Scale 140X.

subsidiary deformation zones and from its internal fabric. As in the Wright Pond fault, the small faults and retrograde foliation dip very steeply to the east. The retrograde foliation probably parallels the fault zone boundary, as the strains are very high with a large component of pure flattening, inferred from the S>>L fabric. Phanerozoic rocks up through Middle Devonian in age are carried in the hangingwall, but the displacement history is unknown.

The gneisses in the footwall of the Reservoir fault show more plastic deformation (related to the retrograde event) than any other country rock. Deformation twins in feldspars, bent and kinked twins, and recrystallization of both quartz and feldspars record a small but penetrative plastic strain in the wall rocks. Short actinolite needles are found as beards on pyriboles and in tension fractures. Fracturing and plasticity appear to be equally important in reducing grain size in the deformation zones. Hornblende is slow to break down to chlorite, which dominates the retrograde assemblage, and odd chlorite-amphibole breccias can be found. Some tectonites have very strong fabrics with both quartz and feldspar exhibiting significant plastic strains, and these tectonites have been classified as quasiplastic mylonites (Figure 21a). Epidote veins often cut across the chlorite-rich zones and record a component of flattening strain across the foliation.

Though the quasiplastic mylonites interlayered with microbreccias are most abundant in the Reservoir fault zone, other tectonites have been found. On the northwest shore of Oak Ridge Reservoir (Figure 6), coarse breccias akin to those along the Wright Pond fault were discovered during the 1985 drought (Figure 21b). Boulder-sized (!) angular fragments of gneiss in a chlorite-rich breccia can be found here, though the cataclasites are usually finer grained and slightly foliated. Fluid-assisted diffusional mass transfer is suggested by chlorite beards on elongate clasts and by anastomosing, opaque-rich seams in the matrix that resemble pressure solution cleavage.

SUMMARY AND DISCUSSION

Medium to high grade mylonite zones have been discovered in the Grenville of New Jersey but the distribution of Proterozoic deformation zones is not known. Phanerozoic deformation zones form a consistent sequence of early chlorite, epidote, and late chlorite. Each of these different types forms a population of faults on the outcrop scale that produces bulk, retrograde deformation of the basement. The resulting strain fields are quite variable both in style and orientation, and only the late chlorite zones show a consistent pattern. Chlorite- and epidote-rich zones can form simultaneously, depending on the bulk composition of the protolith and the physical conditions during deformation.

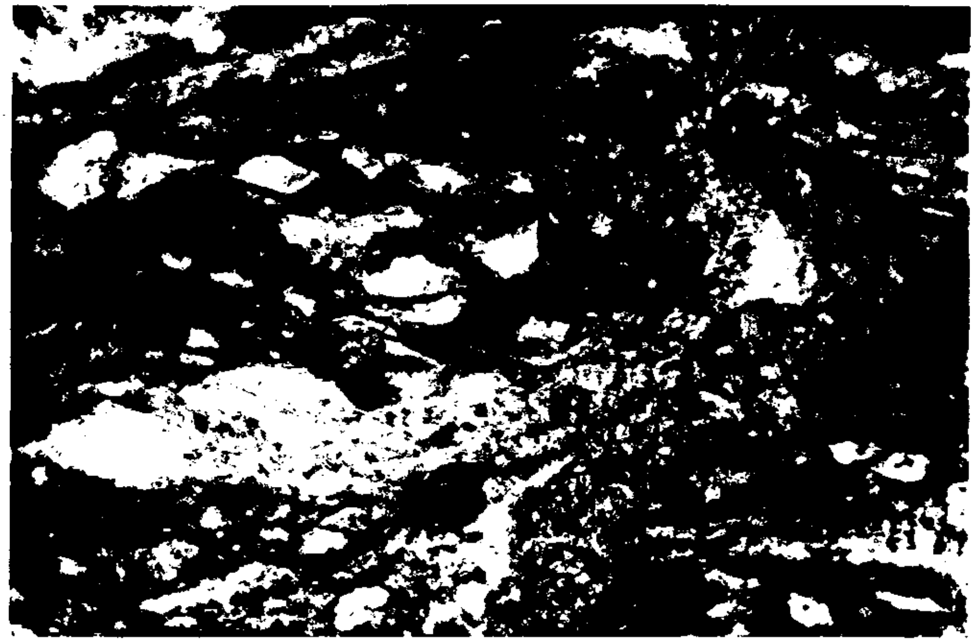


Figure 21a. Strongly foliated, quasiplastic mylonite in the Reservoir fault zone at the north end of Clinton Reservoir. The mylonite is cut by an epidote vein (right side of image) that shows later shortening by buckling. Positive image. Scale 100X.

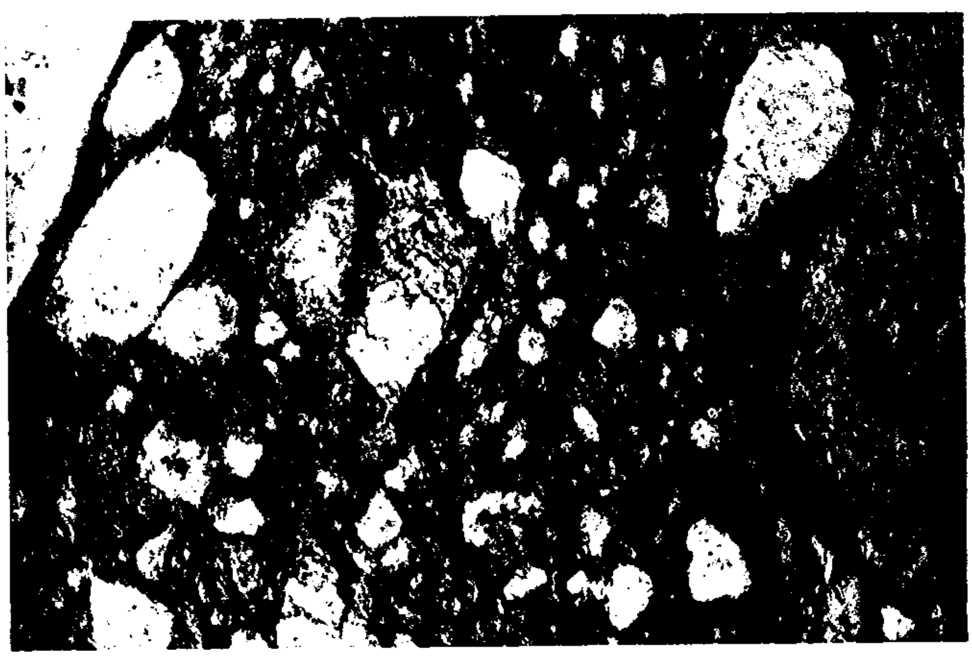


Figure 21b. Foliated breccia and microbreccia from the Reservoir fault at Oak Ridge Reservoir. The foliation is defined by elongate clasts, a preferred orientation to chlorite, and opaque-rich seams. Chlorite beards can be seen on some clasts. Positive image. Scale 100X.

Tectonites from two map-scale deformation zones are similar to those associated with the small chlorite-rich zones. The Wright Pond fault zone contains a suite of cataclasites (breccias through gouge) exhibiting a weak foliation that dips steeply. The most striking tectonite is a coarse grained breccia with large angular clasts of gneiss, which is also seen along the Reservoir fault zone. In addition, the RFZ contains chlorite-rich quasiplastic mylonites showing more plastic deformation than any other retrograde zone. Higher temperatures of deformation are also indicated by syntectonic amphibole fibers. Neither of these fault zones is exposed continuously in cross section, and important tectonites like ultramylonites and incohesive gouge may be hidden.

One of the goals of this study is to relate the sequence and geometry of retrograde deformation zones to the different Phanerozoic orogenies in New Jersey: Taconic, Acadian, Alleghanian and Newark. The intensity of each of these orogenic episodes in New Jersey is a matter of some debate. The work of Ratcliffe et al (1986), supported by our new results, suggests extensive (though perhaps mild) Mesozoic deformation in the Highlands. The presence of Alleghanian deformation is not in dispute, though the youngest Paleozoic formation preserved in New Jersey is Middle Devonian (Skunnemunk Formation of the Hamilton Group). In the Little Mountains thrust belt of the Hudson Valley (Zadins, 1983; Marshak, 1986), folds and faults involving Devonian rocks are considered either Acadian and/or Alleghanian in age.

Structures typically assigned to the Taconic orogeny are also being re-evaluated. Ratcliffe (1981) illustrates cleaved Martinsburg Formation xenoliths in post-Taconic (Late Ordovician) Beemerville intrusives in the Great Valley of western New Jersey. Ratcliffe also reports clasts of cleaved Martinsburg in the basal conglomerates of the Early Silurian-Late Ordovician (?) Shawangunk Formation. However, Epstein and Lyttle (1986) have shown that slatey cleavage in the Martinsburg dies out at the Taconic unconformity, and that the Shawangunk acted as a strain shadow during Alleghanian deformation. The intensity of slatey cleavage and strain magnitude are much higher in Martinsburg siltstones than in comparable lithologies of the Devonian Marcellus Formation (Bizub and Hull, 1986). These differences could be due to either structural or stratigraphic position during Alleghanian folding.

We have previously argued that the retrograde deformation zones are Phanerozoic in age, as they cut Late Proterozoic pegmatites. We have not seen definitive cross-cutting relationships with the few Eocambrian metadiabase (greenstone) and Late Ordovician lamprophyre dikes present in our study areas. Separations on faults cutting these dikes are small where observed. Jurassic basaltic dikes also show small offsets, of both normal and reverse sense. A combined radiometric and structural study of the cross-cutting relationships is needed.

Another method of establishing the ages of the deformations is to compare the metamorphic grade in the cover rocks and basement tectonites. The Martinsburg Formation has a low grade assemblage of chlorite, muscovite, quartz and calcite (Beutner, 1978). About a kilometer higher in the section, the Devonian Bellvale Formation in the Green Pond outlier exhibits a similar assemblage (Bizub and Hull, 1986). In addition, conodonts recovered from the Silurian units just below the Bellvale are black and opaque (Barnett, 1966). A diagnostic paragenesis has not been seen in the retrograde deformation zones in the basement rocks. The variety of textures and abrupt mineralogic changes make it difficult to identify which minerals are in equilibrium. The retrograde tectonites typically contain a high variance combination of chlorite, epidote, quartz, plagioclase and alkali feldspar, consistent with low grade metamorphism. Thus there appears to be little difference in metamorphic grade between retrograde tectonites and cleavage in the cover.

Another approach to dating is to trace the retrograde deformation zones out of the basement and into the cover rocks. Relatively little Phanerozoic strata is preserved in the Highlands, except in the Green Pond Outlier. The structures in the Outlier are thick-skinned and quite similar to basement-involved thrust sheets of the Laramide foreland (Bizub and Hull, 1986). High angle reverse faults propagate out of the basement and root in tip anticlines in the cover. One such example can be seen on Bowling Green Mountain (Figure 6), where a steep fault has been traced continuously from the basement into folded Shawangunk Formation. Unfortunately, few tectonites are exposed here.

During the 1985 drought, previously submerged outcrops of highly retrogressed cover were found adjacent to the Reservoir fault on the southwest shore of Clinton Reservoir (Figure 6). Both the Devonian Bellvale and Skunnemunk Formations are exposed in the fault zone, perhaps as small horses. The retrogressed Bellvale is very similar to the chlorite-rich, foliated quasiplastic mylonites derived from basement gneisses. Small faults in the the Skunnemunk conglomerates and sandstones contain unusual quartzite breccias and ultracataclasites. The faults are intimately associated with folds and other structures formed during the Late Paleozoic, and are not the product of Mesozoic tectonism.

There are many large fault zones along the western border of the Newark basin that involve Mesozoic cover, and extensive studies of tectonites in the border faults have been described by Ratcliffe and others (Ratcliffe, 1980; Ratcliffe et al, 1986). Mesozoic tectonites consist of chlorite-rich foliated cataclasites (mostly breccias) separated from footwall-derived tectonites by black ultracataclasites a few decimeters thick. Delineating footwall (basement or Paleozoic) from hangingwall (Mesozoic) tectonites can be difficult if retrogression is thorough. Tectonites from the Flemington fault zone (confined to the interior of the basin) have been briefly reported by Burton and Ratcliffe (1985). We have begun a study of fault zones in the Newark basin, and recent observations at one exposure are worth mentioning here.

A large quarry near the Delaware River, about 30 km south of the border fault, provides fresh and extensive exposures of the NE-striking Hopewell fault, part of the Hopewell-Chalfont fault system. The Hopewell fault cuts the Quarry Hill diabase, and the resulting sequence of Mesozoic deformation zones is remarkably similar to that seen in the Highlands. A penetrative set of small, chlorite-rich BDZ strikes NW to NE, and is cut by large, widely spaced, epidote-rich BDZ with characteristic polished sliding surfaces. The epidote zones parallel the presumed trace of the Hopewell fault, and dip moderate to steep to the south. Small, highly polished epidote-calcite slickensides (east dipping) cut the large epidote zones.

Though the sequence of Mesozoic deformation zones at Quarry Hill is superficially similar to that of the Highlands, we do not equate the two temporally. This same sequence of early chlorite-late epidote can be seen in gneisses of the Wind River Mountains of Wyoming (produced during a single Archean deformation) and appears to be quite common. The temporal progression from (predominantly) chlorite to epidote may be related to changing fluid composition and/or variable rates of reaction. In the gneisses of the Highlands, pyriboles rapidly convert to chlorite, whereas feldspars react continuously and change composition. Though a consistent sequence of tectonites is found in the gneisses of the Highlands, the pattern is not unique and cannot be reliably used to identify the age of faulting unless bracketing intrusives are present.

ACKNOWLEDGEMENTS

Discussions of New Jersey geology with Jane Gilotti, Greg Herman, Rich Volkert and Ed Zofchak are greatly appreciated. Zofchak and Ann Kutyla prepared the innumerable thin sections and manuscripts, respectively. This work was supported by Sigma Xi student research grants to Koto and Bizub.

BIBLIOGRAPHY

Arthaud, F. (1969), Methode de determination graphique des directions de raccourcissement, d'allongement et intermediaire d'une population de failles; Bull. Soc. geol. Fr., v. 11, pp. 729-737.

Angelier, J., Colletta, B. and Anderson, R. E. (1985), Neogene paleostress changes in the Basin and Range: a case study at Hoover Dam, Nevada-Arizona; Geol. Soc. Am. Bull., v. 96, pp. 347-361.

Ashby, M. F. and Verrall, R. A. (1978), Micromechanisms of flow and fracture and their relevance to the rheology of the upper mantle; Phil. Trans. Roy. Soc. Lon., v. 288A, pp. 59-95.

Aydin, A. (1978), Small faults as deformation bands in sandstones; Pageoph., v. 116, pp. 913-930.

Aydin, A. and Reches, Z. (1982), Number and orientation of fault sets in the field and in experiments; Geology, v. 10, pp. 107-112.

Baker, D. R. and Buddington, A. F. (1970), Geology and magnetite deposits of the Franklin and part of the Hamburg quadrangle, New Jersey; U.S.G.S. Prof. Paper 638, 73 pp.

Barnett, S. G. (1966), Late Cayugan and Helderbergian stratigraphy of southeastern New York and northern New Jersey; Ph.D. thesis, Ohio State Univ., 217 pp.

Bell, G. L. (1952), Geology of the northern Farmington Mountains; Utah Geol. Soc. Guidebook 8, pp. 38-51.

Bell, T. H. (1981), Foliation development: the contribution, geometry and significance of progressive, bulk inhomogeneous shortening; Tectonophysics, v. 75, pp. 273-296.

Bell, T. H., and Etheridge, M. A. (1973), Microstructure of mylonites and their descriptive terminology; Lithos, v. 6, p. 337-348.

Beutner, E. C. (1978), Slaty cleavage and related strain in Martinsburg slate, Delaware Water Gap, New Jersey; Am. J. Sci., v. 278, pp. 1-23.

Bizub, R. and Hull, J. (1986), Shortening of cover and basement in the Green Pond Outlier of northern New Jersey; Geol. Soc. Am. Abst. Prog., v. 18, p. 5.

Boyer, S. E. and Elliott, D. (1982), Thrust systems; Bull. Am. Assoc. Petr. Geol., v. 66, pp. 1196-1230.

Carey, E. (1976), Analyse numerique d'un modele mecanique elementaire applique a l'etude d'une population de failles: calcul d'un tenseur moyen des contraintes a partir de stries de glissement; Ph.D. Thesis, Universite de Paris Sud.

Dallmeyer, R. D., Sutter, J. F. and Baker, D. J. (1975), Incremental 40Ar/39Ar ages of biotite and hornblende from the northeastern Reading Prong: their bearing on Late Proterozoic thermal and tectonic history; Geol. Soc. Am. Bull., v. 86, pp. 1435-1443.

Drake, A. A. (1969), Precambrian and Lower Paleozoic geology of the Delaware Valley, New Jersey-Pennsylvania; in Subitzky, S. (ed.), Geology of selected areas in New Jersey and eastern Pennsylvania, Rutgers Univ. Press, pp. 51-131.

Drake, A. A. (1970), Structural geology of the Reading Prong; in Fisher, G. W. et al (eds.), Studies in Appalachian Geology-Central and Southern, John Wiley, pp. 271-290.

Drake, A. A. (1984), The Reading Prong of New Jersey and eastern Pennsylvania: an appraisal of rock relations and chemistry of a major Proterozoic terrane in the Appalachians; G.S.A. Spec. Paper 194, pp. 75-109.

Drake, A. A. and Lyttle, P. T. (1980), Alleghanian thrust faults in the Kittatinny Valley, New Jersey; in Manspeizer, W. (ed.), Field studies of New Jersey geology and guide to field trips, NYSGA Guidebook #52, pp. 92-114.

Drake, A. A. and Lyttle, P. T. (1985), Geologic map of the Blairstown Quadrangle, Warren County, New Jersey; U. S. Geol. Surv. Map GQ-1585.

Drake, A. A., Lyttle, P. T. and Owens, J. P. (1978), Preliminary geologic map of the Newark Quadrangle, New Jersey and Pennsylvania; U.S.G.S. Open File Rept. 78-595.

Draper, G. (1976), Grain size as an indication of shear strain in brittle fault zones; Tectonophysics, v. 35, pp. T7-T13.

Epstein, J. B. and Lyttle, P. T. (1986), Chronology of deformation along the Taconic unconformity from eastern Pennsylvania to southern New York; Geol. Soc. Am. Abst. Prog., v. 18, p. 15.

Helenek, H. L. and Mose, D. G. (1984), Geology and geochronology of Canada Hill granite and its bearing on the timing of Grenvillian events in the Hudson Highlands, New York; G.S.A. Spec. Paper 194, pp. 57-74.

Higgins, M. W. (1971), Cataclastic rocks: U.S. Geol. Surv. Prof. Paper 687, 97 pp.

Hudleston, P. J. (1980), The progressive development of inhomogeneous shear and crystallographic fabric in glacial ice; J. Struct. Geol, v. 2, pp. 189-196.

Hull, J. (1986), Width-displacement relationships for deformation zones; Geol. Soc. Am. Abst. Prog., v. 18, p. 24.

Kligfield, R., Crespi, J., Naruk, S., and Davis, G. (1984), Displacement and strain patterns of extensional orogens; Tectonics, v. 3, pp. 577-609.

Logan, J. M., Friedman, M., Higgs, N., Dengo, C. and Shimamoto, T. (1979), Experimental studies of simulated gouge and their application to studies of natural fault zones; U.S.G.S. Open File Rept. 79-1239, pp. 305-343.

Marshak, S. (1986), Structure and tectonics of the Hudson Valley fold-thrust belt, eastern New York state; Geol. Soc. Am. Bull., v. 97, pp. 354-368.

Maxey, L. M. (1973), Dolerite dikes of the New Jersey Highlands: probable co-magmatic relation with Mesozoic Palisades sill and dolerite dikes of eastern United States; Geol. Soc. Am. Bull., v. 84, pp. 1081-1086.

Maxey, L. M. (1976), Petrology and geochemistry of the Beemerville carbonatite alkalic rock complex, New Jersey; Geol. Soc. Am. Bull., v. 87, pp. 1551-1559.

McClelland, J. M. (1984), The origin of ribbon lineation within the southern Adirondacks, USA; J. Struct. Geol., v. 6, pp. 147-158.

Mitra, G. (1978), Ductile deformation zones and mylonites: the mechanical processes involved in the deformation of crystalline basement rocks; Am. J. Sci., v. 278, pp. 1057-1084.

Mitra, G. (1979), Ductile deformation zones in Blue Ridge basement rocks and estimation of finite strains; Geol. Soc. Am. Bull. v. 90, 935-951.

Nickelsen, R. P. (1986), Cleavage duplexes in the Marcellus shale of the Appalachian foreland; J. Struct. Geol., v. 8, pp. 361-372.

Puffer, J. H. (1980), Precambrian rocks of the New Jersey Highlands; in Manspeizer, W. S. (ed.), Field studies of New Jersey geology and guide to field trips, NYSGA Guidebook #52, pp. 42-52.

Ramsay, J. G. (1980), Shear zone geometry: a review; J. Struct. Geol. v. 2, pp. 83-99.

Ratcliffe, N. M. (1971), Ramapo fault system in New York and adjacent northern New Jersey: a case study of tectonic heredity; Geol. Soc. Am. Bull., v. 82, pp. 125-142.

Ratcliffe, N. M. (1980), Brittle faults (Ramapo fault) and phyllonitic ductile shear zones in the basement rocks of the Ramapo seismic zones, New York and New Jersey, and their relationship to current seismicity; in Manspeizer, W. S. (ed), Field studies of New Jersey geology and guide to field trips, NYSGA Guidebook #52, pp. 278-311.

Ratcliffe, N. M. (1981), Cortlandt-Beemerville magmatic belt: a probable Late Taconian alkalic cross-trend in the central Appalachians; Geology, v. 9, pp. 329-335.

Ratcliffe, N. M., Burton, W. C., D'Angelo, R. M. and Costain, J. K. (1986), Low-angle extensional faulting, reactivated mylonites and seismic reflection geometry of the Newark basin margin in eastern Pennsylvania; Geology, v. 14, pp. 766-770.

Segall, P. & Simpson, C. (1986), Nucleation of ductile shear zones on dilatant fractures; Geology, v. 14, pp. 56-69.

Sibson, R. H. (1977), Fault rocks and fault mechanisms; J. Geol. Soc. Lon., v. 123, pp. 191-213.

Simpson, C. (1981), Ductile shear zones: a mechanism of rock deformation in the orthogneisses of the Maggia Nappe, Ticino, Switzerland; Ph.D. thesis, ETH Zurich.

