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GEOLOGY OF THE HYDE PARK QUADRANGLE, VERMONT

By

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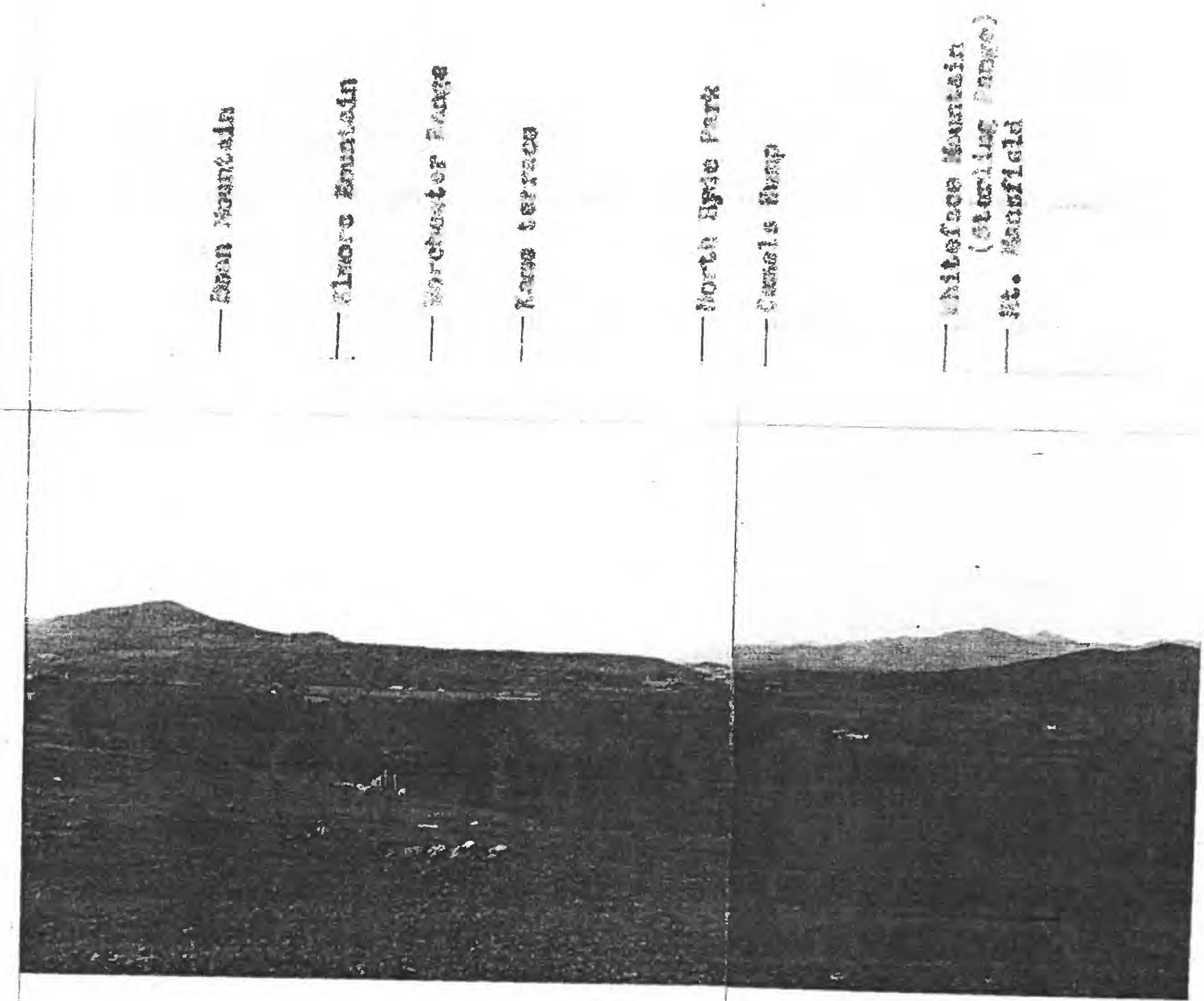

PLATE 5
~~Figure 1.~~ General view of the Hyde Park quadrangle looking south from the south slope of Belvidere Mountain (Eden).

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GEOLOGY OF THE HYDE PARK QUADRANGLE, VERMONT

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ABSTRACT

The axis of the Green Mountain anticlinorium trends north-northeast across the northwest corner of the Hyde Park quadrangle. This anticlinorium, which is the principal structural feature of the bedrock of Vermont, extends north-northeast from the Massachusetts-Vermont border the full length of the state and about 50 miles into Quebec, a total distance of about 210 miles.

The bedrock in the Hyde Park quadrangle is predominantly metamorphosed sedimentary and volcanic rock, which has been intruded by lamprophyre dikes and by ultramafic rocks. The sequence is divided into five formations; its estimated thickness is about 19,500 feet. No fossils have been found, but the rocks are tentatively dated as Cambro-Ordovician by their position between the Precambrian unconformity and rock of probable Middle Ordovician age.

All the rocks in this area except the lamphrophyre dikes have been affected by regional metamorphism. Most of the Hyde Park quadrangle is in the chlorite zone of metamorphism. In the higher grade part of the chlorite zone either chloritoid or biotite may occur, depending primarily upon the aluminum content of the rock. Biotite is rare in the rocks of the Hyde Park quadrangle, but chloritoid is

present in the Stowe formation. The garnet and kyanite zones are confined to the Stowe formation in the southeastern part of the quadrangle. The assemblages of the garnet and kyanite zones have partially or wholly retrograded to assemblages of the chlorite zone. Phase diagrams, based on the observed assemblages, are used to discuss the mineralogical transformations at successively higher grades of metamorphism and the retrograde alteration of the higher grade rocks.

A statistical study of the chemical composition of argillaceous and arenaceous sedimentary rocks forms the basis for determining the protolith (premetamorphic parent rock) of the metamorphic rocks. The weight percent of silica, and the relative molar proportions of Al_2O_3, Na_2O, and K_2O are useful parameters for such comparison. These factors were also calculated for the metamorphic rocks from estimated modes. The common protoliths of the metamorphic rocks of the Hyde Park quadrangle are quartzose sandstone, carbonaceous and noncarbonaceous shale, and subgraywacke. Neither graywacke nor arkose were common.

The Green Mountain anticlinorium is the major structure in the area, and the larger folds throughout the quadrangle parallel it. No evidence of major faults was observed. Minor structures include planar features, folds, rotated porphyroblasts, and linear features. Planar features shown on the map and discussed include bedding, various types of schistosity, and slip cleavage; fracture cleavage, slaty cleavage, and joints, though not mapped, are present locally. Linear features shown on the map include quartz rods, fold and crinkle axes, and the intersections of certain planar features; mineral lineation, streaming, and crenulation, though not shown on the map, are common. Within the Hyde Park quadrangle two differently oriented sets of folds, each with related slip cleavage and schistosity that intersect folded foliation in lines parallel to respective fold axes, may be distinguished. In many places, particularly near the axis of the Green Mountain anticlinorium, one set is superimposed upon the other. The axes of the earlier set of folds are parallel to the bedding foliation and nearly at right angles to the axes of the Green Mountain anticlinorium. The axes of the later set of folds are subparallel to the axis of the Green Mountain anticlinorium. The earlier folds and related schistosity are folded about the north-trending, nearly horizontal axes of broad open folds. It is tentatively suggested that the earlier folds are due to shortening or extension parallel to the major fold axes caused by the formation of plunging folds, salients, and recesses in the Green Mountain anticlinorium.

A talc mine is operated northeast of Johnson Village, and another talc deposit of possible commercial value has been discovered southwest of Bowen Mountain.

INTRODUCTION

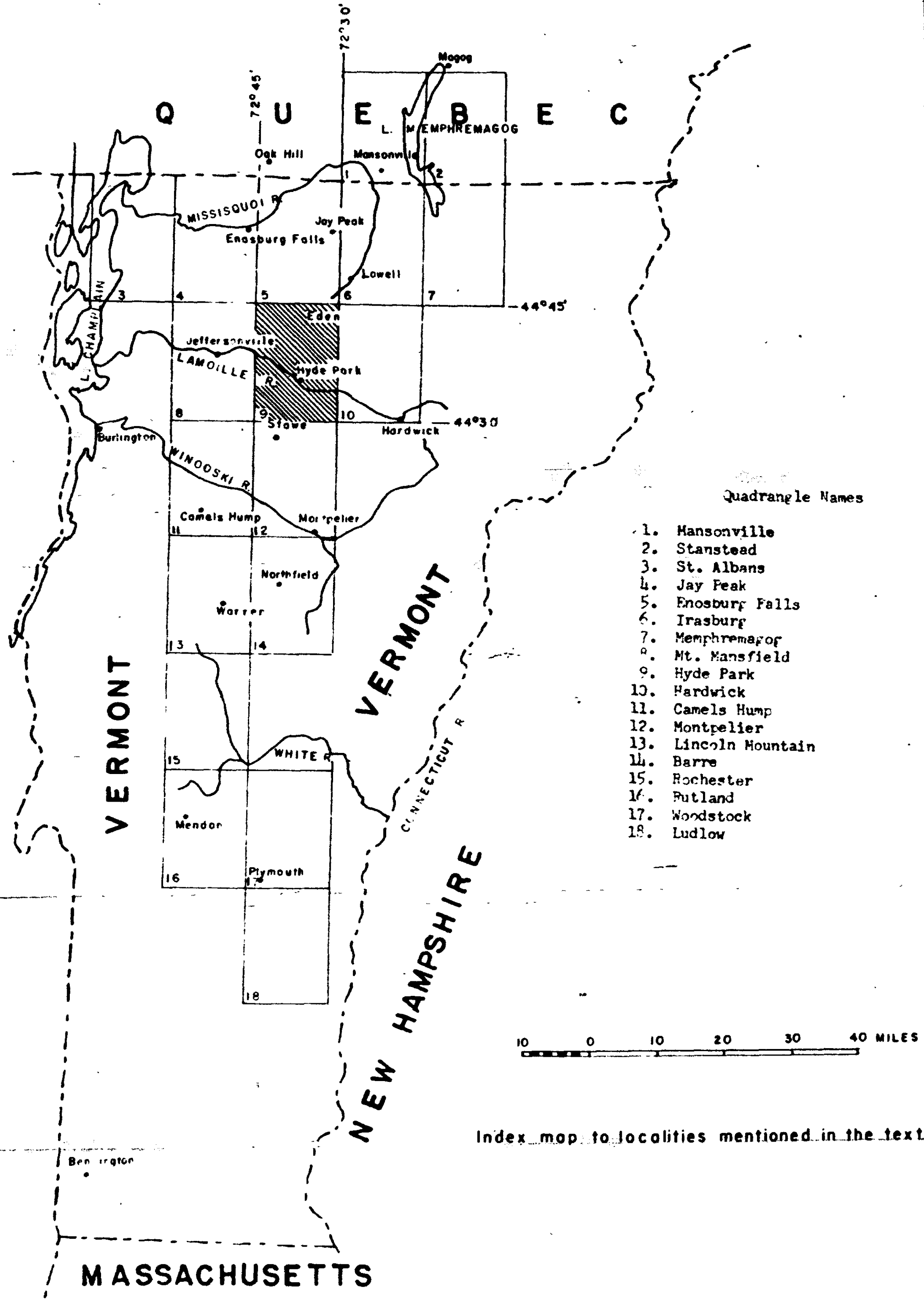
Location and culture

The Hyde Park quadrangle is in the north-central part of Vermont (see index map, Figure 1). It is bounded by parallels 44°45' and 44°30' north latitude and by meridians 72°45' and 72°30' west longitude, and covers about 212 square miles.

The area is primarily an agricultural district; lumbering, the manufacture of forest products, and mining are other major occupations. Saw mills are in North Hyde Park, Morrisville, and Johnson, and a plywood mill is in Morrisville. A talc mine and mill are in Johnson, and an asbestos mine and mill are in Eden, just north of the quadrangle boundary.

Most of the area is readily accessible by good roads. Many back-roads have been abandoned in recent years, but a revised topographic map (Plate 3), which has been published since the completion of the mapping and drafting, shows the status of the roads in 1953. Cultivated land and open pasture make up about 15 percent of the area; the rest is second-growth forest and abandoned, partly overgrown farm lands.

Figure 1



Quadrangle Names

1. Mansonville
2. Stanstead
3. St. Albans
4. Jay Peak
5. Enosburg Falls
6. Irasburg
7. Memphremagog
8. Mt. Mansfield
9. Hyde Park
10. Hardwick
11. Camels Hump
12. Montpelier
13. Lincoln Mountain
14. Barre
15. Rochester
16. Rutland
17. Woodstock
18. Ludlow

Index map to localities mentioned in the text

Topography and drainage

The Green Mountains on the west side of the quadrangle have a maximum elevation of 3,715 feet, whereas Elmore Mountain on the east side has an elevation of 2,608 feet. The Lamoille River at the west side of the quadrangle has an elevation of about 455 feet and is the lowest point in the quadrangle. Average local relief is about 700 feet, but local relief may be as much as 1,800 feet. Systems of north trending ridges on both the east and west sides converge in the northe part of the area, closing a broad central valley. In the northwest an north-central part the usual linear form of the Green Mountains is replaced by a dissected upland with numerous peaks and ridges ranging from 2,000 to 2,700 feet in altitude.

The Lamoille River flows northwesterly across the area, transecting both the Green Mountains and the eastern ridges.

The topography, which has been largely produced by preglacial stream dissection, reflects to some extent the structural pattern, attitude, and varied erosional resistance of the underlying rocks. Large glacial-lacustrine deposits of silt, sand, and gravel have accumulated in the valleys, kame terraces modify the slopes, and ground morraine covers much of the higher slopes. Locally, small streams and the Lamoille River have been superimposed upon rock ridges after cutting through large accumulations of glacial deposits. In general, the glacial deposits have smoothed the preglacial land surface except in the Diggings on the east side of the area.

Almost all the area drains to the Lamoille River and thence to Lake Champlain, but Sterling Brook in the southwest corner flows south into the Winooski River, which also drains into Lake Champlain. The Gihon River, the Green River, and the North Branch of the Lamoille River are the major tributaries of the Lamoille River in the area. Lake Lamoille was formed by damming the Lamoille River for the production of hydroelectric power. Lake Eden, South Pond, and Elmore Pond, which were formed by glacial erosion and by damming by glacial deposits, have been enlarged by the erection of dams to control the supply of water for power dams on the Lamoille River. Green River Reservoir (Plate 3) is another storage area which was formed by the erection of a concrete dam, about 60 feet high, on the Green River. The Hyde Park quadrangle is well drained, except for portions of the Upper and Lower Diggings, where glacial deposits have formed numerous small lakes and swamps. Much of this poorly drained area is now included in Green River Reservoir.

Purpose of study

Although numerous areas in southern Vermont east of the Green Mountains have been studied in detail in recent years, few detailed studies have been attempted in north-central Vermont. The primary purpose of this work is to delineate the stratigraphy and structure of the Hyde Park quadrangle and to relate it to the general geologic framework of Vermont. A further aim is to discuss in detail the petrography and metamorphism of this area.

The work was undertaken for the United States Geological Survey, which has been engaged for a number of years in a general study of the belt of ultramafic rocks in Vermont. This study, which was prompted by domestic shortages of asbestos and certain grades of talc, has included detailed study of individual deposits as well as areal geologic mapping. The relations of the ultramafic rocks to the areal geology and to the metamorphism are being studied by geologic mapping and study of all or part of eight 15-minute quadrangles.

Method of study

About ten man-months were spent in the field within the area from 1951-1954, but the author's other duties during this time involved field mapping in all the bordering quadrangles as well as reconnaissance work in Canada and other portions of Vermont. The study of 125 thin sections from this area was supplemented by the study of thin sections from the Montpelier and Lincoln Mountain quadrangles to the south and the Jay Peak and Irasburg quadrangles to the north.

The author had spent three field seasons mapping the same formations in adjacent areas before he began the present study. In addition he had examined over 450 thin sections taken in that work. From this experience a number of characteristic rock types were noted and given symbols. These symbols, useful for note taking, were particularly valuable for plotting lithologic data upon maps.

The field map was the 1927 edition of the Hyde Park sheet of the United States Geological Survey enlarged to a scale of 1:48,000. Structural and lithologic data were transferred to 1:24,000 enlargements. About a tenth of the area, that which lies in the extreme northeast corner of the quadrangle, was mapped in greater detail on maps (scale 1:24,000) prepared by multiplex methods from aerial photographs taken in 1951. Aerial photographs (scale 1:20,000; 1:24,000) were used extensively in less heavily wooded areas. Most of the stations were located by altimeter traverse, but pace and compass methods were used occasionally. Contacts were transferred from the multiplex map and from aerial photographs to the final compilation by use of a vertical sketchmaster. Thus, although placed in their proper horizontal position, the location with respect to topographic features may be slightly erroneous in a few areas.

The location of specimens which are cited in the tables is given by a convenient reference system. The quadrangle map is divided into its nine 5-minute sections, which are designated northeast, east-central southeast, south-central, etc., and abbreviated NE, EC, SE, SC, etc. Within each ninth the southwest corner is the origin and decimal rectangular coordinates are used to fix any point; the first figures designate miles east and the last figures designate miles north of the origin. Thus, NW-1.05, 2.03 indicates a point 1.05 miles east and 2.03 miles north of the southwest corner of the northwest ninth of the quadrangle. Unless prefixed by a quadrangle name all such locations are in the Hyde Park quadrangle.

Acknowledgments

The writer is indebted to Professors J. B. Thompson, Jr. and M. P. Billings of Harvard University and Dr. W. M. Cady of the United States Geological Survey under whose direction this work was pursued and who gave freely of their time for instruction and criticism. Dr. W. M. Cady has reviewed in the field some of the more complex problems and in his company the author performed reconnaissance work in adjacent areas. The insight into the regional geologic picture given the author by Dr. Cady is chiefly responsible for the section on regional relations.

Part of the laboratory work and of the writing was completed while the author was studying at Harvard University under a National Science Foundation Fellowship. To Dr. Hans Eugster of the Geophysical Laboratory, Carnegie Institute of Washington, D. C., I am grateful for the X-ray identification of numerous "sericite" samples.

The manuscript was critically reviewed in its entirety by Professors M. P. Billings, J. B. Thompson, Jr., and Dr. W. M. Cady, and portions of it were reviewed by A. H. Chidester and Donald Tatlock, both of the United States Geological Survey.

Members of adjacent field parties provided helpful discussion while reviewing common problems in the field. Among these are M. P. Billings, W. F. Brace, W. M. Cady, A. H. Chidester, H. C. Cooke, J. K. Gilmore, P. H. Osberg, J. L. Rosenfeld, J. B. Thompson, and E-an Zen.

Finally, my sincere gratitude goes to my wife, Ann, for her assistance during the statistical study of the analyses of sedimentary rocks, and for typing numerous drafts of the manuscript during its compilation.

Previous work

Edward Hitchcock and others in 1861 published a comprehensive report on the geology of Vermont, including a geological map. They recognized the anticlinal nature of the Green Mountains and delineated the broader lithologic types. C. H. Hitchcock (1877, 1878) produced a geologic map of New Hampshire including adjoining parts of Vermont and Maine with an accompanying report. This showed rather accurately the distribution of the major rock types in the Hyde Park quadrangle. Both of these reports form valuable sources of information, especially for the location of small mineral prospects. No other areal mapping has been done within the quadrangle. All geological maps of areas within the state which were published prior to 1952 have been indexed on a map of Vermont (Boardman, Leona, 1952).

The interpretation of the stratigraphic sequence of central and eastern Vermont evolved to nearly its present form in a series of papers by Richardson (1902 to 1927), Perry (1929), and Currier and Jahns (1941). Richardson traced an unconformity (now recognized as the base of the Shaw Mountain formation) from Canada to Massachusetts. The rocks above the unconformity were described by Currier and Jahns

(1941) and White and Jahns (1950). Those below the unconformity were described in the Plymouth-Bridgewater area by Perry (1929) and by Hawkes (1940, 1941).

Since 1946 much of central and eastern Vermont has been or is being mapped, but very little of this work has yet been published. A generalized map of the southern half of the state, including much of this unpublished mapping, was compiled by Thompson (Billings and others, 1952). The rocks of Perry's section have been traced south and correlated with strata in Massachusetts by Thompson (1950), Rosenfeld (1954), Skehan (1953), and MacDonald (unpublished). They have been traced north by Hawkes (1940, 1941), Thompson (1950), Chang (1950) Brace (1953), Osberg (1952), and Jahns and White (1950) to the southern borders of the Lincoln Mountain and Montpelier quadrangles. Cady, Albee, and Murphy have partially completed the mapping of the Lincoln Mountain quadrangle. Cady has completed the Montpelier quadrangle (Cady, 1956), Albee has completed the Hyde Park quadrangle, and through incomplete and reconnaissance mapping the stratigraphic units have been carried north to the Canadian border by Cady, Albee, and Chidester. This work, as well as work in the southern part of the state, is still in progress.

Regional geologic setting

The axis of the Green Mountain anticlinorium (Plate 4), trending north-northeast, crosses the extreme northwest corner of the Hyde Park quadrangle. This anticlinorium, which is the principal structural feature of Vermont, extends north-northeast from the Massachusetts border the full length of the state and about 50 miles into Quebec, a total distance of about 210 miles. In the southern half of Vermont Precambrian rocks are exposed in the core of the anticlinorium. West of the Green Mountain anticlinorium slightly metamorphosed limestone, dolomite, quartzose sandstone, and some shale are exposed in a broad synclinorium. These rocks rest with pronounced angular unconformity upon the Precambrian of the Green Mountains and fossils indicate a fairly complete section from Lower Cambrian to Middle Ordovician age. These rocks of the western syncline are folded and cut by numerous thrust faults.

On the east the Precambrian rocks are overlain, again with pronounced angular unconformity, by a thick series of metasedimentary and metavolcanic rocks. The upper part of this eastern section is sparsely fossiliferous; it cannot be older than Middle Ordovician nor younger than early Devonian. The lower part of the series is nonfossiliferous, but is believed to include Cambrian and early Ordovician rocks. Although presumably equivalent in age to the slightly metamorphosed carbonate rocks and quartzose sandstones to the west of the Green Mountains, the section east of the mountains consists largely of metamorphosed argillaceous, arenaceous, and volcanic rocks. A generally accepted correlation of the eastern and western sections is not yet possible because of the east-west change in sedimentary facies (Thompson, p. 14, in Billings and others, 1952). The Green Mountain anticlinorium plunges gently northward so that in the northern part of the state the Paleozoic gneisses and schists bridge the anticlinorium in a series of folds. The difficulties of correlation in this area and the detailed regional relations will be described in a later section.

The general features of Vermont are markedly similar to those in the Berkshire Hills of Massachusetts and the Blue Ridge Mountains and Piedmont of the southern Appalachians. In each of these areas there is a crystalline Precambrian core and a relatively unmetamorphosed carbonate and quartzite sequence to the west, resting unconformably upon the Precambrian and cut by numerous thrust faults. East of the Precambrian core in each area there is a metamorphosed argillaceous-arenaceous sequence of possible early Paleozoic age.

STRATIGRAPHY AND LITHOLOGY

General statement

The bedrock in the Hyde Park quadrangle is predominantly metamorphosed sedimentary and volcanic rocks, mostly schist, gneiss, quartzite, phyllite, greenstone, and amphibolite, but also including slate, conglomerate, granulite, and impure calcite marble. These rocks have been intruded by lamprophyre dikes and by ultramafic rocks; the latter have been partly or wholly altered to serpentinite, talc-carbonate rocks and steatite.

All of the rocks except the lamprophyre dikes have been affected by regional metamorphism. Chlorite, garnet, and kyanite have been interpreted as successive indicators of increasing metamorphic grade in the micaceous schist. Similarly, chlorite, actinolite, and hornblende are indicators in the greenstone and amphibolite. Most of the Hyde Park quadrangle is in the chlorite zone. In the higher grade part of the chlorite zone either chloritoid or biotite may occur depending upon the alumina content. Biotite is rare in the Hyde Park quadrangle, but chloritoid is locally abundant in the Stowe formation. The garnet and kyanite zones are confined to the Stowe formation in the southeastern part of the quadrangle.

The primary consideration in defining formations was to utilize mappable rock units which correspond as closely as possible to units already defined in adjacent areas. It was necessary to be familiar with type sections described many miles to the south, especially those in the Woodstock and Ludlow quadrangles (Perry, 1927, 1929). However certain formations, which are characterized by a rather distinctive and uniform lithology in the type section, cannot be distinguished on the same basis in the Hyde Park quadrangle. They have been traced northward from the type sections, attempting to take account of gradual changes in sedimentary and metamorphic facies along the way. Mapping has been partially completed in some of the intervening quadrangles, but in others the units were traced by reconnaissance work. This correlation is summarized in a later section and in Table 15.

The rock units summarized in Table 1 and shown on the geological map and cross sections correspond as closely as possible to the type units. Except for the Umbrella Hill formation, no formation contains a completely distinctive type of rock; nearly the same rocks may be found in each of the units. The lack of distinctive rocks and sequences made correlation difficult between the rocks on the east side of the Hyde Park quadrangle and rocks in a syncline in the Foot Brook area (the Foot Brook syncline) in the northwest part of the quadrangle.

The thicknesses given for the rock units in Table 1 are approximations that depend upon the interpretation of the map pattern. The scarcity of distinctive marker beds made it impossible to estimate the amount of repetition in most parts of the quadrangle. Small folds are extremely abundant and it is apparent from the pattern of the boundaries of the formations that in many parts of the area there must be numerous larger folds. In much of the area the observed dip of bedding and bedding schistosity is very different from that of the average dip. Section A-A' (Plate 1) has been so located that reasonable approximations of the thicknesses of the formations are shown upon it.

The ages assigned to the units are tentative as no fossils have been found in any of them. Southwest of the Hyde Park quadrangle in the Lincoln Mountain quadrangle the Camels Hump group rests upon known Precambrian rocks with pronounced angular unconformity (see also Cady, 1956; and Osberg, 1952, p. 24, 34-38). The Moretown formation is probably Ordovician according to Currier and Jahns (1941, p. 1496, 1508-1509), who based the age designation of its equivalent, the Cram Hill formation, upon their correlation along the strike with graptolitic slates in Magog, Quebec. The Shaw Mountain formation, which overlies the Moretown formation, contains crinoid fragments (Currier and Jahns, 1941, p. 1500-1501) which indicate by their size a Middle Ordovician or younger age. Corals found by Cady (1950) in the Waits River formation, which overlies the Shaw Mountain, suggest a Middle Ordovician or younger age. No marked stratigraphic break has been found from the Camels Hump through the Moretown formation, and hence this part of the section can only be assigned to the Cambro-Ordovician. The upper limit of the Cambrian is placed arbitrarily at the boundary between the Ottauquechee and the overlying Stowe formation to conform with a tentative correlation between sections east and west

of the Green Mountains derived by Cady and the author from the resul[t]
of reconnaissance in Quebec. This correlation has since been confir[med]
by Osberg (1956). Thompson (Billings and others, 1952, p. 18) sugge[sts]
"An alternative interpretation with much to commend it is that the p[re-]
Cram Hill (pre-Moretown in this area) formations are all of Middle
Ordovician age and are to be correlated only with the rocks above th[e]
Middle Ordovician unconformity west of the mountains." This problem
will be discussed in more detail in a later section.

The ultramafic intrusive rocks are assigned an Ordovician age,
inasmuch as such rocks are not known to intrude strata younger than
Ordovician in western New England or adjacent Quebec. The lamprophy[re]
dikes can be dated only as post-metamorphic, but are assigned to the
Mississippian(?) to correspond with the ages assigned to similar dik[es]
along the Connecticut River (White and Billings, 1951, p. 662). The
principal folds in the region, and the metamorphism, are of Middle o[r]
Late Devonian age (Cady, 1945, p. 560; White and Billings, 1951, p.
695).

TABLE 1. Stratigraphic Section

Age	Formation	Thickness
Middle Ordovician	**Moretown formation:** Quartz-sericite-chlorite-albite-epidote granulite with close-spaced micaceous partings, micaceous quartzite, and sericite slate.	over 4000'
Ordovician (?)	**Umbrella Hill formation:** Quartz-pebble and slate-pebble conglomerate with thin interbeds of sericite slate and thicker interbeds of friable calcareous slate; contains abundant chloritoid.	0 - ± 900
	UNCONFORMITY	
	Stowe formation	
	Upper:	± 3300
	Chlorite zone: Predominantly fine grained, thinly laminated, sericite-quartz-chlorite-albite schist containing numerous thin lenticles of quartz. Greenstone, graphitic phyllite, calcareous greenstone, and calcareous phyllite are minor lithologic types. Some impure calcite marble in the Foot Brook area.	
	Garnet and kyanite zones: Coarse-grained, quartz-muscovite-chlorite schist which contains porphyroblasts of chloritoid, garnet, and kyanite, and abundant quartz segregations. Interbeds of coarse-grained amphibolite are present.	
	Middle:	± 1300
	Greenstone and amphibolite; albite-actinolite-epidote-chlorite-carbonate greenstone in northern part of area; hornblende-epidote-albite-quartz amphibolite in southern part of area; includes some interbeds of micaceous schist.	
	Lower:	± 2500
	Lithologic characteristics are similar to the chlorite zone of upper part of the Stowe formation.	

TABLE 1. Stratigraphic Section -- Continued.

Age	Formation	Thickness
Cambrian (?)	Ottauquechee formation: Black graphitic quartz-sericite phyllite and schist and massive dark-gray quartzite in southern two-thirds of quadrangle; graphitic phyllite, pebbly quartzite, and pebbly quartz-sericite schist in northern part. Quartzite with micaceous partings and sericite-quartz-chlorite phyllite are minor rock types. Thickness in Foot Brook area: ± 600 feet.	± 2500
	Camels Hump group: Predominantly graphitic schist and quartzite with interbeds of nongraphitic schist and dark-gray massive quartzite in central and southern part of quadrangle; predominantly graphitic and nongraphitic quartz-sericite-chlorite schist and gneiss with porphyroblastic albite in western and northern part of quadrangle. Silver-green quartz-sericite-chlorite-magnetite-albite schist forms several map units; one is as much as 800 feet thick and lies at the top of the Camels Hump group in the southern half of the quadrangle. The Belvidere Mtn. amphibolite, as much as 850 feet thick, lies at the top of the Camels Hump group in the northern part of the quadrangle, but fingers out southward.	over 5000

BASE NOT EXPOSED

Metamorphosed sedimentary and volcanic rocks

Camels Hump group

General statement.