Simpson, C. and Schmid, S. M. (1983), An evaluation of criteria to deduce the sense of movement in sheared rocks; Geology, v. 94, pp. 1281-1288.

Simpson, C., (1986), Determination of movement sense in mylonites; J. Geol. Ed., v. 34, pp. 246-261.

Tjia, H. D. (1971), Fault movement, reoriented stress field and subsidiary structures; Pac. Geol., v. 5, pp. 49-70.

Weathers, M. S., Bird, J. M., Cooper, R. F. and Kohlstedt, D. L. (1979), Differential stress determined from deformation-induced microstructures of the Moine thrust zone; J. Geophys. Res., v. 84, pp. 7495-7509.

Westgate, L. G. (1896), The geology of the northern part of Jenny Jump mountain, in Warren County, New Jersey; N. J. Geol. Surv. Ann. Rept., 1895, pp. 21-61.

Winkler, H. G. F. (1974), Petrogenesis of metamorphic rocks; Springer-Verlag, New York, 334 pp.

Wise, D. U., Dunn, D. E., Engelder, J. T., Geiser, P. A., Hatcher, R. D., Kish, S. A., Odom, A. L. and Schamel, S. (1984), Fault-related rocks: suggestions for terminology; Geology, v. 12, pp. 391-394.

Wojtal, S. (1986), Deformation within foreland thrust sheets by populations of minor faults; J. Struct. Geol., v. 8, pp. 341-360.

Zadins, Z. Z. (1983), Structure of the northern Appalachian thrust belt at Cementon, New York; M. S. thesis, Univ. of Rochester, 137 pp.

Zeck, H. P. (1974), Cataclasites, hemiclasites, holoclasites, blastoditto and myloblastites--cataclastic rocks; Am. J. Sci., v. 274, pp. 1064-1073.

Road Log for the Field Trip on the Geology of the New Jersey Highlands

By

Richard Volkert
New Jersey Geological Survey

Avery Drake
United States Geological Survey

Joseph Hull
Rutgers University

Robert Koto
Rutgers University

October 25, 1986

Geological Association of New Jersey

COMULATIVE MILEAGE	STOP MILEAGE	
0.0	0.0	The starting point for this trip is the Quality Inn on Route 46 in Ledgewood, New Jersey. Trip leaves from Quality Inn parking lot at 8:00 a.m. Participants will walk eastward about 0.1 mile along the shoulder of the northbound lane of Route 46. We are in the Stanhope 7 1/2-minute quadrangle. Please be exceedingly careful because of the heavy traffic on this highway. STOP 1. CHARNOCKITIC DIORITE ALONG ROUTE 46 (30 Minutes). This typical exposure of diorite displays many characteristics of the unit and of charnockitic rocks in general. The rock is a medium-grained, white- to light-tannish-gray weathering, greenish-gray, greasy-lustered, moderately well-lineated, massive- to locally moderately well-layered diorite to quartz diorite. It contains thin wisps and layers of intercalated amphibolite. Plagioclase (An_{35} of this exposure) is by far the dominant constituent. Overall, the composition of plagioclase in the charnocktic rocks is somewhat variable, ranging from An_{27} to An_{49} (Sims, 1958;

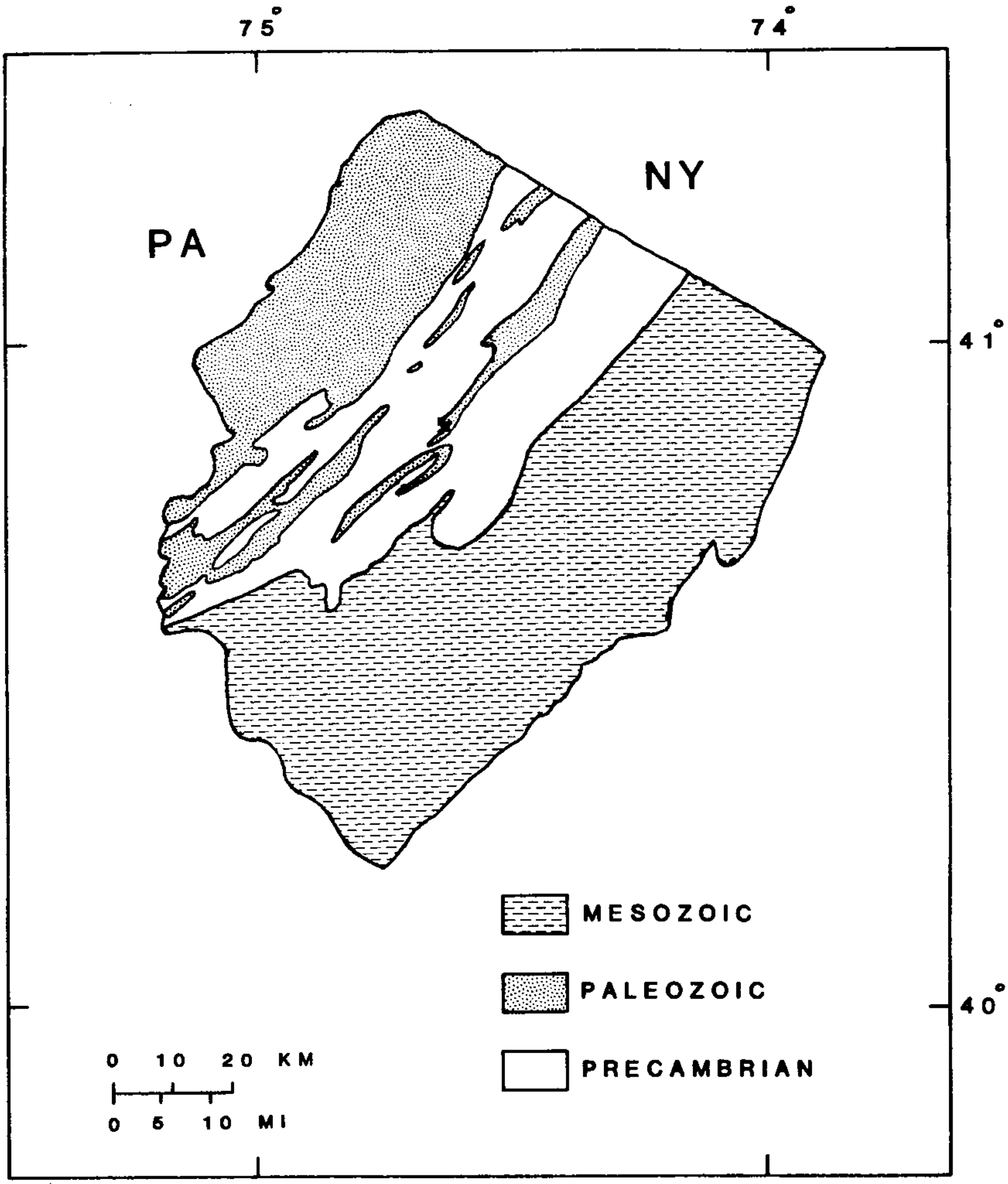


Figure 1. Location map of the Precambrian Highlands and its relation to the Triassic/Jurassic Lowlands and Valley and Ridge Provinces.

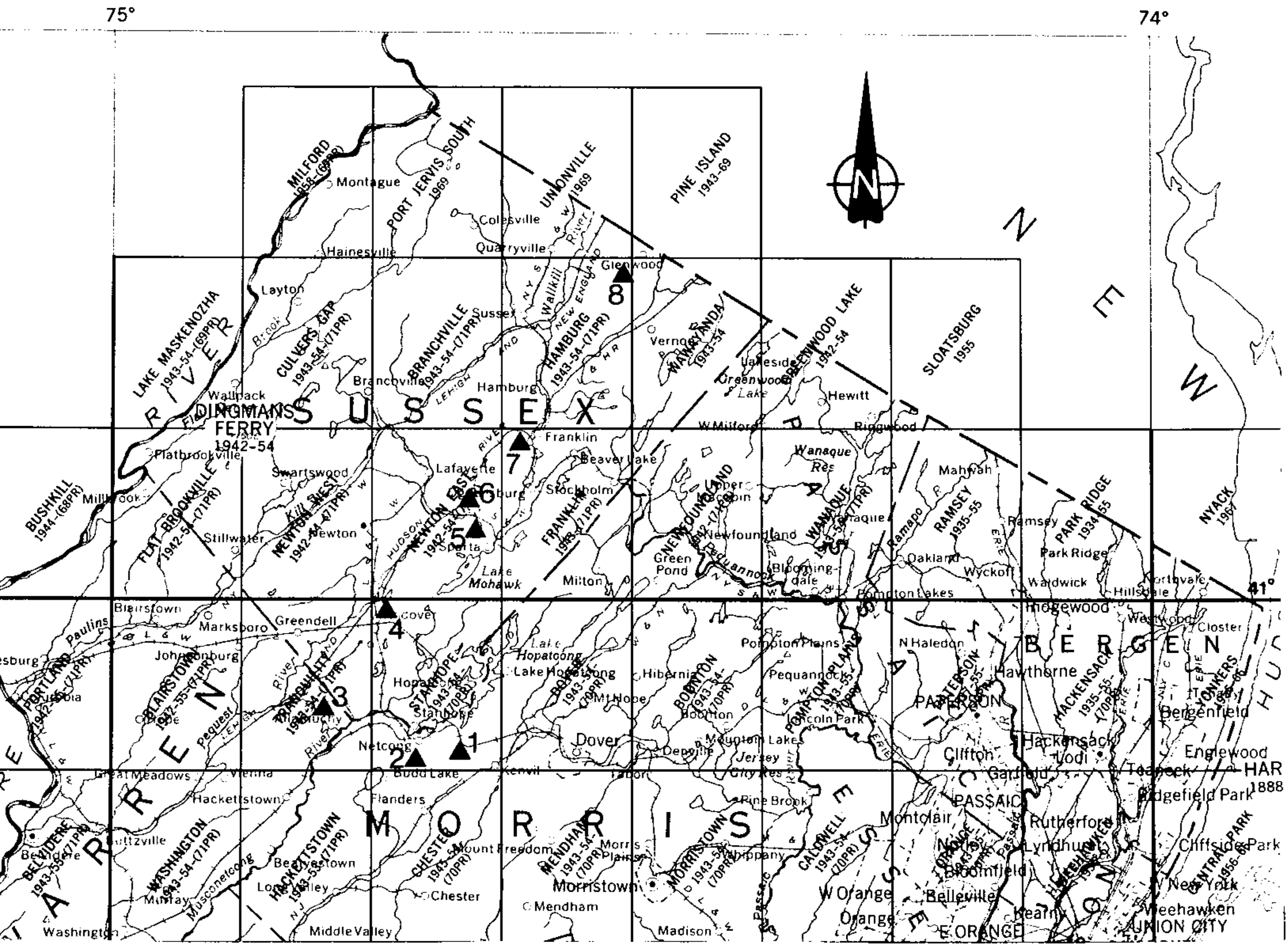


Figure 2. Field trip stop locations.

CUMULATIVE MILEAGE	STOP MILEAGE

Smith, 1969). Accessory minerals include quartz, hypersthene, clinopyoxene, and hornblende; opaques and apatite occur in trace amounts. This outcrop is on the west limb of a northeast plunging upright synform (fig. 3). The foliation has an attitude of N45E 86°SE and a pronounced mineral lineation defined by hornblende prisms, plunges about 30°N52E. The dominant joint set is N45W 81°NE, and subordinate sets are due north, 75°W, N08W 17°NE, and a conjugate to the dominant joint has an attitude of N37W 60°SW. Thin chlorite-coated fractures have attitudes of N18E 44°SE and N30W 56°SW. Evidence for slight slippage is visible on some joint surfaces. Limonite stains resulting from the alteration of constituent mafic minerals occur on some joint surfaces. This is a common feature of many Proterozoic outcrops in the Highlands. The origin of these hypersthene-bearing rocks is problematic. Protoliths for these charnockitic rocks in the N.J. Highlands range from igneous intrusive (Sims, 1958) to metasedimentary (Baker and Buddington, 1970) to metavolcanic (Young,

Figure 3. Geologic map of the Stanhop Quadrangle. Explanation of symbols is on facing page.

EXPLANATIONS

Qd	Quaternary deposits	
Sg	Green Pond Conglomerate	-- Silurian
Or	Rickenbach Dolomite	-Ordovician
Os	Stonehenge Formation	
O€a	Allentown Dolomite	-- Ordovician and Cambrian
€l	Ieithsville Formation	-Cambrian
€h	Hardyston Formation	
Ybh	Hornblende granite	-- Proterozoic
Ybs	Hornblende syenite	
Ybb	Biotite granite	
Ypg	Pyroxene granite	
Yps	Pyroxene syenite	
Ymr	Marble	
Ym	Microcline gneiss	
Ymp	Clinopyroxene-quartz-microcline gneiss	
Yp	Pyroxene gneiss	
Yb	Biotite-quartz-feldspar gneiss	
Ya	Amphibolite	
Yd	Diorite	
Ylo	Quartz-oligoclase gneiss	
Yh	Hypersthene-quartz-andesine gneiss	

.. Contact - Dotted where concealed

Faults - Dotted where concealed: queried where doubtful

.. Tear - Arrows indicate relative horizontal displacement

U D .. High angle - U, upthrown side; D, downthrown side

.. Inclined thrust - Sawteeth on upper plate

Folds of Proterozoic age - All folds are in foliation

Overturned antiform - Showing trace of axial surface, direction of dip of limbs, and direction of plunge

Overturned synform - Showing trace of axial surface, direction of dip of limbs, and direction of plunge

Synform - Showing trough line and direction of plunge

Folds of Paleozoic Age - Folds are in bedding

Syncline - Showing trough line and direction of plunge

Field trip stop locations

CUMULATIVE MILEAGE	STOP MILEAGE	
		1969; Drake, 1969, 1984). For a more detailed discussion of these rocks the reader is referred to Drake (1984). Return to parking lot and load cars. Turn right out of parking lot and head west on Route 46.
.45		Outcrops of pyroxene granite on right.
.90		Wisconsin terminal moraine on left.
1.75		Junction with Route 206.
		Keep left and go south on Route 206.
12.95		Turn right on Gold Mine Road.
12.98	12.98	STOP 2. PYROXENE SYENITE AND GRANITE IN ESC QUARRY (30 Minutes). NOTE: THIS IS AN ACTIVE QUARRY AND PERMISSION MUST BE OBTAINED TO ENTER. PLEASE MAINTAIN CAUTION AND STAY BACK FROM QUARRY FACE AS BLOCKS OVERHEAD ARE UNSTABLE.

This is an excellent exposure of relatively fresh clinopyroxene-bearing syenite and granite. These rocks occur in thick sheets which, to the north of here and to the west of Lake Hopatcong (fig. 3), have been folded into a doubly plunging synform slightly overturned to the northwest (Chapman, 1966; Young, 1969; Volkert and others, in press).

CUMULATIVE MILEAGE | STOP MILEAGE

The rock in this quarry ranges from pyroxene syenite on the east to pyroxene granite on the west. These rocks can be distinguished only by their quartz content and their intimate relation is well displayed within the quarry. The relation of these quartz-rich and quartz-poor pyroxene-bearing rocks is problematic. We feel they are comagmatic on the basis of field, mineralogic, and chemical data. Further work, however, will be necessary to prove this unequivocally.

The rock here is a medium-grained, light-medium gray weathering, greenish-gray, moderately well-foliated syenite to granite, containing clinopyroxene, typically ferrohedenbergite (Baker and Buddington, 1970; Young, 1969), variable amounts of quartz, mesoperthite to microantiperthite, and traces of sphene, opaques, apatite, and zircon. Thin layers of intercalated amphibolite are ubiquitous within the rocks in the quarry. Conformable to locally disconformable seams and pods of granitic to syenitic pegmatite are common and have

CUMULATIVE MILEAGE	STOP MILEAGE	
		gradational contacts with the host rock, suggesting that they formed in situ by local melting. Moderately thick veins of quartz occur oblique to foliation. These veins contain thin veinlets of magnetite near their contacts with the wall rock.
		Foliation ranges from N51E 72°SE on the east to N51E 72°SE to N14E 64°SE on the west. A moderately strong mineral lineation plunges about 54°N51E. A small shear zone can be seen on the quarry wall. It is subparallel to the foliation and has an attitude of about N25E 64°SE. Late brittle deformational features consist of abundant chlorite-coated fractures. The main joint sets are oriented N35W 81°NE, N45W 80°SW, N75W 90°±, N40W 33°NW, and N54E 44°SE. Cars turn around, return to Route 206, turn left, and head north.
15.98		I-80 underpass. Follow signs for Waterloo Village. Approaching tricky trafic circle.
16.43		Trafic circle. Go 1/3-way around. Follow signs for Route 206 North.
16.78		Trafic light in Stanhope

CUMULATIVE MILEAGE	STOP MILEAGE	
16.88		Lake Musconetcong on right
17.16		Outcrop of pyroxene syenite on right
18.11		Outcrop of mylonitic quartz-oligoclase gneiss on right. Abandoned Stanhope magnetite mine is on hill behind the Black Forest Inn.
18.38		Acorn Road on left
18.71		Traffic light. Turn left onto Waterloo Road (Route 604). Kennedys fault (Drake, 1967a, 1967b) parallels the road. Middle Proterozoic rocks are to the right;
18.98		Leithsville Formation (Middle and Lower Cambrian) underlies the valley to the left. Entrance to Waterloo Village
19.28	6.3	Large roadcut on right. Park on side of road. Entrance to Mount Allamuchy Boy Scout Camp. I-80 is just ahead. STOP 3. Separate description by Hull and Koto is found on the five following pages. Return and turn left onto Route 206 North.
22.33		Traffic light at intersection with Route 607. We are crossing the trace of the Kennedys

(cumulative Road Log continues on page 83)

STOP 3.

This stop is located along County Road 604 near the village of Waterloo at the entrance to the Mt. Allamuchy Boy Scout camp in the Tranquility quadrangle. Roadcuts along 604, the camp entrance and Interstate 80 (just to the south) expose most of the major types of deformation zones found within the New Jersey Highlands. PERMISSION must be obtained to examine the cuts along the camp entrance. In addition, because of the limited number of good outcrops here, we strongly request:

NO HAMMERS!

ORTHOGNEISSES. Portions of the 604 roadcuts on either side of the camp entrance are sketched in Figure 4. The roadcuts strike northeast, almost normal to the Grenville (Proterozoic Y) fabric. The southern roadcut exposes amphibolite gneiss (plagioclase, hornblende, diopside, accessories) surrounding small, lens-shaped bodies of relatively undeformed plagioclase metaporphyry (same mineralogy). Both the gneiss and the metaporphyry are cut by undeformed felsic pegmatite dikes, that are often rich in magnetite. The northern roadcut exposes both amphibolite and granitic gneiss (alkali feldspar, quartz, hornblende, plagioclase, garnet and other accessories) intruded by pegmatites. The contact between the two gneisses is exposed in the northern roadcut, and is believed to be a deformed intrusive contact. First, the Grenville foliation is locally oblique to the contact. Second, tabular xenoliths of amphibolite can be found within the granitic gneiss adjacent to the contact. Overall, the contact and gneissosity are parallel to the Grenville schistosity, suggesting very high strains.

GRENVILLE DEFORMATION ZONES. A more detailed sketch of part of the southern roadcut adjacent to I-80 is shown in Figure 5a. This rare exposure illustrates how a rather homogeneous igneous protolith can be transformed to gneiss simply by deformation. The lithons (also augen, lenses) contain relatively undeformed, plagioclase metaporphyry. The cm-sized phenocrysts are compositionally zoned, but other igneous textures are not preserved. The larger lithons are rounded at their terminations and show little internal deformation, whereas smaller augen are tapered and often show a weak fabric. Their boundaries may be sharp or gradational into the adjacent amphibolite gneiss. The metaporphyry is deformed to produce augen gneiss and laminated "pin stripe" gneiss. Individual phenocrysts are deformed into mm-sized plagioclase laminae, indicating the magnitude of the strain. The rock deformed by high temperature intracrystalline plasticity, and the resultant tectonites are thoroughly recrystallized, with a equigranular mosaic texture. These tectonites could be considered high grade mylonites.

The Grenville fabric elements are shown in Figures 5b and 5c. In general the foliation wraps around the lithons. This anastomosing fabric (in three dimensions) is shown by the scatter in orientations in Figure 5b. Where deformation is complete and no lithons are preserved, the Grenville fabric shows a uniform orientation (Figure 5c). The gneiss exhibits a schistosity that is stronger than the mineral/shape lineation (S > L tectonite). In places, the foliation makes a small angle to the lithon boundaries, and the geometry resembles a classic

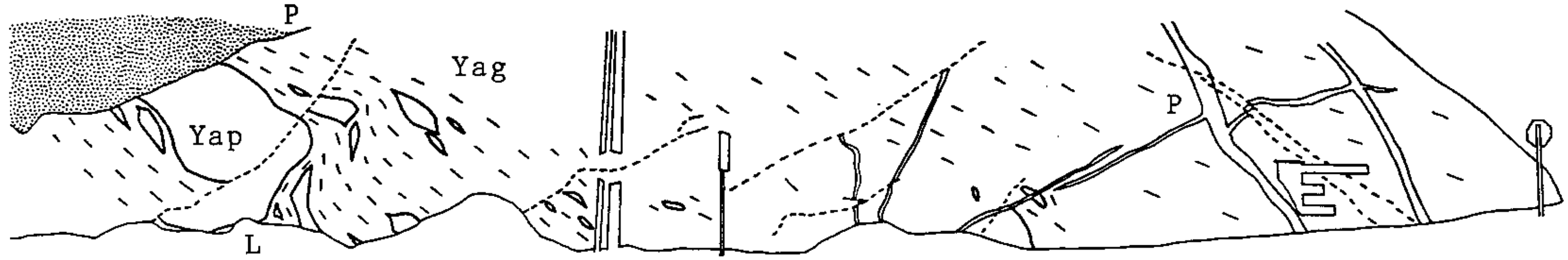
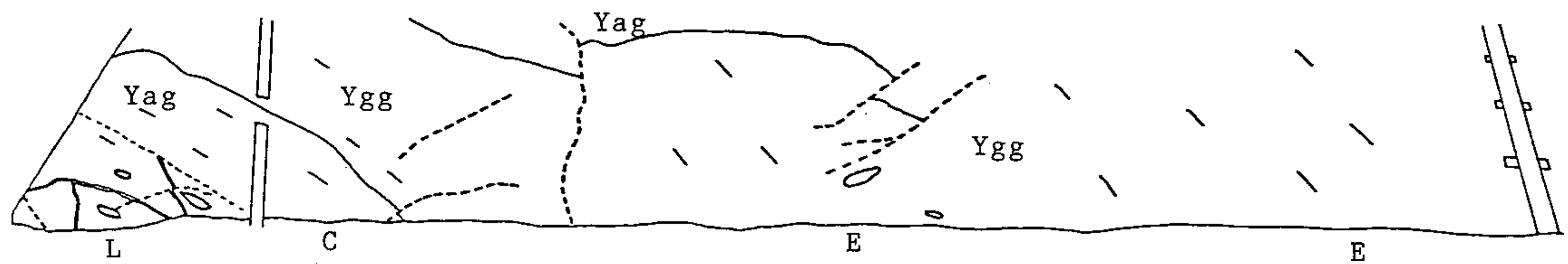


Figure 4. Sketches of the roadcuts along the west side of County Road 605, both to the south (above) and north (below) of the entrance to the Boy Scout camp. View is looking WNW. Scale is 1:200. Yap-amphibolite porphyry; Yag-amphibolite gneiss; Ygg-granitic gneiss, P-pegmatite. Contacts between gneisses are thin solid lines. Short ticks are the trace of gneissosity/foliation. Dashed lines are small faults. Lithons (augen) of metaporphyry may be seen at L. The contact is best viewed at C. Epidote-rich faults can be seen at E. The late chlorite zones are found just inside the entrance to the Boy Scout camp on the northern side.