The Camels Hump group (Cady, 1956) includes the schist, gneiss, quartzite, and other less common rocks between the known Precambrian rocks and the base of the Ottauquechee formation. The base and the lower part of the Camels Hump group are not exposed in the Hyde Park quadrangle. The Camels Hump group includes part of the Mount Holly complex as used by Osberg (1956, p. 21-26), and the Monastery, Grenville, and Pinney Hollow formations of Osberg (1952, p. 42-51) in the Rochester area. The latter three formations occupy about the same stratigraphic position as the Tyson, Hoosac (or Grahamville), and Pinney Hollow formations of Thompson (1950; Billings and others, 1952, p. 40-41), and Brace (1953, p. 45-50) in areas to the south of Roches[ter.] It has been possible to trace northward the general correlation of th[e] formations, but they do not form practical mapping units in the Hyde Park quadrangle because of lithologic change along the strike.

The greater part of the Camels Hump group, comprising albite schist and gneiss, graphitic schist, and quartzite, is a heterogeneous unit with a wide range in composition and grain size within the quadrangle. Three other units, quartz-sericite-chlorite-magnetite-albite schist, greenstone, and the Belvidere Mountain amphibolite, were mapped and are described separately.

Albite schist and gneiss, graphitic schist, and quartzite.

Selected estimated modes of these rocks are given in Table 2, columns 1-14. The Camels Hump group in the northwest corner of the area and north of the White Face Mountain-Caper Hill ridge consists predominantly of quartz-sericite-chlorite schist and gneiss, and graphitic quartz-sericite-chlorite schist both with porphyroblastic albite. A good section of these rocks is exposed from Belvidere Junction, southeastward over Laraway Mountain, and along the Green Mountain Club Trail to the east. A more accessible section is exposed along the Lamoille River valley west from the site of Riverside School (Johnson).[1] Schist and gneiss grade imperceptibly into one another. They are silver-gray, medium- to coarse-grained rocks which form large smooth outcrops. Albite porphyroblasts commonly range from 3 to 10 millimeters and rarely to 25 millimeters. In some beds albite makes up about 25 percent of the rock. Quartz, sericite,[2] albite, and

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[1] This notation will be used throughout the paper to refer to the town in which the feature is located. The word "site" is used because many of the schools shown on the map are now abandoned.

[2] Sericite is used in this report to refer to fine-grained white mica, including muscovite, paragonite, and possibly others. Paragonite was present in 16 of 68 specimens examined by X-ray methods; muscovite was present in all 68. Most of the X-ray study was performed by Hans Eugster of the Geophysical Laboratory, Carnegie Institution of Washington; part of it was performed by the author.

chlorite are the dominant minerals; magnetite, clinozoisite-epidote, and tourmaline are the common accessory minerals. Along the crest of the Sterling Range a little west of the Hyde Park quadrangle garnet porphyroblasts are present. Small euhedral garnet crystals enclosed in albite porphyroblasts were noted in thin sections from the Camels Hump group on the west side of the Hyde Park quadrangle. Locally, friable weathered rock indicates the presence of carbonate. Layers rich in quartz alternate with layers rich in sericite and chlorite, and the platy minerals are oriented parallel to the layers. Lenticular aggregates of quartz parallel to the foliation and irregular quartz veins, that locally contain calcite, chlorite, and albite, are abundant. Numerous beds of black and white banded quartzite, 3 to 12 inches thick, are exposed with the gneiss on the south slope of Laraway Mountain, but are less abundant southward.

The quartz-sericite-chlorite schist and gneiss with porphyroblastic albite is very irregularly interbedded with graphitic quartz-sericite-chlorite schist containing porphyroblastic albite. The thickness of the interbeds range from several inches to several hundred yards. The graphitic schist is similar in general lithologic characteristics to the nongraphitic schist, but is finer grained and more schistose. It is gray to black, weathers rusty, and contains some pyrite, but no magnetite. The porphyroblasts of albite are black because of disseminated graphite; some 50 millimeters in diameter were noted near Belvidere Junction.

In the southern and central parts of the quadrangle the Camels Hump group is much finer grained and quartzite is more abundant than in the northwest part of the quadrangle. Graphitic schist with a wide range in quartz and mica content predominates; it is associated with micaceous quartzite and massive gray quartzite.

The graphitic schist and quartzite are characteristically gray to black, contain sporadic porphyroblasts of pyrite, and are rusty weathering. A single group of outcrops commonly contains the following intergradational lithologic types in the following decreasing order of abundance: (1) graphitic quartz-sericite-chlorite-albite schist that grades into a more micaceous graphitic quartz-sericite-chlorite phyllite or into extremely quartzose graphitic schist with thin mica-rich layers that separate quartz-rich layers; (2) nongraphitic quartz-sericite-chlorite-albite schist; (3) micaceous quartzite, both graphitic and nongraphitic, with discrete paper thin schistose partings comprised of sericite and chlorite; and (4) blue-gray massive quartzites without foliation. Outcrops showing this characteristic interbedding are shown on Plate 6. Both the micaceous and the massive quartzite locally contain coarse detrital grains (granule and pebble size) of quartz. The quartzite beds are tightly folded, but in spite of this intense deformation, there are relatively few veins and segregated lenticles of quartz. The albite is fine-grained and is not readily seen in the field, although thin sections show as much as 15 percent of fine-grained albite. Accessory minerals include graphite, carbonate, sphene, clinozoisite-epidote, pyrite, limonite, tourmaline, apatite, allanite, and zircon. The thickness of the layers of various lithologic types ranges from a few inches to several tens of feet. The range in relative proportions of each type is not consistent enough to define mappable units. Thick blue-gray massive beds of quartzite are abundant at several localities, but they can be traced for only a short distance along the strike. All of these rocks are well exposed in the network of roads between Johnson and Centerville (Hyde Park township) and in Sterling Brook (Morristown).

Grain size increases to the north in the Camels Hump group east of the Foot Brook area and the micaceous quartzite becomes quartz-sericite-chlorite-albite gneiss, the graphitic schist and phyllite becomes graphitic quartz-sericite-chlorite schist with porphyroblasts of black albite, and nongraphitic schist becomes quartz-sericite-albite-chlorite schist and gneiss. Albite occurs as porphyroblasts from 1 to 3 millimeters in diameter and thin sections shown as much as 35 percent albite. Segregated lenticles and veins of quartz are more abundant. In the northern third of the quadrangle the rocks of the Camels Hump group on both sides of the Foot Brook area are coarse-grained. On Belvidere Mountain, 3.6 miles N. 27° W. of the Knob (Eden), a quartz-sericite-albite-chlorite gneiss, which underlies the Belvidere Mountain amphibolite, is similar to gneiss exposed on the summit of Laraway Mountain.

Figures 1 and 2. Outcrops of Camels Hump group showing typical lithologic features. Note the interbedding and quartz segregations, 0.80 mile S. 85° W. of the site of Wiswell School (Hyde Park).

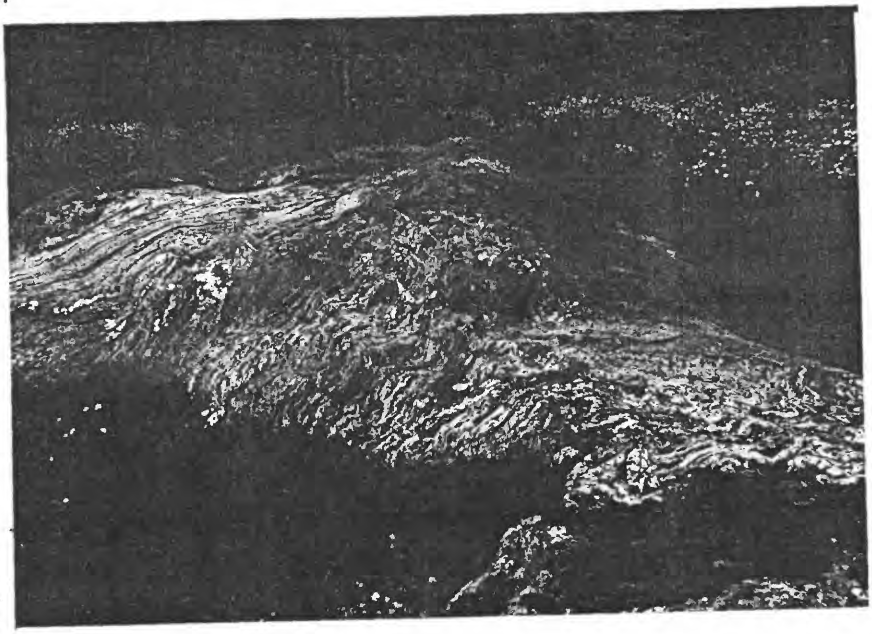


Figure 1

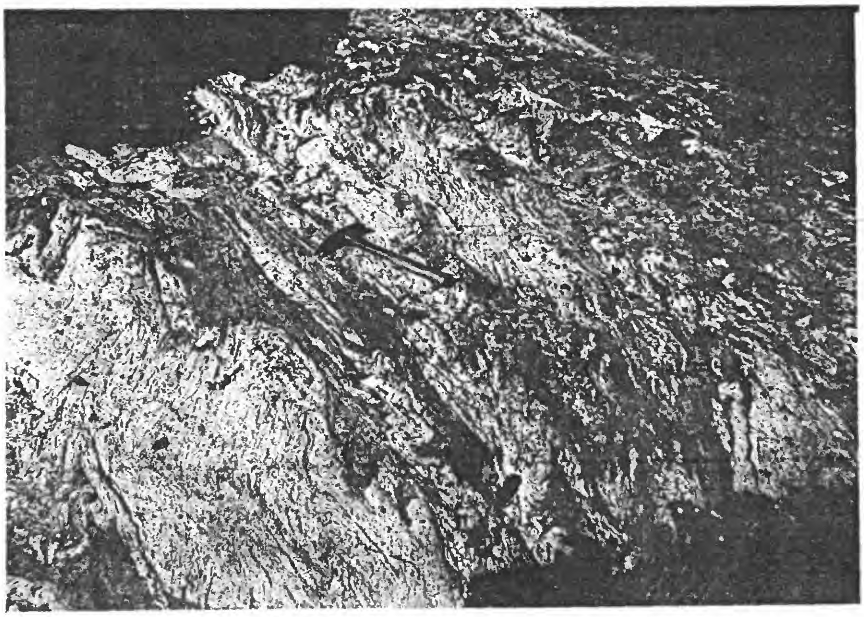


Figure 2

Plate 6

Quartz-sericite-chlorite-magnetite-albite schist.

Selected modes of these rocks are given in Table 3, columns 15-18. Eight bands of quartz-sericite-chlorite-magnetite-albite schist are shown on the geologic map. These will be referred to separately as the Caper Hill-west, Caper Hill-east, Swamp Road, Terrill Hill, Morrisville, Talc Mine, Ober School, and East Johnson-Hyde Park bands. The clean silver-green color of the weathered outcrops and the presence of abundant magnetite octahedra (1 to 3 millimeters across) contrasts sharply with the general dark color, rusty-weathered surface, and graphite-pyrite-limonite content of the rocks that surround these bands. However, quartz-sericite-chlorite-magnetite-albite schist is not confined entirely to these bands; other smaller bands could be mapped if the exposures were better. The porphyroblastic albite-quartz-sericite-chlorite schist and gneiss west of the Foot Brook area are probably in part stratigraphically equivalent to these schist bands.

The rock is comprised of closely alternating layers (one-fourth to 2 inches thick) of mica-rich schist (50-95 percent sericite and chlorite with minor quartz) and quartz-rich granulite with minor sericite and chlorite. The quartz-rich granulite layers are commonly thinner than the micaceous layers, parallel the schistosity of the micaceous layers, and probably represent original bedding. Locally, as in the Terrill Hill and East Johnson-Hyde Park bands, beds of sericite-chlorite quartzite that contain detrital grains of quartz as much as 5 millimeters across are abundant. Lenticular segregations

TABLE 3. Estimated modes of the Camels Hump group.

	15	16	17	18	19	20	21	22
Quartz	50	12	1	42	3			
Albite		3	5	20	57	25	28	15
Sericite	30*	70*	50**	10				
Chlorite	18	10	40	<1	13	28	10	3
Biotite				15	2	<1	<1	
Actinolite						5	25	50
Magnetite	1	5	4	1	1	2	1	
Graphite								
Carbonate				1	5			
Epidote group				<1	15	40	35	30
Pyrite				<1				
Ilmenite								
Sphene	1				4			2
Rutile								
Apatite		tr		<1	tr			
Tourmaline		tr	<1					
Location:***	SC-3.14, 4.34	SC-1.48, 4.83	SC-0.10, 4.70	SC-3.55, 2.63	SC-0.15, 2.75	NE-1.82, 4.10	NE-1.10, 4.23	NE-1.53, 4.98

* Muscovite; paragonite absent.
** 20% of sericite is paragonite.
*** Explanation on page 5.

15. Quartz-sericite-chlorite-magnetite-albite schist - Morrisville band.
16. Quartz-sericite-chlorite-magnetite-albite schist - Terrill Hill band.
17. Quartz-sericite-chlorite-magnetite-albite schist - Swamp Road band.
18. Quartz-sericite-chlorite-magnetite-albite schist - Caper Hill - west band.
19. Albitic greenstone zone in Camels Hump group.

20-22. Belvidere Mountain amphibolite.

of quartz parallel to the schistosity, and irregular pods and veins of quartz are common in all outcrops (Plate 7). Some of the veins are several feet wide and may be traced for as much as 100 feet.

The magnetite octahedra are most abundant in the layers rich in micaceous minerals. The albite content is low (0 to 5 percent) in the Morrisville, Terrill Hill, Swamp Road, and Caper Hill-east bands, but is higher locally (as much as 40 percent) in the Caper Hill-west, and higher generally in the Talc Mine and Ober School bands. Paragonite is present in two of the four specimens studied by X-ray methods. Accessory minerals include rutile, ilmenite, sphene, tourmaline, apatite, zircon, and clinozoisite-epidote. The southern bands are generally fine-grained schist with albite porphyroblasts less than one millimeter in diameter, but the northern thirds of the Talc Mine and Ober School bands are quartz-albite-sericite-chlorite-magnetite schist and gneiss.

Some lithologic features are peculiar to certain of the bands. The Morrisville band locally contains beds of limestone from half an inch to 2 inches thick, and similar beds of limestone as well as beds of greenstone have been noted in some of the other bands. About 35 feet of quartz-albite-biotite-sericite schist is exposed in the Lamoille River a little south of the village of Johnson, and similar rock crops out on Caper Hill in the Caper Hill-east band, the similarity suggesting that the Caper Hill-east and -west bands converge under the cover just south of Waterman Brook (Johnson).

Figures 1 and 2. Outcrop of quartz-sericite-chlorite-magnetite-albite schist in Terrill Hill band. West side of Terrill Hill (Morristown). The white lamina and lenticles are quartz.

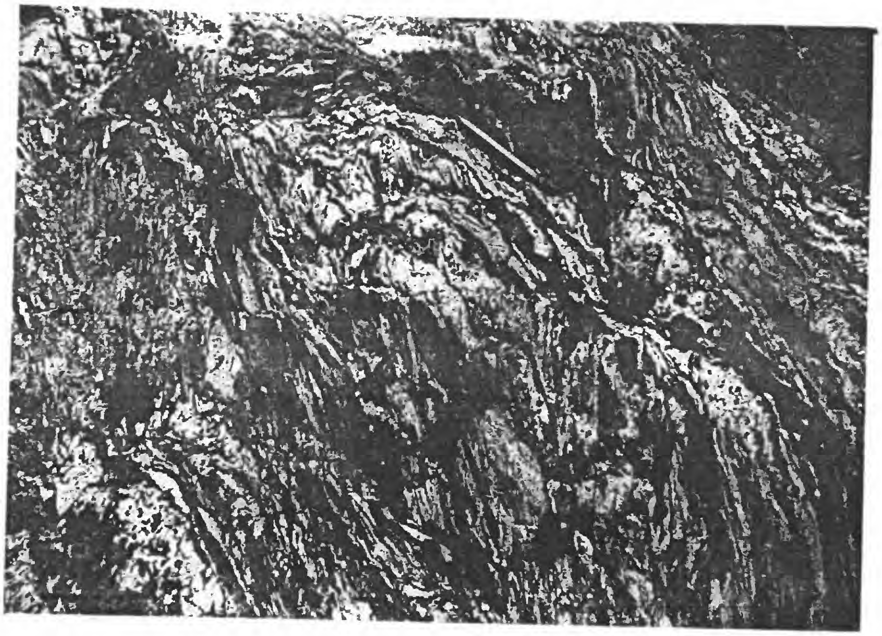


Figure 1

Figure 2

Plate 7

The quartzitic laminae and quartz lenticles are intensely deformed in most of these bands. This deformation is particularly well shown on the west side of Ferrill Hill and in the Lamoille River at Johnson. In these areas the laminae and lenticles of quartz are tightly folded, are commonly thicker in the crests of folds, and are commonly intensely sheared. These shear surfaces (shown on the map as schistosity which transects bedding) are the dominant breakage planes and are commonly subparallel to the long limbs of the folds. Horizontal sections of individual laminae or lenticles are shaped like an elongate "S" or scroll. The folds plunge steeply both north and south and their movement sense is both dextral (offsets to the right in plan) and sinistral (offsets to the left in plan). The overall trend of the bedding is commonly indeterminate. The traces of the laminae and lenticles on the schistosity surface weather out to form bands of quartz parallel to the axes of the steeply plunging folds. The thickened crests and axes of these steep folds and the quartz "bands" are folded about subhorizontal axes that trend a little east of north, parallel to the axis of the Green Mountain anticlinorium (see Figure 2).

Slip cleavage is in places extremely well developed in the micaceous layers, but only poorly developed in the more quartzose layers. In the micaceous layers it is commonly the dominant surface of breakage, and rarely completely obscures the earlier foliation. The pattern of the early compositional layering in such micaceous layers is preserved only by differences in the distribution of quartz and clinozoisite-epidote. Sericite and chlorite are oriented parallel to the slip cleavage and to the axial planes of crinkles in an earlier schistosity.

The pattern formed by these bands, the lithologic similarity, and the lack of sedimentary or structural evidence for the direction of tops of beds suggest that any one of the quartz-sericite-chlorite-magnetite-albite schist bands could be a synclinal outlier of the Morrisville band. On the other hand, minor differences in lithologic characteristics and the apparent gradation of the Terrill Hill band into micaceous quartzite to the south suggests that the bands reflect sedimentary facies variations. The relative movement parallel to the bedding foliation indicated by minor folds on both the east and west margins of the Caper Hill-west band in Waterman Brook (Johnson) and in the Lamoille River is the same, and is consistent with a position on the east limb of an anticline. The bands of quartz-sericite-chlorite-magnetite-albite schist are shown as sedimentary facies of the Camels Hump group on the structure sections, but it is possible that one or more may be one of the other bands repeated by folding.

Greenstone.

Thin bands of greenstone were noted at numerous localities within the Camels Hump group, but only two bands were extensive enough to show on the map. The largest of these is southwest of Bowen Mountain (Eden), where fine-grained, light-green tremolitic greenstone crops out in a band about 6,000 feet long from north to south and about 3,000 feet wide; the actual thickness is much less. This greenstone is probably part of the Belvidere Mountain amphibolite repeated in the Foot Brook syncline. The other greenstone mapped is a band of albite-magnetite greenstone from 10 to 75 feet wide, which was traced for about one mile both north and south of Waterman Brook (Johnson).

Several small masses of highly albitic greenstone were noted in the albite schist and gneiss west of the Foot Brook area. These masses seem to be discontinuous, and are not confined to a single stratigraphic zone. They are characterized by a high content of albite porphyroblasts (Table 3, column 19). Numerous such zones, which range from lenses a few feet wide to bands several hundred feet wide, and which do not form mappable units, crop out in the brooks that flow east and northeast from the Sterling Range.

Belvidere Mountain amphibolite.

The name Belvidere Mountain amphibolite was introduced by Keith and Bain (1932, p. 174) for the rocks exposed on Belvidere Mountain (3.6 miles N. 27° W. of the Knob, Eden). It is here used as the uppermost formation of the Camels Hump group, underlying the Ottauquechee formation composed of black phyllite and pebbly quartzite. On

Belvidere Mountain the amphibolite is probably 700 to 1,000 feet thick. It thins to less than 100 feet north of Belvidere Mountain, but was traced 20 miles north across the Canadian boundary. The Belvidere Mountain amphibolite is poorly exposed for about 2½ miles along the strike in the northeastern part of the Hyde Park quadrangle; but its possible extension farther south is concealed by surficial deposits. The amphibolite fingers out laterally southward into quartz-sericite-chlorite schist and is absent south of the area of surficial deposits.

The amphibolite on Belvidere Mountain is very coarse grained with large blades of dark-green hornblende, porphyroblasts of dark-red almandite garnet, and segregations of epidote. Within the Hyde Park quadrangle it consists of a fine-grained, light-green, albite-epidote amphibolite. Selected modes are given in Table 3. Major mineral constituents are epidote, albite, chlorite, and actinolitic hornblende; minor constituents are magnetite and biotite. The parallel orientation of amphibole and chlorite grains give the rock a rude schistosity and compositional layers parallel the schistosity.

The amphibolite is believed to have formed from a water-laid mafic volcanic detritus. Beds of graphitic quartz-sericite-albite schist as thin as 6 inches are present within the amphibolite. In one outcrop two quartzite beds, each three-fourths of an inch thick, extend for 20 feet parallel to the foliation. The map shows the interbedding and abrupt transition along strike from amphibolite into graphitic quartz-albite-sericite and nongraphitic quartz-sericite-chlorite-albite schist of the Camels Hump group. Where the interbeds of schist in the transition zones are very thin, both the schist and the amphibolite are highly calcareous. Calcareous amphibolite is common at the north edge of the quadrangle, and graphitic amphibolite occurs half a mile farther north on the lower east slopes of Belvidere Mountain.

Ottauquechee formation

<u>General statement</u>.

The Ottauquechee formation, first described by Perry (1929, p. 27) in the Ottauquechee River valley in southern Vermont, is characterized by massive dark-gray quartzite and black graphitic quartz-sericite phyllite and schist. It has been traced northward to the Hyde Park quadrangle by Currier and Jahns (1941), Brace (1953), Osberg (1952), and Cady (1956). Cady, A. H. Chidester, and the author have traced the Ottauquechee north and across the International Boundary into the "Mansonville slate" of Clark (1934, p. 11) in Quebec.

<u>Distribution</u>.

The Ottauquechee formation crops out in a north-trending band about one mile wide in the southern two-thirds of the quadrangle and in a band extending from the northeast corner of the quadrangle into the Upper Diggings (Eden). The latter band, which is 0.3 - 0.5 mile wide in the Upper Diggings, ends a little northeast of Davis Hill (Hyde Park). For several miles the two bands of Ottauquechee are parallel and about two miles apart. These bands are probably connected through folds that plunge southward beneath surficial deposits in the area between North Hyde Park and Eden Mills. The Ottauquechee formation also appears in several doubly plunging anticlines within the Stowe formation east of the main belt of the Ottauquechee, and within a syncline, the Foot Brook syncline, in the Foot Brook area west of the main belt of the Ottauquechee formation.

Description.

Selected modes are given in Table 4. The Ottauquechee formation is characterized by black graphitic quartz-sericite phyllite and schist and by massive beds of dark-gray quartzite.

The phyllite commonly contains pyrite and weathers rusty. The albite content is low and porphyroblasts of albite are rare. Alternation of thin layers of white quartz with layers of graphite and sericite in flakes oriented parallel to the layering produce a schistosity, probably a bedding schistosity. A more prominent later schistosity cuts across the earlier one, and in many places totally obliterates it.

The quartzite is dark gray and composed almost entirely of quartz with a little disseminated graphite; rarely detrital grains can be discerned. The quartzite commonly contains abundant randomly oriented veins of white quartz. Quartzite with thin, closely spaced partings of sericite and chlorite, and sericite-quartz-chlorite phyllite also occur. The phyllite contains thin beds of limestone in a few places.

Many outcrops of the Ottauquechee formation cannot be distinguished from those of the Camels Hump group, but in general, the Camels Hump contains more quartzite with micaceous partings, less black phyllite, and

TABLE 4. Estimated modes of the Ottauquechee formation.

	23	24	25	26	27
Quartz	96	22	50	* 60	** 79
Albite				5	1
Sericite		70	40	21	10
Chlorite			6	10	
Graphite	3	8	2		
Carbonate			2		
Sphene				2	2
Rutile	T			T	
Tourmaline	T				
Limonite	1			2	3
Location:***	Montpelier quadrangle MM-3.88, 3.74	Montpelier quadrangle MM-3.88, 3.74	SC-3.11, 4.40	MB-3.15, 2.86	MB-2.45, 6.77

* Detrital grains as much as 1 millimeter in diameter.
** Detrital grains of quartz, quartzite (<1%), and albite (<1%) as much as 3 millimeters in diameter set in a finer matrix.
*** Explanation on page 5.
23. Massive dark-gray quartzite.
-25. Graphitic quartz-sericite-chlorite schist.
-27. Quartz-sericite schist with coarse detrital grains.

its schist appears more quartzose and more albitic than that of the Ottauquechee. The Ottauquechee on the west side of the Foot Brook syncline is readily distinguished from the coarser grained rocks of the Camels Hump, but on the east side of the syncline it is difficult to distinguish one from the other because of the similarity of the Ottauquechee and the fine-grained rocks of the Camels Hump group and because of the paucity of outcrop.

The Ottauquechee formation of the northeastern band, north of the Upper Diggings includes light-gray quartzite and quartz-sericite schist contain detrital grains of granule and pebble size. Beds of the massive dark-gray quartzite crop out north of the Hyde Park quadrangle. The detrital grains, which are as large as 6 millimeters, are predominantly quartz, but some are black shale and feldspar; the grains constitute as much as 45 percent of the rock. Those of quartz are subrounded to subangular and are undeformed. The coarser sizes are in beds with a higher pebble content. The matrix consists of quartz, sericite, and chlorite, with iron-bearing carbonate weathering out. North of the Hyde Park quadrangle a yellowish-brown quartz-sericite schist with only a small percentage of granules is common; it is interbedded with graphitic phyllite. This lithologic type is not abundant within the Hyde Park quadrangle.

Rocks of the northeastern band of Ottauquechee are well exposed west of South Pond (Eden) and on a hill 0.4 mile south-southeast of Eden Mills. Rocks of the main band are well exposed northwest of Blake Cemetery and east of Ryder Brook (Morristown). The contact of the Ottauquechee formation with the underlying Belvidere Mountain amphibolite is exposed on the west side of a hill 1.15 miles northwest of Eden Mills. The contact between the Ottauquechee and the Stowe formations is exposed at several places between Beaver Meadow Brook (Hyde Park) and a small hill 0.25 mile northwest of Blake Cemetery (Hyde Park) and in numerous places within the Upper Diggings. The Stowe formation immediately above the contact contains a few beds of graphitic phyllite and schist that differ from those of the Ottauquechee only in the lack of associated beds of quartzite.

Stowe formation

General statement.

The Stowe formation (Cady, 1956) comprises "schists and thick interbeds of greenstone and amphibolite, which are typically developed in the southeastern part of Stowe township and adjoining areas." The name Stowe formation replaces the Bethel schist of Richardson (1924, p. 82-83) because the name "Bethel" was preoccupied.

The Stowe formation is characterized by the predominance of a silvery-green, sericite-quartz-chlorite schist with numerous lenticles of granular white quartz parallel to the schistosity (Plate 8). Bedded amphibolite or greenstone are common throughout the Stowe, and range from a foot to a thousand feet or more in thickness. Although schists above and below the large mass of amphibolite shown on the map are similar, it is convenient to describe them separately because the effects of middle-grade metamorphism are confined almost entirely to the upper schist. Estimated modes are given in Tables 5, 6, and 7.

Figure 1. Outcrop of typical sericite-quartz-chlorite schist of the Stowe formation. Note the abundance of quartz segregations. About 400 feet southwest of southwest corner of Collins Pond (Hyde Park).

Figure 2. Typical outcrop of Stowe formation in the Foot Brook area, 1.20 miles S. 65° W. of the site of Davis Neighborhood School (Johnson).

34-17
Plate - 8

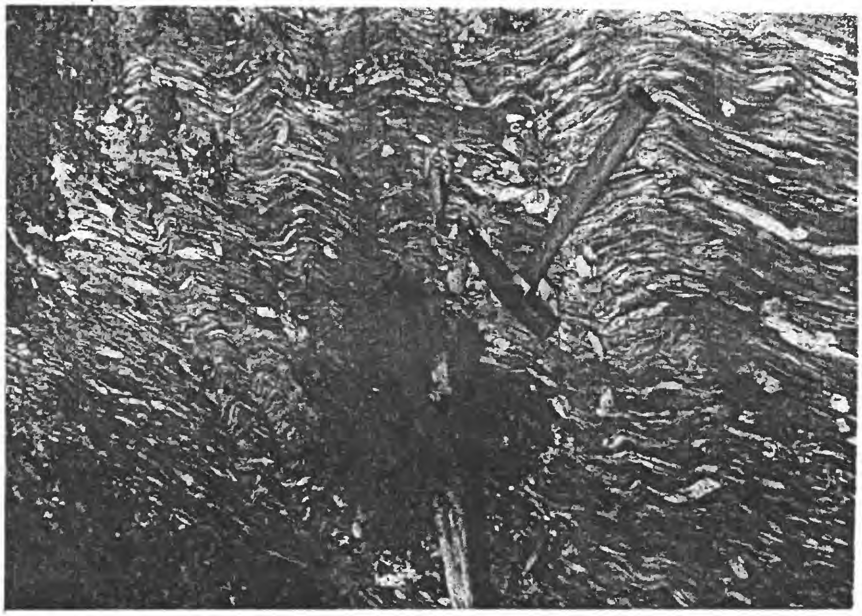


Figure 1

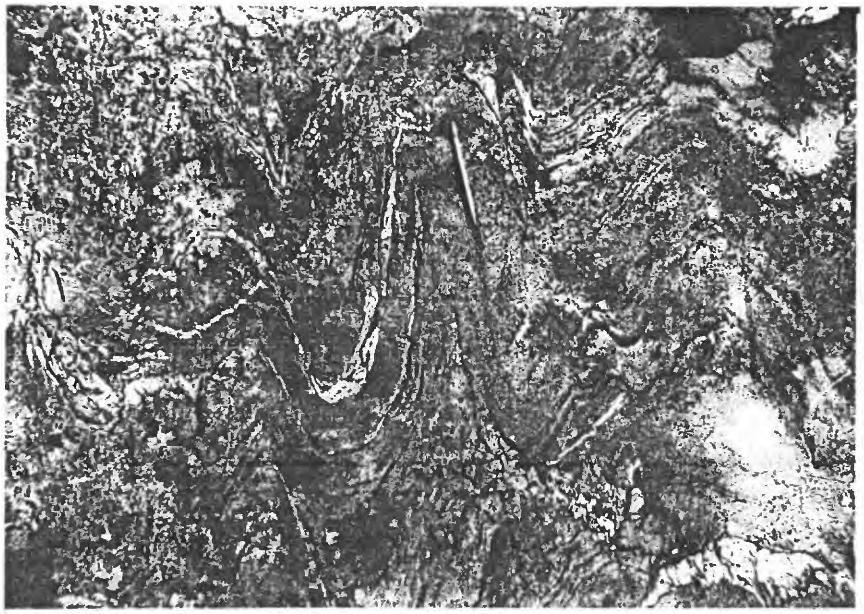


Figure 2

Plate 8

TABLE 5. Estimated modes of the Stowe formation - chlorite zone.

	35	36	37	38	39	40	41	42
Quartz	15	12	50	40	20	25	35	30
Muscovite	44	23	14	5	13	50	38	32
Paragonite			16				12	8
Chlorite	40	40	13	10		5	8	22
Chloritoid					15	14	2	4
Albite		12						
Magnetite	T	8	3					1
Carbonate		2	<1	35	50			
Limonite				10	2	3		<1
Epidote group	T	2	1					
Ilmenite	1							
Rutile			1			3	5	~2
Pyrite					<1			
Tourmaline	T		1			T	T	
Apatite			<1				<1	<1
Location:*	NW-0.77, 5.45	SE-2.32, 3.14	SC-2.35, 1.35	SC-3.76, 1.70	Hardwick quad NW-0.1, 0.3	NE-3.30, 0.45	SC-3.60, 1.88	NC-2.45, 4.75

* Explanation on page 5.
35-37. Silvery-green sericite-quartz-chlorite-albite schist.
38. Ankeritic quartz-sericite-chlorite schist.
39. Ankeritic quartz-sericite-schist with chloritoid porphyroblasts.
40-42. Quartz-sericite-chlorite schist with chloritoid porphyroblasts. No. 42 from Foot-Brook syncline.

Table 6. Estimated modes of the greenstone and amphibolite of the Stowe formation.

	28	29	30	31	32	33	34
Albite	**	55	30	45	14	5	10
Chlorite	25	19	10	20			<1
Actinolite*			15*	20*			
Hornblende*					40*	30*	55*
Epidote group	25	<1	50	12	40	50	30
Biotite				T			
Quartz	37**	7	3		2		
Carbonate	10	25	2	1	<1		
Magnetite			<1		3	15	2
Rutile					1	T	
Sphene	3			2			3
Pyrite	<1	<T	<1	<1			
Limonite		1	T				
Location:***	NE-3.35-4.68	NE-3.27, 1.46	SC-3.76, 1.70	NE-3.66, 0.49	SE-1.17, 4.50	SE-2.27, 0.48	SE-3.20, 1.85

*Hornblende if $Z \wedge c = 25°$.
**Quartz plus albite; quartz predominant.
***Explanation on page 5.
28-29. Ankeritic greenstone from northern part of quadrangle.
30-31. Epidote-actinolite-albite-chlorite greenstone.
32-34. Epidote-albite-quartz amphibolite.

Table 7. Estimated modes of the Stowe formation - kyanite and garnet zones.

	43	44	45	46	47	48	49	50
Quartz	30	68	25		5			35
Albite		2						
Muscovite	50	18	38	45	60		17	55
Paragonite	T		12			57		
Chlorite	10	8	16	20	20		8	5**
Chloritoid				5	7			
Almandite	2R!	T	5R!		R!	35R	15R	2
Kyanite				15R	1R	4R	50R	*
Magnetite	<1	2			7			3
Ilmenite			3	15		4	10	
Rutile	1							
Sphene	<1				<1			
Epidote group	5	1						
Tourmaline	T	1	T	T	T			
Apatite	<1	T	<1		T		T	T
Location:***	SC-3.17, 0.91	SE-0.79, 4.31	SC-3.60, 1.88	SE-2.17, 0.15	SE-2.41, 0.83	SE-1.61, 1.77	SE-2.30, 0.40	SE-1.74, 2.11

R! Completely retrograded.
R Retrograded with relics left.
* Present in hand specimen, not in slide.
** Some biotite present.
*** Explanation on page 5.
43-45. Stowe formation, garnet zone.
46-50. Stowe formation, kyanite zone.

North of the vicinity of Zack Woods Pond (Hyde Park) the Stowe formation is overlain to the east by the Umbrella Hill formation, consisting of conglomerate and slate. Farther south where the Umbrella Hill formation is absent the Stowe grades upward into the Moretown formation by a decrease in the number of quartz lenticles and by the addition of the characteristic "pinstripe" schistosity of the Moretown. This gradational zone, several hundred feet across, is exposed about half a mile south of the Lamoille River.

Lower schist.

The dominant rock beneath the main body of amphibolite is a fine-grained and thinly layered, silver-green sericite-quartz-chlorite-albite schist. This rock is well exposed on Bean Mountain (Eden). The sericite content is generally greater than the quartz content if the quartz lenticles and veins are excluded; if they are included the sericite and chlorite together generally exceed the quartz. The albite, commonly fine grained, is generally less than 5 percent of the rock. Magnetite octahedra are common. Accessory minerals include rutile, carbonate, ilmenite, sphene, limonite, clinozoisite-epidote, tourmaline, and apatite. The schist contains quartz lenticles and laminae oriented parallel to the foliation as well as numerous quartz pods and veins. Bedding is rarely preserved in this rock, and the lenticles, laminae, pods, and veins of quartz are believed to be segregations of quartz derived from the surrounding rocks. The segregations may in part have formed about originally quartz-rich beds;

they would thus parallel bedding, but would not actually be bedding. The dominant schistosity is a bedding schistosity in places, as both studies of thin sections and field observations have shown, but elsewhere the dominant schistosity cuts an earlier foliation. Some porphyroblasts of albite intersect this late schistosity and hence formed after it.

Fine-grained, light-colored epidote-actinolite-albite-chlorite greenstone, brown-weathering calcareous greenstone, and thin beds of graphitic phyllite are interbedded in the schist. Locally the schist weathers brown and contains as much as 15 percent of limonite which seems to have altered from ankerite.

East of the northeastern band of the Ottauquechee formation in the Upper Diggings a silver-green quartz-sericite-chlorite slate or phyllite is abundant. Bedding is poorly preserved and the slaty cleavage apparently generally cuts across the bedding. The slate or phyllite contains porphyroblasts of limonite apparently pseudomorphic after magnetite, and locally weathers brown owing to the limonite and ankerite content. A few beds contain as much as 15 percent of tabular porphyroblasts of chloritoid.

The basal part of the Stowe formation is well exposed north of Blake Cemetery (Hyde Park) and south of the site of North Randolph School (Morristown). It is made up chiefly of sericite-quartz phyllite, similar to that in the Upper Diggings, and of black graphitic quartz-sericite phyllite similar to that of the Ottauquechee formation. Ankeritic sericite-quartz phyllite, ankeritic greenstone,

greenstone, and beds of impure dolomite are common. Impure dolomite, ankeritic quartz-sericite phyllite, graphitic phyllite, and ankeritic greenstone are abundant from Cleveland Corners (Hyde Park) to about half a mile south of the Lamoille River. They are especially well exposed in Rodman Brook.

Amphibolite and greenstone.

The main mass of amphibolite south of the Lamoille River and west of Elmore Mountain is composed of a fine-grained, greenish-black, epidote-albite-quartz amphibolite with hornblende needles barely visible to the unaided eye. Magnetite is an abundant accessory mineral, and sphene and rutile are common. Selected estimated modes are presented in Table 6, columns 30-34. Although the amphibolite appears massive, parallel orientation of hornblende needles and some compositional layering, particularly of epidote and carbonate, produce a poorly developed foliation. However, the rock does not split readily along this foliation. Thin quartz lenticles parallel to the foliation are fairly common, and one thin section of the amphibolite contains 25 percent of quartz. Small pegmatitic aggregates of albite, hornblende, epidote, and quartz were probably derived from the surrounding rock during metamorphism. The origin of large irregular patches with a high content of epidote is not known, but may be the result of original differences in composition rather than of metamorphic segregation. This amphibolite is well exposed at numerous points along the Elmore Mountain road.

Most of the amphibolite in the large area west and southwest of Elmore Pond (Elmore) is correlated with the main mass of amphibolite west of Elmore Mountain. It is similar to the main mass, though somewhat coarser grained. Amphibolite beds as much as 100 feet thick are rather common in the upper schist of the Stowe on Elmore Mountain and south of Elmore Pond. Some of this area is within the kyanite zone, and the amphibolite is coarse grained with stubby hornblende blades as much as 4 millimeters across. Garnet porphyroblasts as much as 10 millimeters in diameter were noted at a contact between amphibolite and schist. Pegmatitic aggregates of albite, hornblende, and epidote are more abundant, coarser grained, and larger than in the main amphibolite mass.

North of the Lamoille River the amphibolite is a fine-grained, light-green albite-actinolite-epidote-chlorite-carbonate greenstone. Table 6, columns 28-29, contains selected, estimated modes of this rock. Actinolite needles are fine grained, and in places are not apparent to the unaided eye. The carbonate content is higher than in the amphibolite, and there are numerous interbeds of highly ankeritic greenstone and of ankeritic, graphitic quartz-sericite phyllite. These rock types are well exposed above and below the Garfield Dam, 0.7 mile northeast of Davis Hill (elevation 1,140 on Green River, Hyde Park).

The protolith, or premetamorphic parent, of the amphibolite and greenstone was probably water-laid mafic volcanic detritus adulterated with other sedimentary material, and the compositional layering probably represents bedding. The compositional layering is commonly parallel to bedding in adjacent schist, to the contact with the schist, and to thin schist interbeds within the greenstone or amphibolite. Even where the dominant schistosity in the schist is not parallel to its bedding on Elmore Mountain, the foliation in the amphibolite commonly parallels the bedding.

Upper schist.

North of Hyde Pond (Hyde Park) the rock above the main mass of amphibolite is similar to the lower schist. Yellowish-gray weathering, fine-grained, and thinly layered sericite-quartz-chlorite-albite schist predominates. This grades into brown-weathering, ankeritic sericite-quartz phyllite, many beds of which contain abundant porphyroblasts of chloritoid. Albite porphyroblasts are locally abundant, but do not seem to be as widespread as in the lower schist. Albite and chloritoid have not been noted together. Numerous interbeds consist of rusty-weathering, patchily graphitic and ankeritic, sericite-quartz phyllite with porphyroblasts of pyrite. Thin quartz lenticles locally resemble pebbles. Actinolitic greenstone and greenstone high in carbonate are common.

The schist north of Hyde Pond (Hyde Park) is fine-grained and contains chlorite and chloritoid, but no garnet. South of Hyde Pond the schist is much coarser grained, and at the south border of the quadrangle muscovite plates are as much as 10 millimeters across; garnet dodecahedrons reach 8 millimeters. Garnet and chlorite aggregates, pseudomorphic after garnet, occur a little south of Hyde Pond and became more abundant to the south. Kyanite occurs from the north end of Elmore Mountain south into the Montpelier quadrangle.

The mineral assemblages noted in the schist of the Stowe formation are included in Table 11. Muscovite and paragonite were identified by X-ray methods. Paragonite was found in rocks containing chlorite and chloritoid, but not in those containing garnet and kyanite. In the latter the muscovite probably contains a substantial amount of sodium.

In the chlorite- and chloritoid-bearing rocks magnetite and carbonate are the abundant accessory minerals, and epidote-clinozoisite, rutile, ilmenite, limonite, tourmaline, and apatite are common. In the garnet- and kyanite-bearing rocks magnetite, rutile, and ilmenite, are abundant; sphene, tourmaline, apatite, and allanite are common.

The mineralogy of these rocks is complicated by the partial alteration of kyanite, garnet, and biotite to minerals characteristic of a lower metamorphic grade. Thus, kyanite is partly replaced by fine-grained muscovite (identified by X-ray methods) and chloritoid, and garnet by chlorite and magnetite. The relict textures, the pseudomorphic replacement of kyanite by muscovite rather than pyrophyllite, and the known pro-grade mineral assemblages in adjacent areas suggest that the mineral assemblage almandite-kyanite-biotite-muscovite-quartz existed in the schist before the retrograde alteration.

In the kyanite zone partly altered porphyroblasts of garnet and kyanite respectively form as much as 35 percent and 50 percent of the rock. Lenticles and irregular shaped masses of coarsely crystalline quartz are abundant. These seem to have been segregated from the surrounding rocks, and thin sections from localities in the kyanite zone where quartz lenticles and masses are especially abundant show that the adjacent rock commonly contains little or no quartz. The schistosity is poor owing to the coarseness of grain. Bedding is rarely preserved in the coarse-grained schist on Elmore Mountain, although north of the Lamoille River compositional layering commonly indicates the bedding. However, the attitude of bedding is indicated in many places by the contact between coarse-grained amphibolite and schist interbeds.

Schist in the Foot Brook area.

The rock of the Stowe formation exposed in a syncline in the Foot Brook area, the Foot Brook syncline, is predominantly fine-grained quartz-sericite-chlorite schist with interbeds of graphitic quartz-sericite-chlorite schist and phyllite. Greenstone and impure calcite marble also occur. The quartz-sericite-chlorite schist has a high sericite content (both muscovite and paragonite) and a shiny weathering silver-green appearance (Plate 8). It is generally non-albitic, but thin beds containing porphyroblasts of albite are present. Fine-grained chloritoid porphyroblasts occur locally throughout the Stowe formation in this area. Magnetite octahedra are very abundant; some beds contain as much as 5 percent magnetite with octahedra as much as 15 millimeters across. Porphyroblasts of pyrite occur in some beds and are abundant in the graphitic interbeds. Tourmaline, apatite, rutile, and ilmenite are common accessory minerals. The schist contains numerous thin quartz lenses, and commonly has well-developed slip cleavage and crinkles. Quartzite beds are absent in both the quartz-sericite-chlorite schist and the graphitic schist.

The graphitic schist and phyllite is highly micaceous, thinly foliated, and appears to contain very little albite. The greenstone beds are highly albitic and have a marked layering caused by differences in albite content. A thin section from the band of greenstone shown on the map contains more than 75 percent albite. Both the quartz-sericite-chlorite schist and the graphitic phyllite are well exposed in the cliffs 0.55 mile east of the site of Riverside School (Johnson) and in the upper tributaries of Foot Brook north of the site of Davis Neighborhood School (Johnson).

This area of quartz-sericite-chlorite schist is correlated with the Stowe formation on the basis of general lithologic similarity (including the common appearance of chloritoid and paragonite). Also the bordering zone of non-albitic black phyllite and quartzite which separates the quartz-sericite-chlorite schist from the albitic schist and gneiss of the Camels Hump group closely resembles the Ottauquechee formation, and the normal sequence--Camels Hump, Ottauquechee, and Stowe--is suggested. The quartz-sericite-chlorite-magnetite-albite schist bands of the adjoining Camels Hump group are readily distinguishable from the rocks of the Stowe formation in the Foot Brook area because they contain numerous micaceous quartzite beds and have a higher quartz and albite content.

Umbrella Hill formation

Conglomerate and interbedded slates, which are typically exposed on and near Umbrella Hill (Hyde Park) along the east-central border of the quadrangle are here designated the Umbrella Hill formation. A small

Figure 1. Umbrella Hill conglomerate. Slate fragments outlined by chalk; the white fragments are quartz, 0.15 mile S. of height of land on Eden-Craftsbury road, Hardwick quadrangle.

Figure 2. Relation of slaty beds in Umbrella Hill conglomerate to schistosity. Contact between slate and conglomerate outlined with chalk. West side of Umbrella Hill (Hyde Park).

45-A
Plate 9

Figure 1

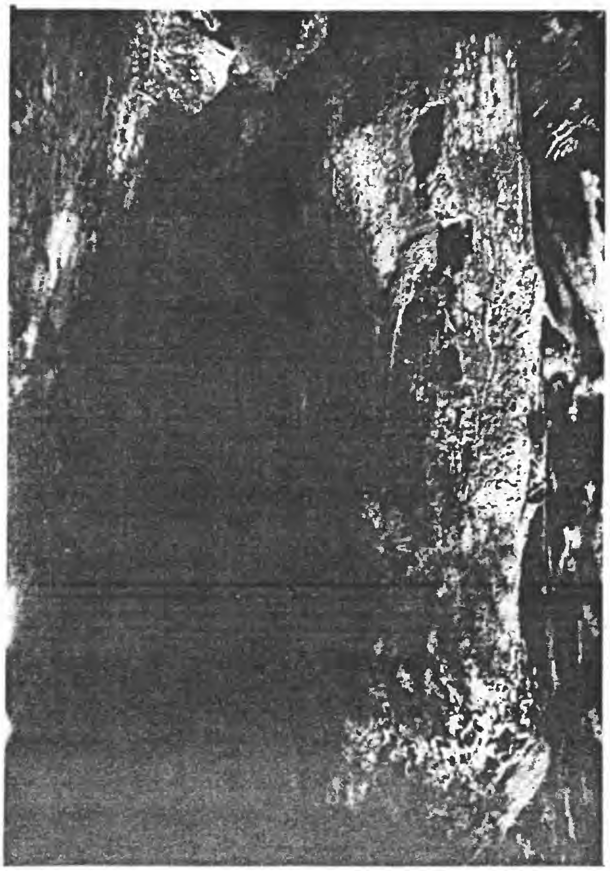


Figure 2

Plate 9

mass of the conglomerate, which is isolated in a tight syncline downfolded in the Stowe formation on the west side of Umbrella Hill, rests with angular unconformity upon amphibolite beds within the Stowe. The unconformity may be seen at several places. At one of these, on the west side of the isolated mass, vertical beds of slate within the conglomerate strike north. Twenty feet west is an outcrop of amphibolite, the compositional layering and foliation of which strike N. 50° E. and dip 44° NW.; this attitude is general over a considerable area. The Umbrella Hill formation is overlain, apparently conformably, by the Moretown formation. The conglomerate has not been traced south of the vicinity of Zack Woods Pond. In the southern part of the Hyde Park quadrangle and in the Montpelier quadrangle (Cady, 1956) the Stowe formation grades upward into the Moretown formation without apparent stratigraphic break. The Umbrella Hill formation is readily accessible in the Hardwick quadrangle, 2.55 miles southeast of Eden Mills, at the height of land on the road from Eden Mills to Craftsbury. W. M. Cady and the author have traced the conglomerate about 19 miles to the north-northeast from the point where it leaves the east side of the Hyde Park quadrangle. It apparently thins to extinction both to the north and south. In several places north of the Hyde Park quadrangle interbeds of greenstone are present within the Umbrella Hill formation, and the conglomerate thins to extinction in greenstone northward.

Bands of conglomerate at the east and west borders of the formation are separated by a calcareous grey slate. The slate is poorly exposed, but there are several outcrops a little north of Collins Pond (Hyde

Park). The eastern band of conglomerate thins southward to extinction just east of Collins Pond. A little south of Collins Pond 100 feet of conglomerate are exposed in the western band, but 0.75 mile farther north it is only 4 feet wide.

Subrounded white quartz pebbles are most abundant in the conglomerate, but red, yellow, and blue-gray slate fragments are common (Plate 9, figure 1). The largest pebbles in most outcrops range from 2 to 3 inches in diameter, but quartz pebbles as much as 5 inches in diameter have been noted. The pebbles are undeformed. The matrix is composed of quartz, sericite (both muscovite and paragonite in some specimens) and fine-grained opaque minerals; it ranges in color from blue-gray to light tan. Porphyroblastic plates of greenish-black chloritoid, 1 to 3 millimeters in diameter, are abundant in both the matrix and the slate pebbles. The overall chloritoid content seldom exceeds 5 percent, but much of this is concentrated in certain slate pebbles and in thin zones in the matrix. Estimated modes are given in Table 8.

Thin beds of greenish-yellow sericite slate with as much as 30 percent chloritoid porphyroblasts are interlaminated with the conglomerate. Slaty cleavage cuts across a color layering and across thin pebbly beds in similar slate in the isolated mass of the conglomerate on the west slope of Umbrella Hill (Plate 9, figure 2).

TABLE 8. Estimated modes of Umbrella Hill formation and Moretown formation.

	51	52	53	54	55
Quartz	55	35	55	P*	75
Albite		7±	2		T
Sericite	35	45	25	P*	19
Chlorite		5		P*	3
Chloritoid	1				
Biotite			15		
Rutile		T		P*	T
Graphite		?			
Magnetite	9	8	3		<1
Epidote group					2
Tourmaline	T	<1	T		T
Apatite		T			T
Limonite				P*	
Location:**	MC-4.05, 3.30	MC-3.80, 3.33	Hardwick quad. SW-0.6, 5.1	Hardwick quad. MC-0.5, 3.2	Montpelier quad. ME-1.26, 0.96

* Too fine grained to estimate mode.
** Explanation on page 5.
51-52. Matrix of Umbrella Hill formation - quartz and albite are in part unrecrystallized detrital grains. Quartz and slate fragments are present in hand specimens. Paragonite was found in 2 of 6 specimens.
53-55. Granulite from Moretown formation.

Moretown formation

The Moretown formation (Cady, 1956) comprises quartz-albite-sericite-chlorite granulite, quartzite, phyllite, greenstone, and slate; it is typically exposed in eastern Moretown township in the Montpelier quadrangle. In the Hyde Park quadrangle it includes quartz-sericite-chlorite-albite-epidote granulite with numerous close-spaced micaceous partings (the "pinstripe"), micaceous quartzite, and sericite slate, all rather poorly exposed. Estimated modes are given in Table 8. These rocks are all light green and locally contain porphyroblasts of biotite (0.5 millimeter across). Several beds of calcareous greenstone were noted. The Moretown overlies both the Umbrella Hill conglomerate and the Stowe formation conformably, and, as already pointed out, grades upward from the latter by interlamination. Only the lower part of the Moretown formation appears within the Hyde Park quadrangle.

Intrusive rocks

Serpentinite, talc-carbonate rock, and steatite

Bodies of serpentinite or its metamorphic derivatives, talc-carbonate rock and steatite, occur at fifteen localities known to the author. They are commonly elongate parallel to the foliation of the enclosing rocks, and most dip very steeply. Serpentinite is dark green on a fresh surface, but weathers to pale greenish-white or buff. Talc-carbonate rock is mottled greenish-gray and weathers brown. The carbonate, predominantly magnesite, ranges in content from about 25 to 45 percent. Steatite is white or green, weathers grayish tan, and is nearly pure talc. Smaller bodies may be completely altered to talc-carbonate rock and steatite, but where in larger bodies the alteration is incomplete talc-carbonate rock and steatite form successive outer zones. The adjoining country rock is altered to chlorite schist called "blackwall."

Serpentinite, or its derivatives, or both were noted at the following localities; those with possible economic interest will be described in a later section.

1) 1.15 miles N. 6° E. of the Knob (Eden).

2) 2.20 miles N. 48° W. of the Knob (Eden).

3) 0.50 mile N. 40° E. of the summit of Bowen Mountain (Eden) on Green Mountain Club Trail.

4-9) Within area of greenstone shown on map southwest of Bowen Mountain (Eden), three exposed in brooks on map, three exposed on ridges.

10-11) Talc Mine area (Johnson).

12) 1.20 miles N. 57° W. of Christy School (Johnson).

13) 0.39 mile S. 53° W. of bridge over Lamoille River (Johnson village).

14) 0.27 mile S. 58° W. of Red Bridge (Morristown).

15) 1.6 miles N. 30° W. of the Knob (Eden).

Lamprophyre dikes

Two dikes were noted within the Hyde Park quadrangle and one a little west of the quadrangle. The dikes are unmetamorphosed, unlike the rocks they intrude, and are commonly intruded into joint systems that are transverse to the regional structure. Similar dikes are found throughout the state. Estimated modes of dikes from the Hyde Park and Montpelier quadrangles are given in Table 9.

The rock is black and aphanitic. Phenocrysts of hornblende, magnetite octahedra, and aggregates of analcite and carbonate occur in a fine-grained groundmass of plagioclase feldspar and brown hornblende. The amphibole seems to be an oxyhornblende near the common hornblende end of the series, which indicates that these dikes should be classified as spessartites.

TABLE 9. Estimated modes of post-metamorphic dikes

	56	57	58	59
Hornblende	37	25	5	30
Biotite	10		10	
Chlorite				
Augite			10	
Plagioclase	40 (An55)	60 (An5)	55 (An45)	50
Magnetite	10	10	10	5
Analcite	3			
Carbonate		5	10	15
Location:*	C-0.65, 5.23	SW-1.35, 5.35	Montpelier quad. EC-0.70, 0.66	Montpelier quad. EC-3.0, 5.6

* Explanation on page 5.

56-57. Post-metamorphic dikes from Hyde Park quadrangle.

58-59. Post-metamorphic dikes from Montpelier quadrangle.

STRUCTURE

Major structural features

The Green Mountain anticlinorium is the major structural feature in this area. Its axis crosses the extreme northwest corner of the Hyde Park quadrangle in a north-northeasterly direction, but farther south, a little west of the area, it trends more northerly. The bedding dips gently east near the crest of the anticlinorium, but steepens farther eastward on the east limb. The general east dip is reversed by a number of subsidiary anticlines (see cross sections) that trend subparallel to the main anticlinorial axis. There is no evidence for major faulting in the Hyde Park quadrangle.

The general areal pattern of the various mapping units indicates broad, shallow open folds, but, as discussed in the section on minor structural features, many of the minor structures fail to reflect the attitude of the axial planes or the open nature of the larger folds. The cross sections are inferred primarily from the general areal pattern.

Northwest and north of Johnson village part of the Ottauquechee and Stowe formations appear in the Foot Brook syncline where they are downfolded in the Camels Hump group, west of their main belts of exposure. These formations crop out in the Foot Brook syncline over an area about 11 miles long from north to south and about three miles wide; they extend northward into the Jay Peak quadrangle.

An anticline must be present east of Johnson village between the Foot Brook syncline and the main belt of exposure of the Ottauquechee formation, but its axis could not be located. The discontinuity of the bands of quartz-sericite-chlorite-magnetite-albite schist in the central part of the quadrangle are shown on the cross-sections as the result of sedimentary facies changes, but, as discussed previously, some of them may be isolated, downfolded synclines.

East and northeast of Centerville (Hyde Park) and in the Upper Diggings several folds are shown by the pattern formed by the Ottauquechee-Stowe contact. These folds just east of Centerville are doubly plunging, whereas those in the Upper Diggings plunge gently southward. Other folds are shown by the pattern of the amphibolite unit south of Garfield, and by minor structures southeast of Garfield. In the vicinity of Elmore Mountain several folds are shown by the pattern of the amphibolite units. All of these rather open folds have nearly horizontal, north-south trending axes. Similar folds are found in the Montpelier quadrangle and even farther south, as well as north of the Hyde Park quadrangle, along the trend of the Northfield, Worchester, and Lowell Mountains.

Minor structural features

Introduction

The outcrop pattern of the formation boundaries and a knowledge of the stratigraphic sequence commonly provides sufficient evidence to outline the major structural features. However, the minor structural features are also of great importance in understanding the nature of the major folds and the sequence of tectonic events. These minor structures include planar and linear features, which may be observed megascopically, and textural features, such as rotated porphyroblasts and fractured grains, which may be readily observed microscopically. Preferred lattice orientation and preferred dimensional orientation so slight that it may be recognized only by statistical analysis are not included here. Both the character and the spatial relations of minor structures are used in the final interpretation.

Megascopic directional properties of a rock may be outlined as follows:

Planar features (S planes)

1) Preferred planar dimensional orientation of elongate, platy, or ellipsoidal grains.
2) Nonequant aggregates of grains, that is, a layering or lamination.
3) Fractures, independent of preferred planar dimensional orientation of grains or grain aggregates.
4) Fractures and joints.

Linear features (lineation)

1) Preferred linear dimensional orientation of elongate grains.
2) Elongate aggregates of grains.
3) Lines of intersection of planar features.
4) Fold axes.

The following features are commonly readily observable microscopically:

1) Rotation of grains marked by helical planes of inclusions or by the orientation of pressure shadows around porphyroblasts (Mugge, 1930, p. 32, Fairbairn, 1949, p. 52-57).
2) Fractures and strain shadows in grains.

3) Granulated grain boundaries.

Planar structures include foliation, fracture cleavage, slip cleavage, and joints. Foliation describes a parallelism of the planar and linear fabric elements of the rock, which enables it to part along nearly parallel planes. These fabric elements include aggregates of grains in lenses, pods, laminae, and layers as well as individual nonequant mineral grains. This usage of foliation corresponds rather closely to that of Mead (1940, p. 1009), Fairbairn, (1949, p. 5), Turner (1951, p. 559), Williams, Turner, and Gilbert (1954, p. 169), Billings (1954, p. 336), and Kemp and Grout (1940, p. 222). Primary foliation includes bedding fissility in sedimentary rocks and flow planes or layers in igneous rocks; secondary foliation includes secondary layering or lamination and schistosity, as they are defined here. The angular relationship of secondary foliation to primary lamination or foliation may be expressed by the terms, bedding or parallel, and transverse. Layering or lamination, which may be primary or secondary, describes a parallelism of layers distinguished by their composition, texture, or color. Schistosity, though also used for a specific type of foliation, describes the parallel preferred orientation of nonequant minerals. The perfection of a schistosity, i.e., the extent to which nonequant minerals in parallel preferred orientation make up or characterize the rock, depends upon both the amount of nonequant minerals and the degree of preferred planar orientation of these minerals. The term flow cleavage has been used in much the same sense by Leith (1905, p. 12) and Cloos (1937, p. 61), but others (Fairbairn, 1949, p. 239) have raised objections to this term.