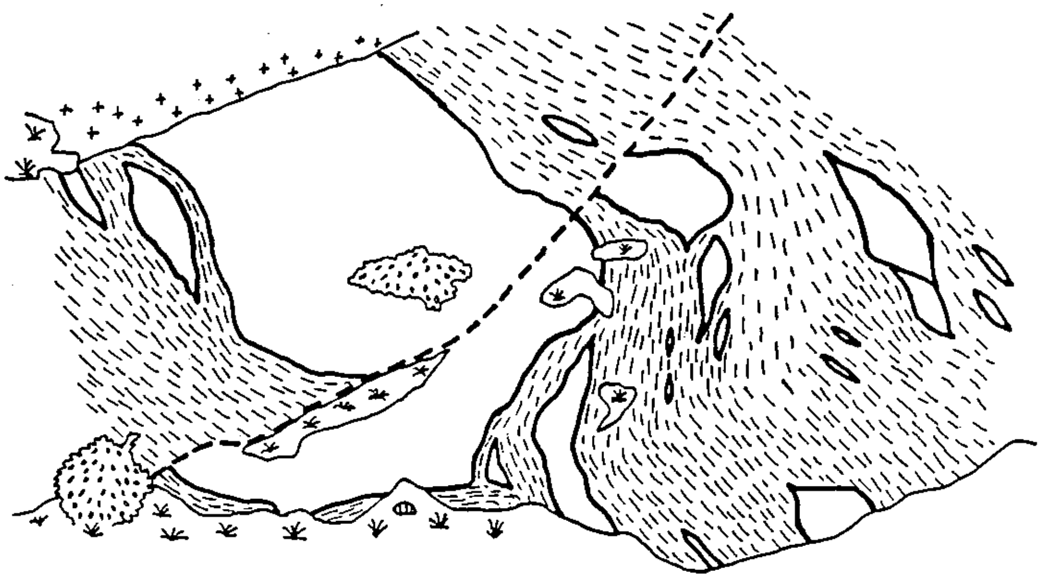


Figure 5a. Enlarged portion of the southern outcrop showing undeformed lithons of plagioclase metaporphyry (blank) surrounded by augen gneiss. Both the lithons and gneiss are cut by pegmatite (pluses) and a small fault (dashed line). The scale is 1:50.

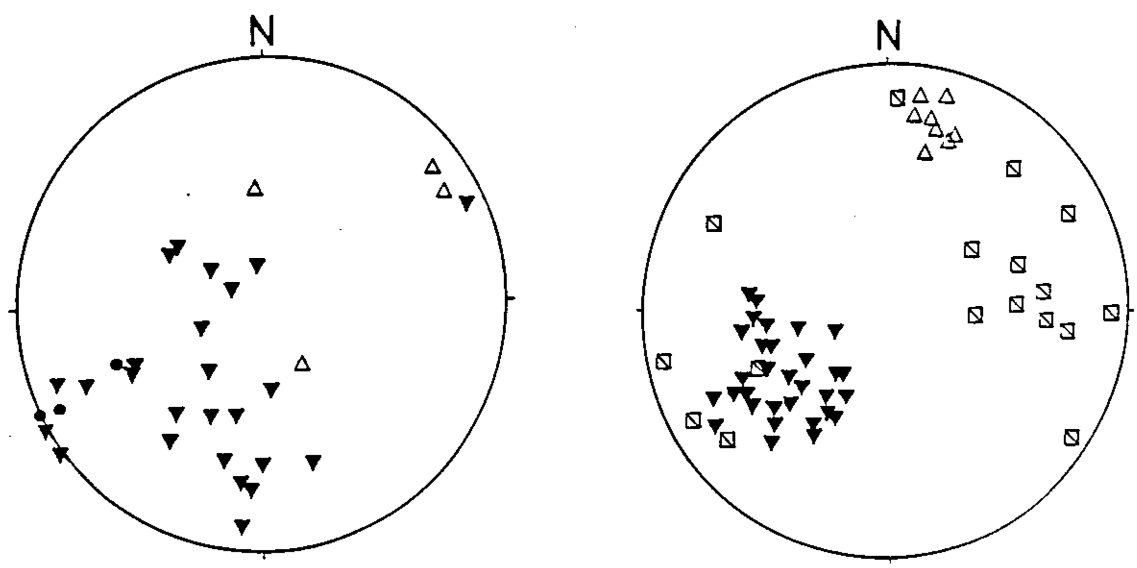


Figure 5b. Fabric elements from the boundaries of the lithons. Foliation (solid triangles) and lineation (open triangles) are scattered. Foliation is subparallel to lithon boundaries (solid dots). Figure 5c. Proterozoic elements from rest of roadcuts. The fabric is uniformly oriented. Pegmatites (squares) also shown. Lower hemisphere Schmidt.

simple shear zone. However, stereographic analysis reveals that the stretching lineation is subparallel to the rotation axis. These deformation zones were most likely produced by heterogeneous flattening, not simple shear.

RETROGRADE DEFORMATION ZONES. The retrograde deformation zones cut all Precambrian rocks and structures (including pegmatites) as seen in Figures 4a and 4b, however their absolute ages are not really known. Based on cross-cutting relationships (most of which can be seen in these exposures), we have established a relative chronology of (1) early chlorite-rich zones, (2) epidote-rich zones and (3) late chlorite zones. There is good evidence elsewhere in the Highlands that epidote- and chlorite-rich zones can form simultaneously in different rock types.

Little is known about the early chlorite-rich brittle deformation zones (BDZ) along 604 as they are poorly developed and obscured by later deformation. Good cross-cutting relationships with later epidote zones can be seen at several localities in the northern roadcut (Figure 4b). The epidote-rich BDZ vary from a few mm to several cm in width: displacement/width ratios average 100. Grain size reduction by fracturing (cataclasis) produced breccias and microbreccias, consisting of feldspar and quartz clasts in a fine-grained matrix dominated by epidote. Plastic deformation is limited. The breccia zones are bounded by and contain sliding surfaces (slickensides) associated with fine-grained ultracataclasites. These slickensides are often highly polished and exhibit tool-and-groove type slickenlines. The epidote zones are abundant, especially in the granitic rocks, and form a "fault population" that produces bulk deformation of the gneisses. Slip on these BDZ can be determined from slickenlines and separation of pegmatite dikes (Figure 6a). The geometry of the epidote BDZ at Stop 3 is one of the most complex we have encountered in the Highlands (Figure 6b). Further study is required to determine the principle strains associated with this retrograde deformation.

Late chlorite-rich BDZ are exposed along the entrance to the Boy Scout camp on the north roadcut (Figure 4b). There are two broad chlorite zones, each about a half meter wide, but a network of small BDZ can be found in both the hangingwall and footwall. The broad zones contain a wide variety of tectonites, including quasiplastic mylonites, cataclasites (microbreccias) and ultracataclasites. The latter is best exemplified by black, glassy pseudotachylite that has been subsequently deformed during late movement on the zones. The chlorite zones are transverse to the regional trend and dip moderately northeast subparallel to gneissosity (Figure 6c). Both the direction and sense of movement on the large zones are difficult to determine, but oblique normal slip is indicated.

It is tempting to correlate the three retrograde events with the three Phanerozoic orogenies (Taconic, Alleghanian, Mesozoic). Such a chronology is only speculative without dated, bracketing intrusives. Eocambrian, Ordovician and Mesozoic dikes are present elsewhere in the Highlands, and the cross-cutting relationships are discussed in Hull et al (this volume).

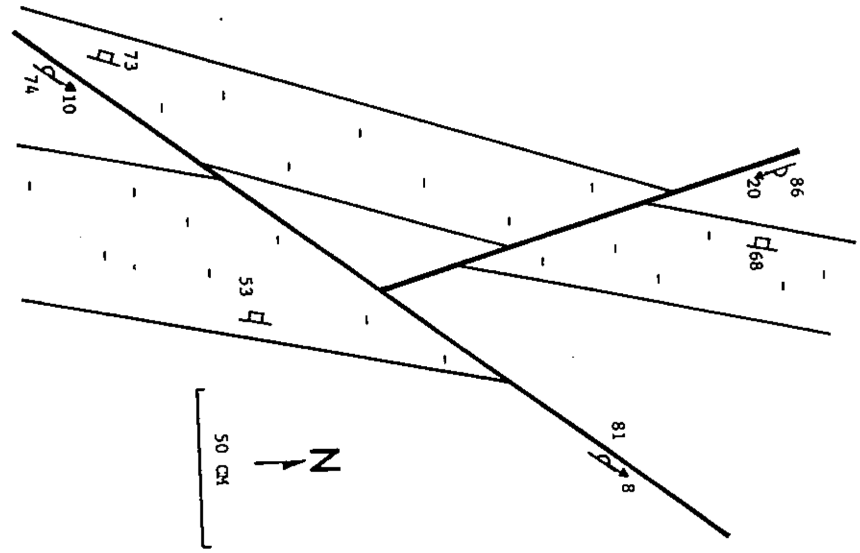
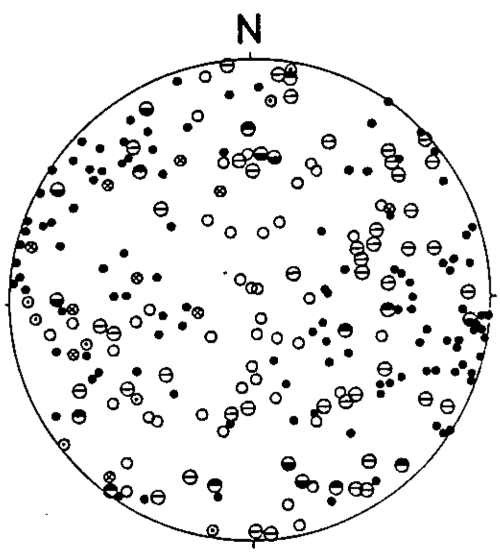


Figure 6a. Cross-cutting relationships among a pegmatite dike (pattern) and epidote-rich faults (heavy lines). The orientations are also shown.



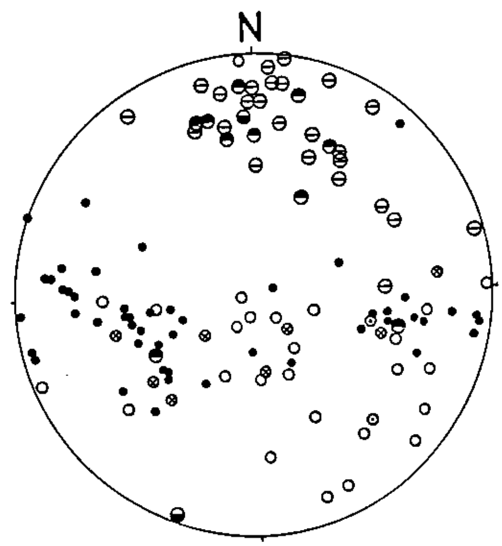


Figure 6b. Epidote fault zones. A NE-striking set dominates but the orientations are highly scattered. Figure 6c. Structural elements from the late chlorite zone, including both footwall and hangingwall zones, the main zone itself, and late sliding surfaces. Lower hemisphere Schmidt. For further explanation, see Hull et al (this volume).

CUMULATIVE MILEAGE	STOP MILEAGE	
		fault. A small tectonic lens of Hardyston Quartzite (Lower Cambrian) and Leithsville Formation is exposed on the north side of Route 607.
23.03		Entrance to Byram Crushed Stone Quarry on right. As we continue north on Route 206 we are travelling about parallel to the axial trace of a quartz-oligoclase gneiss cored antiform. Rocks in the quarry are hornblende syenite. Outcrops of charnockitic rocks and amphibolite crop out farther to the north.
23.92		Crossing the trace of the Wright Pond fault (Hague and others, 1956).
24.23		Traffic light. Cranberry Lake on left.
24.64		Crossing the axial trace of the northeast-trending Cranberry Lake synform that is cored by amphibolite. We have just passed outcrops of amphibolite and hornblende syenite. Cliff face to the northwest is Allamuchy Mountain.
25.08		Outcrop of pyroxene granite on right.
25.33		Outcrop of hornblende syenite on right.
25.48		Outcrop of charnockitic gneiss on right.

CUMULATIVE MILEAGE	STOP MILEAGE	
25.7		Outcrop of quartz-oligoclase gneiss and amphibolite on right. Outcrops from here north to Andover are microcline gneiss.
26.25		Railroad underpasses
26.75		Trafic light in Andover. Turn right onto Route 517.
26.83		Turn left into A and P parking lot and park. Proceed to Stop 4 on foot.
26.9		Road intersection. Bear left on Route 517 North. Entrance to abandoned sand and gravel pit on right.
26.99	7.71	STOP 4. MICROCLINE GNEISS IN ABANDONED SAND AND GRAVEL PIT OF ANDOVER INDUSTRIES, ANDOVER (60 Minutes). This exceptional exposure lies close to the contact of a thick fold-repeated sequence of quartzofeldspathic gneiss to the north and east and lower Paleozoic sedimentary rocks of the main outcrop belt to the west (fig. 3). The Proterozoic unconformity is obscured here by Quaternary deposits, but is exposed a short distance to the north and will be seen later at Stop 7 of this trip.

CUMULATIVE MILEAGE | STOP MILEAGE

The rock here is a light-weathering, pinkish-white-to-light-tan, well-layered and foliated, medium-grained gneiss containing quartz and microcline, subordinate plagioclase, and locally biotite, garnet, and sillimanite. Thin, conformable, garnetiferous, biotite-rich layers occur locally, probably reflecting compositional variation within the protolith. In outcrops visible to the immediate east, biotite decreases and hornblende becomes the dominant accessory. A likely protolith for this unit is an arkosic sandstone, in which the biotite-rich layers reflect more argillaceous (Al-rich) beds. Abundant, conformable to locally discordant seams of quartzofeldspathic pegmatite cut the gneiss and locally have mafic rims. These pegmatites are interpreted to be locally generated anatectites.

Compositional layering and foliation have an attitude of about N15E 35°SE, and a strong lineation defined by biotite plates and elliptical pods of magnetite plunges about 23°N57E. Small intrafolial folds with a sinistral rotation sense deform the regional

CUMULATIVE MILEAGE	STOP MILEAGE	
		foliation and have a less well-defined axial surface foliation. This suggests at least two separate phases of folding, an interpretation supported by the dome and basin type fold interference pattern clearly visible on dip slope surfaces of the outcrop. Recent geologic mapping in the Stanhope quadrangle (Volkert and others, in press) shows that these outcrops are on the west limb of a broad northeast-plunging antiform that is overturned to the northwest. The three main joint sets are N75W 90°±, N74E 70°SE, and due E 69°S. Numerous brittle fractures crosscut all earlier fabric elements, and while variable in their orientation seem to have dominant trends of N59E to N66E and especially due E. (Hull, Koto, and Volkert, unpublished data). Movement has occurred along these fractures and is shown by the offset of a mafic marker (fig. 7). Movement sense is not consistent, being south wall up on some fractures and north up on others, displacements are as great as 37 cm. Thin, 2-3 cm wide zones of healed breccia trend about due E

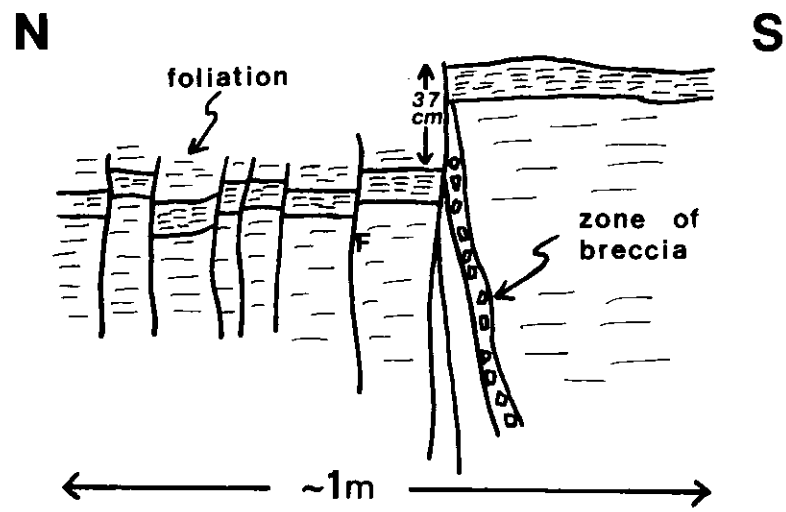


Figure 7. Outcrop sketch showing offset of biotite-rich marker along late brittle fracture and zone of healed breccia.

CUMULATIVE MILEAGE	STOP MILEAGE	
		and appear to be contemporaneous with the dominant fractures. Fault population analysis by Hull and others (fig. 8) suggests that these rocks have undergone mixed pure and simple shear having north-south horizontal extension and west-directed shortening.
		These rocks have obviously been deformed under conditions of relatively high strain, with an early developed ductile fabric parallel to layering and foliation which is probably Proterozoic. Quartz, garnet, and sillimanite bands are stretched and flattened, pegmatite pods are flattened and rotated, and, locally, augen are developed. Tails on the augen and rotation of garnets suggest a sinistral (south over north) sense of shear. Seams of pegmatite locally wrap around the more mafic layers, display "pinch and swell" structure of boudins. Considerable folding (and consequent shortening) of these pegmatitic layers have occurred and is well shown in places (fig. 9).
		Return to cars by foot and reload.
		Turn left and continue north on Route 517.

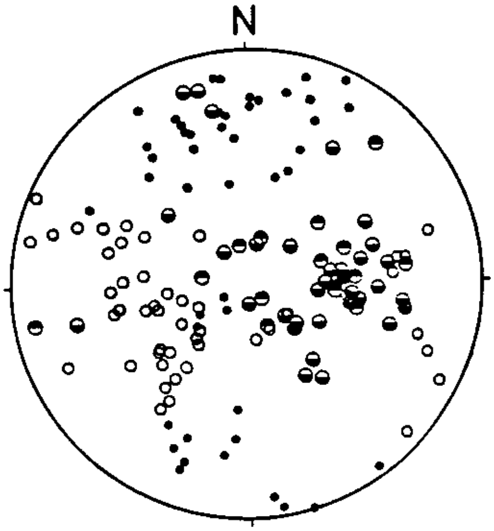


Figure 8a.

- ● pole to BDZ
- ○ slickenline
- rotation axis-up
- rotation axis-down

Figure 8b.

X Extension direction

Y Intermediate direction

Z Shortening direction

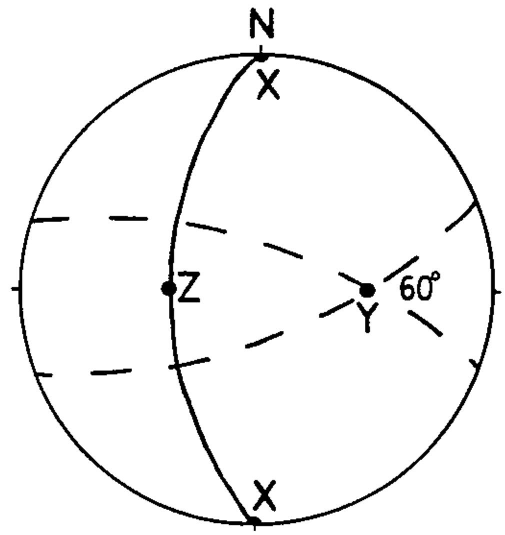


Figure 8. (a). Retrograde, brittle deformation zones (BDZ), including slickensides (both tool-and-groove and fibrous) and small faults. BDZ form steep, N- and S-dipping conjugates (the S-dipping set dominates). Slip is mostly oblique (moderately west-plunging) and normal. Rotation vectors of both polarities (defined by the right hand rule) plunge moderately to the east. (b). Strain geometry formed by the BDZ. The deformation is two dimensional (plane) strain, produced by pure shear (Anderson conjugates at 60°). A component of simple shear (a single orientation) may also contribute. Extension (X) is horizontal-NS and shortening (Z) plunges moderately west, bisecting the conjugate faults (dashed lines). Stereonets are lower hemisphere Wulff. For a complete explanation of the fault population analysis, see Hull et al (this volume).