Several terms are used to describe foliation characterized by certain limited ranges of relative development of schistosity and lamination.

Slaty cleavage characterizes the planar structure of a fine-grained rock with good schistosity, but no or poor lamination.

Schistosity characterizes the planar structure of a fine- to medium-grained rock with good schistosity and poor to fair lamination.

Gneissic foliation (gneissosity) characterizes the planar structure of a medium- to coarse-grained rock with poor to fair schistosity and fair to good lamination.

The decreasing perfection of slaty, schistose, and gneissic foliation is due to larger grain size and to a larger proportion of equant as compared to nonequant minerals.

Cleavage, in accord with its etymological derivation, describes a cleaving or splitting—an actual parting surface. Fracture cleavage, in accord with common usage (Billings, ; Leith, 1923, p. 148) is essentially a close-spaced fracture or "near-fracture", which is independent of preferred planar orientation of nonequant minerals.

Slip cleavage (Dale, 1896, p. 560) was first used to describe small microfaults that develop on the limbs of small crinkles, but its usage has gradually been extended to a variety of transitional structures. Some slip cleavage is merely the striking "herring bone" symmetry shown by tight crinkles in highly micaceous rock aligned along parallel axial planes; no surface of discontinuity may be seen although the rock will commonly part. As the crinkles become tighter actual surfaces of shear appear along the thinned and aligned fold limbs of the folds. Commonly the first-formed surfaces are discontinuous, passing from actual surfaces to tight crinkles and back to actual surfaces of discontinuity (Plate 13, figure 2). In this stage the slip cleavage could actually be called fracture cleavage except that, as pointed out by White (1949, p. 592), it is questionable that the slip cleavage is entirely independent of mineral orientation. With further development platy minerals are present in the surfaces, oriented parallel to them. Some schistosity and slaty cleavage consists of such surfaces, closely spaced, and containing abundant platy minerals oriented parallel to the surfaces; the surfaces are separated by crests of crinkles in the earlier foliation. Slip cleavage is dependent upon the existence of an earlier foliation for recognition; were this to be obliterated close-spaced slip cleavage would be described as a slaty or schistose foliation rather than a slip cleavage. Such a transition has been described by White (1949, p. 591).

Thin layering and sufficient platy minerals so that slippage can occur parallel to the surfaces of the layers seem to be essential for the formation of close-spaced slip cleavage. It may be absent in granoblastic layers of a schist and yet be extensively developed in adjacent micaceous layers. In fine-grained rocks rich in micaceous minerals the individual surfaces are smooth and continuous, but in coarser-grained rock they are quite irregular and discontinuous.

Folded planar structures are a very useful structural element because the pattern and spatial relations of a fold indicate the relative movement at the point of observation. The attitude of an imaginary plane tangent to the crests of a series of folds gives the average dip and the trend of folded planar features. The fold patterns of folds with limbs of unequal length are distinguished as <u>dextral</u> when they offset to the right in plan and <u>sinistral</u> when they offset to the left. A fold is <u>neutral</u> when its limbs are of equal length.

Differences in the geometry of folds appear to be closely related to differences in lithology and hence to differences in competency (see for example, Dale, 1896, p. 564; Thompson, 1950, p. 94; Osberg, 1952, p. 79; Grace, 1953, p. 94). Thick-layered, nonmicaceous units such as quartzite form open folds of larger wave length than do more micaceous, thinly laye ed schistose units. The schistose units have small tight folds and may have a transverse schistosity parallel to the axial plane of the more ope folds in the quartzite. Moreover, folded micaceous beds have changed sha more than have adjacent quartzite beds; the limbs generally being thinned with respect to the crest of the folds.

Character and spatial relations of minor structures

<u>General statement</u>.

Planar features shown on the map (Plate 2) include bedding, various types of schistosity, and slip cleavage; fracture cleavage, slaty cleavag and joints, though not mapped, are present locally. Linear features show on the map include quartz rods, fold and crinkle axes, and the intersections of certain planar features; mineral alignment, streaks on the folia tion surfaces, and elongate quartz lenses, though not shown on the map, a common.

Within the Hyde Park quadrangle linear and planar structures with different spatial relations are superimposed. Their time relationships are clearly shown by their superimposition in many outcrops, particularly near the axis of the Green Mountain anticlinorium. Hand specimens may be found in which the axial plane, axis, and limbs of an earlier fold are clearly refolded about a fold axis nearly at right angles to the earlier fold axis. The later fold axes trend nearly north-south, subparallel to the axis of the Green Mountain anticlinorium. Throughout much of the area the two sets of folds may be distinguished by inference from their orientation. It is convenient, and common practice, when numerous minor structures are described to indicate the planar structures as S_1, S_2, S_3, ... and the linear features as L_1, L_2, L_3, ..., numbering them in the probable order of formation.

In the Hyde Park quadrangle the minor structures seem to have formed in the following order:

S_1 Bedding.

S_2 Secondary bedding foliation and secondary compositional layering.

L_1 Lineations nearly at right angles to the axis of the Green Mountain anticlinorium; includes fold and crinkle axes; quartz rods, elongate quartz lenses, mineral alignment, intersection of S_2 and S_3 planes, streaking on foliation surface.

S_3 Schistosity and fracture cleavage and slip cleavage which transects the crests of L_1 folds.

L_2 Lineations nearly parallel to the axis of the Green Mountain anticlinorium; includes fold and crinkle axes, intersections of S_4 planes with S_3 planes.

S_4 Slip cleavage, slaty cleavage, schistosity, fracture cleavage (?).

L_3(?) Near vertical fold axes and intersection of S_2(?) planes with a fracture cleavage, S_5(?).

S_5(?) Fracture cleavage.

S_6 Joints.

S_1 and S_2 planar features.

Definite bedding (S_1) in the form of quartzite beds, calcareous beds, and thick layers of distinctive composition may be seen throughout the Hyde Park quadrangle (Plate 10, figure 1). Thinner compositional and textural layering can usually be attributed to a primary origin in this area. However, such layering has probably been accentuated during metamorphism and deformation by such processes as Eskola's "concretion principle" (Turner, 1948, p. 139). Local migration of material into layers parallel to the foliation seems to be responsible for the thin conformable quartz lenticles (Plate 10, figure 2).

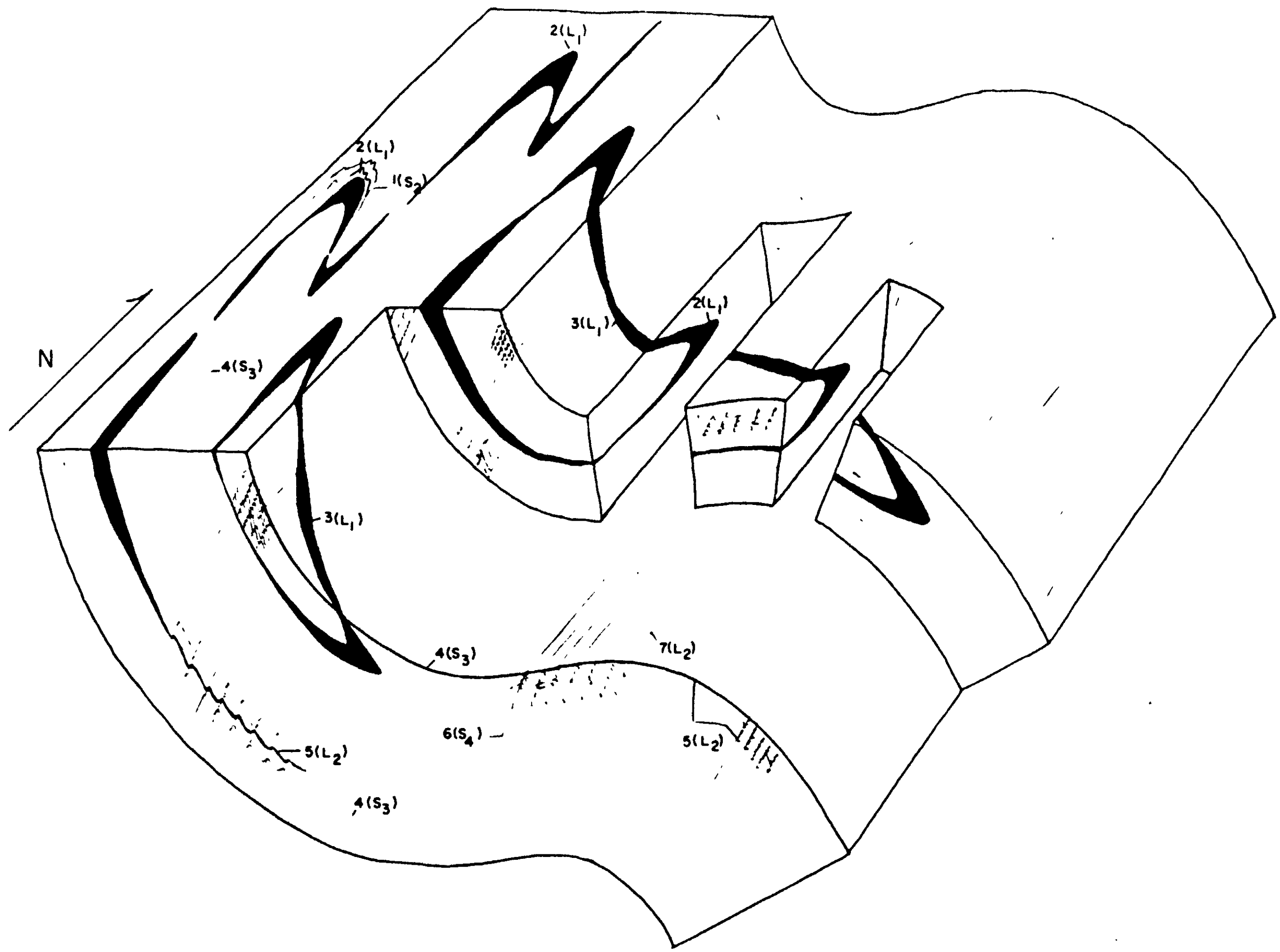
Throughout the area both bedding and compositional layering parallel a schistosity (S_2) (Plate 10, figure 1; Plate 11, figure 2). The perfection of this bedding schistosity depends both upon the grain size of the rock and upon the proportion of granular and micaceous minerals. Gneiss and quartzite have a poorer schistosity than schist, although

Explanation to Figure 2.

1. Schistosity (S_2) parallel to quartzose beds (S_1) and quartz segregations (shown in black).
2. Folded quartzose beds and quartz segregations; fold axes (L_1) plunge nearly down dip. "U" or "S"-shaped rods are formed by shearing or stretching out of fold limbs.
3. A lineation (L_1) formed by the intersection of quartzose beds and quartz segregations with the S_3 schistosity surface.
4. A schistosity (S_3) which transects crests of L_1 folds and is subparallel to the long limbs. The S_3 schistosity forms the surface of the outcrop because of the ease with which the schistosity parts.
5. Open folds and small crinkles (L_2) which fold S_2 and S_3 schistosities and L_1 fold axes.
6. A slip cleavage (S_4) associated with small crinkles (L_2) and parallel to the axial planes of the open folds (L_2).
7. A lineation (L_2) formed by the intersection of slip cleavage (S_4) and schistosity (S_3).

61-A

Figure 2



Diagramatic representation of minor structural features west of the Foot Brook syncline

Figure 1. Interbedded micaceous quartzite and schist in Camels Hump group. The schistosity (S_2) is parallel to the bedding (S_1).

Figure 2. Quartz segregations in quartz-sericite-chlorite-magnetite-albite schist, top of Camels Hump group, west side of Terrill Hill (Morristown).

61-B

Plate-10

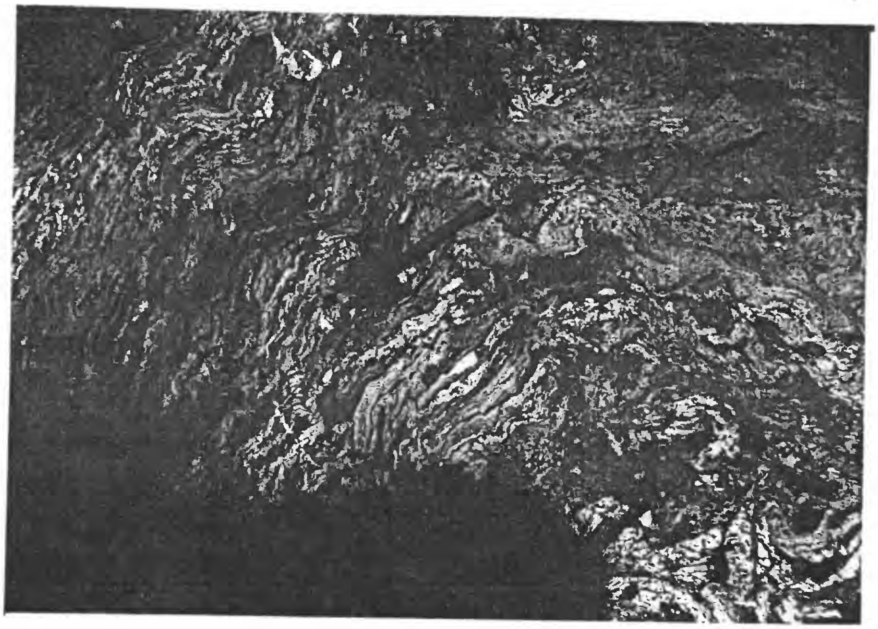


Figure 1

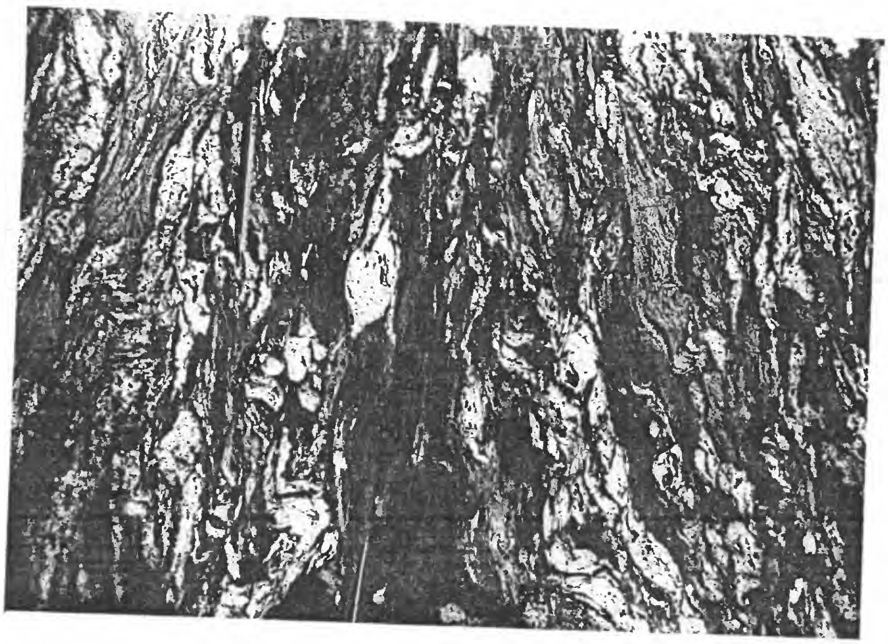


Figure 2

Plate 10

Figure 1. Interbedded micaceous quartzite and graphitic schist in Ottauquechee formation, 0.50 mile S. 10° W. of Centerville (Hyde Park). The folds (L_1) plunge steeply south, away from the camera. The quartzite bed near the hammer has been rotated and its ends stretched or sheared off.

Figure 2. Detail from Figure 1. Cleavage (S_3) transects bedding; it is parallel to schistosity (S_3) in adjacent schist. Schistosity (S_2), parallel to the quartzite beds, may be seen at top and bottom of the photograph.

61 - C

Plate - 11

Figure 1

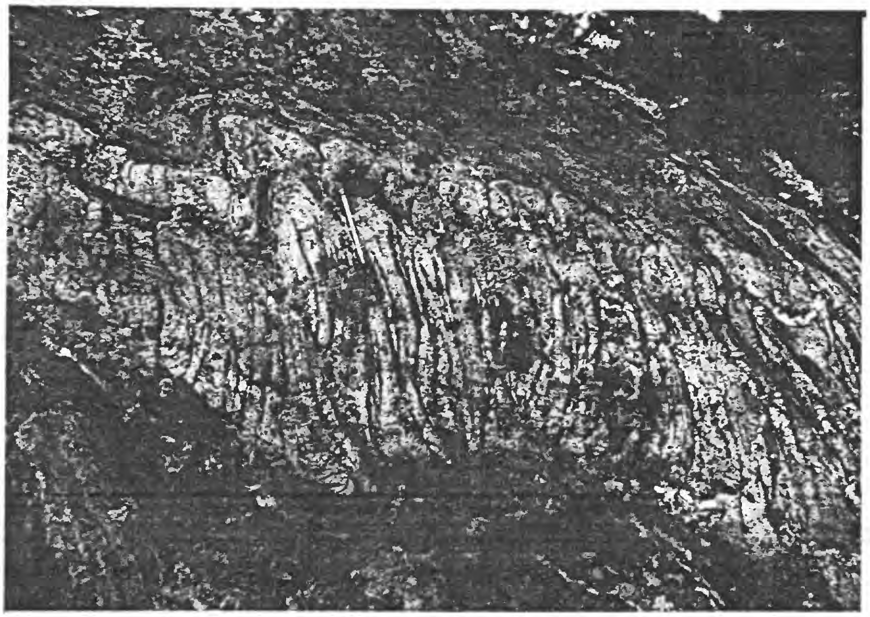


Figure 2

Plate 11

Figure 1. Another part of outcrop shown in Figure 1, Plate 11. Bedding (S_1) and bedding schistosity (S_2) transected by cleavage (S_3). Folds (L_1) plunge steeply south away from the camera.

Figure 2. Detail of Figure 1.

61-D.

Plate 12

Figure 1

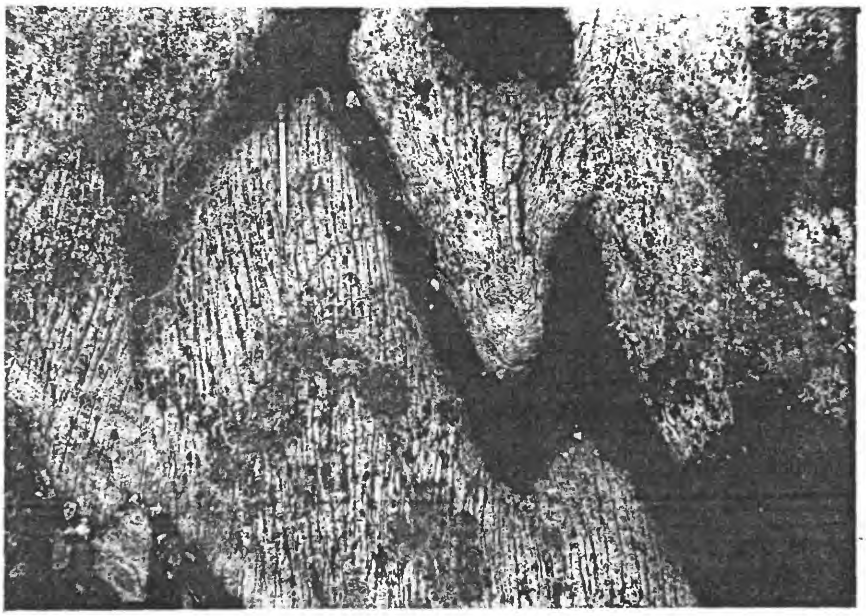


Figure 2

Plate 12

Figure 1. Close-spaced cleavage (S_3) parallel to axial planes of L_1 folds. Bedding schistosity (S_2) is nearly obliterated. Stowe formation, 1.20 miles S. 65° W. of the site of Davis Neighborhood School (Johnson).

Figure 2. S_3 planar structures transecting bedding foliation (S_1 and S_2). In various parts of the photograph the S_3 planar structure could be called a slip cleavage, a fracture cleavage, or a schistosity. Note the transition from aligned crinkles into a well-developed planar structure.

61-E

Plate-13

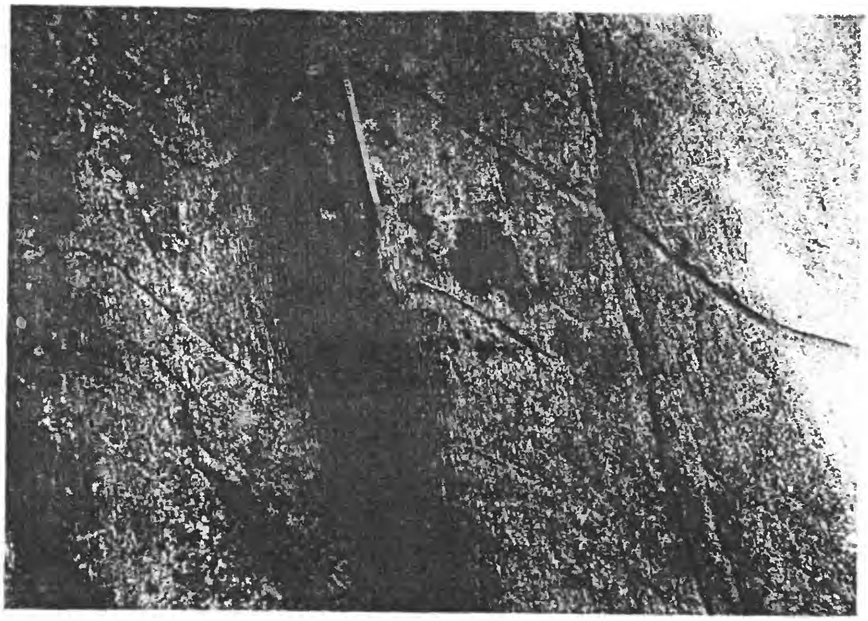


Figure 1

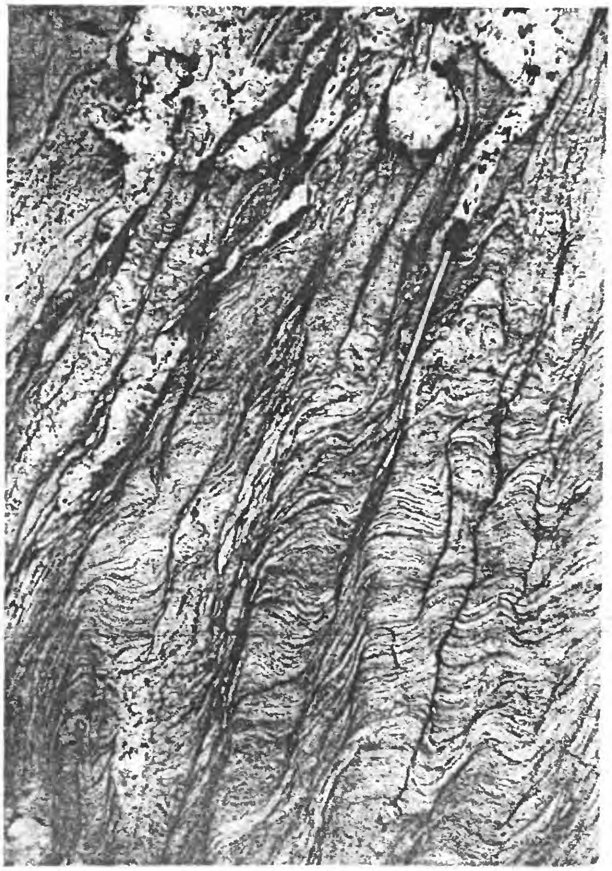


Figure 2

Plate 13

Figure 1. Schistosity (S_3) on left side of photograph in micaceous beds is parallel to cleavage (S_3) transecting the bedding foliation (S_1 and S_2) of the micaceous quartzite bed in the center of the photograph. The quartzite bed is cut off at each end by S_3 schistosity. Camels Hump group, west side of Terrill Hill (Morristown).

Figure 2. Typical development of S_3 planar structures in quartzose units of the Camels Hump group. West side of Terrill Hill (Morristown).

61-F

Plate-14

Figure 1

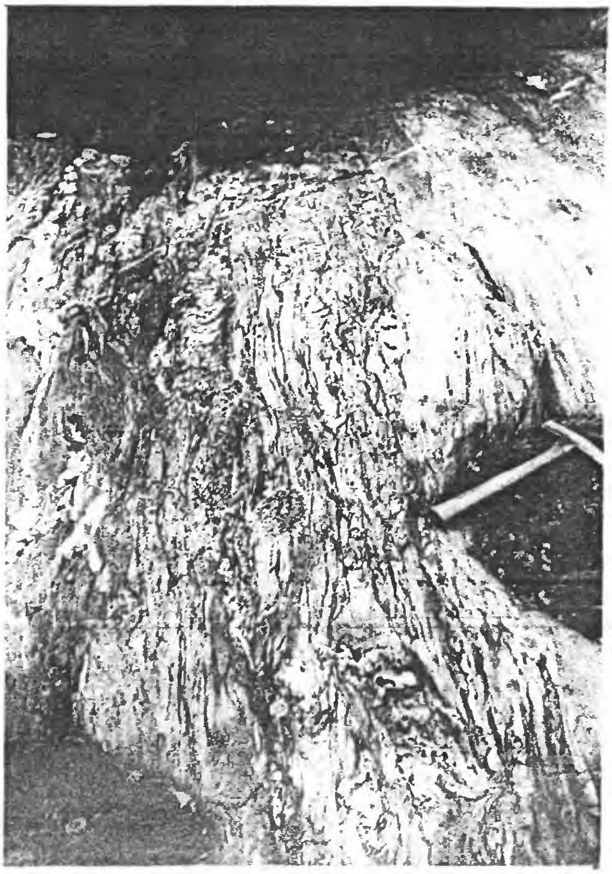


Figure 2

Plate 14

Figure 1. The light-colored quartz marks the intersection (L_1) of bedding foliation (S_1 and S_2) with a schistosity (S_3) which transects the bedding foliation and which forms the surface of the outcrop. The pencil is parallel to fine crinkles and larger folds (L_2) which deform the other structures. Camels Hump group, 0.72 mile N. 80° W. of the site of Riverside School (Johnson).

Figure 2. A fold (L_2) in quartzite extends through the specimen with its fold axis parallel to the mechanical pencil and the cigarette. The crest of the fold is transected by a cleavage (S_3) parallel to the schistosity (S_3) which forms the upper surface of the specimen. The cleavage and schistosity (S_3) and the fold (L_2) are folded about an axis parallel to the yellow pencil. Camels Hump group, 0.75 mile N. 80° W. of Riverside School site (Johnson).

61-J.
Plate - 15

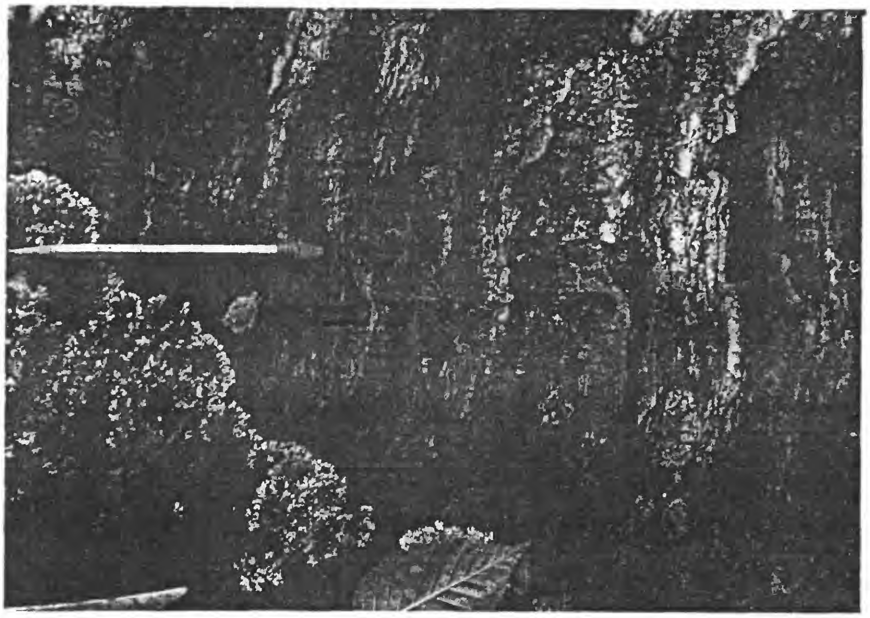


Figure 1

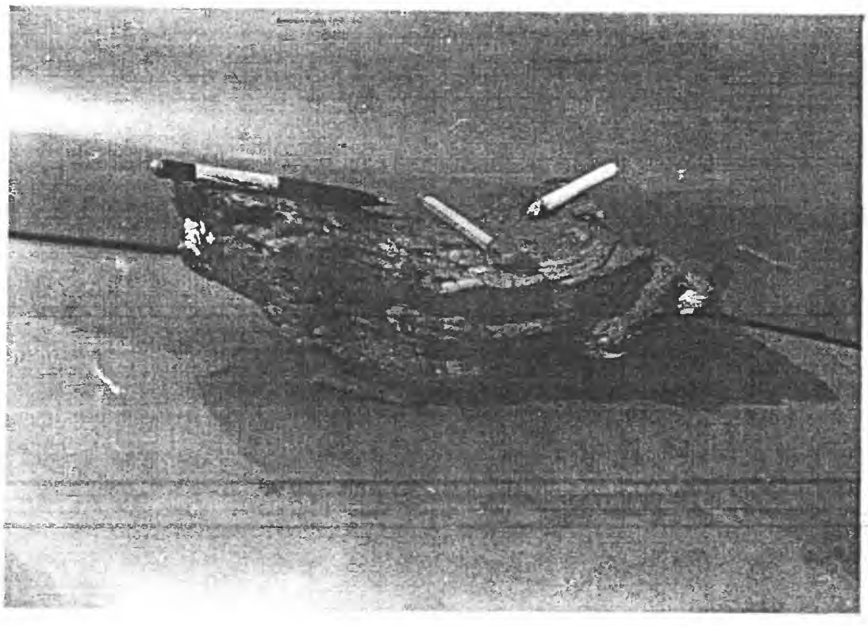


Figure 2

Plate 15

Figure 1. Dark pencil parallel to axes of L_1 folds in massive gray quartzite plunging southeast. Light pencil parallel to axes of L_2 folds in schistose beds. Camels Hump group, 0.80 mile S. 45° E. of the site of Hillside School (Johnson).

Figure 2. Cleavage (S_4), outlined by white chalk, transecting L_2 folds. The cleavage is associated with L_2 crinkles elsewhere in the outcrop. Belvidere Mountain amphibolite, 1.50 miles N. 50° E. of the site of Wescome School (Eden).

61-H
Plate - 16

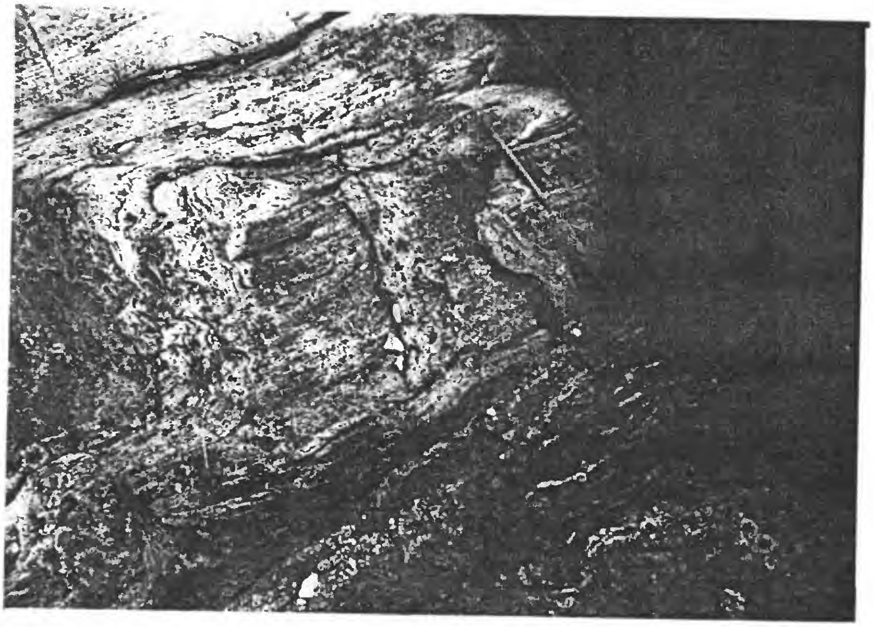


Figure 1

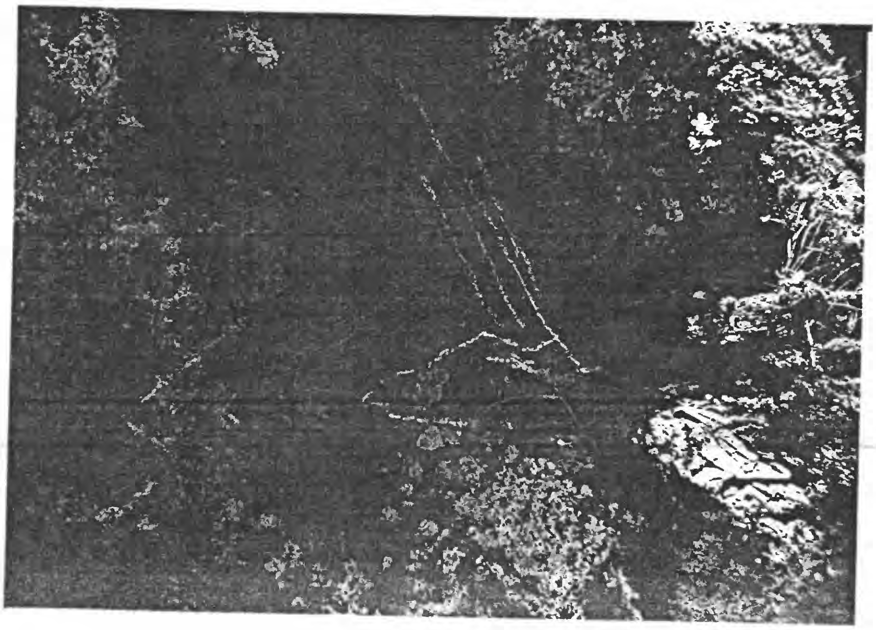


Figure 2

commonly quartzite cleaves readily due to the segregation of the small amount of micaceous minerals into thin parallel surfaces parallel to the bedding (Plate 13, figure 2). The schistosity (S_2) is locally nearly completely obscured where the S_3 or S_4 planar features are well developed (Plate 13, figure 1; Plate 14, figure 1).

L_1 linear features.

The S_1 and S_2 foliations are folded about axes nearly at right angles to the axis of the Green Mountain anticlinorium. These L_1 fold axes and linear features parallel to them, plunge east where the bedding foliation (S_1 and S_2) dips gently east (Plate 16, figure 1) and plunge nearly vertically--generally steeply south--where the dip of the bedding foliation is steep (Plate 11, figure 1). There is no consistent movement sense. In the central part of the area both dextral and sinistral folds, whose axes plunge steeply both north and south, may be observed in a single outcrop.

These folds (L_1), which are best shown by quartzite beds in the schist, are very tight with long limbs subparallel to the foliation in the surrounding schist. Thick massive quartzite beds form more open folds, whose short limbs and, less commonly, crests may be sliced up by closely spaced, thin micaceous layers along which some displacement has occurred (Plate 11, figure 2; Plate 12, figure 2). The limbs of folded quartzite beds and quartz lenticles in the schist are commonly broken--either stretched or sheared off--so that the crests of the folds are preserved as rods whose cross section is "U" or "S" shaped (Figure 2).

These rods are especially abundant and well exposed on the west side of the Hyde Park quadrangle between Whiteface Mountain and the Lamoille River. Although the L_1 folds are numerous, the trend of the long limbs of these folds corresponds more closely to the trend of the formation boundaries than does the overall trend of the schistosity and compositional layering.

Prominent lineations parallel to the L_1 fold axes are formed by the intersection of the schistosity (S_3) with the quartzose beds and lenticles (S_2), streaks on the foliation surface (S_2), and alignment of mineral grains. Quartz lenticles, which are thought to have formed by metamorphic differentiation, are also commonly elongate parallel to the L_1 fold axes.

S_3 planar features.

The S_3 planar feature can be identified only where it transects bedding foliation (S_1 and S_2) in L_1 folds (Plates 11, 12, 13, and 14). Elsewhere it parallels the bedding foliation and cannot be distinguished from it. Generally it is a schistosity or a slip cleavage, locally a fracture cleavage. In the crests of thick quartzite beds the S_3 planar structure consists of parallel fracture cleavage surfaces containing micaceous minerals oriented parallel to the surface (Plate 12, figure 2), but in adjacent micaceous beds it consists of a schistosity. The limbs of quartzose beds and laminae are sheared off or stretched out along planes which parallel the foliation surfaces; these planes commonly contain schistose layers or thin layers of platy minerals oriented parallel to the planes (Plate 13, figure 2; Plate 14). All of these somewhat different features are believed to be related, the differences being due to differences in the mechanical properties of the rocks and in the amount of shear which had to be accommodated.

<u>L_2 linear features.</u>

All of the earlier minor structures (S_1, S_2, S_3, L_1) are folded about north-trending fold and crinkle axes (L_2) which are subparallel to the axis of the Green Mountain anticlinorium (Plates 15 and 16, and Figure 2). These folds correspond to the larger folds of the area. In general, the movement sense is "east side up", corresponding to the east limb of the Green Mountain anticlinorium, but this is reversed on the west limbs of subsidiary anticlines. In the western and eastern part of the Hyde Park quadrangle these minor folds and the subsidiary larger folds are rather open flexures, whereas in the central part of the area both the minor folds and the larger folds are believed to be rather tight.

Most of the minor drag folds in the central part of the area are inferred from their spatial relations to be earlier folds (L_1). The later set of folds (L_2) in the central and eastern parts of the quadrangle is represented by fine crinkles with an associated slip cleavage, and by gently west-dipping foliation that occurs locally on the west limbs of larger anticlines.

The intersection of slip cleavage (S_4) with the earlier foliations (S_1 and S_2, except in L_1 fold crests, and S_3) forms a prominent lineation parallel to the L_2 fold axes.

Although only two distinct sets of folds are described here, it is possible that this is too simple a picture. Superimposition of the two sets is clearly shown only in the western part of the Hyde Park quadrangle. It is possible, particularly in the central part of the quadrangle where minor structural relations are not as clear, that the earlier set of folds (L_1) was preceded by folds having the same orientation as the later set of folds (L_2), even though such structures have not been recognized.

S_4 planar features.

Slip cleavage is associated with small crinkles (L_2) which are subparallel to the axis of the Green Mountain anticlinorium (Figure 2). It is best developed in the western and eastern parts of the Hyde Park quadrangle. It generally trends nearly north-south and dips steeply west, parallel to the axial planes of the larger L_2 folds. The spacing of the surfaces ranges from a fraction of a millimeter to several centimeters. The intersection of the slip cleavage (S_4) with the S_3 foliation planes consistently parallels the axes of the associated folds and crinkles and the spatial relations of the two planar features indicates the relative movement on the S_3 foliation plane (Figure 2).

In the Stowe formation slaty cleavage and schistosity are transitional into slip cleavage. In some cases the crinkle, that is commonly associated with slip cleavage, is absent or has been completely recrystal lized; in others it is present. The slaty cleavage and the schistosity parallel the axial planes of folds, where these may be observed. This slaty cleavage and schistosity transects an earlier foliation and seems to parallel nearby slip cleavage. As indicated on the cross sections these planar features seem to be subparallel to the axial planes of the larger folds. The schistosity (S_3) in the central part of the area bears a similar spatial relation to the inferred major folds, but its relations with L_1 folds and the fact that it is locally transected by slip cleavage (S_4), makes it possible to differentiate S_3 and S_4 planar features in this central area.

Fracture cleavage was noted locally in competent beds, but it is not common in the Hyde Park quadrangle. The general spatial relations indicate that it is probably related to the other S_4 planar structures, but no conclusive evidence has been found.

L_3 (?) linear features and S_5 (?) planar features.

In the Moretown formation and locally within the upper part of the Stowe formation, sigmoid flexures (L_3) and cleavage bands (S_5), that resemble small faults, offset the schistosity. Commonly the schistosity passes unbroken through the cleavage bands, but in some an actual fracture is present. These structures may be seen 0.60 mile S. 80° E. of the village of Elmore and along the shore of Great Pond (Hyde Park). The pattern is generally dextral, and the cleavage band commonly strikes northwest and dips vertically. These seem to have formed later than the two sets of folds described previously, but no conclusive evidence has been found. The same general movement sense—rocks to the east moving south relative to those to the west—is widespread in the Montpelier quadrangle (Cady, 1956).

Joints.

No systematic study of the joints was undertaken as they are younger than the major deformation and the metamorphism, and thus seemed to offer little help in understanding the deformation. Prominent joint systems were noted along the crest of the Green Mountain anticlinorium and upon the crest of Elmore Mountain (Elmore). The general trends of these were northeast and northwest, dipping nearly vertically.

Rotated porphyroblasts.

Albite porphyroblasts commonly show an S-shaped alignment of inclusions in thin sections, which is assumed to indicate a rotation coincident with growth (Mugge, 1930, p. 32). The arrangement of these lines indicates the rotational sense. The pattern can be seen only in thin sections oriented about normal to the axis of rotation, which is parallel to the fold axes. The inclusions of clinozoisite-epidote, graphite, rutile, sericite, chlorite, ilmenite, and quartz are fine grained and show preferred planar orientation. It is presumed that this preferred orientation was originally part of the schistosity, and that the grains were included as the porphyroblast grew. The radius of curvature reflects the velocity of shear relative to the rate of growth of the porphyroblast. Lack of curvature indicates growth without rotation. Inclusions aligned in a straight line at an angle to the schistosity indicate rotation after growth had ceased.

The inclusions indicate that a crinkled schistosity was present before the porphyroblast formed. Some porphyroblasts enclosed very tight crinkles with an associated slip cleavage. In some specimens the crinkles are so tight that they must have formed when the rock was finer grained than at present. Commonly the porphyroblasts were rotated after completion of growth, but indication of growth during rotation and of lack of rotation have also been observed. A single thin section contains albite porphyroblasts with the following four different paragenetic sequences: 1) growth of porphyroblast, followed by rotation with growth continuing during rotation and with no further rotation after cessation of growth; 2) growth of porphyroblast, followed by rotation without further growth, followed by growth without further rotation; 3) growth of porphyroblast followed by rotation accompanied by growth slow enough to exclude inclusions; and 4) growth of porphyroblast including small crinkles followed by rotation without further growth. Such combinations suggest that no simple picture of the movement involved can be formed, although no systematic investigation of the movement sense was undertaken.

<u>Fabric</u>.

Certain features of the fabric of these rocks support the idea of "post-tectonic recrystallization" after gliding, fracture, and grain rotation has ceased. Undeformed tabular porphyroblasts of chloritoid are oriented at a large angle to the schistosity, and garnet and kyanite porphyroblasts contain very few fractures. Inclusions in garnet indicate that no rotation occurred, and kyanite blades do not have a preferred orientation. Evidence of rotated albite porphyroblasts has been cited previously.

A mosaic pattern characterizes the fabric of quartzites and of quartz-rich layers in schist. It consists of intersections of grain boundary at about 120 degrees between three grains. Wide variation in grain size, variable extinction, and evidence of cataclastic structure are rare in quartz, and twinning laminae are rare in carbonate and albite grains. It is believed that these features were destroyed by annealing or recrystallization which continued beyond deformation.

Retrograde textures are abundant in the garnet and kyanite zones in the Stowe formation. Garnet is altered to pseudomorphic aggregates of fine-grained chlorite and magnetite with relics of garnet; kyanite is altered to pseudomorphic aggregates of fine-grained muscovite and coarse plates of muscovite.

Relation of minor structural features to major structural features

All the minor structural features are compatible in a general way with the major structures in the Hyde Park quadrangle. However, while the second stage folds (L_2) and slip cleavage (S_4) aided the synthesis of the major structures, the earlier folds (L_1) did not prove useful in this synthesis. In much of the area the second stage folds (L_2) are uncommon, and the observed folds and foliations seem to bear little direct relation to the major structures. This difficulty is illustrated diagrammatically on the cross sections (Plate 1).

The easterly dip of the bedding schistosity increases easterly from the axis of the Green Mountain anticlinorium to the exposures of the

Ottauquechee formation in the Foot Brook syncline. In this area broad rolls, crinkles, and slip cleavage, with nearly horizontal linear elements closely parallel to the axis of the Green Mountain anticlinorium, indicate that younger beds are to the east—drag sense is east side up relative to west side. The early stage of deformation is represented by quartz rods and sparse tightly folded quartzite beds with axes subperpendicular to the axis of the Green Mountain anticlinorium. The relationship of various minor structures near the axis of the Green Mountain anticlinorium are shown in Figure 2.

The prominent minor structural feature in the Stowe formation exposed in the Foot Brook syncline is a slip cleavage and slaty cleavage (Plate 1, section A-A'). The bedding schistosity commonly trends northeast to east; the fold axes commonly plunge steeply south. However, the syncline, as interpreted from the map pattern, must be rather shallow with a nearly horizontal axis. The attitudes of quartzite beds and of the tremolitic amphibolite unit indicates that the east limb of the syncline is overturned to the west. Minor structures in the Camels Hump group south of the Lamoille River give little indication of the southward extension of the axis of the Foot Brook syncline.

Similarly, the axis of the anticlinorium between the Foot Brook syncline and the Ottauquechee formation in the eastern homoclinal sequence could not be located. Folded quartzite beds are abundant in the Camels Hump group in this central area, but second stage minor features are uncommon. The long limbs of folds and a transverse schistosity commonly trend northerly and dip steeply east. The bedding trend ranges widely, but the fold axes generally plunge steeply south. In the narrow quartz-sericite-chlorite-magnetite schist band at the top of the Camels Hump group the long limbs of folds and the transverse schistosity are more nearly parallel to the trend of the band than the bedding trend.

As shown in section A-A' a steep schistosity is the dominant minor structure in the Ottauquechee and Stowe formations on the east side of the quadrangle. In some parts of the Stowe formation (Upper Diggings) this approaches a slaty cleavage, which parallels a slip cleavage (S_4) and associated gently plunging crinkles (L_2). Southeast and east of Garfield (Hyde Park) major fold axes were located by the relation of these structures to the earlier schistosity. In the Ottauquechee and in part of the Stowe formation the steep schistosity, a S_3 planar structure, is believed to be earlier than the slip cleavage. Throughout the Stowe formation the attitudes of the greenstones and amphibolites conformed to the major structures. Such attitudes were particularly useful in the vicinity of Elmore Mountain (Elmore) where the schist of the kyanite zone is intensely deformed (Plate 1, section B-B').

Origin of minor structures

Several explanations seem possible at first for the occurrence, one upon the other, of two sets of folds with the movement sense at large angles to each other. The later set of folds, parallel to the axis of the Green Mountain anticlinorium, is clearly related to the folding of the Green Mountain anticlinorium. One might try to explain the folds subperpendicular to the axis of the Green Mountain anticlinorium by either the action of a major horizontal shearing couple younger than the north-south axis or by earlier folding about east-trending axes that were later folded with the anticlinorium. The first explanation would require a more consistent pattern of shear sense and a more consistent orientation of the fold axes than is shown in this area, and also that the strata were nearly vertical before rotation in the horizontal plane took place. It would not explain the gently east plunging fold axes found near the crest of the anticlinorium. The second explanation would require compression at right angles to that responsible for the Green Mountain anticlinorium. Moreover, folding which could produce the numerous tight minor folds of the earlier set seen in this quadrangle might be expected to affect markedly the pattern of the map, producing wide departures from the rather simple northerly trend of the formations. The earlier folds were not formed penecontemporaneous with sedimentation inasmuch as quartz lenses of metamorphic origin are similarly folded.

In some areas minor folds are found in thrust fault zones with axes parallel to the thrust movement. These are believed to be formed by differential and diverging movements of individual rock masses within the fault zone (Balk, 1936, p. 738-739; Fairbairn, 1949, p. 222). Balk (1952, p. 418) indicates that folds parallel to the direction of movement associated with thrusts are small "with wave length measured in inches rather than in feet" and that most of them are nearly isoclinal. As a result of detailed structural studies relating the minor structures to thrust blocks in the Bergsdalen quadrangle Kvale (1947, p. 205) concluded: "in metamorphic rocks the lineation which represents the direction of least resistance may under certain conditions be formed at any angle with the principle direction of movement of the rocks."

Lengthening perpendicular to the major compressive force may produce plunging folds, salients, or recesses. The development of such features is accompanied by components of extension parallel to the major fold axes in a given rock layer (Billings, 1942, p. 234-239; Cloos, 1946, p. 26-29). This component of extension may cause folds or crinkles at a large angle to the major fold axes--and subparallel to the direction of movement. These folds may form by the actual buckling of a rock layer with greater competency than the surrounding rock (Kuenen, 1938; Godfrey, 1954) or by the drag produced by differential extension in adjacent rock layers. Cloos (1946, p. 27) indicates that this process should be particularly effective in incompetent rocks and that the folds produced should be of a small order of magnitude in comparison with folds resulting from the major flowage direction.

Irregularities in the "basement rocks" or the presence of intrusive rock masses might also be responsible for deviations from the major flowage direction on the limb of a fold or might cause shortening or extension perpendicular to the major compressive forces.

Brace (1953, p. 68-72) has described folds similar to those in the Hyde Park quadrangle parallel to a strong streaming and to cobble elongation and subperpendicular to major fold axes on the east flank of the Green Mountain anticlinorium in the Rutland quadrangle, Vermont. He considered that they developed subperpendicular to the major fold axes during the formation of the Green Mountain anticlinorium. "Whereas the dominant movement during the deformation was in an east-west direction ... resistant masses in the basement interferred with a uniform movement of layered units westward up the eastern limb of the anticlinorium" (Grace, 1953, p. 69).

No thrust faults were found within the Hyde Park quadrangle, and it is not believed that all the characteristics of the first stage minor structures could be produced by thrust faults. It is believed that when the Green Mountain anticlinorium first began to form that flexural slip occurred along the stratification (S_1) causing the secondary bedding foliation (S_2) to form. Plunging folds and recesses developed (see Plate 4), indicative of extension parallel to the fold axes, and minor folds formed, nearly at right angles to the axis of the Green Mountain anticlinorium. Further flexural slip was accommodated by slip along the bedding foliation (S_2), by rotation of the folds (L_1), and finally by shearing or stretching off of the folds on their short limbs and fold crests. The planes along which flexural slip occurred in this stage of the deformation were generally parallel to the bedding but locally transected the bedding. The actual folds (L_1) were formed by extension nearly parallel to the axis of the Green Mountain anticlinorium, but they are rotated and tightened and the transverse schistosity (S_3) was formed by relative movement between the layers nearly at right angles to the axis of the Green Mountain anticlinorium.

Further flexural slip caused crinkles and folds (L_2) to form subparallel to the axis of the Green Mountain anticlinorium, and eventually caused the slip cleavage, slaty foliation, and schistosity (S_4) to form.

The latest folds (L_3?) and the related fracture cleavage (S_5?) are probably not directly related to the formation of the Green Mountain anticlinorium.

METAMORPHISM

General statement

All of the rocks of the Hyde Park quadrangle, except the post-metamorphic dikes have been affected by regional metamorphism. Different assemblages of minerals are stable in rocks of a given chemical composition under different conditions of metamorphism. In this area chlorite, almandite, and kyanite are successive general indicators of increasing metamorphic grade in the argillaceous and arenaceous rocks. Chlorite, actinolite, and hornblende are similar indicators in the mafic rocks. The mineral assemblages noted in the rocks of the Hyde Park quadrangle will be described in detail and will be used to determine the successive mineral transformations which occurred during the metamorphism.

Mineralogy and petrography of the schist

Estimated modes of typical rock types are tabulated with the descriptions of the individual stratigraphic units. Quartz, "sericite" (both muscovite and paragonite), chlorite, and albite are by far the dominant minerals in all the metasedimentary formations, and there is a wide range in their relative proportions. Quartz ranges from 0 to 97 percent, "sericite" from 2 to 70 percent, chlorite from 0 to 40 percent, and albite from 0 to 40 percent. Even within a single formation the range in the dominant constituents is very great. Almandite, kyanite, graphite, calcite, ankerite, biotite, and magnetite each constitute as much as 5 percent of some rocks. Minor minerals include tourmaline, apatite, pyrite, limonite, rutile, ilmenite, sphene, allanite, zircon, and clinozoisite-epidote.

All of the minerals occur in habits characteristic of their occurrence in low-grade and middle-grade schists and no detailed discussion of individual habit is necessary. In general the grain size of the major constituents increases with the grade of metamorphism.

Graphite, limonite, and pyrite commonly occur together, producing the characteristic dark-colored, rusty-weathering outcrops of the graphitic rocks. Jacobs (1938, p. 52) reports 1.09 weight percent of carbon in a graphitic schist from Hyde Park town. Analyses of carbonaceous shales show a sulfide content appreciably higher than that of noncarbonaceous shales. The sulfide content, the carbon content, and the ferrous nature of the iron all probably are primarily due to strongly reducing conditions at the time of deposition. Those rocks in which magnetite is abundant have a shiny, silvery-green weathered surface. Magnetite is not found in the graphitic schists, probably due primarily to the low ferric iron content of carbonaceous sediments and possibly due in part to reduction of ferric iron by carbon during metamorphism.

$$\text{Ferric minerals} + C \longrightarrow \text{Ferrous minerals} + CO_2 \uparrow$$

Moreover the chlorite in the graphitic rocks commonly contains less ferrous iron than magnesium, as indicated by a positive optic sign, a negative sign of elongation, and low index of refraction (see Figure 4). Inasmuch as analyses indicate comparable total iron contents in both carbonaceous and noncarbonaceous shales, it seems likely that the low iron content of the chlorites in the absence of magnetite or biotite is related to the high pyrite content of the graphitic rocks.

The graphitic rocks commonly contain sphene, while non-graphitic rocks commonly contain ilmenite or rutile. The other minor minerals are present in both graphitic and non-graphitic rocks; tourmaline, apatite, and clinozoisite-epidote being present in nearly every thin section.

Small euhedral crystals of almandite garnet enclosed in large porphyroblasts of albite have been noted in the coarse-grained schist of the Camels Hump group in the western part of the Hyde Park quadrangle.

The various assemblages which have been noted in thin sections are tabulated in Table 11. Dr. Hans Eugster of the Geophysical Laboratory, Carnegie Institution of Washington, kindly identified many specimens of "sericite" by use of the X-ray spectrometer. Other specimens were identified by the author. Both muscovite and paragonite have been identified, but no pyrophyllite was found. In some specimens as much as 55 percent of the "sericite" is paragonite.

The iron and magnesium content of the chlorite has been interpolated from the optic indices. Winchell (1951, p. 385) and Hey (1954) have published graphs for this purpose, but it was felt that it would be better to use a chart specifically constructed for determining the FeO:FeO + MgO ratio of chlorite from low-grade and middle-grade metamorphic rocks.

Although the optic indices of chlorite depend most strongly upon the FeO:FeO + MgO ratio, they are also affected by the content of Al, Fe''', Ti, Mn, Cr, etc. It seems probable that the variation of these components is more restricted in chlorites of low-grade and middle-grade metamorphic rocks than in chlorites as a whole. Thus, a diagram relating the

FeO:FeO + MgO ratio and the optic indices for these particular rocks should be more accurate than one prepared for chlorites in general.

The ideal structural formula for pinninite indicated by Pauling (1930, p. 578) is:

$$\left[(Fe,Mg)_2^{VI}\ Al^{VI}(OH)_6\right]\ \left[(Fe,Mg)_3^{VI}(Al^{IV}Si_3{}^{IV}O_{10})(OH)_2\right]$$

However, most natural chlorites have an excess of alumina over this formula and contain some ferric iron. Excess alumina over the Pauling formula may be divided equally between the six-fold coordination and the four-fold coordination positions in the following fashion:

$$\left[(Fe,Mg)_2^{VI}\ Al^{VI}(OH)_6\right]\ \left[(Fe,Mg)_{3-X}^{VI}\ Al_X{}^{VI}\ (Al_{1+X}^{IV}Si_{3-X}^{IV}\ O_{10})(OH)_2\right]$$

Ferric iron in six-fold coordination may be compensated by alumina entering four-fold coordination in place of silicon, or may result from oxidation of ferrous iron and loss of water (Winchell, 1926; Hey, 1954).

About 55 chlorite analyses, from published and unpublished sources and accompanied by optical data, were recalculated assuming that Si plus ½ (Fe''' + Al''') = 4 (mole proportions). It was found that ½ (Fe''' + Al'''), the number of tetrahedral positions filled by aluminum, ranged from 0.76 to 1.99, the median being about 1.40. Chlorites from argillaceous schists and greenstones (Table 10, #1-12) have ½ (Fe''' + Al''') ranging from 1.27 to 1.63; the lower range coming from greenstones and the higher range from argillaceous schists. The bonds in the chlorite structure would be best satisfied by a ½ (Fe''' + Al''') value of about 1.33 (J. B. Thompson, written communication). Most values above that figure are from analyses of chlorites separated from argillaceous schists. It is possible that imperfect separation of muscovite may be responsible for some of these higher values. This apparent range would be reduced if part of the ferric iron is due to the oxidation of ferrous iron. Other chlorites (Table 10, 13-42) were chosen for a low ferric iron content (less than 3.5 weight percent) and a limited range of values of $\frac{1}{2}$ (Fe''' + Al'''). For each of these

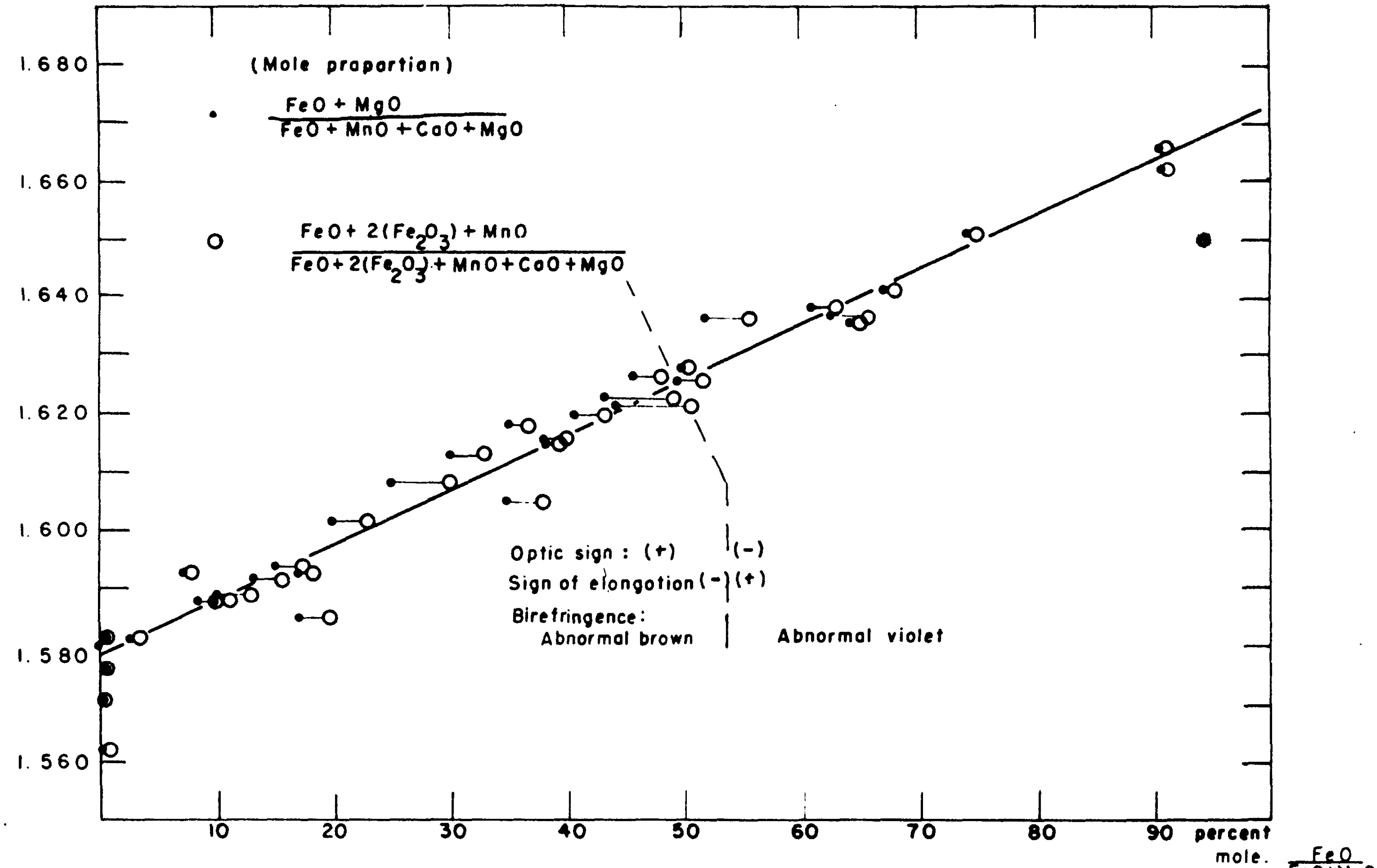
$$4\left[\frac{Al_2O_3}{Al_2O_3 + SiO_2}\right],\quad 4\left[\frac{Al_2O_3 + Fe_2O_3}{Al_2O_3 + Fe_2O_3 + SiO_2}\right],\quad \frac{FeO + MnO}{FeO + MnO + MgO + CaO},$$

and $\frac{FeO + 2(Fe_2O_3) + MnO}{FeO + 2(Fe_2O_3) + MnO + MgO + CaO}$ in mole proportions were plotted against the Y index of refraction. The factor of four has been added so that the ratio gives the number of Al''' atoms in four-fold coordination; the value can then be substituted directly into the chlorite formula.