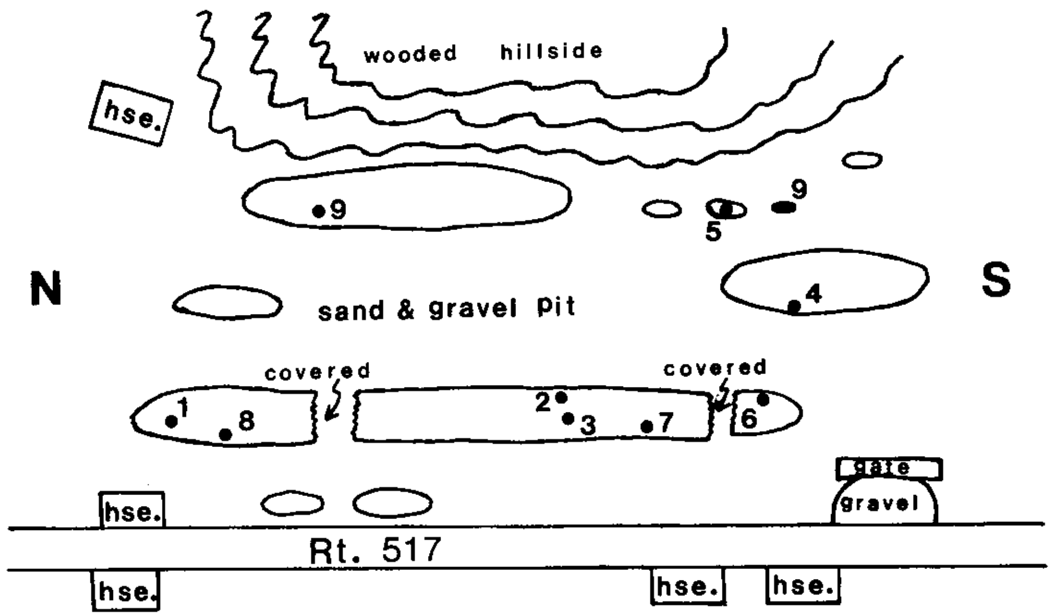
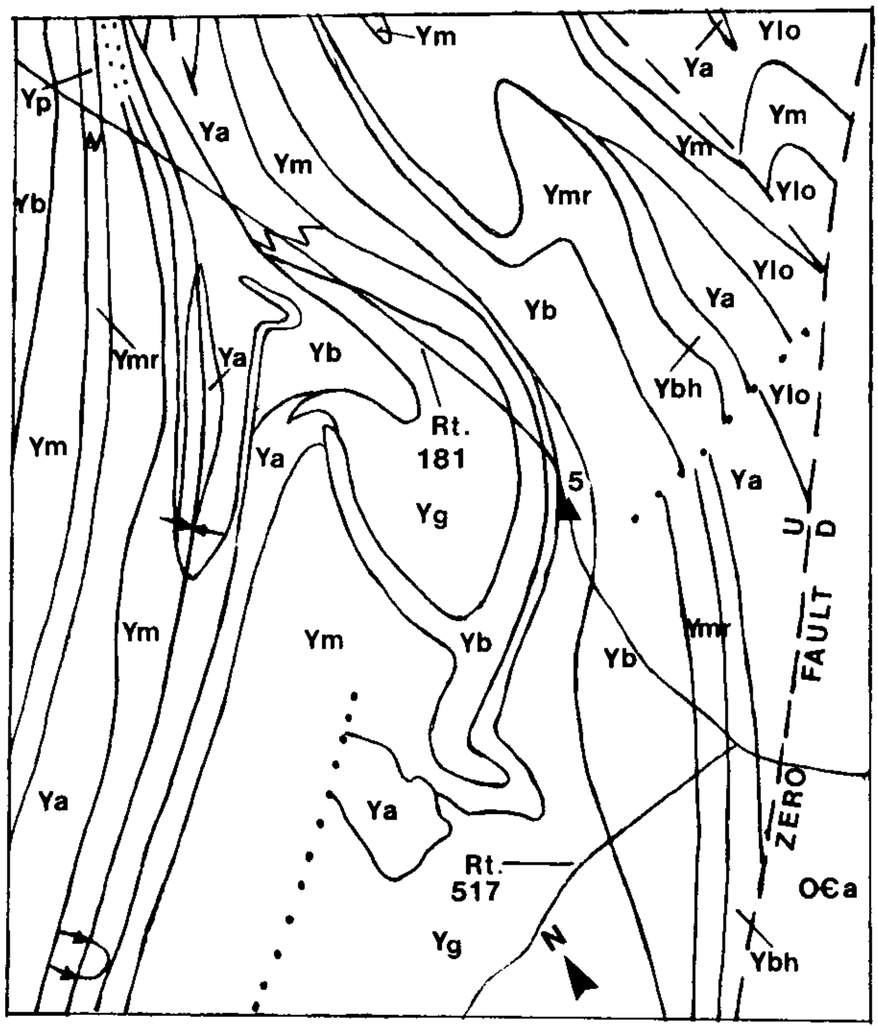


Figure 9. Sketch showing outcrops in sand and gravel pit in plan view. Numbers refer to locations of interesting features: 1. ductile shear zone, fold closure, asymmetric augen, flattened garnets; 2. elliptical magnetite pods; 3. dome and basin fold interference patterns; 4. sheared out hematite layer; 5. nice fold closures; 6. zones of healed breccia; 7. offset of marker along subvertical fracture; 8. offset of marker and healed breccia; 9. folded pegmatitic seams.

CUMULATIVE MILEAGE	STOP MILEAGE	
26.9		Road intersection. Keep to left on Lenape Road - Route 517 North
26.99		Passing sand and gravel pit which was Stop 4.
27.69		Enter Newton East 7 1/2-minute quadrangle
28.29		Lake Lenape on left.
29.77		Outcrop of biotite-quartz-plagioclase gneiss on right.
30.72		Perona Farms Restaurant on right.
30.77		Outcrop of microcline gneiss on left.
32.52		T-intersection. Keep right on 517 North.
33.17		Outcrops of "granodiorite gneiss" on right.
33.87		Blinking light. Keep left at intersection onto Route 181 North.
34.52	7.53	STOP 5. "GRANODIORITE" GNEISS ALONG ROUTE 181 (Optional), (30 Minutes). Be extremely careful in crossing highway. Walk up left side of road. DO NOT step back to admire the outcrop or you might create a vacancy for an unemployed geologist! This rock, as named and defined by Hague and others (1956), is a "medium-light to medium-dark, medium-grained gneiss which contains biotite, hornblende, and local pyroxene." As mapped, it occurs in two disproportionate and detached bodies, the larger is a northeast

striking sheet which extends from east of Lake Lenape to Fox Hollow Lake and the smaller is a heart-shaped refolded body north of Sparta (fig. 10). At this locality we are on the extreme northern tip of the main body near its contact with the enclosing biotite gneiss. Sample 3 (Table 1) is from this body. The second body is surrounded on all sides by biotite gneiss. Sample 1 (Table 1) is from the interior of that body.

The rock here is a medium-gray weathering, greenish-gray, moderately well-layered, well-foliated parallel to layering, well-lineated, medium-grained gneiss which ranges in composition from biotite amphibolite on the south to biotite-quartz-oligoclase gneiss on the north. Concentrations of biotite on foliation surfaces make this rock appear much darker and biotite-rich than it actually is. A few thin, conformable seams of albite-oligoclase granite occur in gradational contact with the host rock. Both rocks have a similar mineralogy. Constituent minerals in the gneiss are: plagioclase, quartz, hornblende, and biotite, with traces of opaques and apatite. Plagioclase



SCALE

0 1 MILE

Figure 10. Geologic map of the Sparta area, Newton East 7½-minute quadrangle (modified after Hague and others (1956)). Yg is granodiorite gneiss. Otherwise, map symbols are the same as in Figure 3.

CUMULATIVE MILEAGE	STOP MILEAGE	

is oligoclase with an average composition of An_{24}. Foliation here averages N32E 56°SE and a well-developed mineral lineation plunges about 23°N42E. The dominant joint set is N50W 61°SW, and a subordinate set is N66E 29°NW.

Table 1 lists chemical analyses of "granodiorite gneiss" (A.A. Drake, Jr. and R.A. Volkert, unpublished data) from localities nearby, and for comparison a sample of quartz-oligoclase gneiss from nearby (A.A. Drake, Jr. and R.A. Volkert, unpublished data) and "average" granodiorite chemistry from Nockolds (1954). It is apparent from the modes reported by Hague and others (1956) and from the chemistry of sample 3 that at least some of the "granodiorite gneiss" contains appreciable K-feldspar. Elsewhere it appears to be virtually devoid of K-feldspar and to be similar to rocks of the Losee Metamorphic Suite (sample 1). This unit is chemically similar to rocks of the Losee Suite where the "granodiorite gneiss" is interlayered with biotite gneiss, whereas it is much more potassic to the south (sample 3) where it is in conformable contact with microcline gneiss and hornblende granite to syenite.

TABLE 1. CHEMICAL ANALYSES OF "GRANODIORITE GNEISS," QUARTZ-OLIGOCLASE GNEISS, AND GRANODIORITE

	1	2	3	4	
		Chemical Analyses (weight percent)			
SiO_2	71.9	73.2	66.3	66.88	(62.64 - 70.47)
Al_2O_3	14.8	15.3	15.4	15.66	(15.47 - 15.82)
Fe_2O_3	0.5	0.37	0.80	1.33	(0.63 - 1.63)
FeO	1.2	0.56	2.7	2.59	(2.05 - 4.31)
MgO	1.1	0.29	2.2	1.57	(0.65 - 2.83)
CaO	2.9	2.8	2.0	3.56	(1.91 - 4.72)
Na_2O	4.3	4.7	3.5	3.84	(3.37 - 4.12)
K_2O	1.2	1.4	4.8	3.07	(2.62 - 3.59)
H_2O+	0.69	0.20	0.54	0.54	(0.42 - 0.70)
H_2O-	0.08	0.18	0.22	-----	---------------
TiO_2	0.36	0.15	0.41	0.57	(0.30 - 1.32)
P_2O_5	0.07	0.04	0.27	0.21	(0.16 - 0.27)
MnO	0.04	0.07	0.10	0.07	(0.03 - 0.09)
CO_2	0.09	0.05	0.04	-----	---------------
Total	99.0	99.0	99.0	100	

1. "Granodiorite gneiss" from outcrop along Rt. 15 about .85 km southeast of intersection of Rt. 15 and New York Susquehana and Western railroad tracks, Newton East 7 1/2-minute quadrangle, N.J., analyst, Leung Mei, U.S. Geological Survey.

2. Quartz-oligoclase gneiss from outcrop along Rt. 15 about .25 km east of Lk. Saginaw, Franklin 7 1/2-minute quadrangle, N.J., analyst, Leung Mei, U.S. Geological Survey.

3. "Granodiorite gneiss" from outcrop about .85 km N35W of Valentines Pond, Newton East 7 1/2-minute quadrangle, N.J., analyst, Leung Mei, U.S. Geological Survey.

4. Average of 137 granodiorite analyses from Nockolds (1954). Ranges are listed in parenthesis.

CUMULATIVE MILEAGE	STOP MILEAGE	
		Additional chemistry, as well as detailed petrography on representative samples of "granodiorite" from along strike and transverse to strike is critical to an understanding of this enigmatic unit. Perhaps in some areas it does approach a meta-granodiorite. Field relationships are ambiguous since this unit is conformably layered with both paragneiss and orthogneiss. Further work is certainly required; perhaps a worthy and interesting topic for a thesis?
35.57		Traffic light. Turn left onto Route 517 North.
36.22		Outcrops of garnetiferous biotite-quartz-plagioclase gneiss on right.
36.32		Flashing light. Turn right onto Houses Corner Road.
36.52		P. Michelotti and Sons Sand and Gravel Pit on left.
37.22	2.7	STOP 6. FOLDED QUARTZ-OLIGOCLASE GNEISS INTRUDED BY LAMPROPHYRE DIKE IN GRINNELL INDUSTRIES SAND AND GRAVEL PIT, SPARTA. NOTE: THIS IS PRIVATE PROPERTY AND PERMISSION MUST BE OBTAINED TO ENTER (90 Minutes, Lunch Stop!).

CUMULATIVE MILEAGE	STOP MILEAGE

This locality exposes an outcrop that is areally small, but significant in terms of the features to be seen. We are presently on the west limb of the Lake Lenape Syncline of Hague and others, (1956), which is overturned to the northwest (fig. 11). Here we will see evidence that this structure has been modified by at least one and possibly two additional phases of folding. The trace of the Hamburg fault follows along the base of the Pimple Hills that is to our immediate east. The rock here is a white weathering, medium-grained, moderately layered, moderately well-to indistinctly-foliated quartz-oligoclase gneiss that contains minor accessory biotite and traces of chlorite, altered pyroxene, and opaques. Some thin, conformable layers of biotite gneiss occur, as well as thin, discordant seams of Byram pegmatite. Foliation is somewhat undulose, reflecting gentle warping and broad buckle folding. In places, the foliation is folded into small, tight, isoclinal, intrafolial folds. A second, less well-defined foliation passes through the hinges of these folds, suggesting at least two phases of folding.

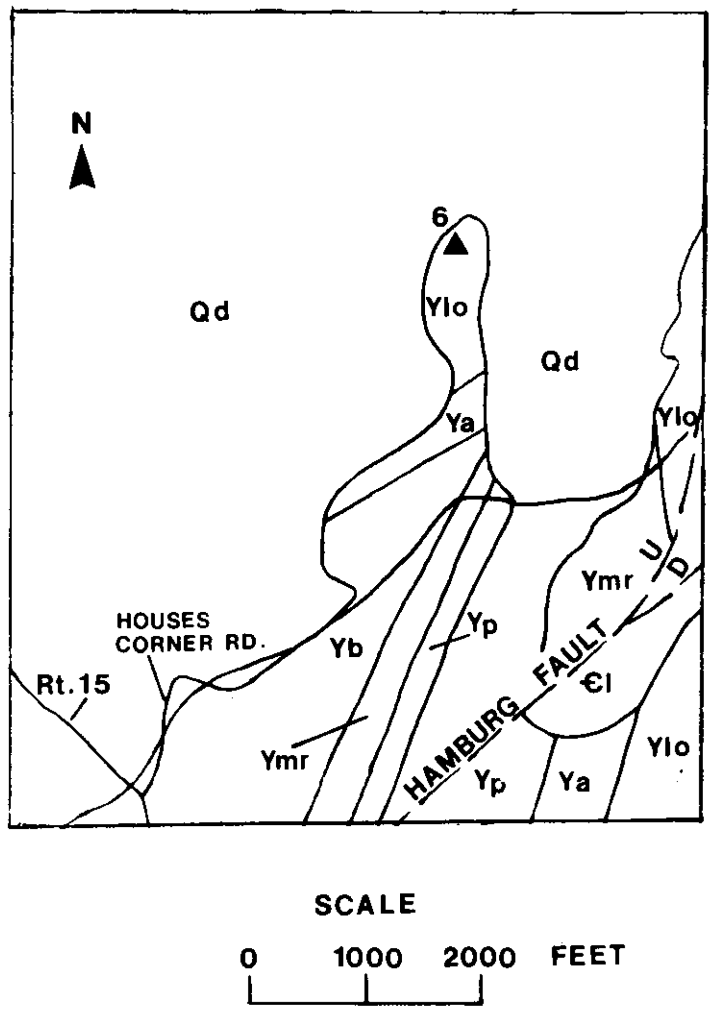


Figure 11. Geologic map of the Woodruffs Gap area, Newton East $7\frac{1}{2}$-minute quadrangle (Volkert, unpublished data). Map symbols are the same as in Figure 3.

CUMULATIVE MILEAGE | STOP MILEAGE

Fold interference patterns, however, are difficult to see here. Interpretation of the local deformational history is hampered by the thickness and widespread distribution of Quaternary deposits and sparse to nonexistant exposure of Proterozoic rock in the immediate area. Regionally, however, all mapped folds are of foliation. Here the foliation has an average attitude of N16E 40°SE. A good lineation as defined by plates of biotite plunges about 28°S25W. This parallels the axial line of the early northeast-trending fold (F_2?). Axes of intrafolial folds (F_3?) plunge about 44°N73E and reflect modification of the early phase. the latest event produced rodding lineations which plunge about 45°S65E parallel to the trend of cross folds (F_4?). Joints in the gneiss trend N32W 73°NE, N64W 75°SW, N24E 67°NW, and due E76°S. The quartz-oligoclase gneiss is intruded by a one meter thick, medium-dark gray, fine-grained- to aphanitic, biotite-bearing, highly undersaturate lamprophyre dike, probably a minette or sadaminette, which exhibits rather interesting chemistry (Table 2). The contact

TABLE 2. CHEMICAL ANALYSES OF LAMPROPYRES FROM NORTHERN NEW JERSEY

	1	2	3	4	5
			Chemical Analyses (weight percent)		
SiO_2	32.6	24.7	34.63	40.71	40.47
Al_2O_3	10.2	7.4	10.21	19.46	11.86
Fe_2O_3	6.4	4.6		7.46	
FeO*			12.56		17.44
FeO	8.7	10.1		6.83	
MgO	6.2	4.9	3.5	6.21	3.1
CaO	12.9	18.9	17.96	11.83	16.8
Na_2O	1.8	1.5	2.88	1.8	1.9
K_2O	3.1	1.6	3.06	3.26	4.21
H_2O+	2.4	1.7	-----		
				1.53	3.6
H_2O-	0.36	0.27	-----		
TiO_2	4.7	3.3	2.87	-----	-----
P_2O_5	3.2	4.2	-----	-----	-----
MnO	0.19	0.29	0.33	0.18	-----
CO_2	6.6	15.2	10.83	0.74	-----
Total	99.0	99.0	98.83	100.01	99.38

1. Stop 6, this trip. Analyst, Leung Mei, U.S. Geological Survey.

2. Dike intruding Ordovician dolomite, 1.5 km S30W of Ross Corner, Newton East 7 1/2-minute quadrangle, N.J. Analyst, Leung Mei, U.S. Geological Survey.

3. Matrix rock from Rutan Hill (lamprophyre micromelteigite), Branchville 7 1/2-minute quadrangle, from Maxey (1976).

4. Minette dike from Franklin, Franklin 7 1/2-minute quadrangle, N.J., L. G. Eakins, analyst, from Iddings (1898).

5. Ouachitite from Rutan Hill, Branchville 7 1/2-minute quadrangle, N.J., J. F. Kemp, analyst, from Kemp (1889).

CUMULATIVE MILEAGE	STOP MILEAGE	
		between gneiss and dike is sharp and the dike has chilled margins. Thin lamellar lines are barely visible on the weathered surface of the dike and probably are flow structures. The age of the dike is problematic. It follows the Beemerville trend, striking N30W, and is quite similar chemically to dikes which intrude Ordovician carbonate rocks to the north and west, and to some dikes in the Beemerville area (Table 2). Biotite from the main nepheline syenite body and a plug at Rutan Hill was dated at 434 Ma (Late Ordovician) by conventional K-Ar and Rb-Sr techniques by Zartman and others (1967) Two prominent joints cut the dike and are oriented at N36E 67°NW and N81E 80°SE, attitudes similar to two of the joint sets observed in the gneiss. The dike apparently was intruded along the dominant joint set in the gneiss.
		Reload cars and exit sand and gravel pit. Turn left and continue east on Houses Corner Road.
37.62		Pimple Hills to right.
38.27		Sand and gravel pit on left. Has exhumed brittlely deformed amphibolite along the trace of the Hamburg fault (Hague and others, 1956).

CUMULATIVE MILEAGE	STOP MILEAGE	
38.92		Y-intersection. Bear left.
38.97		Stop sign. Turn left on West Mountain Road.
39.12		Turn right on Prospect School Road
39.47		Outcrops of amphibolite on right
39.52		Turn right onto Davis Road
39.72		Enter Franklin 7 1/2-minute quadrangle
39.97		Crossing trace of Hamburg fault and passing onto carbonate rocks of the Wallkill Valley.
40.92		Outcrops of Allentown Dolomite (lowest Lower Ordovician and Upper Cambrian) on right.
40.97		Turn right on Maple Road
41.47		Outcrops of Leithsville Formation on right.
41.57		Stop sign. Turn left on to Wildcat Road. Outcrops of K-feldspar gneiss on all sides.
41.67	4.45	STOP 7A. PROTEROZOIC-PALEOZOIC UNCONFORMITY ALONG FRANKLIN GOLF COURSE. NO HAMMERING PLEASE, THIS CONTACT IS RARELY EXPOSED (30 Minutes). The exposures of Cambrian Hardyston Quartzite and Leithsville Formation we will be seeing here occur in a downfaulted, wedge-shaped block bounded on the west by the Hamburg fault (fig. 12). To our east, lower Paleozoic rocks are repeated in another down-faulted block bounded on the west by the Zero fault. We will pass over the trace of this fault and see (from the

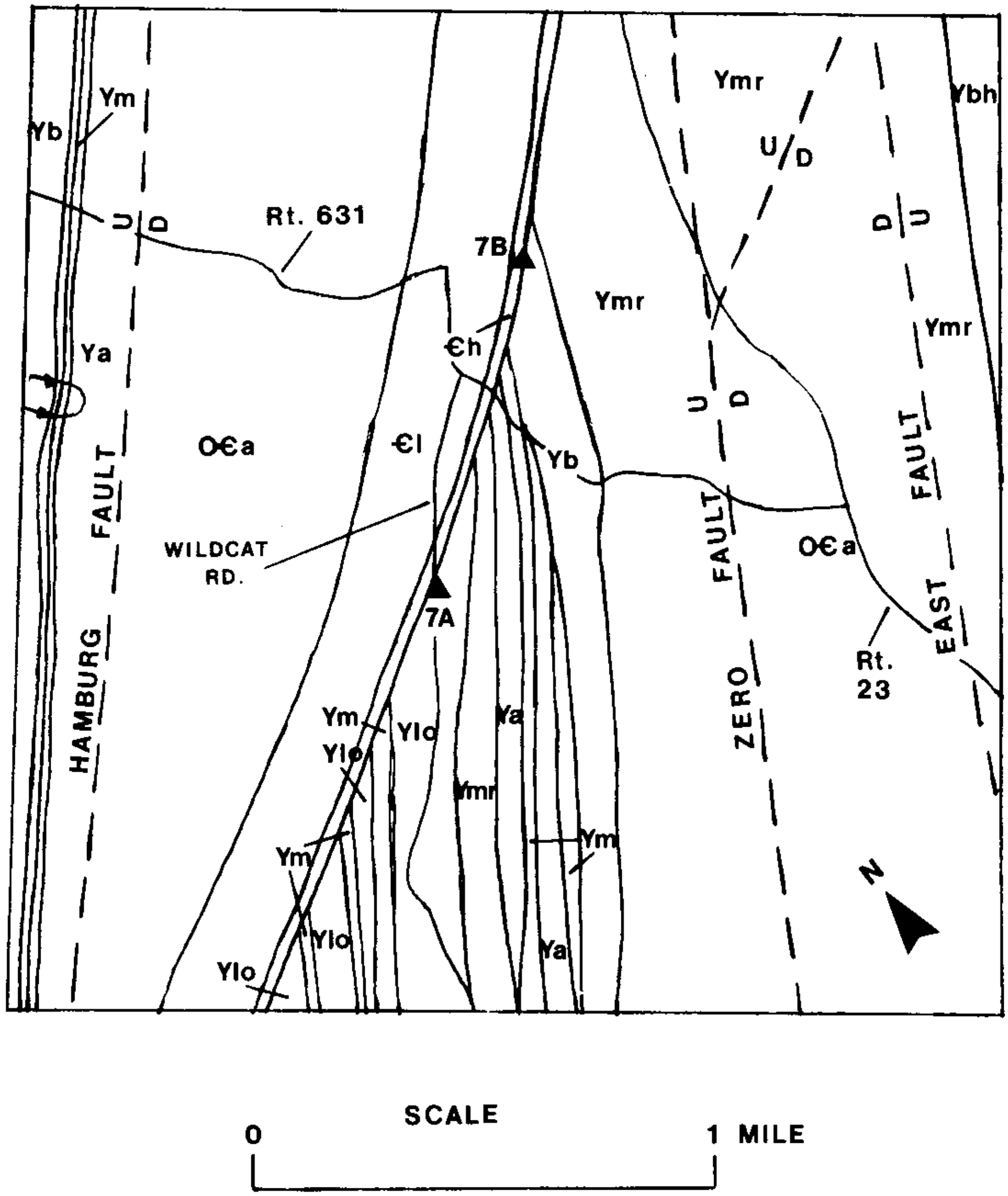


Figure 12. Geologic map of the Franklin area, Franklin 7½-minute quadrangle (modified from Hague and others (1956)). Map symbols are the same as in Figure 3.