Examination of Figure 3 indicates that the iron content is chiefly responsible for an increase in the optic indices and that the deviation from a straight line is little greater than the error in the determination of the optic indices. The effect of a variation in alumina and ferric iron within the range considered is apparently less than the variation due to the combined analytical and index determination error. Yoder's (1952, p. 576) synthetic pure magnesium chlorite has a Y index of 1.580, lying upon the curve.

The optic sign changes and the birefringence passes through zero at a Y index of about 1.627. On either side of the sign-change abnormal interference colors are visible. The low index side (i.e., magnesium-rich) has a positive sign, a negative sign of elongation, and abnormal

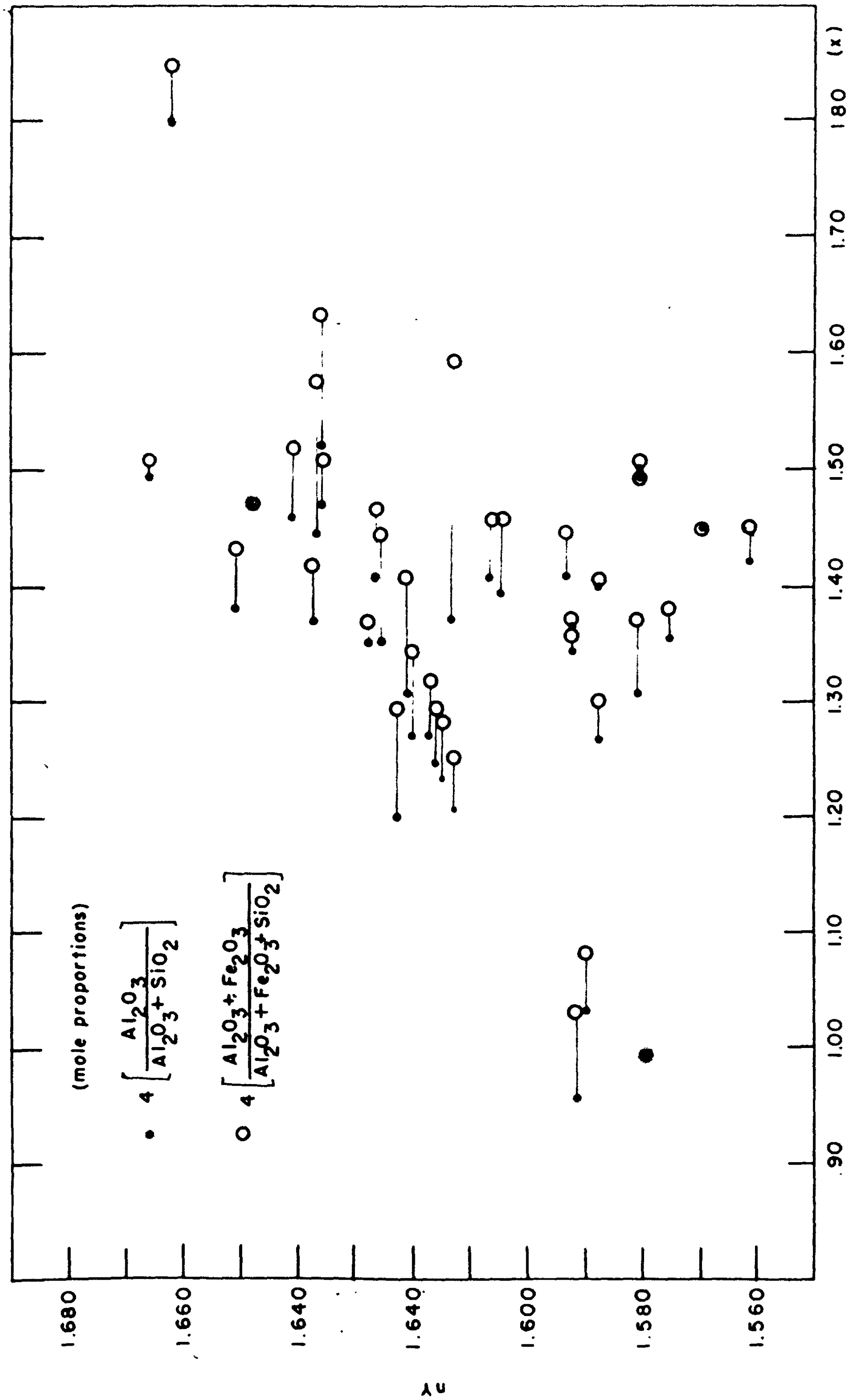
Figure 4



Relation of beta (nY) index of refraction and Fe-content of chlorite.

80-A

Figure 3



Relation of beta (nY) index of refraction and tetrahedrally coordinated aluminum(x) in chlorite $[(Fe,Mg)_2\,Al(OH)_6][(Fe,Mg)_{4-x}Al\ (Al_xSi_{4-x}O_{10}(OH)_{12}]$

80-B

brown interference color, the birefringence increasing with increasing magnesium content. The high index side of the sign change (i.e., iron-rich) has a negative sign, a positive sign of elongation, and abnormal violet interference color, the birefringence increasing with increasing iron content. This point corresponds very closely to the similar point on Winchell's chart (1951, p. 385), but not to Hey's chart (1954, p. 284). A similar sign change has been noted at a Y index of 1.575 in "chlorite" from an ultramafic body in Lowell, Vermont. However, these have not been positively shown to be true 14Å chlorite.

In a few thin sections chlorite grains associated with limonite show peripheral alteration to a yellow-brown, more highly birefringent mineral, bearing a resemblance to stilpnomelane. X-ray investigation failed to show the presence of stilpnomelane and it was concluded that the alteration is the result of oxidation of ferrous iron in the chlorite, probably during weathering.

The content of iron and magnesium in biotite has been estimated from refractive indices using the chart of Winchell (1951, p. 374). Very few analyses of biotite from low-grade or middle-grade schists are available and the following formula suggested by J. B. Thompson has been used as representing the most likely alumina content of the biotite.

$$K(Fe,Mg)^{VI}_{2\ 2/3}\ (Al + Fe''')^{VI}_{1/3}\ (Al^{IV}_{1\ 1/3}\ Si^{IV}_{2\ 2/3}\ O_{10})(OH)_2$$

The variation in the iron and magnesium content of chloritoid is apparently small; seven specimens had a Y index of 1.720, two had a Y index of 1.719, and one had a Y index of 1.718. A typical specimen had

The following optical characteristics:

chloritoid (AA-272b) (+) 2V = 65° 001 lamellar twinning

n_X—1.717

n_Y—1.720

n_Z—1.727

Analyses of chloritoid from other localities suggest that the indices indicate a ratio of iron to iron plus magnesium plus manganese between .80 and .90.

The schists in the kyanite and garnet zones show retrograde alteration. Kyanite, garnet, and biotite are partially or completely altered to muscovite, chlorite, chloritoid, and magnetite. In a few thin sections only a peripheral alteration of kyanite and garnet grains was noted. More commonly, a fine-grained aggregate of muscovite with coarser plates of chloritoid encloses irregular relict grains of kyanite, which are in parallel optical orientation. The chloritoid plates are commonly oriented in rosettes. Six specimens of altered kyanite proved to contain muscovite with no paragonite or pyrophyllite. Fine-grained aggregates of chlorite, commonly with sericite, magnetite, and rutile, enclose garnet relics. Irregular relict grains, which are in parallel optical orientation, of coarse muscovite in a fine-grained aggregate of sericite were also noted. In other thin sections the fine-grained aggregates are present in a rock containing coarse-grained muscovite and quartz, but no relict grains of kyanite or garnet have been preserved.

No pseudomorphs or relict grains of staurolite or of sillimanite have been noted, but several thin sections contain irregular plates of biotite partially altered to chlorite, or aggregates of chlorite possibly pseudomorphic after biotite. Some chloritoid is clearly an alteration product, but it appears to be part of the original assemblage in other specimens.

Petrology of the micaceous schists

Zones of progressive regional metamorphism

In 1893 Barrow, working on the Dalradian schists of Scotland, mapped zones of progressive regional metamorphism. These zones were based on mineralogical transformations in derivatives of argillaceous sediments and were correlated in a general way with increasing temperature and pressure. Each zone of progressive metamorphism was defined by an index mineral, the first appearance of which (in passing from low to higher grades) marks the outer limit of the particular zone. The trace of this surface on the earth's surface was termed an isograd (line of equal metamorphic grade) by Tilley (1925) and was interpreted by both Tilley (1925) and Harker (1932, p. 186) as being also an isotherm. Since the occurrence of a mineral depends in part on composition, it was recognized early that rocks of comparable chemical composition must be selected for zonal mapping in terms of index minerals. In many areas pelitic rocks have been chosen because of their widespread occurrence and because of the numerous mineral changes which take place in them with increasing metamorphic grade. The zonal sequence: chlorite, biotite, almandite, staurolite, kyanite, and sillimanite (in order of increasing metamorphic grade)- is typical of many areas.

In the Hyde Park quadrangle the almandite and kyanite isograds have been mapped on the basis of the first appearance——that is, the garnet isograd surrounds all occurrences of garnet in argillaceous schist, etc. The garnet and kyanite zones are confined to the southeast corner of the area, lying entirely within the Stowe formation.

In the Mt. Mansfield quadrangle along the crest of the Green Mountain anticlinorium and just west of the edge of the Hyde Park quadrangle garnets were also observed. Furthermore, six thin sections from the coarse-grained schist and gneiss of the Camels Hump group, lying on the west side of the area, contained small euhedral porphyroblasts of garnet within larger albite porphyroblasts. Many albite porphyroblasts were rotated and the texture of the rock modified after the porphyroblasts ceased their growth (page 68). It is believed that the small garnets are a relict feature, all garnets not protected by albite having been altered to chlorite. This implies that the garnet zone originally covered much more of the quadrangle than shown on the map, and that the retrograde alteration has destroyed nearly all evidence of its existence.

A biotite isograd has not been mapped within this area. The paucity of biotite and the abundance of chloritoid in this area prompted the author (Albee, 1952) to calculate numerous analyses of argillaceous-arenaceous sediments to determine what minerals would be present if the sediment were metamorphosed. It was found that the range of chemical composition is such that many argillaceous sediments would not contain biotite until the garnet, kyanite, or even the sillimanite zone of metamorphism. It was also found that many of the argillaceous rocks that do not contain biotite should contain chloritoid. The presence of chloritoid has commonly been ascribed to abnormal conditions, generally "special pressure conditions" (eg. Barth, 1952, p. 336). However, investigation of the chemical composition has made it clear that the presence of chloritoid and the absence of biotite in argillaceous rock of low grade areas such as the Hyde Park quadrangle is not abnormal; it is due to a composition different, but not anomalously different, from that present in areas in which the biotite zone is well defined.

Petrology of the micaceous schists of the Hyde Park quadrangle

Introduction.

The retrograde alteration of the rocks of the garnet and kyanite zones has made it impossible to determine the exact sequence of mineral transformations in this area. The sequence presented here is based in part upon investigations in the Lincoln Mountain and Montpelier quadrangles and upon general knowledge of low-grade and middle-grade mineral assemblages. The mineral assemblages noted in the chlorite, garnet, and kyanite zones are presented in Table 11.

It is convenient to show the observed mineral assemblages diagrammatically. Niggli (1954, p. 391-403) and Thompson (unpublished lecture notes) have discussed approximations in accordance with the phase rule which permit a reduction in the number of components which must be shown graphically. Such diagrams "facilitate the obtaining of a first comprehensive view," "aids the discussion of the possible relationships of the phases among themselves," "are a useful aid to visualization," "help in

TABLE 11. Mineral assemblages of micaceous schist in the Hyde Park quadrangle.

Quartz and muscovite are present as additional phases in all assemblages. Apatite, tourmaline, zircon, calcite or clinozoisite-epidote, sphene, or rutile or ilmenite, and pyrite may be present in all assemblages. Armoured grains of two or more titanium minerals are common.

Chlorite zone

albite-graphite
chlorite-magnetite-albite-paragonite
chlorite-magnetite-paragonite
chlorite-magnetite-albite (garnet enclosed in albite)
chlorite-magnetite
chlorite-albite-graphite (garnet in albite)
chlorite-paragonite
biotite-chlorite-magnetite-albite
biotite-chlorite-albite-graphite
chloritoid-chlorite-magnetite
chloritoid-chlorite-magnetite-paragonite
chloritoid-chlorite

Garnet zone (partially retrograded to chlorite zone assemblages)

chlorite-magnetite-albite
chlorite-magnetite-albite with relics or pseudomorphs of garnet
chlorite-magnetite-paragonite with relics or pseudomorphs of garnet
chloritoid-chlorite-magnetite-paragonite
chloritoid-chlorite-magnetite with relics or pseudomorphs of garnet
chloritoid-chlorite-paragonite
chloritoid-magnetite-paragonite
chloritoid-paragonite
chloritoid
biotite-chlorite-magnetite with relics or pseudomorphs of garnet

Kyanite zone (partially retrograded to chlorite zone assemblages)

chlorite-magnetite with relics or pseudomorphs of garnet and kyanite
chlorite with relics or pseudomorphs of garnet and kyanite
chlorite with relics or pseudomorphs of garnet, kyanite, and biotite(?)
chloritoid-chlorite-magnetite
chloritoid-chlorite-magnetite with relics or pseudomorphs of garnet and kyanite
chloritoid-chlorite with relics or pseudomorphs of garnet and kyanite
chloritoid-chlorite with relics or pseudomorphs of garnet, kyanite, and biotite(?)
biotite-chlorite-magnetite with relics of garnet

TABLE 12. Optical data on chlorite, biotite, and chloritoid

Zone	Specimen Number	Chlorite nY ± .003	Chlorite Fe:Fe + Mg	Biotite nY ± .003	Biotite Fe:Fe + Mg
Chlorite zone	Biotite-chlorite-magnetite-albite				
	AA-845	1.631	53%	1.657	70%
	AA-1028	1.631	53%	1.651	65%
	AA-1031b	1.643	67%	ndt	
	Biotite-chlorite-albite				
	AA-741	1.625	48%	1.640	50%
	101	1.618	40%	ndt	
	Chloritoid-chlorite-magnetite			Chloritoid	
	AA-641	1.631	53%	1.719	
	Chloritoid-chlorite-magnetite-paragonite				
	AA-718	1.631	53%	1.720	
	Chloritoid-chlorite				
	AA-583	1.629	52%	ndt	
Garnet zone	Chloritoid-chlorite-magnetite-paragonite				
	AA-535a	1.626	48%	ndt	
	Chloritoid-chlorite-magnetite with relics or pseudomorphs of garnet				
	AA-30	1.631	53%	1.720	
	Chloritoid-chlorite-paragonite				
	AA-272a	1.635	58%	ndt	
	Chloritoid-magnetite-paragonite				
	AA-82			1.720	
	Chloritoid-paragonite				
	AA-272b			1.720	
	Biotite-chlorite-magnetite with relics or pseudomorphs of garnet				
	AA-239	1.629	52%		
Kyanite zone	Chloritoid-chlorite-magnetite				
	AA-22	1.621	43%	1.719	
	Chloritoid-chlorite-magnetite with relics or pseudomorphs of garnet and kyanite				
	AA-10b	1.625	48%	1.720	
	AA-231	1.631	53%	1.720	
	Chloritoid-chlorite with relics or pseudomorphs of garnet and kyanite				
	AA-20	1.615	37%	1.720	
	Biotite-chlorite-magnetite with relics of garnet			Biotite	
	AA-75	1.623	45%	ndt	
	AA-110b	1.625	48%	1.643	55%

correctly formulating the problems encountered," and "fulfill the important function of discouraging the purely speculative approach and curbing flights of fancy" (Niggli, 1954, p. 403).

The argillaceous schists contain phases which to a first approximation lie in the system SiO_2-Al_2O_3-FeO-MgO-K_2O-H_2O. Quartz-bearing assemblages in this system may be shown graphically in a tetrahedron with the apices (Al_2O_3-K_2O), $2K_2O$, $1/3FeO$, and $1/3\ MgO$. With the further restriction that muscovite be present the assemblages may be projected onto a plane; graphically a projection through muscovite to the plane between the (Al_2O_3 - K_2O) apex and biotite (see Figure 5a). In such a projection the composition of the phases may be located as shown on Figure 5b.

In accordance with the phase rule the three phase assemblages in the above diagram may contain one additional phase for each of several other components such as Na_2O, CaO, Fe_2O_3, and TiO_2. Two phase assemblages on the diagram may contain two additional phases for one of these other components. Such phases in observed assemblages may be noted on the diagram.

The diagrams shown on Figure 6 show the mineral assemblages found at successively higher metamorphic grades. These are based upon the observed assemblages given in Table 11, the interpretation of the retrograde textures, and upon assemblages found in nearby areas. Assemblages that were actually observed in the Hyde Park quadrangle are connected by solid lines and the Fe:Fe/Mg ratios of biotite and chlorite that were determined optically are shown with solid circles.

Explanation to Figures 5 and 6.

The construction and the corners of the triangular diagrams in Figure 6 are shown in Figure 5. Each diagram represents the mineral assemblages stable with quartz and muscovite over a limited range of P, T, and μ_{H_2O}. Paragonite is present with assemblages above and albite with assemblages below the heavy dashed line in Figures 6b and 6c.

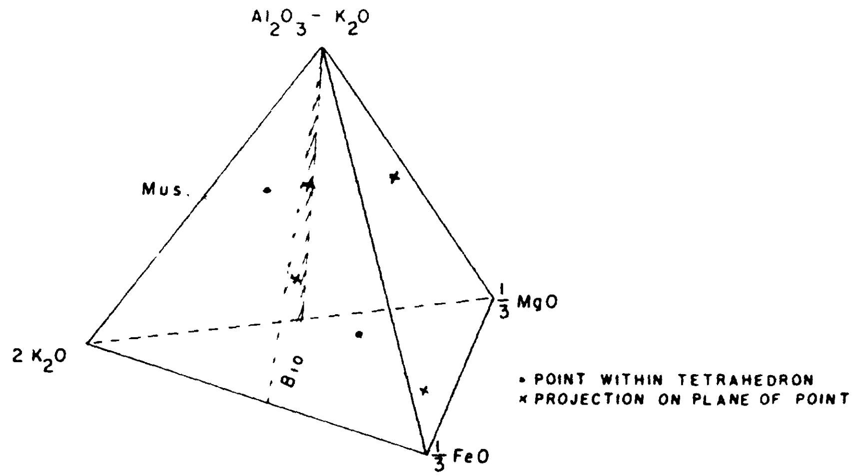
Mineral assemblages actually observed in the Hyde Park quadrangle are shown with solid lines. For these assemblages solid circles indicate optical determination of Fe:Fe+Mg for biotite and chlorite. Hollow circles indicate a variable Fe:Mg ratio.

Abbreviations used on the diagrams:

Alb	albite
Alm	almandite
Bio	biotite
Chl	chlorite
Ctd	chloritoid
Kao	kaolinite
Ksp	potassium feldspar
Kyt	kyanite
Mus	muscovite
Par	paragonite
Pyp	pyrophyllite

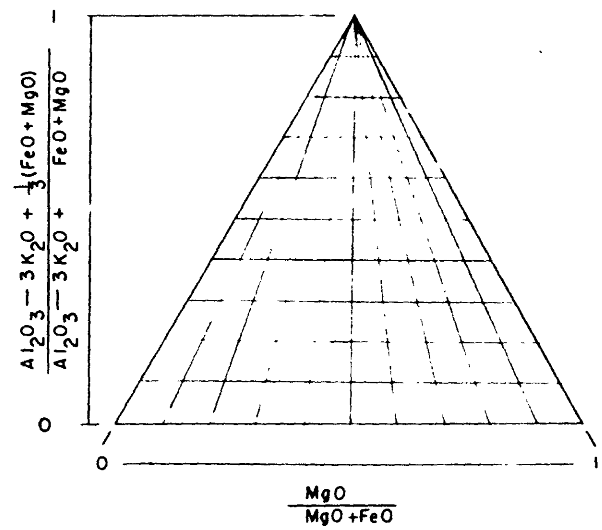
88-17

Figure 5



(a)

Location in the tetrahedron of the projection plane used to show phase assemblages of quartz-and muscovite-bearing schist



(b)

Method of locating composition of phases upon the triangular projection plane

Figure 6a.

The assemblage chlorite-muscovite-quartz is very common throughout the center of the Hyde Park quadrangle. It may in part in this area be an assemblage of the chlorite zone prior to the appearance of chloritoid or biotite, although the assemblage does persist into higher metamorphic grades.

It seems likely that chlorite is one of the first stable minerals to appear in argillaceous-arenaceous sediments in the lowest grades of metamorphism (diagenesis). The variation in the aluminum content and Fe:Mg ratio of chlorite (see page 79) and in the K:Al ratio of muscovite (Yoder and Eugster, 1955, p. 257) is great enough that the muscovite-chlorite field on Figure 6a includes the composition of the most common shales. In addition, if oxygen is "mobile" (Thompson, 1955, p. 81), the chlorite-muscovite-quartz-magnetite field would be considerably larger than the chlorite-muscovite-quartz field. Rocks with compositions below the muscovite-chlorite join commonly contain grains of potassium feldspar which are apparently of detrital origin. It might be assumed that such rocks retain the disequilibrium assemblages of the sediment, but Chayes (1955, p. 80) has cited evidence for a stable potassium feldspar-chlorite-muscovite-quartz assemblage. Rocks with compositions above the muscovite-chlorite join may contain kaolinite, aluminum-montmorillonite, or pyrophyllite, but X-ray work on samples from the Hyde Park quadrangle and from northwest Maine have not yet confirmed this.

Figure 6b.

Although biotite is uncommon both chloritoid-chlorite-muscovite-quartz and biotite-chlorite-muscovite-quartz assemblages are present in the Hyde Park quadrangle. Chlorite is higher in magnesium than its co-existing chloritoid or biotite and the trends of the tie lines in the two phase fields are indicated on the diagram. There is no evidence as to whether biotite or chloritoid formed earlier. A careful search of the literature and of specimens in the Harvard collections disclosed only four occurrences of chloritoid and biotite together; moreover, several of these may be disequilibrium assemblages. This suggests that the chloritoid-biotite-chlorite-field shown on Figure 6b is very small, and to be found only in rocks with an unusually high FeO:MgO + FeO ratio. It is also possible that the chlorite field cuts it out until just prior to the formation of garnet. As indicated on the diagram both paragonite and albite are found with the chlorite-muscovite-quartz assemblage; but albite is not found with chloritoid and paragonite is not found with biotite in this area. However, the assemblage chloritoid-biotite-chlorite-albite-muscovite-quartz has been reported elsewhere (Cloos and Hietanen, 1941, p. 101-146). The assemblage chloritoid-pyrophyllite-chlorite-muscovite-quartz is uncommon in bedded rocks, and it is believed that kaolinite rather than pyrophyllite probably persists into this grade of metamorphism.

Figure 6c.

This diagram is marked by the appearance of almandite garnet. In the garnet zone in the southeastern part of the area garnet-chloritoid-chlorite-quartz-muscovite seemed to have been the stable assemblage at the peak of the metamorphism before the retrograde alteration. The same assemblage is common in the Lincoln Mountain area to the south. Also in the same area garnet-chlorite-biotite-muscovite-quartz is an assemblage common in stratigraphic units which contain chlorite-albite-muscovite-quartz or chlorite-biotite-albite-muscovite-quartz in the Hyde Park quadrangle. One specimen of the latter assemblage did contain small garnet porphyroblasts enclosed in the albite grains; probably as armoured relics (page 84).

If the biotite-chloritoid-chlorite field is present the transition from Figure 6b to Figure 6c is:

$$\underset{\text{chloritoid}}{3(Al_2SiO_5)(Fe(OH)_2} + 1\underset{\text{biotite}}{\left[K(Fe,Mg)_3(AlSi_3O_{10})(OH)_2\right]} + \underset{\text{quartz}}{3SiO_2}$$

$$\underset{\text{almandite}}{2\left[Fe_3Al_2(SiO_4)_3\right]} + 1\underset{\text{muscovite}}{\left[KAl_2(AlSi_3O_{10})(OH)_2\right]} + 3H_2O$$

In this transition the field chloritoid-almandite-biotite-muscovite-quartz would probably appear briefly on the extreme iron-rich side of the diagram.

When the field chloritoid-chlorite-biotite-muscovite-quartz does not appear a nearly isochemical transition of iron-rich chlorite to almandite is indicated by the common compositions, as compared to the ideal formulas, of chlorite and biotite. Commonly the aluminum content of chlorite is high enough that the following transition can occur nearly isochemically with very little biotite being formed.

$$7\left[(Fe,Mg)_{4\frac{2}{3}}\ Al_{1\frac{1}{3}}(Al_{1\frac{1}{3}}Si_{2\frac{2}{3}}O_{10})(OH)_6\right] + 1\left[KAl_2(AlSi_3O_{10})(OH)_2\right] + 11\ SiO_2 =$$

chlorite muscovite quartz

$$10\left[Fe_3Al_2(SiO_4)_3\right] + 1\left[K(Fe,Mg)_{2\frac{2}{3}}\ Al_{1\frac{1}{2}}\ (Al_{1\frac{1}{3}}Si_{2\frac{2}{3}}O_{10})(OH)_2\right] + 21\ H_2O$$

almandite biotite

<u>Figure 6d.</u>

The next transition would probably be the appearance of kyanite. The fields would probably remain similar to Figure 6c, except that kyanite would appear in place of kaolinite or pyrophyllite. These transitions would be:

1 pyrophyllite $\longrightarrow$ 1 kyanite + 3 quartz + 1 H_2O

or

1 kaolinite $\longrightarrow$ 1 kyanite + 1 quartz + 2 H_2O

The size of the kyanite-chloritoid-chlorite field would probably not change too much from that of the kaolinite (or pyrophyllite)-chloritoid-chlorite field. Neither kaolinite nor pyrophyllite were found in the X-ray work. Thus, it is believed that the kyanite zone as mapped in the Hyde Park quadrangle is represented by the next diagram and that this first appearance of kyanite would be rare or absent.

Figure 6e.

This stage is marked by the presence of kyanite and garnet as compatible phases and the incompatibility of chloritoid and chlorite. The transition from Figure 6d to Figure 6e is:

4 chloritoid + 1 chlorite + 4 quartz $\longrightarrow$ 2 kyanite + 3 almandite + 7 H_2O

Interpretation of the retrograde textures suggests that kyanite-garnet-chlorite-muscovite-quartz was a common assemblage before the retrograde alteration. There is no suggestion in the textures that staurolite was ever present.

Neither paragonite nor albite were observed in the rocks of the kyanite zone suggesting both that the sodium content of these rocks is low and that it is contained in solid solution in muscovite rather than bringing about a distinct phase. Muscovite is known to carry as much as 30 percent paragonite in solid solution in association with paragonite in high-grade rocks (Zen, 1955, p. 49; Eugster and Yoder, 1955, p. 125).

Retrograde metamorphism.

Retrograde metamorphism (in the broad sense) is the alteration of a mineral assemblage to an assemblage characteristic of lower metamorphic grade. It may be caused by cooling slowly enough to allow the minerals to partly change to an assemblage stable at lower metamorphic grade, by a later lower grade metamorphism, by an increase in the partial vapor pressure (chemical potential) of water and/or carbon dioxide, or by some combination of these factors. Hsu (1955) has distinguished between "monometamorphic diaphthoresis," retrograde alteration in response to falling temperature, and "polymetamorphic diaphthoresis," retrograde alteration in response to either rising or falling temperature during a later period of metamorphism. However, he did not discuss the possibility of retrograde metamorphism in response to a sudden increase in the chemical potential of volatile components, quite possibly independent of temperature change.

The textures described previously make it clear that in this area mineral assemblages of the kyanite zone have been altered to assemblages of the chlorite zone. Although minor structural features of two and perhaps three different orientations are superimposed, it is believed that these belong to one period of deformation. Both the high-grade areas and the retrograde features are found along a belt of well-defined folds but these folds seem to be of the same age as the Green Mountain anticlinorium. In the Lincoln Mountain quadrangle retrograde features are present in this same belt, whereas kyanite-bearing rocks along the crest of the Green Mountain anticlinorium show no retrograde features. There is no clear evidence of polymetamorphism in the Hyde Park quadrangle.

One puzzling feature is the alteration of kyanite to muscovite rather than pyrophyllite, which would be expected if the alteration were essentially a hydration of kyanite. However, the six specimens of altered kyanite, which were checked by X-ray, contained muscovite but no pyrophyllite. This might suggest that potassium had been added from an external source during the retrograde alteration. However, another alternative would be to assume that the assemblage kyanite-garnet-biotite-muscovite-quartz, as shown in Figure 6f, was present before the alterat occurred. As noted on page 83 there is some suggestion in the relict textures that biotite was present. Moreover, in southern Vermont in similar rocks this assemblage is common (J. B. Thompson, oral communication, 1955), and in the Monadnock quadrangle of southern New Hampshire K. F. Billings (1949, p. 1262) reports a sillimanite-garnet-biotite-muscovite rock altered to a chloritoid-chlorite-muscovite assemblage. The textures described are similar to those observed in the Hyde Park quadrangle. If a kyanite-garnet-biotite-muscovite-quartz rock having a composition above chlorite on Figure 6f were hydrated, it is probably that the kyanite would alter to muscovite rather than pyrophyllite. Thus, it is felt that no addition of potassium during the alteration needs to be postulated.

The common assemblage kyanite-staurolite-biotite-muscovite-quartz would also be affected in the same manner by hydration. Earlier the possibility that this assemblage had been present was dismissed because of the lack of any relict textures indicative of staurolite. However, the author has since studied altered sillimanite, andalusite, and staurolite rocks in the Front Range of Colorado and noted that it is very difficult to differentiate altered relict textures of these three minerals. However, although staurolite may have been present, it seems most likely that kyanite-biotite-garnet-muscovite-quartz was the highest grade assemblage formed in this area.

All such retrograde reactions involve hydration of the higher grade assemblages. It seems impossible to determine whether this hydration occurred because of cooling or because of an increase in chemical potential of water, or some combination thereof.

Premetamorphic nature of the micaceous schists

General statement

Sedimentary features such as bedding and clastic grains, and gross composition indicate that the metamorphic rocks in the Hyde Park quadrangle, excluding the greenstone and amphibolite, were derived from predominantly argillaceous-arenaceous sediments. Chemical composition is one of the few possible lines of evidence as to the exact nature of the "protolith" or parent rock where the sedimentary fabric and mineralogy has been altered. If the metamorphic processes in this area did not involve important changes in chemical composition, then chemical composition should be as useful in determining the sedimentary protolith as in distinguishing the different sedimentary rocks from each other.

No such study of the chemical composition has been published and it was necessary to undertake a study of the chemical composition of argillaceous and arenaceous sedimentary rocks.

The composition of argillaceous and arenaceous sedimentary rocks

Analyses of argillaceous and arenaceous rocks were culled from the literature, rejecting all analyses which did not report all of the following: SiO_2, Al_2O_3, Fe_2O_3, FeO, MgO, CaO, Na_2O, K_2O, and H_2O. Analyses of rocks from contact zones or with any vein material were also rejected. Finally all analyses were rejected which could not be reasonably classified from the published description into the following groups: orthoquartzite or quartzose sandstone, subgraywacke or normal sandstone, graywacke, arkose, and shale or slate. Cumulative frequency curves were then constructed for various oxides and oxide-ratios to determine which were most characteristic of the various groups. The results are summarized in Table 13.

Absolute values of weight percent are not too useful. Two rocks, whose detrital constituents have the same composition, but which contain different amounts of calcite, dolomite, or iron oxide cement would have quite different absolute weight percentages and the close relationship between the two rocks would be obscured. In addition the gain or loss of water, carbon dioxide, and other constituents during metamorphism would affect the absolute values. Relative values of characteristic constituents, expressed as ratios, are much more useful.

It was found that the relative values of Al_2O_3, K_2O, and Na_2O, and the absolute weight percent of SiO_2 afford a rather good index to the sedimentary rock type. The ratio $SiO_2 : SiO_2 + Al_2O_3$ would provide a better separation between several rock types, but it is more difficult to point out its relation to the rock classifications used by various sedimentary petrographers. Other indices were less valuable; for example, the $FeO : FeO + MgO$ ratio, which is important in determining the mineralogy of metamorphic rock is quite variable.

In order to simplify the comparison with metamorphic rocks the sandstones and representative shale and slate analyses were plotted on triangular diagrams to show simultaneously the relative values of the mole percent of Al_2O_3, K_2O, and Na_2O. These diagrams, ranked by the median values of the weight percent of silica, are shown in Figure 7 in which weight percent of silica is the axis of a triangular prism. Other

Table 13. Chemical characteristics of sedimentary rock types.

		Quartzose sandstone	Subgraywacke	Graywacke (G)	Arkose (A)	Shale-slate (S)
Number of analyses		11	3	20	11	130
SiO_2	Mn	61.70	74.43	52.34	51.65	18.56
	1Q	91		61	63	56
	2Q	94		64	69	59
	3Q	98		69	76	62
	Mx	99.45	82.15	72.62	82.14	84.14
Al_2O_3	Mn	.30	5.37	10.15	2.55	2.48
	1Q	.5		12.5	7.5	14.5
	2Q	3.0		13.5	10.5	17.5
	3Q	3.5		15.0	14.0	19.0
	Mx	6.92	11.76	18.33	15.50	26.22
Na_2O	Mn	.00	1.63	1.48	.19	.01
	1Q			2.2	.3	.3
	2Q			2.7	.5	.6
	3Q			3.4	2.3	1.4
	Mx	.34	2.57	6.03	4.01	3.97
K_2O	Mn	.00	1.09	1.26	1.28	.44
	1Q			1.5	2.3	2.5
	2Q			1.8	3.1	3.2
	3Q			2.1	4.3	3.9
	Mx	.61	1.74	4.33	6.16	10.85
CaO	Mn	T	.70	.72	.15	.02
	1Q	.1		1.85	1.85	.25
	2Q	.3		2.6	2.85	.65
	3Q	.6		4.0	77.0	1.5
	Mx	21.00	1.85	12.80	20.09	10.77
MgO	Mn	0	1.30	1.22	T	0
	1Q	.1		1.90	.05	1.55
	2Q	.2		2.25	.15	2.05
	3Q	.35		3.3	.65	3.3
	Mx	1.18	2.22	4.06	2.95	6.43
Fe_2O_3	Mn	T	.55	.30	0.61	0.00
	1Q	.05		.88	1.35	1.6
	2Q	.14		1.15	3.3	3.1
	3Q	.45		1.50	5.5	5.8
	Mx	1.50	1.47	2.58	7.85	16.88
FeO	Mn	.11	1.08	2.36	0.31	.17
	1Q	.3		3.5	.4	1.5
	2Q	.6		4.0	.7	3.9
	3Q	1.0		4.8	2.2	6.0
	Mx	1.83	3.88	7.63	3.39	26.03

aximum (Mx), minimum (Mn), and quartile values (1Q, 2Q, 3Q) in weight ercent of uncorrected analyses for argillaceous-arenaceous sediments. ote the following characteristics:

Al_2O_3: $3Q_{G,A} \cong 1Q_S$
Na_2O : $Mx_G >> Mx_{S,A}$; $3Q_S < Mn_G$
K_2O : $3Q_G < 1Q_S$

CaO : $2Q_S << 2Q_{G,A}$
MgO : $3Q_A < 1Q_{G,S}$; $Mx_A < 3Q_{G,S}$
Fe_2O_3 : $3Q_G < 1Q_S$; $3Q_G > 1Q_A$

indices are also summarized in this figure and the location of commo sedimentary and metamorphic minerals is indicated. Certain well-kno "average" or "composite" analyses of sedimentary rocks are also plot

The quartzose sandstones and subgraywackes are rather well distinguished by their high SiO_2- and low Al_2O_3-content. Arkose is distinguished by having the highest alkali to alumina ratio, commonly accompanied by a high K_2O-content. The three points which indicate a l alkali to alumina ratio suggest either that part of the feldspar was calcic plagioclase or that the feldspar has been extensively altered clay minerals. The graywackes and the shales are distinguished from each other by their $K_2O:Na_2O$ ratio and the alkali-alumina ratio. Admittedly there is rather large overlap, but, considering the loose classification of sedimentary rocks and the poor descriptions accompanying the analyses, the observed grouping seems significant enough to be useful.

It is instructive to correlate these results with the mineralogy of sedimentary rocks and with the observed stability of various minerals during weathering and sedimentation. Different minerals vary greatly in their stability during weathering, transportation, and diagenesis. Olivine, pyroxene, amphibole, biotite, and calcic plagioclase are relatively unstable during sedimentary processes and do not appear as important constituents of sedimentary rocks. Sodic plagioclase, orthoclase, muscovite, clay minerals, and quartz are progressively more stable. A sediment may be said to be "compositionally mature" when it contains only the most stable minerals, and "texturally mature" when it is well sorted, lacking a wide range of grain sizes.

Graywacke and arkose are two highly compositionally immature sediments whose mineral composition, because of their immaturity, reflects two contrasting source areas (Dapples and others, 1953, p. 297-302; Folk, 1954, p. 353-354; Pettijohn, 1954, p. 360-361). Graywacke consists of minerals characteristically derived from a predominantly metamorphic source area and is particularly characterized by the presence of rock fragments. Arkose consists of minerals derived from a predominantly granitic or gneissic source area and is particularly characterized by the presence of potassium feldspar. Both of these sediments are commonly texturally immature, containing a wide range of grain sizes.

Quartzose sandstone, shale, and chemical precipitates of dissolved material are both compositionally and texturally mature. Other sedimentary rocks lie between these five end-members. Inasmuch as compositional and textural maturity do not necessarily accompany one another, some sediments may be compositionally mature, but texturally immature (ie. a rock composed of quartz grains in a matrix of fine-grained quartz and clay minerals or a quartzose sandstone with a calcite matrix).

In sedimentary rocks sodium is principally found in plagioclase feldspar ($Na_2O = Al_2O_3$, mole percent) and to a much smaller extent in clay minerals, especially those of the montmorillonite group ($Na_2O < Al_2O_3$, mole percent). Although sodium-bearing clay-minerals are known, the potash-content of the clay minerals found in sedimentary rocks is much higher than the soda-content and most of the sodium in sedimentary rocks is in albite. Thus, a sedimentary rock with a relatively high soda-content will also have an alkali-alumina ratio approaching 1:1 indicative of the high albite-content. On the other hand a sedimentary rock with a high potash-content may be indicative of either a high feldspar or a high illite-muscovite content and the alkali-alumina ratio may range widely.

Quartzose sandstones are obviously characterized by their high silica content, while shales are characterized by their low alkali-alumina ratio and low soda-content, reflecting the low feldspar and high clay content. The abundance of potash feldspar in arkoses is reflected in a higher alkali-alumina ratio and in potash being more abundant than soda. It was noted earlier that graywackes are characterized chemically by their high soda content and by an alkali-alumina ratio less than that of arkoses but higher than that of shales. Low-grade metamorphic rocks commonly contain nearly pure albite, whereas igneous rocks commonly contain more calcic plagioclase. Albite is less apt to be broken down chemically during sedimentation than is a more calcic plagioclase. Thus a graywacke, characteristically derived from a metamorphic terrane, generally contains more albite—and hence more soda—than an arkose.

The effect of a carbonate or iron oxide cement does not mask the above characteristics.

Explanation to Figure 7.

1Q, 2Q, 3Q are quartile values.

● Individual analyses

○ Composite or averaged analyses

1. Composite analysis of 253 sandstones (Clarke, 1924, p. 547).
2. Composite analysis of 371 sandstones used for building stone (Clarke, 1924, p. 547).
3. Tyrrell's composite analysis of 30 graywackes (Pettijohn, 1949, p. 250).
4. Unweighted average of 10 graywacke analyses (Pettijohn, 1949, p. 250).
5. Unweighted average of 33 Precambrian slates (Nanz, 1951, p. 1466).
6. Composite analysis of 51 Paleozoic shales (Clarke, 1924, p. 552).
7. Composite analysis of 27 Mesozoic and Cenozoic shales (Clarke, 1924, p. 552).
8. Unweighted average of 30 roofing slates (Albee).
9. Unweighted average of 35 ceramic shales (Lamborn and others, 1939, p. 20).

Supplementary characteristics of shales and slates:

C is reported in many shales, but in very few sandstones.

Black shale has a high C and high S and FeS_2 content with $FeO >> Fe_2O_3$ (Wt. %).

Red shale has a low C and low S and FeS_2 content with $Fe_2O_3 >> FeO$ (Wt. %).

Comparison of the composition of the micaceous schists to that of sedimentary rocks

The comparison of the chemical composition of the micaceous schists with the chemical compositions of argillaceous and arenaceous sedimentary rocks is complicated by the loss (or gain) of certain constituents during metamorphism. The most important of these are water and carbon dioxide. The segregation of quartz into lenses and veins may leave a rock less rich in silica. This process is quite troublesome, inasmuch as the silica-content is one of the most useful indices to the type of sedimentary rock, and can probably be overcome only by an estimate of the amount of quartz in lenses and veins in each individual outcrop.

These two effects can be shown by a plot of Al_2O_3 (weight percent) against the weight percent of all other constituents except silica, ($100-SiO_2$). The weight percent of alumina and silica were calculated graphically from estimated modes of the micaceous schists from the Hyde Park quadrangle. The great variation in the composition of these schists made this sufficiently accurate. The sediments have a Al_2O_3: ($100-SiO_2$) ratio with a median of about .45, most ranging between .35 and .55. Those points with a ratio less than .35 were found to contain rather large quantities of carbonate, iron oxide, or graphite and pyrite. The schists have a higher ratio, ranging around .55. This is slightly exaggerated as no estimate could be made of uncombined water in the schist. The loss of silica cannot affect the ratio of Al_2O_3 to ($100-SiO_2$). This loss is reflected by higher alumina and lower silica values in the schists than are found in any sediment. The shift toward higher alumina and lower silica values is clear in Figure 8.

Comparison of silica-content, and of any other absolute values, must be made with considerable discretion. However, they do provide a useful qualitative guide. The percentage of albite in a rock (assuming all the sodium is in albite) is slightly greater than ten times the weight percent of soda. The graywackes, whose analyses were cited, would have an albite content ranging from 15 to 60 percent with a median at about 27 percent, while the shales, whose analyses were cited, would have a median albite content of about 6 percent and a maximum of about 40 percent. The third quartile value for shale is only about 15 percent, which is about the minimum for graywacke. All of these percentages would be increased in a metamorphic rock due to loss of volatiles and silica.

The metamorphic rocks can be plotted on the Al_2O_3-Na_2O-K_2O triangle from the relative percentages of the minerals which contain these constituents. Estimated modes from various formations in the Hyde Park quadrangle have been plotted on Figure 9. The alumina and silica values are also given. Comparison of Figure 7 and Figure 9 makes it clear that arkoses and graywackes were not abundant in this area, and that the sediments must have been predominantly shales, quartzose sandstones, and subgraywackes. The Stowe formation was predominantly nongraphitic shale, the Ottauquechee formation was made up of interbedded quartzite and black shale, while the sediments of the Camels Hump group were more varied, consisting predominantly of shale, but probably also subgraywacke and some quartzose sandstone and graywacke. One thin section from the Moretown formation suggests an arkose composition.

REGIONAL RELATIONS

General statement

The Green Mountain anticlinorium is the major structural feature of Vermont, extending north-northeast from the Massachusetts border the full length of Vermont and about 50 miles into Quebec, a total distance of about 210 miles.

In the southern part of Vermont Precambrian rocks are exposed in the core of the anticlinorium. Slightly metamorphosed rocks of Lower Cambrian to Middle Ordovician (with the possible exception of the Middle Cambrian) age, the Western Vermont sequence, rest with pronounced angular unconformity upon the west side of the Precambrian core. To the east the Precambrian core is overlain, again with pronounced unconformity by a second thick series of metasedimentary and metavolcanic rocks, the Eastern Vermont sequence. The upper part of the eastern sequence is sparsely fossiliferous, including rocks of Middle Ordovician or younger age. The lower part of the series is nonfossiliferous and is believed to include Cambrian and early Ordovician rocks. A third sedimentary series, the Taconic sequence, also of Cambrian and Ordovician age, rests with structural discordance upon Ordovician rocks of the Western Vermont sequence.

Although presumably equivalent in age the composition and degree of metamorphism is very different in the three sequences. The Eastern Vermont sequence consists largely of metamorphosed argillaceous, arenaceous, and volcanic rocks; the Western Vermont sequence consists largely of slightly metamorphosed dolomite and limestone with minor quartzose sandstone and shale; and the Taconic sequence consists largely of [illegible]ly metamorphosed argillaceous rocks with minor arenaceous and carbonate rocks.

The Eastern and Western Vermont sequences are rather uniform along the strike and individual formations can be traced from the Massachusetts border the full length of Vermont and some distance into Canada. The major problems confronting geologists working in Vermont are the correlation of these three sequences and the relation of the Eastern Vermont sequence to the New Hampshire sequence. Rapid east-west facies changes, extensive unconformities, and thrust faults of unknown extent have all been utilized in the correlation of these sequences (Osberg, 1952, p. 68; Billings and others, 1952, p. 18; Brace, 1953, p. 54-57; Billings, 1956, p. 89-99). However, at the present time there is no generally accepted correlation.

J. B. Thompson (Billings and others, 1952, p. 14-21) has summarized the geology of southern Vermont (south of 44° north latitude) and has discussed these major problems. No similar summary and map are available for nothern Vermont. The geology of north-central Vermont, insofar as it is known, will be summarized, the correlations from north-central Vermont north into Quebec and south to southern Vermont will be indicated, and the many problems which still remain unsolved will be pointed out.

Much of what is known of the geology of north-central Vermont is a result of the work of W. M. Cady, the author, and A. H. Chidester, all of the United States Geological Survey; published sources are indicated on Plate 4. The Western Vermont sequence in northern Vermont rather well known due to the work of Cady (1945), Booth (1950), Shaw (1949), Stone (1951), and numerous earlier workers. Areas in Quebec, just north of the International Boundary, have been mapped by Clark (1934, 1936), Ambrose (1942, 1943), and Cooke (1950, 1954). Unfortunately, the work in Quebec was not detailed and the interpretations and stratigraphic names vary widely. Several of the formation boundaries recognized in Vermont are interpreted as faults in Quebec. Cady and the author have visited Cooke and Clark in the field several times and, although their stratigraphic names will be utilized, the sequence of formations is revised to fit the writer's correlation with the Vermont sequence.

Stratigraphy of Northern Vermont

The Precambrian core of the Green Mountain anticlinorium plunges north under the Paleozoic schists in the area southwest of Warren and Precambrian rocks do not appear anywhere to the north along the axis. In northern Vermont the Precambrian rocks do not form a natural division between the eastern and western sequences. The Western Vermont sequence in northern Vermont can, for convenience, be defined as those rocks above but excluding the lowest fossiliferous rocks—the Cheshire quartzite and its equivalents. A strip, which lies just west of the main chain of the Green Mountains, is not included in either sequence. It includes the basal Cambrian (Ꞓb on Plate 4), the Tibbet Hill schist (Ꞓbt), the "Jay Peak series" (JPS), and the Precambrian rocks (PꞒ) west of Warren. In this report this strip will for convenience be designated as the North-central Vermont sequence. The Camels Hump group and the rocks east to the eastern limit of the Gile Mountain formation may be defined as the Eastern Vermont sequence in northern Vermont.

Eastern Vermont sequence in northern Vermont

The gradual development of the standard section and the north-south correlation of the Eastern Vermont sequence in northern Vermont is summarized in Tables 14 and 15. The units will be described briefly here; many of them are present in the Hyde Park quadrangle and have been described in detail.

Richardson (1902, 1906, 1919a, 1919b, 1924, 1927), Perry (1927, 1929), Currier and Jahns (1941), and Doll (1945) recognized and named the major rock units of Eastern Vermont (see Table 14). Richardson traced an unconformity, now recognized as the base of the Shaw Mountain formation, the whole length of Vermont. This corresponds to the base of the higher "clay slate" of Hitchcock (1861). The Shaw Mountain formation Currier and Jahns, 1941, p. 1496-1499; White and Jahns, 1950, p. 186) consists of quartz conglomerate, sericite schist, and limestone. The formation is thin and is absent at some points. The Shaw Mountain formation can probably be correlated with the Peasely Pond conglomerate and the uppermost part of the Sherbrooke group in Quebec. Throughout this istance the Shaw Mountain formation separates the noncalcareous rocks in the west from the calcareous rocks to the east and forms an extremely important reference point in the north-south correlation of the Eastern Vermont sequence. Currier and Jahns (1941, p. 1500-1501) assigned the

Shaw Mountain formation to Middle Ordovician or younger on the basis of the size of crinoid columnals. Billings (1948; 1956, p. 92) and Boucot and others (1953) have suggested that the Shaw Mountain formation is of lowermost Silurian age.

The rocks overlying the Shaw Mountain formation consist of interbedded impure limestone, gray to black phyllite or slate, and quartzite. In higher metamorphic grades the phyllite becomes a schist containing biotite, garnet, staurolite, and/or andalusite. These rocks have been subdivided on the basis of the relative amount of the various rock types. A rather thin band of black slate or phyllite directly overlying the Shaw Mountain formation has been called the Northfield formation. The Waits River formation consists of impure limestone with interbedded gray to black phyllite while the Gile Mountain formation consists of gray to black phyllite and micaceous quartzite with thin beds of limestone. In southern Vermont the Standing Pond amphibolite lies in the Waits River formation near the contact with the Gile Mountain, but this unit has not yet been traced through northern Vermont and does not seem to be continuous. The rocks above the Shaw Mountain formation correlate with the St. Francis group (Cooke, 1950) and possibly with the upper part of the Glenbrooke group (Clark, 1934; Ambrose, 1942, 1943) in Quebec.

The rocks between the Shaw Mountain formation and the Precambrian rocks are predominantly argillaceous, arenaceous, and volcanic rocks with very few carbonate rocks. The Cram Hill formation (Currier and Jahns, 1941, p. 1495-1496) consists of gray to black, rusty-weathering slate or phyllite, and quartzite with some interbedded metamorphosed volcanics. The Moretown formation (Cady, 1956) consists of a gray-green micaceous quartzite, quartz-muscovite-chlorite-albite-epidote granulite, and greenstone. Porphyroblasts of biotite and garnet have been noted in areas of higher metamorphic grade.

In southern and central Vermont the Cram Hill formation seems to overlie the Moretown formation, but Cady (personal communication, 1954) indicates that in northern Vermont the two types of rock are merely interfingering sedimentary facies and do not maintain a definite stratigraphic position. The Cram Hill (and Moretown) formation, as suggested earlier by Currier and Jahns (1941, p. 1496, p. 1508-1509), seems to be the equivalent of the Beauceville (Magog) slate in Quebec, which contain graptolites of Normanskill age (Clarke, 1934; Ambrose, 1942, 1943).

East of Mansonville, Quebec and extending southward into the United States are large areas of volcanic rock (Bolton group of Cooke, 1950) which have been variously mapped as interbedded volcanic rocks, intrusive rocks, and as post-metamorphic volcanics resting unconformably on all the other rocks. Cady (personal communication, 1954) indicates that such volcanic rocks are interbedded with the Cram Hill and Moretown formations just west of Lake Memphremagog. Therefore, these rocks have been included with the Cram Hill and Moretown formations on Plate 4.

Throughout most of the state there appears to be a gradational contact between the Stowe formation and the Moretown formation. However, in northern Vermont, east of Lowell, these rock units are separated by the Umbrella Hill conglomerate for a distance of about 26 miles. This formation consists of quartz- and slate-pebble conglomerate with interbeds of sericite slate and greenstone. Porphyroblasts of chloritoid are abundant in both the conglomerate and the slate. The formation thins to extinction to the south and to the north gives way to greenstones mapped as the upper part of the Stowe formation. In the Hyde Park quadrangle a probable angular unconformity has been noted, but this relation has not been seen elsewhere.

The Stowe formation (Cady, 1956) replaces the "Bethel schist" of Perry (1929) as the former name was preoccupied. This formation has been traced north, as shown on Plate 4, to the town of Stowe and then north into the Lowell area. The Stowe formation consists of silver-green, quartz-sericite-chlorite schist and phyllite, and greenstones, with some interbedded black phyllite in the upper and lower parts. In higher metamorphic grades the schists contain chloritoid, garnet, and kyanite, and the greenstones are metamorphosed to epidote amphibolites.

The Ottauquechee formation (Perry, 1929, p. 27) has been traced in a narrow band north from south of Warren to the Hyde Park quadrangle. It consists of interbedded black phyllite, massive blue-gray quartzite, and quartz-sericite-chlorite phyllite. Near Eden the Ottauquechee formation is covered by an extensive area of glacial deposits. Its equivalent north of this covered area consists of interbedded black phyllite, pebbly quartzite, and pebbly quartz-sericite-chlorite phyllite. The Ottauquechee formation, which north of Lowell is in part probably equivalent to the Stowe formation, has been traced north to the type locality of the Mansonville slate (Ambrose, 1942) at Mansonville, Quebec and for some distance farther north.

In southern Vermont the Ottauquechee formation is underlain by the Pinney Hollow formation (Perry, 1929, p. 26). This formation is lithologically similar to the Stowe formation; silver-green quartz-sericite-chlorite schist or phyllite with some greenstone and black phyllite. In part of southern Vermont an amphibolite, the Chester amphibolite, lies at the top of the Pinney Hollow formation. The Pinney Hollow formation has been traced northward into the vicinity of Warren (with considerable uncertainty as to the location of its base), but no such unit can be distinguished underlying the Ottauquechee formation for some distance north of the Winooski River. Near Morrisville a narrow band of quartz-sericite-chlorite schist underlies the Ottauquechee formation, but north of the extensive area of cover the Belvidere Mountain amphibolite underlies the Ottauquechee formation. The Belvidere Mountain amphibolite has been traced northward into Quebec, and may possibly be an equivalent of the Chester amphibolite.

In southern Vermont the Hoosac formation (Thompson, in Billings and others, 1952) underlies the Pinney Hollow formation. The Hoosac formation consists of gray or black schist with abundant porphyroblasts of albite, and contains quartz-sericite-chlorite schist, quartzite, dolomite, greenstone, and microcline augen gneiss in various areas. The underlying Tyson formation consists of conglomerate, schist, and "graywacke."

The Precambrian plunges under this sequence west of Warren and the Pinney Hollow, Hoosac, and Tyson formations have not been distinguished in northern Vermont. The Camels Hump group (Cady, 1956) has been defined to include those rocks between the base of the Ottauquechee formation and the Precambrian. The Camels Hump group consists predominantly of a monotonous sequence of interbedded graphitic and non-graphitic, quartz-sericite-chlorite schist and gneiss with porphyroblastic albite and some thin quartzite. In a few areas it contains garnet and biotite. The eastern exposures of the Camels Hump group consist of interbedded graphitic schist, micaceous quartzite, nongraphitic schist, and massive quartzite. In certain areas it is difficult to distinguish the latter lithologic assemblage from the overlying Ottauquechee formation. Detailed petrographic work indicates that there is no significant compositional difference between the two apparently dissimilar lithologic assemblages of the Camels Hump group, and indicates that the difference is predominantly textural. Near Eden the Belvidere Mountain amphibolite is directly underlain by the coarse-grained porphyroblastic albite schist and gneiss. The Sutton schist along the axis of the Green Mountain (Sutton Mountain) anticlinorium in Quebec consists predominantly of albite schist and gneiss for some distance north of the International Boundary. No reliable mapping units have been found in the Camels Hump group in the northern part of the state, although bands of silver-green quartz-sericite-chlorite-magnetite schists have been mapped for limited distances in the Hyde Park area.

Near Lake Memphremagog in Quebec, the Beauceville slates are overlain unconformably by two synclinal troughs of fossiliferous slates (the Glenbrooke group) of Silurian and Devonian age (Clark, 1934; Ambrose, 1942). Cady (written communication, 1955) indicates that lithologic similarities strongly suggest that these rocks are the equivalent of the

Shaw Mountain and succeeding rocks in Vermont.

Western Vermont sequence in northern Vermont

Geologic work in the Western Vermont sequence has been in progress for nearly a century. Much of western Vermont has recently been re-examined in greater detail by Cady (1945), Fowler (1950), Shaw (1949), Stone (1951), Flower, Rodgers, Thompson (Billings and others, 1952), Zen (1955), and MacFadyen (1956).

The Western Vermont sequence in southwestern and west-central Vermont ranges in age from lowest Cambrian to Middle Ordovician. The lower units are predominantly of detrital origin, including quartzite, phyllite, graywacke, and conglomerate, with dolomite in the upper part of the lower Cambrian. The central and largest portion of the sequence consists largely of carbonate rocks with dolomitic limestones dominant in the older formations and calcitic limestones dominant in the younger. Slates and phyllites of Middle Ordovician age rest unconformably upon the carbonate rocks and locally truncate the Paleozoic rocks so that they rest upon the Precambrian basement.

The correlation of the Western Vermont sequence from Massachusetts and New York north into Quebec is rather well established and is summarized in Table 16. Several of the units which Clark (1934, 1936) and Booth (1950) placed in the Lower Cambrian have been shown as Middle or Upper Cambrian to correspond with Shaw's (1949, 1954) work. Many of the units grade laterally northward into units of different lithology (Shaw, 1949; Stone, 1951); several carbonate units gaining slate members or being replaced completely by slate. The whole sequence appears to thicken northward.

The Dunham dolomite and Cheshire (Gilman) quartzite, containing fossils of Lower Cambrian age, are the lowest fossiliferous units which have been traced the length of the state. Indeed, the type locality of the Dunham dolomite is in Quebec and that of the Cheshire quartzite is in Massachusetts.

North-central Vermont sequence

In west-central Vermont the Mendon formation consists from bottom to top of the Nickwacket member (graywacke, quartzite, and conglomerate), Forestdale member (dolomite), and the Moosalamoo member (phyllite and quartzite) (Osberg, 1952, p. 26-36; Brace, 1953, p. 30-34). The Nickawacket member seems to correlate with beds mapped as Cheshire quartzite near North Adams, Massachusetts which contain Olenellus (Osberg, 1952, p. 35; Brace, 1953, p. 33). The Nickawacket, Forestdale, and Moosalamoo members (Mendon "series") correlate with the Pinnacle graywacke, White Brook dolomite, and West Sutton slate respectively of Clark's (1934) Oak Hill series and of Booth (1950).