CUMULATIVE MILEAGE	STOP MILEAGE	
		vehicles) the contact between the Middle Proterozoic Franklin Marble and Leithsville Formation enroute to stop 8.
		Walking south along the western side of Wildcat Road we see an outcrop of white-to light gray-weathering, medium-fine-to medium-grained, well-layered and foliated, well-lineated quartz-oligoclase gneiss containing abundant accessory biotite, partly altered clinopyroxene, and opaques. Plagioclase forms the majority of the rock here, but along strike the quartz content is variable and, locally, actually exceeds the plagioclase content. Compositional layering averages about N30E 61°SE and a well developed biotite lineation, seen clearly on the foliation surface facing the road, plunges an average of 36°N34E. The dominant joint set is N49W 68°SW, with subordinate sets of N78E 70°NW, N25W 72°NE, and N48E 31°NW.
		Walking north we see exposures of Leithsville dolomite on both sides of the road. Along the western edge of the golf course Middle

CUMULATIVE MILEAGE	STOP MILEAGE	
		Proterozoic rocks are unconformably overlain by basal Hardyston Quartzite. This part of the Hardyston consists of poorly-sorted, dominantly arkosic conglomerate and lesser quartz-pebble conglomerate. These rocks grade upward into quartzose sandstone that is transitional with the lower Leithsville. In all, we estimate the thickness of the Hardyston here to be about 15 feet. For discussion of the origin and petrology of the Hardyston the reader is referred to Aaron (1969).
		Bedding in the Hardyston and Leithsville here averages about N47E 54°NW. This clearly demonstrates the angular nature of the unconformity because Proterozoic foliation trends about N25E.
		Reload cars and continue north on Wildcat Road.
41.72		Excellent outcrops of Leithsville Formation on right.
41.82	.15	STOP 7B. SAME AS STOP 7A. NO HAMMERS PLEASE (30 Minutes). This is a continuation of stop 7A. Here we have an actual hands-on contact of the unconformity seen at the previous stop. Notice the southeast dipping foliation in the quartz-

CUMULATIVE MILEAGE	STOP MILEAGE	
		oligoclase gneiss and the overlying northwest dipping basal Hardyston conglomerate. The Proterozoic rock clearly is not very deeply weathered and, in fact, appears moderately fresh. To our immediate north, a cut along the abandoned railroad tracks, on the eastern side of Wildcat road, exposes about 160 feet of Leithsville. As one walks down section (eastward) the transition to the sandier facies is seen, however, this exposure ends just short of the contact with the underlying Hardyston.
		Reload cars continue north on Wildcat Road.
41.87		Outcrops of Leithsville Formation on right
42.02		Outcrops of Leithsville Formation on right
42.12		Excellent section of Leithsville Formation in railroad cut on right.
42.17		Stop sign. Turn right on Church Street (Route 631 East).
42.62		Franklin Pond on left
42.72		Outcrops of Franklin Marble on left
44.85		Crossing trace of Zero fault (Hague and others, 1956).
44.95		Outcrops of Leithsville Formation on left

CUMULATIVE MILEAGE	STOP MILEAGE	
45.15		Traffic light. Turn left on Route 23 North.
45.45		Outcrops of Allentown Dolomite on left.
45.7		Crossing trace of Rutherford cross fault (Hague and others, 1956). Outcrops of Franklin Marble to left of road and of Allentown Dolomite to right of road.
46.4		Entering Hamburg 7 1/2-minute quadrangle.
46.75		Outcrops of Franklin Marble on right
48.15		Outcrops of Allentown Dolomite on right
48.55		Traffic light. Turn right on Route 94.
49.7		Glacial delta complex on right
50.2		Outcrops of microcline gneiss on right
50.45		Outcrops of Franklin Marble on left.
51.15		Outcrops of Franklin Marble on both sides of road.
51.25		Marble quarry and entrance to Great Gorge Resort on right.
51.85		Prominant ridge on right is Hamburg Mountain.
53.35		Turn left on Drew Mountain Road which runs along a saddle in Pochuck Mountain.
56.65		Stop sign. Turn right on Route 565 North toward Glenwood. Outcrop of amphibolite on left near barn.

CUMULATIVE MILEAGE	STOP MILEAGE	
58.35	16.53	STOP 8. "PILLOW STRUCTURES" IN AMPHIBOLITE ON POCHUCK MOUNTAIN (30 Minutes). This exposure, on the west limb of the Glenwood syncline of Hague and others (1956), (fig. 13) is one of several localities identified by N. J. Zinc Company geologists as containing structures interpreted to be relict pillows. These structures occur within a unit which they termed type I hornblende gneiss (amphibolite). According to Hague and others (1956), type I amphibolite typically is a black-to greenish-black, fine-to medium-grained rock, locally containing small white plagioclase augen, and which in places grades into biotite gneiss. Veins and masses of pyroxene-scapolite rock occur adjacent to the "pillows" and calcite occurs sparsely, weathering out of the "pillows" to form irregular cavities. Hague and others (1956) interpreted type I amphibolite to be a possible metavolcanic rock which formed either by the "metamorphism of two different types of rocks, namely volcanic flows and argillaceous sediments of similar bulk composition, which formed indistinguishable end products" or the "interfingering of volcanic flows and calcareous sediments."

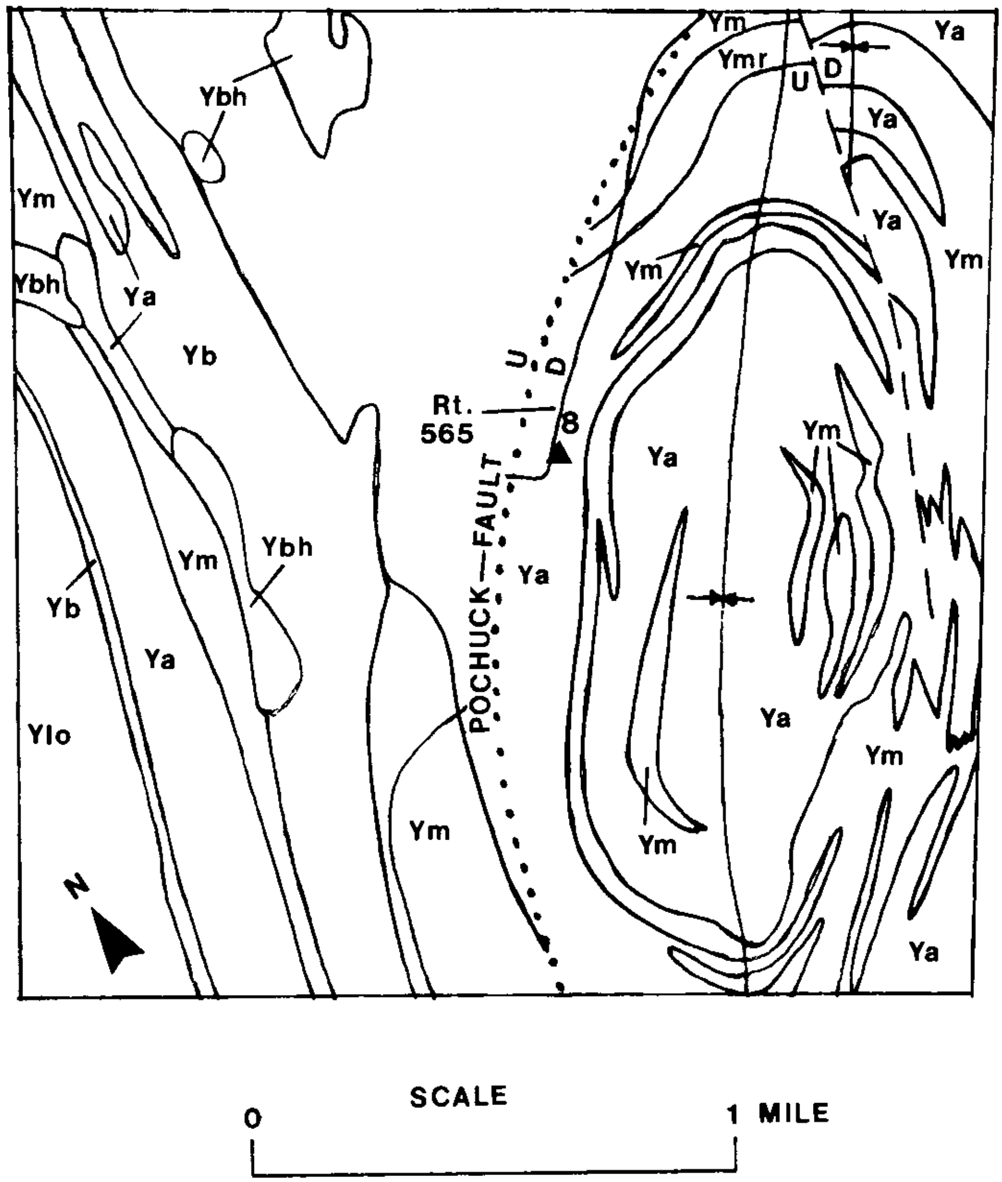


Figure 13. Geologic map of the Glenwood area, Hamburg $7\frac{1}{2}$-minute quadrangle (modified from Hague and others (1956)). Map symbols are the same as in Figure 3.

CUMULATIVE MILEAGE	STOP MILEAGE

At this locality, the rock is a dark gray- to grayish-black, medium-fine- to medium-grained, moderately layered and foliated amphibolite. The essential minerals are hornblende and plagioclase, with minor local clinopyroxene and biotite. Parts of this unit are quite mafic; elsewhere it contains alternatingly light and dark layers because of an increase in plagioclase content. Small plagioclase augen are well developed here imparting a sheared, ductile-appearing fabric to the rock. Curious flattened, elliptical cavities occur ranging in size from about 2 cm to 300 cm along a subhorizontal axis. These cavities, or vugs, contain some secondary mineralization and superficially resemble the vugs between pillows in the Orange Mountain Basalt.

Maxey (1971) chemically analyzed 56 samples of amphibolite from various locations within the New Jersey Highlands. He concluded that 47 were similar in composition to tholeiitic basalts and the remainder similar to alkali basalts. This makes the idea of possible relict pillow structures more plausable.

CUMULATIVE MILEAGE	STOP MILEAGE

Major and trace element chemistry was obtained on a sample of amphibolite from this locality. While a conclusive argument cannot be made on the basis of a single sample, the similarity between this analysis, the analyses reported by Maxey (1971), and those in the literature on oceanic tholeiites is striking (Table 3). Trace element plots on discimination diagrams of Zr versus Ti, Cr versus Ti, Zr-Ti/100-Sr/2, and SiO_2 versus total alkalis (Fig.14) are virtually identical to modern olivine tholeiites from a ridge/ocean floor setting. Rogers and others (1984) described some diagnostic properties of basalts from different environments. Properties of mid-ocean ridge and small ocean basin basalts are compared with this type I amphibolite sample in table 4. Except for the more mobile elements like Ba, Sr, and Rb, which often show enrichment due to metamorphism and weathering, there is remarkably good agreement between this sample and typical oceanic ridge/floor basalts. It must be remembered, however, that the discrimination criteria cited above are hardly infallible.

TABLE 3. CHEMICAL ANALYSES OF SOME NEW JERSEY HIGHLANDS AMPHIBOLITES AND COMPARATIVE ANALYSES OF OCEANIC THOLEIITES

	1	2	3	4
SiO_2	48.4	50.14	49.21	49.56
Al_2O_3	15.8	15.43	15.81	16.09
FeO*	10.03	11.54	9.40	11.29
MgO	8.11	6.36	8.53	7.69
CaO	11.0	9.35	11.14	11.34
Na_2O	3.12	2.14	2.71	2.80
K_2O	0.29	1.27	0.26	0.24
TiO_2	1.53	1.72	1.39	1.42
MnO	0.17	0.21	-----	-----
Total	98.45	98.16	98.45	100.43
		(parts per million)		
Cr	231	112.6	296	-----
Ni	91.3	46.78	123	-----
Cu	22.3	39.34	87	-----
Co	-----	25.39	41	-----
P	27	------	---	-----
Rb	10	------	---	-----
Ba	195	------	12	-----
Sr	487	------	123	-----
V	169	------	289	-----
Zr	85.1	------	100	-----
Sc	26.5	------	---	-----

1. Stop 8, this trip, M. Kaeding, analyst.

2. Average of 56 analyses, from Maxey (1976).

3. Average of 33 basalts from the mid-Atlantic ridge, from Melson and Thompson (1971).

4. Average of 75 ocean-floor basalts, from Pearce (1976).

*FeO as total iron.

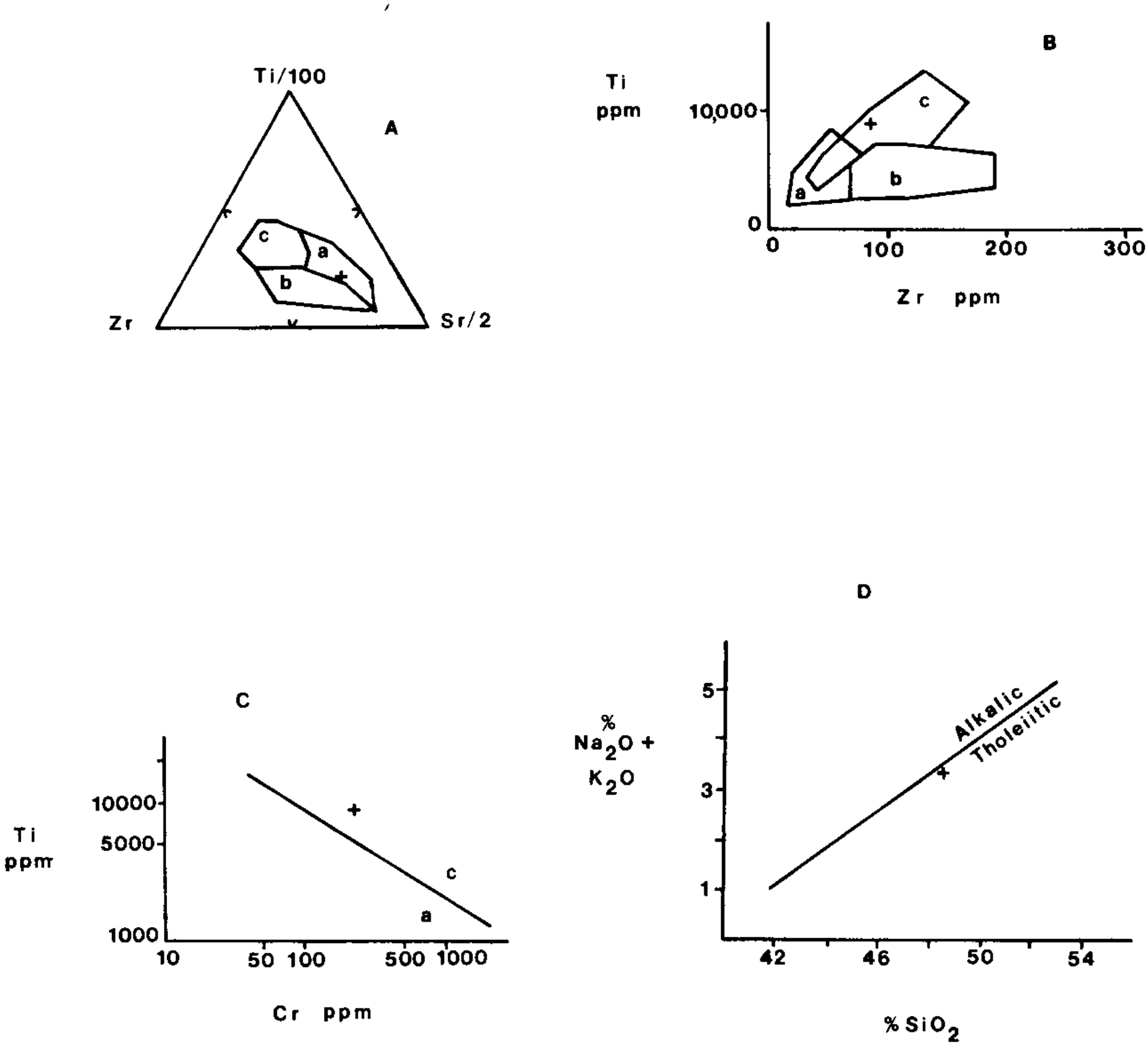


Figure 14. "Pillow" amphibolite sample from stop 8 plotted on tectonic discrimination diagrams. A) Ti-Zr-Sr relationship on plot of Pearce and Cann (1973). B) Ti-Zr relationship on plot of Pearce and Cann (1973). C) Ti-Cr relationship on plot of Pearce (1975). D) SiO_2-total alkalis on plot of Macdonald and Katsura (1964). Low-potassium tholeiites plot in fields a, calc alkali basalts in fields b, and ocean floor basalts in fields c.

TABLE 4. PROPERTIES OF BASALTS FROM MID-OCEAN RIDGES AND SMALL OCEAN BASINS COMPARED WITH AMPHIBOLITE FROM STOP 8

	1	2
SiO_2 range	46 - 52%	48.4%
K_2O	0.1 - 0.2%	0.29%
Rb	2 ppm	10 ppm
Ba	15 ppm	195 ppm
Sr	100 ppm	487 ppm
TiO_2	1.5%	1.5%
Zr	100 ppm	85.1 ppm
Cr	300 ppm	231 ppm
Ni	100 ppm	91 ppm
FeO*/MgO	<1.5	1.24

1. Values listed are from Rogers and others (1984).

2. Amphibolite from stop 8, this trip.

*FeO as total iron.

CUMULATIVE MILEAGE	STOP MILEAGE	
		More systematic sampling for chemistry here, as well as in adjacent quadrangles, is necessary for comparison with amphibolites from elsewhere in the Highlands. It seems possible that at least some New Jersey Highlands amphibolites are metabasalt and, as suggested by Hague and others (1956) and Drake (1984), some amphibolites have probably evolved from a sedimentary protolith. Reload cars and continue north on Route 565.
59.55		Vernon Township High School. Cars enter parking lot and turn around. Return to Route 565 and turn left heading south.
63.05		Turn left on Route 641, Drew Mountain Road.
64.3		Turn right onto Route 517 south.
66.8		Traffic light. Continue straight on Route 94 south.
69.0		Traffic light. Continue south on Route 94.
69.9		Traffic light. Continue south on Route 94. Outcrops of Allentown Dolomite on right.
71.6		Bear to right and continue south on Route 94.
75.75		Outcrops of Jacksonburg Limestone (Middle Ordovician) on right.
75.7		Traffic light. Turn left on Route 15 south.

CUMULATIVE MILEAGE	STOP MILEAGE	
77.15		Outcrops of Bushkill Member of Martinsburg Formation (Middle Ordovician) on left.
77.65		Outcrops of Allentown Dolomite on left.
79.3		Traffic light. Continue south on Route 15. For a description of outcrops of Middle Proterozoic rocks see Puffer (1980) and Volkert (1984).
80.45		Crossing trace of Zero fault.
80.55		Outcrops of Allentown Dolomite that contain one substantial and abundant minor back thrust faults.
81.70		Crossing trace of Morris Lake East fault (Hague and others, 1956).
88.95		Entering Green Pond synclinorium
89.65		Outcrops of Green Pond Conglomerate (Silurian) on left. Outcrop displays strong evidence of brittle deformation.
91.95		Turn right onto I-80 west
95.65		Big outcrops of charnockitic quartz diorite on right.
97.35		Outcrops of pyroxene syenite on right contain a Mesozoic diabase dike.
97.75		Take Exit 27A North toward Netcong.
98.45		Traffic circle. Bear right and go east on Route 46.
100.25		Turn left into Quality Inn parking lot.

END OF FIELD TRIP

Editors'note: Complete references for Road Log may be found at the end of the manuscripts in this volume by Volkert and Drake and Hull et al.

GEOLOGIC ASPECTS OF RADON OCCURRENCE IN NEW JERSEY

BELL, Christy, New Jersey Geological Survey, CN-029, Trenton, NJ 08625

Radon 222, the isotope of radon which presents the greatest hazard when allowed to accumulate in an enclosed area, is a radionuclide in the decay chain of uranium 238. An important factor in understanding the radon hazard is the distribution and chemistry of uranium and its daughters in rock, soil and ground water.

Many occurrences of uranium in the rocks of New Jersey have been known since the first efforts of the Atomic Energy Commission to locate uranium deposits in the United States (U. S. Atomic Energy Commission and U. S. Geological Survey, 1969). Geologic investigations over the last 30 years have shown elevated levels of uranium in a number of rock units.

In the Reading Prong, uranium-bearing minerals which could act as radon sources are locally concentrated in Precambrian igneous and meta-sedimentary rock units. High concentrations are rarer or non-existent in Precambrian meta-volcanic units and Paleozoic quartzite and limestone. Paleozoic dolomite, sandstone and black shale may display moderate local enrichment or low, broad anomalies. Because the high concentrations in Precambrian rock are local and in upland areas, high radon levels are expected to be local and most common on uplands. However, extremely high levels of radon have been measured under unusual circumstances in areas of Paleozoic dolomite.

The Valley and Ridge province is underlain in New Jersey by Paleozoic sedimentary units equivalent in age and lithology to those in the Reading Prong. Cambrian sandstone, Cambrian and Ordovician dolomite and Ordovician black shale, present the greatest possibility of being problematic radon source rocks. Red shale, quartzite, and limestone are not expected to present radon problems. Because the uranium concentration of the dolomite, sandstone and black shale do not reach the high levels found in the Precambrian units of the Reading Prong the radon hazard is expected to be less.

In the Newark Basin, the lower part of the Brunswick (Passaic) Formation and the Lockatong and the Stockton Formations are known to be locally enriched in uranium (Turner-Peterson, 1980; Szabo and Zapecza, personal communication; Turner-Peterson and others, 1985). Turner-Peterson (1980) reports thin beds in black mudstone of the Lockatong Formation which contain as much as 0.01 - 0.02 percent U_3O_8.

In the Coastal Plain province glauconitic units may contain elevated levels of uranium in areas where the units are rich in phosphate but this has not been demonstrated at specific locations. At this time there is no evidence that any other units in the province contain any significant levels of uranium or radium.

Soil also may be enriched in uranium and radium and act as a source for radon. In addition, soil can play a passive role in radon dispersal by acting as a conduit for radon emitted from underlying rock. Factors which affect a soil's radon-bearing and dispersal capabilities are permeability, moisture content, parent material, and location of the radionuclides on the soil particles. A number of reports discuss the migration of radon in the ground (Tanner, 1980; Soonawala and Telford, 1980; Pearson, 1965).

Ground water is a third major factor in the distribution of naturally occurring radon. Water percolating through an aquifer can dissolve uranium and radium. Dissolution and transport depend on the redox state and pH of the water. Radon can become enriched in ground water through the decay of dissolved radium or by ejection from the surrounding grains during decay of radium close to a grain surface (Tanner, 1980).

New Jersey has initiated a "Statewide Scientific Study of Radon" to better understand the distribution of indoor radon in the state. In its initial stage, the geologic aspect of the study will be to compile existing geologic, soil, and ground water data. The data collected will include aerial radiometric and geochemical data collected from 1974 to 1980 during the National Uranium Resource Evaluation Program of the U. S. Department of Energy.

REFERENCES

Pearson, John E., 1965, Radon-222: A study of its emanation from soil, source strength, and use as a tracer, final report: Research Reports, Department of General Engineering, University of Illinois, Urbana, 73 p.

Soonawala, N. M., and Telford, W. M., 1980, Movement of radon in overburden: Geophysics, v. 45, p. 1297-1315.

Tanner, Allan B., 1980, Radon migration in the ground: A supplementary review, in Gesell, Thomas F., and Lowder, Wayne M., eds., Natural radiation environment III, symposium proceedings: U. S. DOE Rept. CONF-780422, NTIS, v. 1, p. 5-56.

Turner-Peterson, Christine E., 1980, Sedimentology and uranium mineralization in the Triassic-Jurassic Newark Basin, Pennsylvania and New Jersey, in Turner-Peterson, Christine E., ed., Uranium in sedimentary rocks: Application of the facies concept to exploration: S.E.P.M Short Course Notes, p. 149-175.

Turner-Peterson, Christine E., Olsen, P. E., and Nuccio, Vito F., 1985, Modes of uranium occurrence in black mudstones in the Newark Basin, New Jersey and Pennsylvania, in Robinson, Gilpin R., Jr., and Froelich, Albert, eds., Proceedings of the second U. S. Geological Survey workshop on the early Mesozoic basins of the eastern United States: U. S. Geological Survey Circular 946, p. 120-124.

U. S. Atomic Energy Commission and U. S. Geological Survey, 1969, Preliminary reconnaissance for uranium in Connecticut, Maine, Massachusetts, New Jersey, New York, and Vermont, 1950 to 1959: RME-4106, TID UC-51.

Workshop For High School Earth Science Teachers: Metamorphism and Metamorphic Rocks

By

Fredric Goldstein
Department of Geosciences
Trenton State College
Trenton, NJ 08625

Raymond Talkington
Geology Department-NAMS
Stockton State College
Pomona, NJ 08240

Jonathan Husch
Department of Geosciences
Rider College
Lawrenceville, NJ 08648

October 25, 1986

Geological Association of New Jersey

PART I

METAMORPHIC ROCKS

Metamorphic rocks are formed below the Earth's surface by the solid state transformation (metamorphism) of pre-existing rocks by heat, pressure, chemical fluids or a combination of these factors.

There are two classes of metamorphic rocks: *foliates* and *nonfoliates.*

Foliated metamorphic rocks are characterized by a lineation of its included minerals. These rocks are classified on the basis of their textures and compositions.

NAME	TEXTURE AND COMPOSITION	PARENT ROCKS
Slate	Fine grained clay minerals	Shale
Phyllite	Medium grained micaceous minerals, small garnets	Slate
Schist	Coarse grained micaceous minerals, quartz, feldspars and large garnets	Phyllite and fine grained igneous rocks including rhyolite, trachyte, andesite and basalt
Gneiss	Coarse grained, quartz, feldspars, micas, amphibole and pyroxene	Coarse grained igneous rocks including granite, syenite, diorite and gabbro, impure conglomerate and breccia
Amphibolite	Medium grained, amphibole and plagioclase feldspar	Basalt and diabase
Serpentinite	Coarse grained, olivine, Ca-plagioclase and talc	Gabbro

Nonfoliated (massive) metamorphic rocks are characterized by their homogenous appearance. These rocks are classified on the basis of their composition.

NAME	COMPOSITION	PARENT ROCKS
Marble	Calcite ($CaCO_3$)	Limestone
Quartzite	Quartz (SiO_2)	Sandstone, siltstone and conglomerate

NAME	COMPOSITION	PARENT ROCKS
Anthracite Coal	Carbon (95%)	Bituminous coal
Graphite	Carbon (100%)	Anthracite coal
Soapstone	Talc (90+%)	Serpentine
Hornfels	Clay minerals with scattered garnets	Shale

DESCRIPTIONS OF METAMORPHIC ROCKS

Amphibolite	Nonfoliated, medium grained, amphibole and plagioclase feldspar.
Anthracite Coal	Nonfoliated, dark black, shiny luster, conchoidal fracture.
Gneiss	Foliated, coarse grained, parallel layers of quartz, feldspar, micas, amphibole and pyroxene give a distinctly banded appearance.
Graphite	Nonfoliated, black to steel gray, metallic luster, hardness 1-2, marks paper.
*Hornfels	Nonfoliated, usually light to dark gray, may contain scattered garnet or andalusite crystals.
Marble	Nonfoliated, light to dark gray, coarse calcite and dolomite crystals predominate, hardness 3, effervesces when acid is added.
Phyllite	Foliated, light to dark gray, hazy micaceous luster caused by barely visible micaceous minerals, small garnets may be present.
Quartzite	Nonfoliated, light to dark gray or red, densely packed fused quartz grains, hardness 7.
Schist	Foliated, light to dark gray, coarse grained micaceous minerals, quartz, feldspars and large garnets are common.

DESCRIPTIONS OF METAMORPHIC ROCKS

Serpentine	Foliated, shades of green with gabbroic minerals dominating the assemblage.
Slate	Foliated, usually gray, green or red; fine grained clay minerals, "slaty cleavage", high pitch when tapped.
Soapstone	Nonfoliated, shades of green, rich in talc.

*Hornfels is generally formed as a result of the contact metamorphism of shale.

USES OF METAMORPHIC ROCKS

Anthracite Coal	Highest rank of coal
Gneiss	Building and monument stones
Graphite	Lead pencils, lubricant, refractory crucibles in the brass, bronze and steel industries
Marble	Interior decoration, statutes, building and monument stones
Slate	Roofing materials
Soapstone	Sculpting materials and acid resistant surfaces for laboratory tables

METAMORPHIC ROCK IDENTIFICATION CHART

No.	COLOR	COMPOSITION	CLASS	PARENT ROCK	NAME	USE
1.						
2.						
3.						
4.						
5.						
6.						
7.						
8.						
9.						
10.						
11.						
12.						

PART II

METAMORPHISM

Introduction

Metamorphic rocks are simply vestiges of pre-existing igneous, sedimentary or metamorphic rocks which have been transformed into 'new' crystalline rocks. The important task for the geologist, aside from determining the type of metamorphic rock, is to determine how the metamorphic transformation took place. In order to grasp how the pre-existing solid rock changed, it is important to define and answer the following questions: 1. What is the bulk chemical composition of the rock?; 2. Have any relict fabrics survived the metamorphic event?; 3. What is the nature of the path of metamorphism(i.e., changes in temperature, pressure, fluid composition, etc.)?; and 4. Has equilibrium been attained during metamorphism. Simply stated, what did the rock start out as; how did it get metamorphosed; and has equilibrium been achieved.

The principal objectives of this exercise are three-fold. 1. to briefly examine the types of metamorphism; 2. to introduce the concepts of metamorphic grade and metamorphic facies; and 3. in an idealized fashion show plate tectonic-metamorphic rock associations.

What is Metamorphism and Types of Metamorphism

What is Metamorphism?

Metamorphism can be defined as follows. "Metamorphism is the process of mineralogical and structural changes of rocks in their solid state in response to physical and chemical conditions which differ from the conditions prevailing during the formation of the rocks; however, the changes occurring within the domains of weathering and diagenesis are commonly excluded"(Winkler, 1979).

From this definition it is apparent that metamorphism involves mineralogical and structural changes in rocks and minerals. These transformations take place in the solid state. That is to say that metamorphic reactions occur outside the realms of magmatic(=igneous) and sedimentary processes. Hence, in the strict sense, at the instant the first drop of magma is produced from the melting of a rock, the process is igneous. On the other hand, the boundary between sedimentary and metamorphic processes is difficult to accurately define and is dependent upon the bulk chemical composition and fluid content of the parent rock. The best(?) accepted evidence that marks the beginning of metamorphism is the first appearance of a metamorphic mineral which is unknown in sedimentary rocks. These minerals are laumontite, lawsonite, glaucophane, paragonite, or pyrophyllite(Winkler, 1979). As there is no unique temperature or pressure which defines the boundary between diagenesis and metamorphism, conditions of 200°C and 1 kilobar are not an unreasonable starting point(Figure 1).

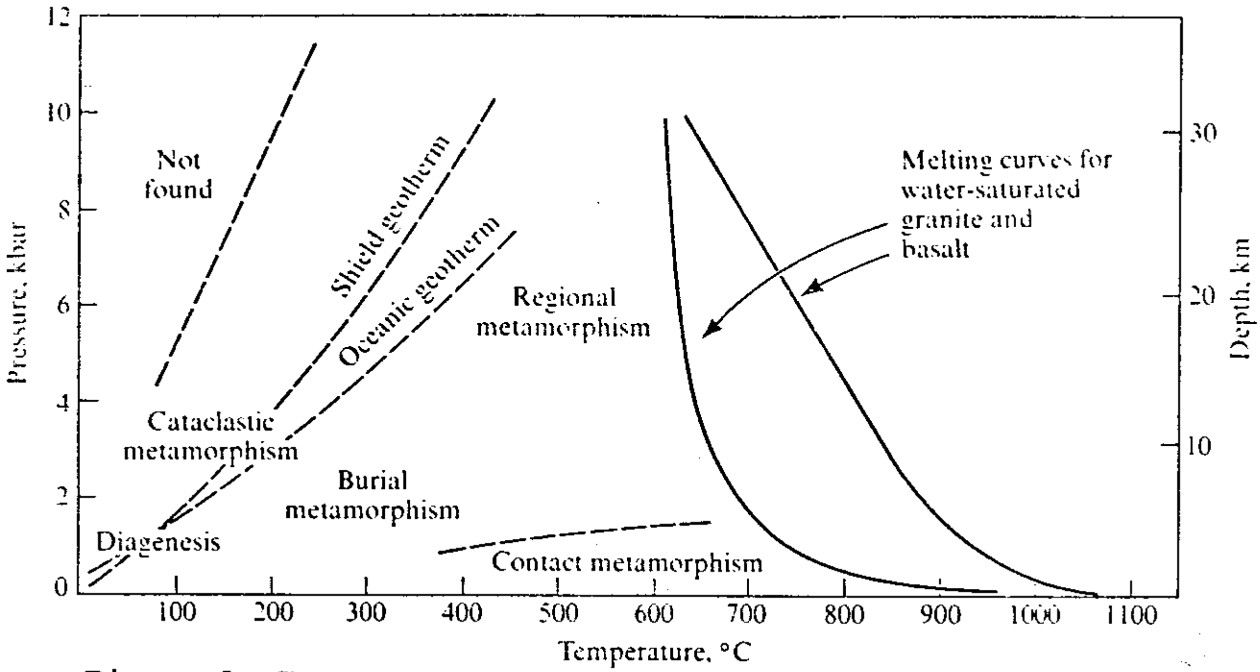


Figure 1. Temperature and pressure ranges for the various types of metamorphism(from Ehlers and Blatt, 1982; Fig. 17.2).

Types of Metamorphism

There are six types of metamorphism(refer to Figure 1): 1. contact or thermal; 2. regional; 3. sub-sea-floor; 4. burial; and 5. shear, dynamic, or cataclastic; and 6. shock or impact.

1. Contact metamorphism occurs when magma or hot igneous rocks come in contact with surrounding igneous, sedimentary, or metamorphic wall rocks. The effects are principally thermal and locally compositional('ionic' transport by hydrothermal fluids), but with very little change in pressure.

2. Regional metamorphism is the most widespread of any type of metamorphism and begins as soon as pre-existing rocks are buried to a sufficient depth at variable temperature to cause deformation and recrystallization. This type of metamorphism is closely associated with plate tectonic activity, especially convergent lithospheric plate boundaries(= subduction zones).

3. Sub-sea-floor metamorphism occurs along divergent plate boundaries where hot, newly-formed oceanic crust interacts with 'cold' ocean water. Principally, this is a form of contact metamorphism with associated metasomatism(= hydrothermal alteration).

4. Burial metamorphism occurs in thick sequences of volcanic and sedimentary rock due to changes in pressure and temperature unrelated to plate tectonic or magmatic activity(Best, 1982). This is synonymous with high-grade diagenesis.

5. Shear, dynamic, or cataclastic metamorphism results from the fracturing and granulation of rocks in faults. The general effect is a decrease of grain size. Temperature influence is minimal; whereas the pressure effect is variable.

6. Shock or impact metamorphism occurs in rocks subjected to meteorite impact. The passage of shock waves through the rock produces extremely high temperatures and pressures for a few microseconds(Mason, 1981).

Metamorphic Grade and Metamorphic Facies (excerpted from Metamorphic Rock Lab; Hozik, Parrott, & Talkington, 1984)

Metamorphic Grade

The degree of metamorphism is referred to as the grade of metamorphism. The higher the grade, the higher the temperature and pressure to which the rock has been subjected. The determination of metamorphic grade is not a simple process because of the variability in chemical composition of the rock, temperature, and pressure conditions, and in some instances, the presence of chemically active fluids(= hydrothermal fluids).

One consequence of changing conditions is a change in the texture of the rock. For example, as metamorphic grade increases, shale is successively modified to: slate --> phyllite --> schist --> gneiss.

Another measure of increasing metamorphic grade is the first appearance of certain minerals sensitive to changing temperature and pressure conditions. These minerals called index minerals, only occur in rocks relatively rich in aluminum and silicon. Rocks of that composition, however, are sufficiently widespread to make this a useful tool. The index minerals, in order of increasing metamorphic grade, are: chlorite --> biotite --> almandine garnet --> staurolite --> kyanite --> sillimanite(see Figure 2).

In mapping rocks on the surface of the earth, a line may be drawn which marks the first appearance of one of these index minerals going up metamorphic grade. This line is known as an isograd. The area over which a characteristic mineral is dominant is called its metamorphic zone. Therefore, the upgrade edge of the chlorite zone is marked by the biotite isograd, beyond which lies the biotite zone(Figure 3).

Metamorphic Facies

In order to determine metamorphic grade in rocks of various compositions, a more general system is necessary. The concept of metamorphic facies has been developed to refer to groups of minerals which occur together, known as assemblages. Each assemblage occurs over a defined range of temperature and pressure conditions. Figure 4 shows the commonly used metamorphic facies and their respective temperature and pressure ranges. Because facies are based on mineral assemblages, only some of the characteristic minerals need to be present to permit the assignment of the rocks to the correct facies.

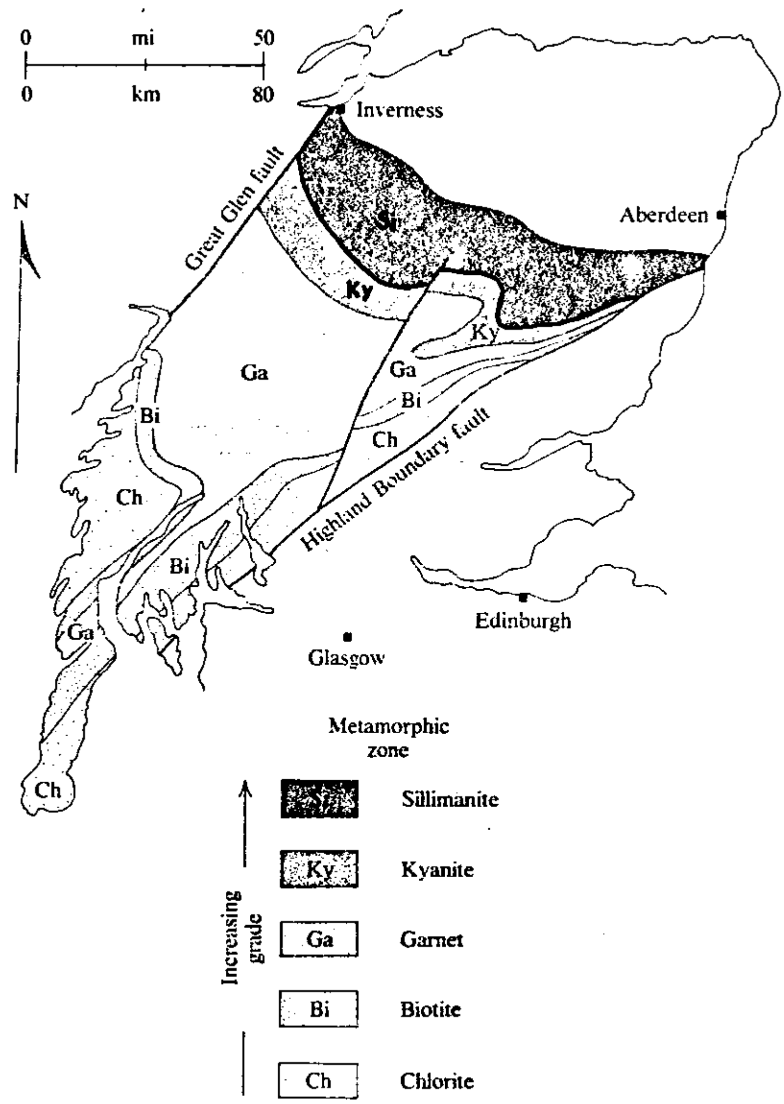


Figure 2. Distribution and relationships of metamorphic index minerals and metamorphic grade(from Best, 1982; Fig. 10.25).

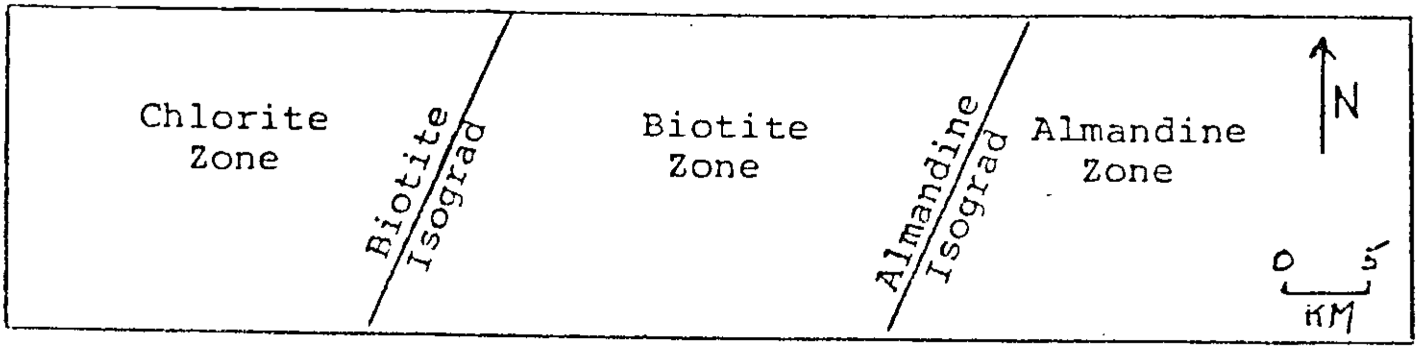


Figure 3. Relationship of metamorphic isograd to metamorphic zone(from Hozik et al., 1984; Fig. V-1).

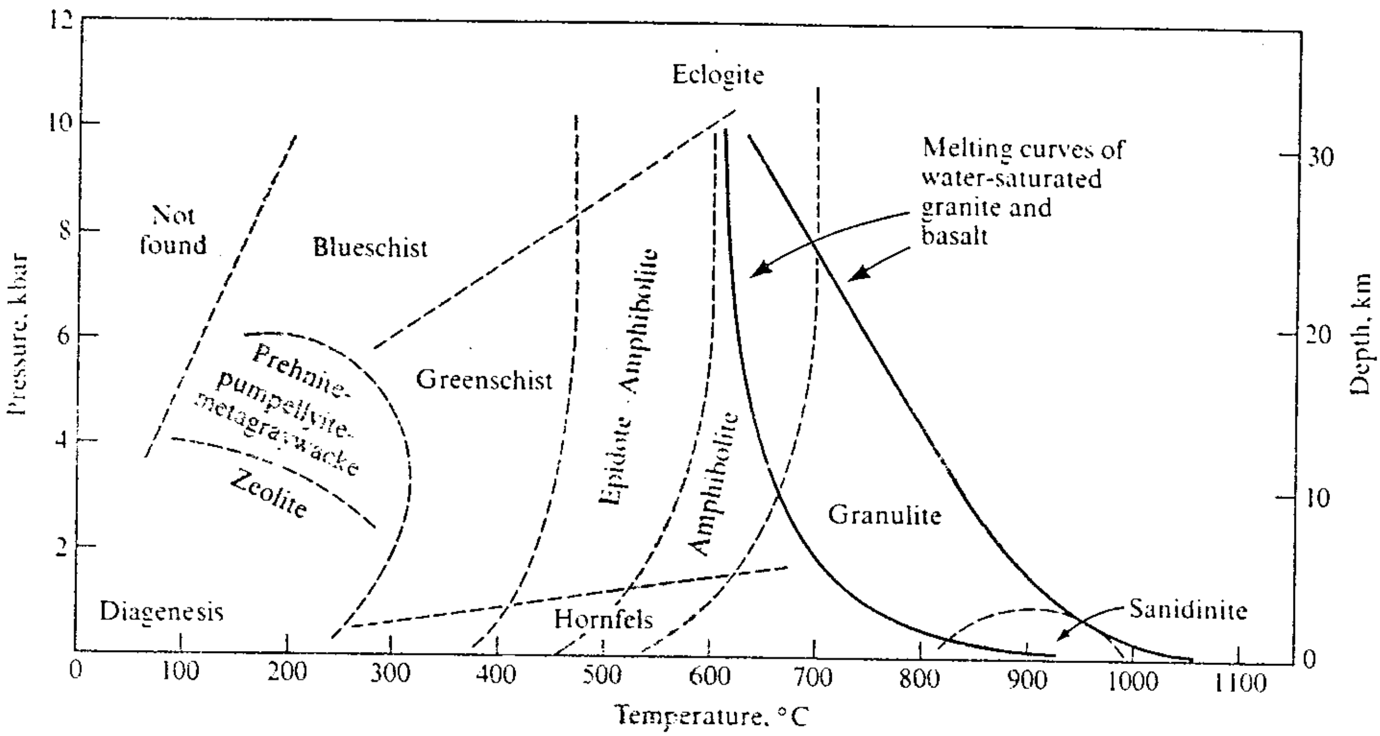


Figure 4. Metamorphic facies diagram(from Ehlers and Blatt, 1982; Fig. 18-1).

All of the measures of metamorphic grade and facies discussed above are functions of bulk chemical composition of the rock, temperature, and pressure, and therefore interrelated. For example, the lowest grade index mineral, chlorite, is a green, platy mineral occurring in slates, phyllites, and schists of the greenschist facies.

Metamorphism and Plate Tectonics

One of principal goals for any study of igneous, sedimentary, or metamorphic rocks nowadays is to relate their formation to a regional, plate tectonic(= orogenic) event. We have attempted to introduce this during the above discussion, but additional explanation is necessary. Figure 5 is a cartoon showing the general relationships between the type of metamorphism and plate tectonics. It is apparent that regional metamorphism is nearly exclusively associated with convergent plate boundaries; whereas contact metamorphism occurs at this type of boundary as well as divergent and intracontinental. Hence, types of metamorphism are not mutually exclusive of one another.

We have only briefly explained metamorphism, its usefullness, and applications. For an in depth treatment of metamorphism, the reader is referred to the text books listed in the references.

Exercises

1. Examine Figure 6, Metamorphic Mineral Localities of the Curved Mountain Quadrangle, West Dakota. On this figure, 43 outcrops are shown by symbols of the metamorphic index minerals found at each. NOTE that 2 localities have more than one of these minerals in the same outcrop!

a. In pencil, draw the isograds which show the areas of stability of the index minerals.

b. Label the isograds.

c. Label the zones.

2. a. Using Figure 7, determine the minimum temperature and minimum pressure for the metamorphism of the rocks in the outcrop located 12 km north of Benjamin City. (HINT: Where, on Figure 7, can the two minerals from the outcrop coexist?).

____

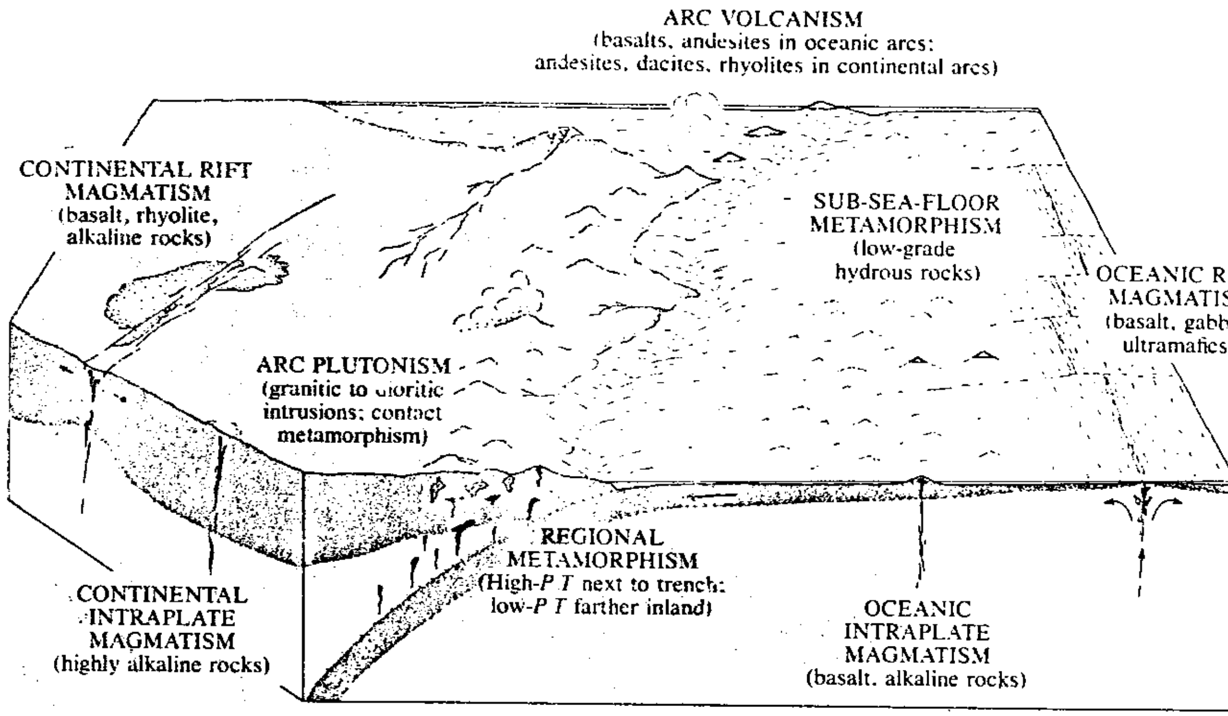


Figure 5. Generalized relationships between plate tectonic setting and the occurrence of igneous and metamorphic rocks (from Best, 1982; Fig. 1.17).

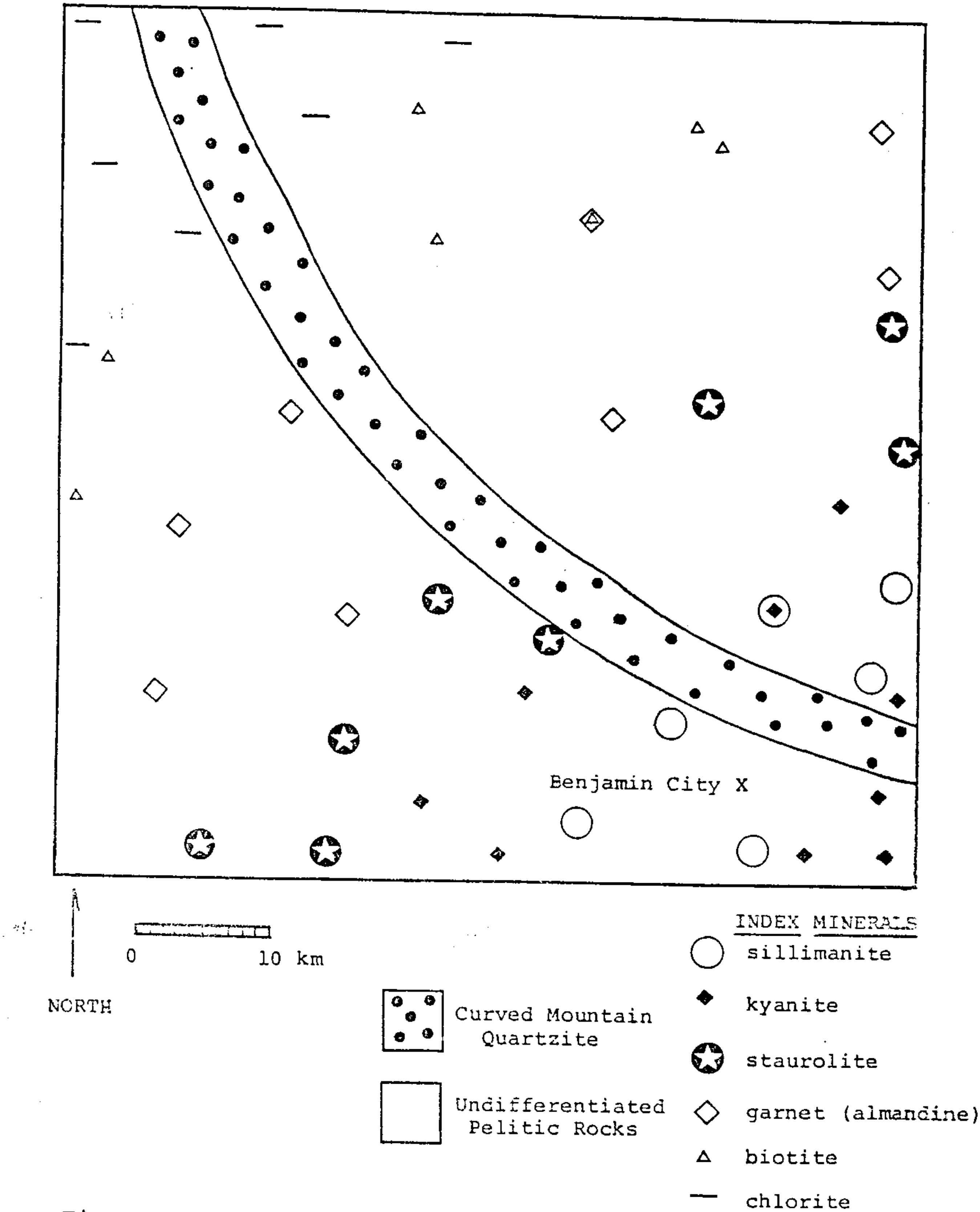


Figure 6. Metamorphic mineral localities, Curved Mountain quadrangle, West Dakota(from Hozik et al., 1984; Fig. V-3).

b. To what depth have the rocks in this outcrop been buried, in miles? (1.6 km = 1 mile)

___

3. a. The reaction kyanite --> sillimanite is possible in which of the facies shown on Figure 8? (HINT: there is more one.)

___

b. The dashed line in Figure 8 represents the average geothermal gradient, which is the average relationship between temperature and pressure in the earth's interior. Assuming the rocks from the outcrop in question # 2 have followed the average geothermal gradient and no melting has occurred, in what facies has the reaction kyanite --> sillimanite occurred?

___

References (Part II Only)

BEST, M.J., 1982. IGNEOUS and METAMORPHIC PETROLOGY. W.H. Freeman and Co., San Francisco, 630p.

EHLERS, E.G., and BLATT. H., 1982. PETROLOGY. W.H. Freeman and Co., San Francisco, 732p.

HOZIK, M.J., W. PARROTT, and R.W. TALKINGTON. 1984. Metamorphic Rock Lab. In: Exercises in Physical Geology, second edition. Stockton State College Publication, V1-V10.

MASON,R. 1981. PETROLOGY OF METAMORPHIC ROCKS. George Allen and Unwin, Ltd., London, 254p.

WINKLER, H.G.F., 1979. PETROGENESIS OF METAMORPHIC ROCKS, fifth edition. Springer-Verlag New York Inc., 348p.

PART III

THE PROTOLITHS OF METAMORPHIC ROCKS

The final question that a geologist can ask after identifying and classifying a metamorphic rock (Part I) and determining the conditions under which it was metamorphosed (Part II), is what was the nature of the rock prior to the metamorphic event(s)? This requires that either the rock's original textures or chemistry were not destroyed or altered to any great extent by the metamorphic process. This is not always the case and, therefore, the identification of a metamorphic rock's parent rock or protolith can be uncertain at best. However before throwing up our hands in complete despair, keep in mind that in many instances the textural and/or chemical characteristics of the protolith have been preserved and a rather precise identification can be made. The determination of the protolith gives a great deal of information on the conditions existing prior to metamorphism and may, in fact, provide additional clues to the causes and mechanisms of the metamorphism itself. Thus, protolith identification is an integral an important part of the study of any metamorphic rock(s).

As shown on the identification table on pages 120 and 121, many protoliths can be identified just on the basis of the mineral content and overall composition of the observed metamorphic rock. This is particularly the case for marbles, quartzites, and slates. Again this assumes that the chemistry of the rocks have not been changed. That is, the metamorphic process was isochemical. However, often protoliths of entirely different type can have similar chemistry and their resulting metamorphic equivalents may be almost identical in appearance and composition. The question then is how do we determine the protolith when it is not immediately apparent?

Rock Associations

In cases where a protolith determination cannot be made immediately, a knowledge of igneous and sedimentary rock associations and environments of formation may be useful. The recognition of a particular rock association may involve studying relict textural features and chemical compositions of different metamorphic rock types over many square miles and utilizing this information to identify their various igneous and/or sedimentary parentages. Results from such studies may allow limits to be placed on the range of protoliths that are possible for a given metamorphic rock, although a unique determination may still not be possible. For example, one would not expect a deep water marine protolith if all associated (same region and age) rocks had their origins in shallow marine and terrestrial sediments (you can remove the boy from the farm but you can't remove the farm from the boy). Common rock associations are given in Table 1.

Relict Textural Features

The identification of relict textural features in igneous and sedimentary rocks that have been metamorphosed requires that the original features be preserved. If relict textural features are recognized and identified, metamorphic rocks which can have both igneous or sedimentary protoliths, such as amphibolites and gneisses, can be placed in their correct rock association. Distinctive and useful relict sedimentary and igneous features are given in Tables 2 and 3, respectively.

In general, the preservation of relict textural features decreases with increasing metamorphic grade because of the increasing amount of recrystallization. Thus, high-grade metamorphic rocks of the amphibolite, granulite, or eclogite facies may not be able to have their protoliths determined by the recognition of relict features because of the simple fact that none have been preserved. In these cases, the rock chemistry may be most useful in determining the protolith. On the other hand, contact metamorphic rocks may have their relict features perfectly preserved and the rock association can be determined unequivocally as a result. REMEMBER, RELICT FEATURES CAN ONLY BE USED IF THEY ARE STILL OBSERVABLE IN THE METAMORPHIC ROCK ITSELF!

Rock Chemistry

When original textural features have been obliterated by the metamorphic process, protolith determinations may be made by comparing the chemical composition of the metamorphic rock with those of its posssible sedimentary or igneous protoliths. In many cases, this gives obvious results. For example, a comparison of the chemistry of a marble with that of a limestone indicates that no other common protolith is possible. The same could be said for a quartzite-sandstone comparison. In addition, a detailed analysis of rock chemistry may allow the tectonic or rock association to be determined. An excellent example of this is the identification of the tectonic environment under which the basaltic protoliths of greenstones were produced (fig. 9). Similar types of determinations also can be made for the sandstone protoliths of quartzites (fig. 10). Such studies require that the metasomatic alteration of rock chemistry was minimal. Otherwise, one is comparing apples to oranges and the results are meaningless.

The Determination of Amphibolite Protoliths:
A Real Problem

Amphibolite is a metamorphic rocks whose origins may be vey difficult to determine. In part, this is because it may have more than one possible protolith (Table 4). However, the major problems arise because original textural features often have been destroyed by the fairly high level of metamorphism associated with generation of an amphibolite and by the subtle chemical differences between a number of the possible protoliths. Fortunately, over the last two decades numerous studies have determined the chemical characterisitics needed to distinguish among the alternatives. An example of just such a study is one currently investigating the origins of late Archean amphibolites in the Ruby Range Mountains of southwestern Montana (Husch and Hennigan, 1984; Bender and Husch, 1985). We will use some of the results of this work to illustrate how an amphibolite protolith can be chemically determined.

There are a variety of chemical methods used to distinguish whether an amphibolite has a sedimentary or igneous protolith. Ratio of elements (Gorbachev, 1974), discriminant functions (Shaw and Kudo, 1965), and variation diagrams (Moine and LaRoche, 1968; Van de Kamp, 1969). None are foolproof and conclusions should not be drawn by utilizing only one method. However, when a variety of variation diagrams are used, the Ruby Range amphibolite rock compositions plot well within the basalt (or igneous) field (figs. 11 and 12). These results are in complete agreement with conclusions drawn from elemental ratios and discriminant functions.

Not only can the basaltic parentage of the Ruby Range amphibolites be determined, but the tectonic environment in which these basalts were generated also can be estimated. Figures 13 and 14 (among others) indicate that two distinct types of basaltic protolith were originally present: 1) ocean floor tholeiitic basalt associated with mid-ocean ridges; 2) calc-alkaline basalt probably produced in a volcanic arc associated with a subduction zone. Field relations and relict textural features suggest that these basalts were lava flows and/or ash deposits and were located on a shallow continental shelf. These conclusions allow us to paint a much more complete picture of the geologic conditions in existence prior to the metamorphism which transformed these basalts into amphibolites and provides some constraints on the causes of that metamorphism, even when it occurred 2.7 billion years ago!

Study Questions and Exercises

1) In what fields in Figure 9 would Ruby Range amphibolite compositions plot?

2) In what fields in Figure 10 would possible associated sandstone (now quartzites) compositions plot?

3) How would you expect an amphibolite formed from a sedimentary protolith to plot in Figures 11 and 12?

4) What rock associations (Table 1) might you find in the Ruby Range Mountains near the area where the amphibolites were collected?

TABLE 1

MAJOR ROCK ASSOCIATIONS

Ophiolites--peridotite, gabbro, basalt, chert, and limestone

Volcanic Arcs--basalt, andesite, dacite, volcaniclastic sediments, and graywackes

Batholiths--granitic plutons and regional metamorphic rocks

Subvolcanic Batholiths--calderas, ash flows and falls, and continental sediments

Continental Rifts--alkaline igneous rocks, basalts, layered intrusions, continental sediments, and evaporites

Stable Continental Margins--carbonates, sandstones, shales, and other mature sediments

Modified from Hyndman (1985)

TABLE 2

RELICT SEDIMENTARY FEATURES

Bedding--variable composition of sequential layers

Graded Bedding--grain size varies from layer to layer and within a single layer (caution-may be reversed by metamorphism)

Cross-bedding--bottom edge is tangential and top edge is truncated

Ripple marks--small repetitive undulations in bedding planes

Bedding-surface features--mud cracks, cut-and-fill channels, load and flute casts

Other features--pebbles, mud chips, sedimentary breccias, fossils, and concretions and nodules

Modified from Hyndman (1985)

TABLE 3

RELICT IGNEOUS FEATURES

Cross-cutting relationships--xenoliths, veins, and dikes

Cumulus textures--similar to various sedimentary features

Zoning--minerals with varying compositions within a single crystal

Hydrothermal activity--amygdules and filled miarolitic cavities

Lava flows--pillows, pahoehoe tops, flow breccias

Crystallization textures--phenocrysts, spherulites, glass

Modified from Hyndman (1985)

TABLE IV

Possible origin	Criteria used for determination
Metamorphism of basalt or gabbro sill, dike, or pluton	Commonly homogeneous and unlayered, but thin layering may be due to metamorphic differentiation. May show contacts discordant to layering in country rocks; may show relict igneous textures, cumulus layers, zoning in plagioclase phenocrysts, subophitic or ophitic textures; may contain relict minerals such as augite, hypersthene, or olivine; hornblende and plagioclase are subequal in abundance, + almandine + epidote + sphene + apatite + ilmenite and other opaques; + minor biotite, quartz, or diopside. Chemical trend, such as the MgO/(MgO + FeO) ratio in samples across the body, follows that for differentiation of diabase or gabbro; Cr, Ni may be high (250 ppm), as may Ti, Cu; Rb/Sr ratio is low (0.3–0.33). La/Ce ratio is low (< 0.4); high (+) ϵ_{Nd}
Metamorphism of basalt flow (see Fig. 11-14)	May show discordant lower contact unconformity; may show relict volcanic flow textures, zoning in plagioclase, phenocrysts, amygdules, monolithologic breccia, pillow structures; hornblende and plagioclase in subequal abundance, + almandine + epidote + sphene + apatite + opaques. Most commonly chemically equivalent to tholeiitic basalt but may be calc-alkaline; low oxidation ratio [$(2Fe_2O_3 \times 100)/(Fe_2O_3 + FeO)$] of about 30; Cr, Ni, high. High (+) ϵ_{Nd}
Metamorphism of basalt tuff or tuff mixed with carbonate	Constant composition, persistence along strike; layered, relict lapilli or agglomerate textures; tuffs of basalt composition forming thin persistent layers rare in unmetamorphosed sedimentary sequences. High oxidation ratio [$(2Fe_2O_3 \times 100)/(Fe_2O_3 + FeO)$ averages 68]; chemistry does not follow pelitic rock-carbonate trend. Intermediate ± ϵ_{Nd}
Metamorphism of shaly limestone or calcareous shale	Layered but not monotonously the same for tens of kilometers along strike and in successive beds for hundreds of meters across strike. Layering can probably also develop on a regional scale by metamorphic differentiation (Walker and others, 1960, p. 155, however, disagree); concordantly interlayered on a centimeter scale with marble, pelitic schist, and other clearly metasedimentary layers; widely varying mineral percentages; hornblende generally more abundant than plagioclase; biotite, quartz, diopside, or epidote may be abundant; sphene and apatite may be present; almandine is generally absent; the chemical trend of different samples lies at a large angle to the differentiation trend of basalt on CaO: MgO/MgO + FeO or on Al + alkalis; Mg/(Mg + Fe): CaO; K > Na; lower in Ni, Cr, Ti, Sc, Cu; negative correlation of Cr, Ni with Mg/(Mg + Fe): positive correlation of Ti with Mg/(Mg + Fe); higher in Ba, Pb, Au. Low (−) ϵ_{Nd}
Metasomatic replacement of marble	Transition to parent carbonate rock from a rock of suitable composition; may have relict carbonate; marble parent makes sense in the parent stratigraphic sequence
Metasomatic replacement of pyroxene granulite (metasediment)	Pyroxene granulite adjacent to metagabbro; Sr 1000–1450 ppm in gabbroic plagioclase of composition An_{56-67}; Sr 250–900, with one 1100, ppm in metasomatic amphibolite plagioclase of composition An_{40-82}

(After Hyndman, 1985)

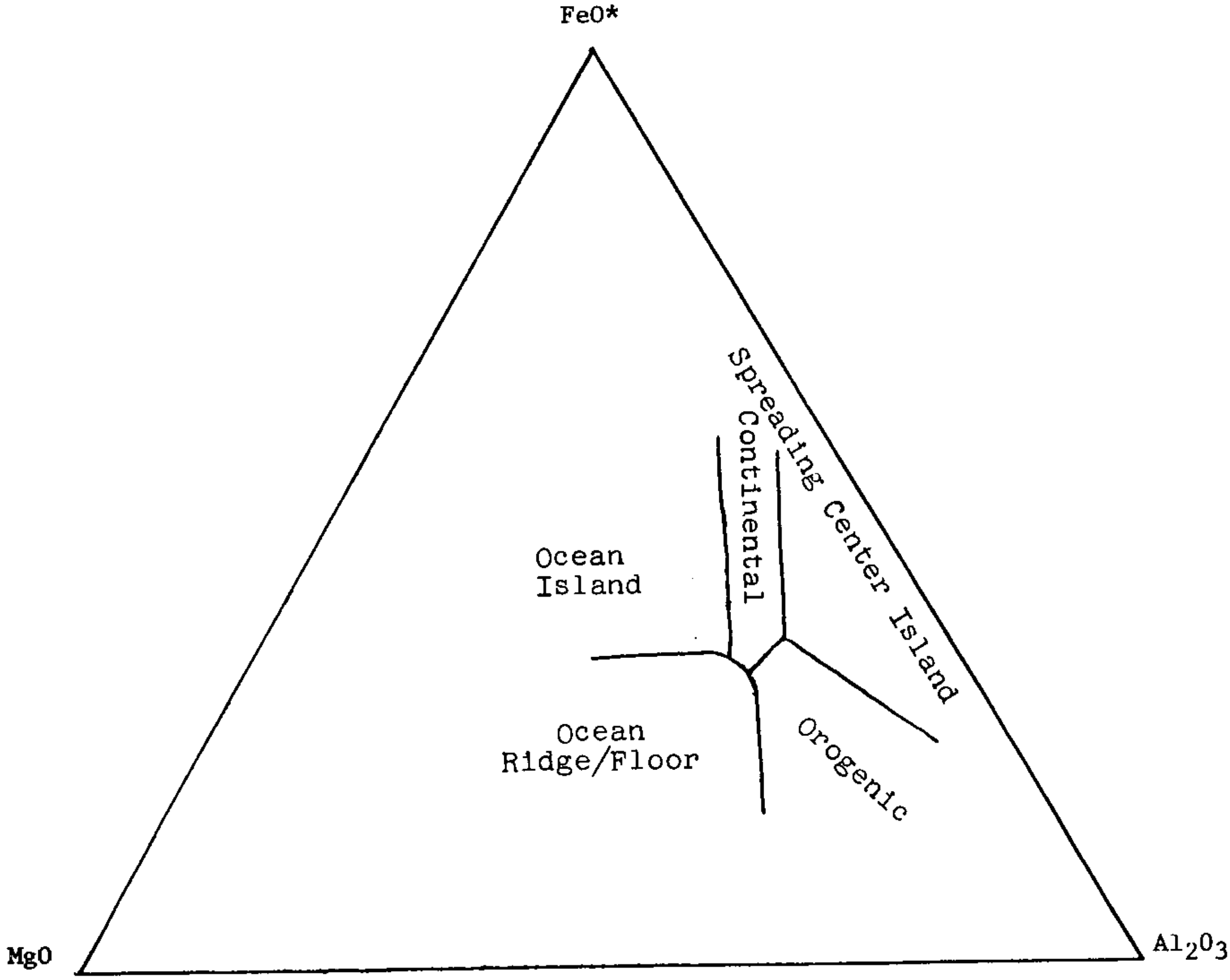


Figure 9. FeO^*-MgO-Al_2O_3 discrimination diagram showing the fields for basaltic rocks produced in different tectonic environments. (after Pearce et al., 1977).

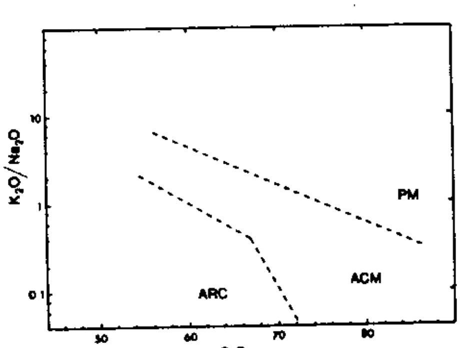


Figure 10. K_2O/Na_2O versus SiO_2 discrimination diagram for sandstones and argillites (shales). PM = Passive Continental Margin (e.g. Atlantic Coast); ACM = Active Continental Margin (e.g. West Coast of North and South America); ARC = Oceanic Island Arc Margin (e.g. Aluetians and Japan). After Roser and Korsch (1986).

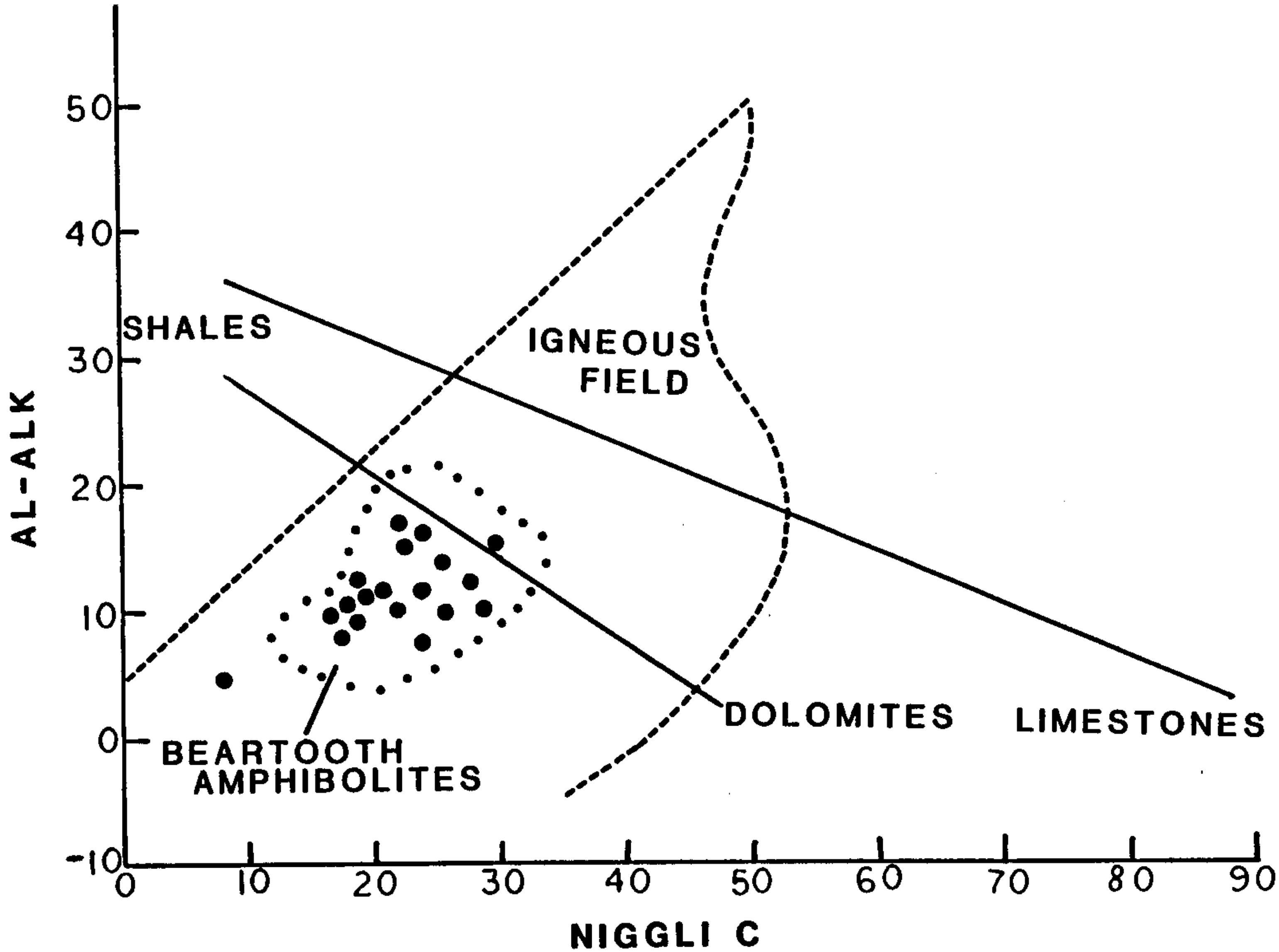


Figure 11. Plot of (Niggli values) Al-Alk versus C. Igneous field is outlined by dashed curve. Solid lines show trends of shale-carbonate mixutres. Diagram after Evans and Leake (1960).

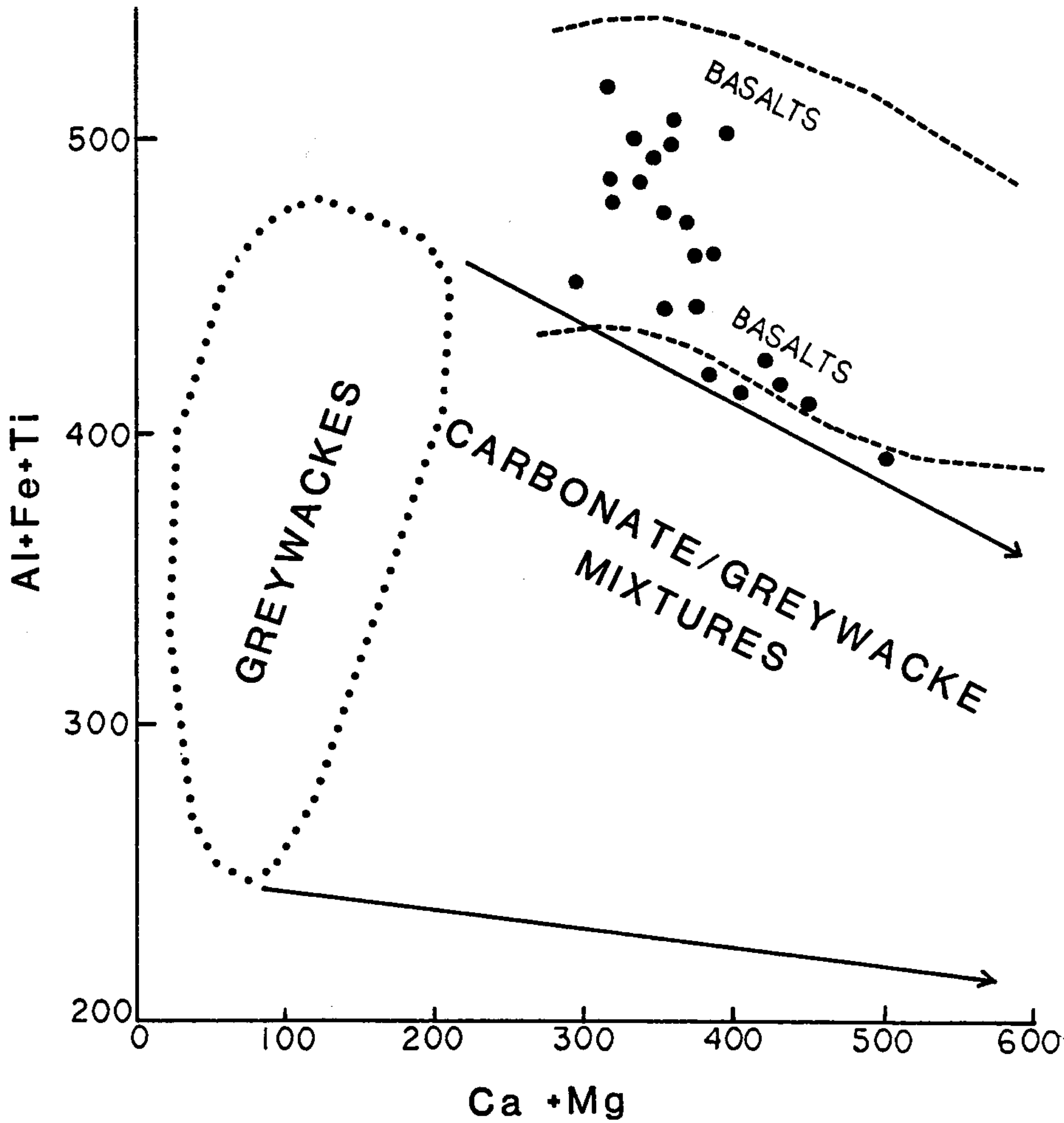


Figure 12. Al+Fe+Ti versus Ca+Mg (in millimoles per gram) discrimination diagram for possible amphibolite protoliths (after Moine and LaRoche, 1968).

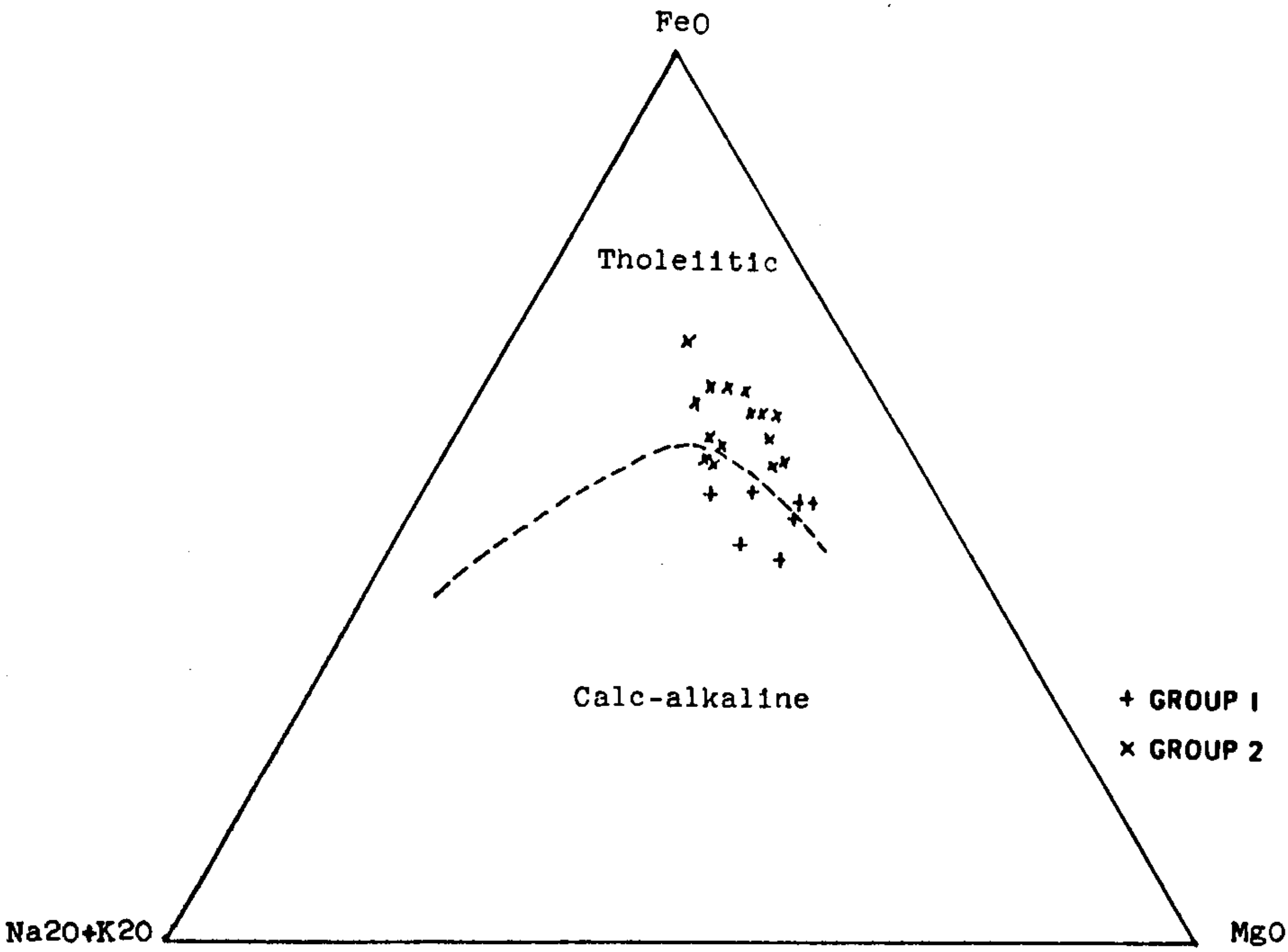


Figure 13. AFM diagram showing division between tholeiitic and calc-alkaline fields (after Irvine and Baragar, 1971).

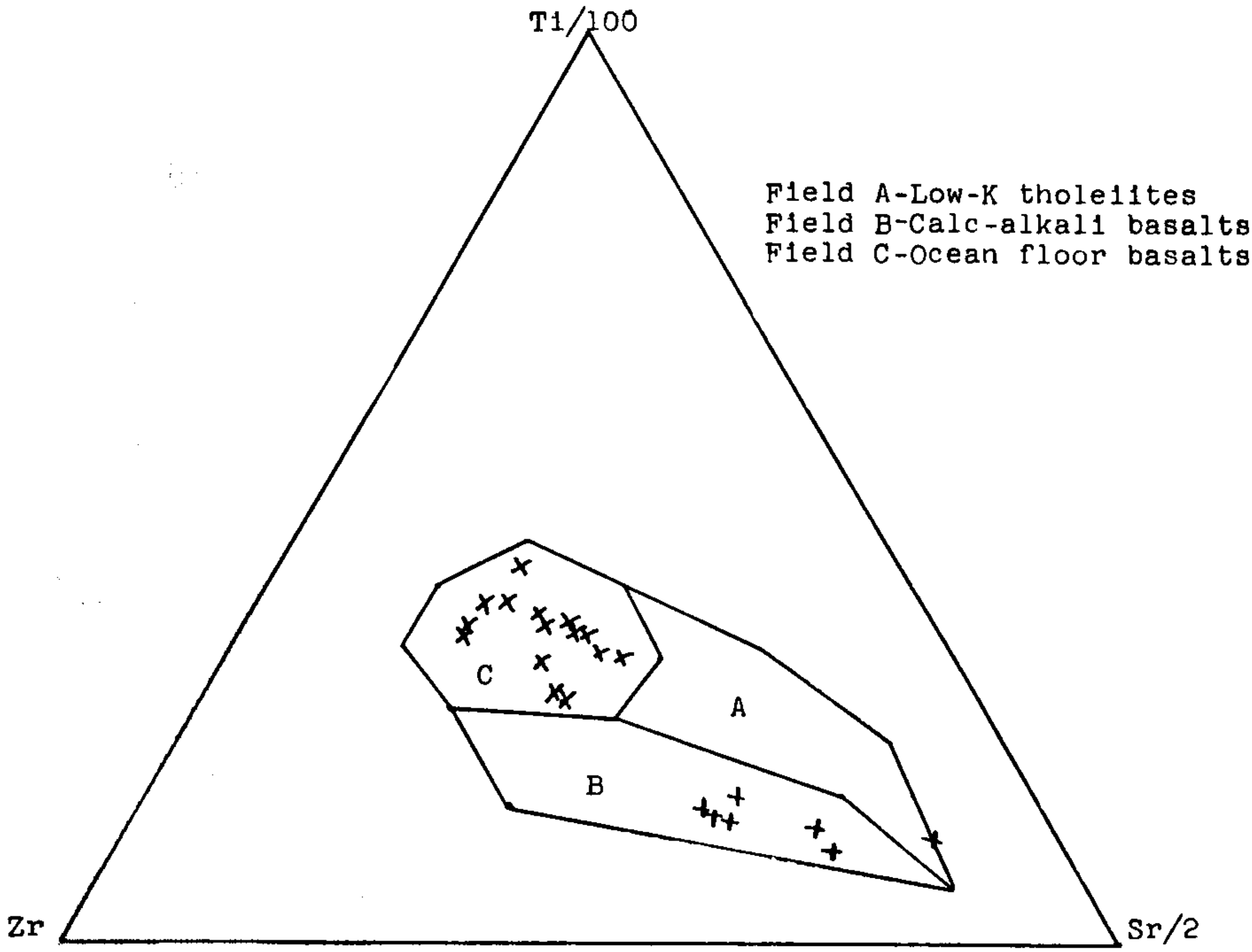


Figure 14. Ti/100, Sr/2, Zr discrimination diagram (after Pearce and Cann, 1973) showing fields for calc-alkaline, ocean floor (tholeiitic), and low-K basalts.

REFERENCES
(Part III Only)

BENDER, E. E. and HUSCH, J. M., 1985, Geochemistry and petrology of a late Archean amphibolite layer, Ruby Range, southwestern Montana. Bull. New Jersey Acad. Science, vol. 30, no. 1, p. 51.

EVANS, B. W. and LEAKE, B. E., 1960, the composition and origin of the striped amphibolites of Connemara, Ireland. Jour. Petrol., vol. 1, p. 337-363.

GORBACHEV, O. V., 1974, Ratios of pairs of similar elements in amphibolites as related to its origin. Doklady Acad. Nauk SSSR, vol. 208, no. 2, p. 432-434.

HUSCH, J. M. and HENNIGAN, T. J., 1984, the geochemistry and origins of late Archean amphibolites from the Ruby Range, Montana. Geol. Soc. America Abs. with Programs for 1984, vol. 16, no. 6, p. 547.

HYNDMAN, D. W., 1985, Petrology of Igneous and Metamorphic Rocks. McGraw-Hill, Inc., New York, 786 p.

IRVINE, T. N. and BARAGAR, W. R. A., 1971, A guide to the chemical classification of the common volcanic rocks. Canadian Jour. Earth Sciences, vol. 8, p. 523-548.

MOINE, B. and LAROCHE, H., 1968, Nouvelle approche du problem de l'origine des amphibolites a partir de leur composition chemique. Acad. Science Paris Comptes Rendus, vol. 267, p. 2084-2087.

PEARCE, T. H. and CANN, J. R., 1973, Tectonic setting of basic volcanic rocks determined using trace element analysis. EPSL, vol. 19, p. 290-300.

PEARCE, T. H., GORMAN, B. E., and BIRKETT, T. C., 1977, The relationship between major element chemistry and tectonic environment of basic and intermediate volcanic rocks, EPSL, vol. 36, p. 121-132.

ROSER, B. P. and KORSCH, R. J., 1986, Determination of tectonic setting of sandstone-mudstone suites using SiO_2 content and K_2O/Na_2O ratio. Jour. Geol., vol. 94, p. 635-650.

SHAW, D. M. and KUDO, A. M., 1965, A test of discriminant function in the amphibolite problem. Mineralogical Mag., no. 268, p. 423-435.

VAN DE KAMP, P. C., 1969, Origin of amphibolites in the Beartooth Mountains, Wyoming and Montana: New data and interpretations. Bull. Geol. Soc. America, vol. 80. p. 1127-1136.