In northern Vermont and north into Quebec the Pinnacle formation is believed by some to be underlain by a greenstone unit, the Tibbit Hill schist. Clark (1934, p. 10; 1936) and Booth (1950, p. 1155) believe that the Tibbit Hill schist is the oldest unit exposed, but that it is of Paleozoic age. Reconnaissance by Cady and the author suggests the possibility that the Tibbit Hill schist lies within rather than under rocks of typical Pinnacle lithology.

Booth (1950) has attempted to distinguish these lowermost Cambrian formations in the area between the International Boundary and the Winooski River. The West Sutton formation and the Pinnacle formation approac lithologic similarity in this area, consisting of sandy slate or phyllit graywacke, and conglomerate; moreover they merge and are indistinguishab where the White Brook is absent. "This thinning of the (White Brook) do omite westward and southward creates a problem because it is an excellen horizon marker; without it the separation of the West Sutton and Pinnacl is impossible in many places" (Booth, 1950, p. 1147). The eastern limit of Booth's mapping was where "the Pinnacle terrane seems to pass imperceptibly eastward into the more metamorphosed rocks that flank the Green Mountains" (Booth, 1950, p. 1154).

The author's and Cady's reconnaissance indicate that Booth's descri tion is quite appropriate. The author has mapped the western contact of the albite schist of the Camels Hump group north of the Lamoille river. Between that contact and the eastern edge of Booth's mapping (Plate 4) the major rock type is a fine-grained quartz-sericite-chlorite schist or phyllite with abundant highly-chloritic "graywacke", greenstone, and mino black phyllite and quartzite. The "graywacke" is abundant only in the western part of the area. On Plate 4 the areas in which "graywacke" is abundant have been designated as basal Cambrian (Ꞓb on Plate 4); the schistose areas to the east have been designated the "Jay Peak series" (JPS on Plate 4), the term being used without any stratigraphic significance. The line between may represent the contact between the West Sutton and Pinnacle formations, but is at best merely an approximation.

Northwest of Jeffersonville is an area, as yet undefined, of black phyllite, black limestone, and black quartzite (JPSb on Plate 4) which appears to be in a syncline surrounded by the quartz-sericite chlorite schist. Farther northwest of Jeffersonville massive highly chloritic "graywacke" and greenstone are common and are difficult to distinguish from one another. It is likely that this is an anticline of Tibbit Hill greenstone. South of the Lamoille River the "graywacke" is coarser grained and much less chloritic.

Primary evidence for tops of beds--and in places even for bedding--is not abundant and detailed mapping is required in this area. However, a tentative picture is that this whole area consists of the West Sutton and Pinnacle formations with a broad anticline and a broad syncline northwest of Jeffersonville.

Igneous rocks

Numerous bodies of ultramafic rocks are present on the east limb of the Green Mountain anticlinorium, only the largest of which can be shown on the map; a few are also known just west of the Green Mountain axis. Several of the bodies contain dunite cores, but most have been extensively or completely altered to serpentinite, talc-carbonate rock, and steatite. These have been assigned an Ordovician age inasmuch as such rocks are not known to intrude strata younger than Ordovician in western New England or adjacent Quebec.

Most of the granitic rocks in the eastern part of the map (those designated Db on Plate 4) are biotite-muscovite granite, quartz, monzonite, or granodiorite. These rocks are clearly younger than the metasedimentary rocks and apparently late-tectonic. They are similar to the granitic rocks of the New Hampshire magma series (Billings, 1934).

Several alkalic stocks (designated Mwh on Plate 4) are clearly younger than both the deformation and the regional metamorphism. These are largely syenite and granite and are similar to the rocks of the White Mountain magma series of New Hampshire (Billings, 1934). Some post-metamorphic, fine-grained lamprophyre dikes are present, but are too small to be shown on the map.

Structure

It is evident from the preceding discussion that the Green Mountain anticlinorium is the major structure in this area. Over much of its length the anticlinorium is a broad structure with several important axes, increasing the difficulties of correlation across the anticlinorium. The location of the axis is not known in detail in the area just north and south of the Winooski River. The shiny quartz-sericite, chlorite schists of the "Jay Peak series" cross the axis in an axial depression near Montgomery Center. Structural relations on the southern edge of this axial depression suggests the presence of a local thrust fault, and the syncline, containing rocks of the "Jay Peak series" east of the anticlinorial axis seems to be overturned to the west. This syncline, the Foot Brook syncline, contains rocks tentatively identified as the Ottauquechee and Stowe formations in the Hyde Park quadrangle. The rocks of the "Jay Peak series" apparently thin to extinction in the southern part of the Foot Brook syncline.

West of Lowell generally east-west trending and vertically dipping bedding schistosity is prevalent over a considerable area. This attitude is well exposed at Hazens Notch in cliffs over 300 feet in height, but the nature of the major structure is not yet known.

A general anticlinal zone must be present in the area between the Foot Brook syncline and the main band of the Ottauquechee formation, but the axis has not been accurately located. A group of large folds greatly broadens the outcrop breadth of the Stowe formation in much of northern Vermont. Northeast of Lowell the Stowe formation crops out in a large north-plunging anticline.

The folds in the Hyde Park quadrangle are demonstrated by the complexity of the Ottauquechee-Stowe contact. The large ultramafic body in the southeast corner of the Jay Peak quadrangle is folded so that its exposed width is much greater than its actual thickness.

The great variation in the width of the Cram Hill-Moretown formations northwest of Hardwick is as yet unexplained. It is probably due principally to the dying out of folds, but stratigraphic convergence and the disconformity at the base of the overlying Shaw Mountain formation may be partially responsible. The structure above the Shaw Mountain formation in northern Vermont is as yet poorly known. The Gile Mountain formation is present in a broad syncline west of the main belt of its exposure (Dennis, 1956, p. 35-39).

In Quebec the base of the Mansonville formation and the top of the Sherbrooke group (Shaw Mountain formation in part) are interpreted as faults. In the opinion of the author and of others this is definitely not the case in Vermont nor does it appear to be so at several areas visited in Quebec.

West of the axis of the Green Mountain anticlinorium the same general picture of broad shallow, but highly plicated, folds appear. Reconnaissance westward from the axis of the Green Mountain anticlinorium in the Hyde Park quadrangle to the small area of black phyllite, black limestone, and black quartzite (JPSb) northwest of Jeffersonville indicates that the beds grow successively younger from the axis westward. Farther west it is believed that the rocks are successively older for about two miles west to the area of greenstone. Booth (1950, p. 1154) indicates that this greenstone is part of the Tibbit Hill greenstone exposed in an anticline. The rocks west of this greenstone have been grouped with the West Sutton and Pinnacle formations by Booth (1950).

Both Booth (1950, p. 1155) and Clark (1934, p. 10-11) concluded that the Tibbit Hill greenstone was the oldest rock exposed and that the rocks east of the Tibbit Hill greenstone band were more highly metamorphosed and finer grained equivalents of the Oak Hill series. Clark interpreted the eastern border of the Tibbit Hill greenstone as a fault (the Brome thrust), but Booth (1950, p. 1155) says "... the Tibbit Hill --Green Mountain rocks contact indicates the 'Green Mountain schist terrane' passes into that of the Pinnacle formation without any break." However, at no place has the relation of the Tibbit Hill greenstone to the schists east of it been definitely shown. It is entirely possible, as suggested by minor structures, that the Tibbit Hill greenstone is underlain by rock of Pinnacle - West Sutton lithology.

The Pinnacle - West Sutton facies is limited on the west by the fossiliferous Gilman and Dunham formations. The fossiliferous sequence of northwestern Vermont is cut by several thrust faults (Plate 4).

Problems of the geology of Northern Vermont

Before attempting a discussion of possible correlations between the Eastern and western Vermont sequences, it is necessary to mention several problems within the areas which are rather well mapped.

Foremost among these problems is that of the validity of the north-south correlation of the Eastern Vermont sequence, uncontrolled by paleontologic evidence as it is. It is recognised, of course, that boundaries of lithic units need not necessarily correspond with time surfaces. There can be no doubt of the north-south correlation of the disconformity at the base of the Shaw Mountain formation and it seems likely that it corresponds rather closely with a time surface. The rather distinctive Stowe formation contains volcanic rocks throughout the state and its upper and lower contacts correspond rather closely to those defined for the "Bethel schist" in the type locality. The absence of the Stowe formation in Quebec has long been a problem in attempts to correlate the section in Quebec with that in southern Vermont. However, its gradual disappearance in northernmost Vermont is rather well understood now.

The absence of a continuous Pinney Hollow type of rock below the Ottauquechee has made it difficult to delineate the base of the Ottauquechee formation north of the Winooski River. It is believed that this has been accomplished successfully in the main band of the Ottauquechee formation, but it is recognized that it has not been possible to delineate accurately between the Ottauquechee formation and the Camels Hump group in the Foot Brook syncline. An additional problem is the correlation of the Ottauquechee formation across the extensive area of cover near Eden. However, detailed mapping in this area has convinced the author that the correlation is correct.

The correlation of the Stowe and Ottauquechee formations in the Foot Brook syncline with the main bands of those formations has been discussed previously. A large amount of time has been spent trying to show conclusively the relation between the rock mapped as Stowe formation in the Foot Brook syncline and the two south-extending "prongs" of the quartz-sericite-chlorite schist of the "Jay Peak series" east of Montgomery Center. However, the work was inconclusive and the relations shown on Plate 4 are based primarily upon lithologic comparisons rather than upon actual traced contacts. The rock in the "prongs" is a silver-green quartz-sericite-chlorite-magnetite schist, some beds approaching a micaceous quartzite; while the Stowe formation in the Foot Brook syncline is a silver-green, quartz-sericite-chlorite-magnetite schist, commonly highly micaceous and containing porphyroblasts of chloritoid.

The major structural features are known generally, but have not yet been fully delineated. This is especially true of the unmapped areas east of the Shaw Mountain formation. The variation in breadth of outcrop of the Cram Hill - Moretown formations is not yet explained. Problems in the unmapped areas in the North-central Vermont sequence will be discussed in the next section.

Correlation between the Eastern and Western Vermont sequences

Any attempt to correlate the Eastern and Western Vermont sequences in northern Vermont should wait upon the completion of geologic mapping in the rock designated here as the North-central Vermont sequence. However, consideration of possible correlations does provide a useful guide to direct further work.

The pre-Shaw Mountain rocks of the Eastern Vermont sequence are Cambrian and/or Ordovician. The Cram Hill - Moretown formation is probably of Middle Ordovician age, hence, the base of the Middle Ordovician must lie somewhere to the west. The Shaw Mountain, Northfield, and Waits River formations are not older than Middle Ordovician nor younger than late Devonian (White and Jahns, 1950, p. 191-192), but are probably Middle Ordovician to Silurian (Boucot and others, 1953).

The pre-Shaw Mountain rocks of the Eastern Vermont sequence may represent the whole section from basal Cambrian (or even late Precambrian) to Silurian. However, given the proper combination of unconformities, these rocks may represent only basal Cambrian or only Middle Ordovician. In the Western Vermont sequence extensive unconformities are known above the Precambrian and at the base of the Whipple marble and Hortonville slate of Middle Ordovician age. Extensive unconformities are present in the Eastern Vermont sequence at the base of the Shaw Mountain formation and above the Precambrian core. Unconformities, probably important only locally, have been noted at the top of the Stowe formation and at the base of the Northfield slate (White and Jahns, 1950, p. 187).

Three general correlations seem possible on the basis of the paleonlologic evidence alone. The interpretation most generally accepted is that the pre-Shaw Mountain formations are equivalent in age to the Cambro-Ordovician sequence west of the Green Mountains. Such an interpretation implies a rapid facies change from the carbonate sequence west of the Green Mountains to the argillaceous-arenaceous sequence with intercalated volcanic rocks on the east. Support for this facies change is found in the Ludlow area in the presence of quartzite and dolomite which are lithologically similar to parts of the Lower Cambrian sequence west of the Green Mountains. Further support is found in the fact that the Northwestern Vermont sequence is intermediate in lithologic character between the eastern and western sequences of central Vermont.

An alternative interpretation suggested by Thompson (Billings and others, 1952, p. 18) is "that the pre-Cram Hill formations are all of Middle Ordovician age and to be correlated only with the rocks above the mid-Ordovician unconformity west of the mountains." Such an interpretation requires that an unconformity, as yet unrecognized, extends north from the point west of Warren where the Precambrian rocks plunge under the younger rocks. Thompson (personal communication, 1954) has suggested that such an unconformity may easily have remained unnoticed within the "graywacke" considered as Pinnacle formation just west of the Green Mountains, inasmuch as little detailed work has been within this critical strip.

A third possible interpretation is that the lowest Cambrian clastic rocks of the Western Vermont sequence are the equivalents of most of the pre-Cram Hill eastern sequence and that the carbonate rocks of higher

Cambrian and lower Ordovician are represented by an unconformity, as yet unrecognized, somewhere in the eastern sequence. Similarly it is also possible that the basal Hortonville unconformity is located within the Eastern Vermont sequence rather than at its base or at the base of the Shaw Mountain formation.

A more detailed correlation can be attempted only by the comparison of detailed lithologic characteristics, or by tracing units along strike to points where they "bridge" the Green Mountain anticlinorium in an axial depression. The Oak Hill series of Clark (1934, 1936) has been traced by Cooke (personal communications and field conferences from 1950 to 1954; 1954) nearly 50 miles north of the International Boundary to the St. Francis River. In the vicinity of Richmond, Quebec, on the St. Francis River, the Oak Hill series "bridges" the Green Mountain anticlinorium. Reconnaissance by Cady and the author has suggested that the Sweetsburg slate, the Scotsmore quartzite, and possibly the Oak Hill slate and Dunham dolomite, are the equivalents of the Mansonville formation. If this correlation is valid, the Mansonville (and the Ottauquechee) formation extends from Lower Cambrian to Upper Cambrian in age. Recently more detailed work by P. H. Osberg (personal communication, 1956; 1956) has confirmed this tentative correlation.

The Foot Brook syncline offers another possible "bridge" for a more detailed east-west correlation. As noted earlier the rock in the Foot Brook syncline has been correlated with the Stowe and Ottauquechee formations to the east, and reconnaissance suggests that the "Jay Peak series" is the equivalent of the West Sutton schist and the Pinnacle "graywacke." However, it has not been possible to demonstrate conclusively the relation between the "Jay Peak series" in the north end of th Foot Brook syncline and the "Stowe formation" in the southern end. The rock in these two areas is lithologically somewhat dissimilar, but in any possible interpretation rather rapid facies changes must be allowed.

The difficulties in the Foot Brook syncline are compounded by the presence of two possible correlatives of the Ottauquechee formation on the west flank of the Green Mountain anticlinorium. One is a narrow band (JPSg on Plate 4) of black phyllite and thin massive quartzite with an underlying greenstone, separated from the albite schist of the Camels Hump group by a silver-green quartz-sericite-chlorite-magnetite schist. The other is the "syncline" of black phyllite, black quartzite, and black limestone northwest of Jeffersonville (JPSb on Plate 4). The correlation of either of these with the "Ottauquechee formation" in the Foot Brook syncline requires that the underlying silver-green quartz-sericite-chlorite-magnetite schist of the "Jap Peak series" thin to extinction eastward as such a unit does not appear in the Foot Brook syncline. Although on the basis of lithologic comparison the black phyllite, black quartzite, and black limestone in the syncline northwest of Jeffersonville (JPSb on Plate 4) may possibly correlate westward with the Hortonville formation, it seems more likely that they correlate westward with the Skeels Corners slate (Sweetsburg slate), Rugg Brook formation (Scotsmore quartzite), and Parker slate (Oak Hill slate). If both this correlation and the northward correlation of the Ottauquechee are correct, then the rocks in the syncline near Jeffersonville also correlate with the Ottauquechee formation.

Another possibility is that the correlation of the rocks within the Foot Brook syncline with the Stowe and Ottauquechee formation is incorrect. The Pinney Hollow formation is lithologically similar to the Stowe formation. The rock within the Foot Brook syncline may be the equivalent of the Pinney Hollow, thinning to extinction eastward, so that it does not appear in the eastern part of the Hyde Park quadrangle. Such a correlation would suggest that the Ottauquechee formation is absent in the Foot Brook syncline, but appears in the "syncline" (JPSb) northwest of Jeffersonville. It would also suggest that the unit shown as Ottauquechee in the Foot Brook syncline correlates westward with JPSg.

Several different correlations are possible in attempting to correlate westward from the Hyde Park quadrangle by comparison of lithologic characteristics. No one of these can be favored on the basis of what is now known. It is believed that detailed work around the two "bridges" will eventually disclose the proper correlation. Further detailed mapping within the North-central Vermont sequence will resolve many of these problems. It is likely that the White Brook dolomite will be recognized within the "Jay Peak series" and will greatly aid the correlation. Osberg has clarified the structure and stratigraphy near Richmond, Quebec, but more work is needed just north of the International Boundary before this knowledge can be completely utilized in northern Vermont.

MINERAL RESOURCES OF THE BEDROCK

Talc

Talc is the most important mineral resource of the bedrock of the Hyde Park quadrangle. Since about 1900 several small mines have been worked intermittently in a large body of ultramafic rock, 2.6 miles N. 39° E. of the center of Johnson village. One of these, the Johnson Mine (Chidester and others, 1951, p. 17-20; Billings and Chidester, 1944) of the Eastern Magnesia Talc Company, as of September 1954 was producing ground talc at the rate of about 22,000 tons annually. The ultramafic body is about 3,500 feet long and 200 feet wide. It is almost completely altered to talc-carbonate rock, but several small masses of serpentinite remain at the south end of the mass. The carbonate, predominantly magnesite, ranges from 25 to 45 percent. Some steatite occurs along the borders. The ore, which is suitable for use as ground talc, is ground at a mill in Johnson village.

A small abandoned talc mine (Chidester, and others, 1951, p. 20) 0.75 mile S. 60° W. of the center of Johnson village, contains talc-carbonate rock of good quality. The caved open pit is about 125 by 40 feet, and underground workings are known to have extended several hundred feet to the south.

Talc-carbonate rock is exposed for about 200 feet along the bed of a brook, 0.45 mile S. 30° W. of the summit of Bowen Mountain (Eden). About 0.90 mile S. 18° W. of Bowen Mountain talc-carbonate rock is exposed for about 150 feet in a brook and in scattered outcrops for 450 feet north from the brook. The talc-carbonate rock is very white, of good quality, and resembles that at the Johnson Mine. The area between the two groups of exposures is covered with surficial deposits, but both occupy approximately the same stratigraphic position. About 700 feet upstream from the second body, talc-tremolite rock within a tremolitic greenstone contains small quantities of tremolite asbestos ($nZ = 1.631$). The exact extent of this rock could not be determined because of cover, but it does not appear to be very great.

Other mineral resources

Two small limestone quarries and two abandoned lime kilns (Dale, 1915, p. 12-14) are located 1.20 miles N. 11° W. of Hillside School (Johnson). The beds of impure marble are thin and can not be traced far along the strike.

Small veinlets of galena and sphalerite and associated pyrite and barite are exposed in a small prospect pit, 1.55 miles S. 5° W. of Cleveland Corners (Hyde Park). Veinlets of pyrite were noted at several places in the calcareous schists in this vicinity.

Disseminated pyrite and some chalcopyrite occur over a width of about 15 feet in greenstone and a black and white banded quartzite in the western part of Wolcott township, 0.89 mile N. 88° E. of the summit of Toothacher Hill (Hyde Park). Pits, trenches, and shafts opened in a search for copper, are now abandoned and filled with water.

Ilmenite and kyanite are exposed in and around the vicinity of a small prospect pit, 1.25 miles S. 65° E. of Delano School (Elmore) and a little northeast of a wood road shown on the map. A band of schist about 200 feet wide contains about 35 percent of partly altered kyanite and about 15 percent of ilmenite along with chlorite, muscovite, partly altered garnet, and sericite and chloritoid formed during retrograde alteration. The ilmenite also occurs in veins; these are exposed in the pit which is believed to have been opened for iron. Beneficiation tests run on samples from this area indicated that the kyanite cannot be concentrated by methods found applicable to commercial kyanite ores, chiefly because inclusions of ilmenite are locked in the kyanite crystals, and also because the kyanite is more or less altered to sericite.

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Table 2

TABLE 2. Estimated Modes of the Camels Hump Group

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Quartz	42	40	50	50	50	10	45	35	60	75	94	7	30	
Albite	10	20	13**	9	15	10		20	10	5		8	15	10
Sericite	25*	25'	25	25	35	55'	50	25	13	7	2	65	35	66
Chlorite	20	10	10	10	10	15		15	13	3		18	15	20
Biotite			< 1	1										
Magnetite	3	1	1							2			< 1	
Graphite				3	1	4	5	3	3		4			
Carbonate		< 1				< 1				8			4	
Epidote group		3	1					1	< 1	tr				
Pyrite			tr	2	2				< 1				< 1	1
Ilmenite					} 2			} 2				2		3
Sphene						1								
Rutile	tr			< 1				< 1				tr		
Apatite	tr	tr	< 1	< 1				tr	< 1	< 1		tr	tr	1
Tourmaline	tr	< 1	< 1	< 1		5		< 1			tr			1
Locations***	W-0.27, 2.05	MC-2.3, 3.2	C-1.2, 4.2	MC-1.05, 4.68	M-3.35, 4.90	MC-2.24, 1.58	C-3.45, 3.47	MC-3.08, 4.71	C-0.47, 3.36	M-3.52, 1.13	-3.62, 1.96	MC-3.03, 3.72	C-0.9, 5.3	MC-2.92, 3.61

* About 25% of sericite is paragonite.
** About 1% of almandite garnet enclosed in albite porphyroblasts.
*** Explanation on page 5.
' Muscovite only, no paragonite present.
1-3 Quartz-sericite-chlorite schist or gneiss with porphyroblastic albite.
4-6 Graphitic, quartz-sericite-chlorite schist with porphyroblastic albite.
7-9 Graphitic quartz-sericite-chlorite-albite schist.
10 Micaceous quartzite.
11 Massive dark-gray quartzite.
12-14 Quartz-sericite-chlorite schist.

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Table 16

	Southwest Vermont Fowler (1950)	Westcentral Vermont Cady (1945) Brace (1953)	Northwest Vermont Shaw (1949, 1954) Booth (1938, 1950)	Southern Quebec - Clark (1934): Oak Hill slice	Southern Quebec - Clark (1934): Rosenberg slice	Southern Quebec - Clark (1934): Phillipsburg slice
Middle Ordovician	Hortonville slate	Hortonville slate			Various shale units of the Rosenberg slice	Various carbonate units of the Phillipsburg slice
	Whipple marble	Glens Falls limestone				
	Orwell limestone	Orwell limestone				
	Middlebury limestone Beldons formation Burchards limestone	Middlebury limestone Beldons formation Crown Point limestone				
		Bridport dolomite	? ? ?			
Lower Ordovician	Bascom formation	Bascom formation Cutting dolomite	Boundary slate			
	Boardman formation	Shelburne marble	Grandge fm. Highgate fm. Morses Line sl.			
			Gorge formation			? ? ?
Upper Cambrian	Clarendon Springs dol.	Clarendon Springs dol.		? ? ?		
	Danby formation	Wallingford member Danby formation	Hungerford slate Rockledge cgl. Saxe Brook dol. Skeels Corners slate Mill River conglomerate	Sweetsburg slate	Dolomite and shale	
Middle Cambrian			St. Albans shale Rugg Brook fm. Boucher ls.	Sootamore quartzite		
Lower Cambrian	Winooski dolomite Monkton quartzite	Winooski dolomite Monkton quartzite	Parker slate	Oak Hill slate		
	Dunham dolomite Cheshire quartzite	Dunham dolomite Cheshire quartzite	Dunham dolomite Gilman quartzite	Dunham dolomite Gilman quartzite	? ?	
	Mendon "series"	Moosalamoo phyllite Forestdale dolomite Nickwocket graywacke and conglomerate	West Sutton slate White Brook dolomite Pinnacle graywacke	West Sutton slate White Brook dolomite Pinnacle graywacke Call Mill slate		
			Tibbit Hill schist	Tibbit Hill chlorite schist		
	Precambrian	Precambrian	? ? ?	? ? ?		

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Table 16. Correlation of the Western Vermont Sequence

Figure 8
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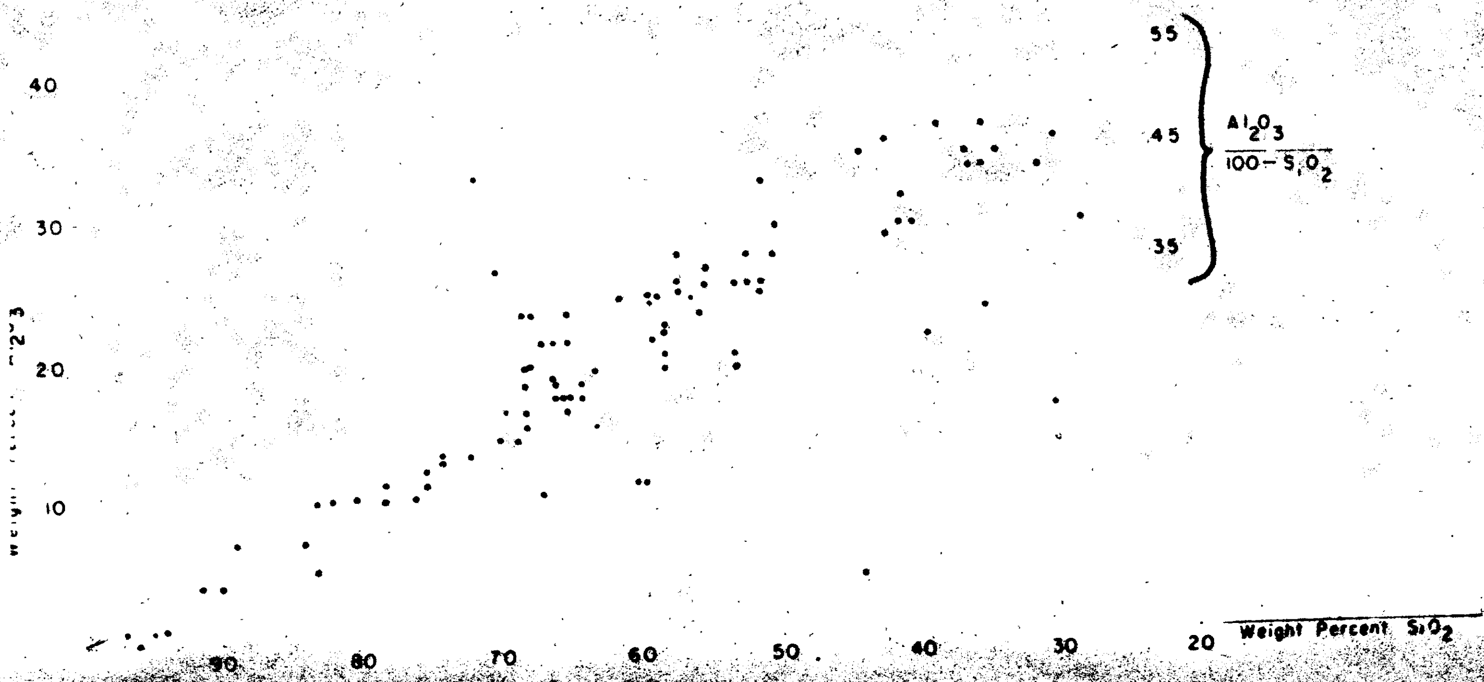
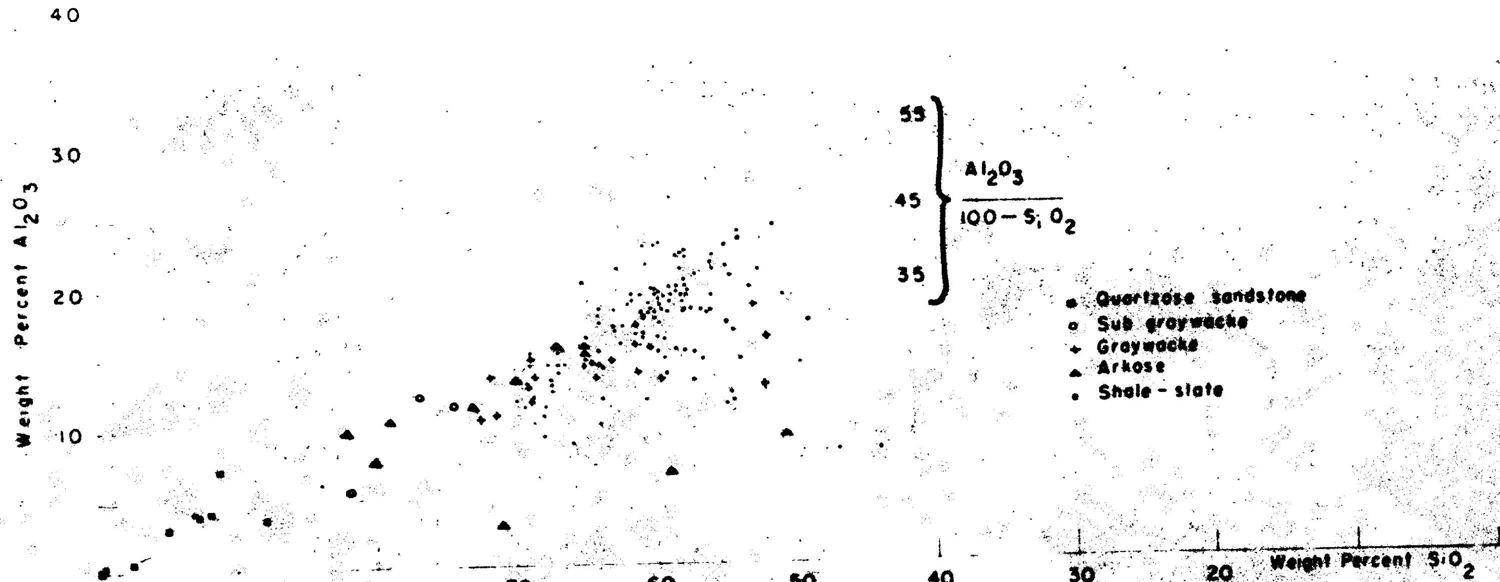


Micaceous schists from Hyde Park quadrangle.



Argillaceous - arenaceous sedimentary rocks

Relation of weight percent SiO_2 and weight percent Al_2O_3 for sedimentary rocks and for micaceous schists from the Hyde Park quadrangle

Figure 6
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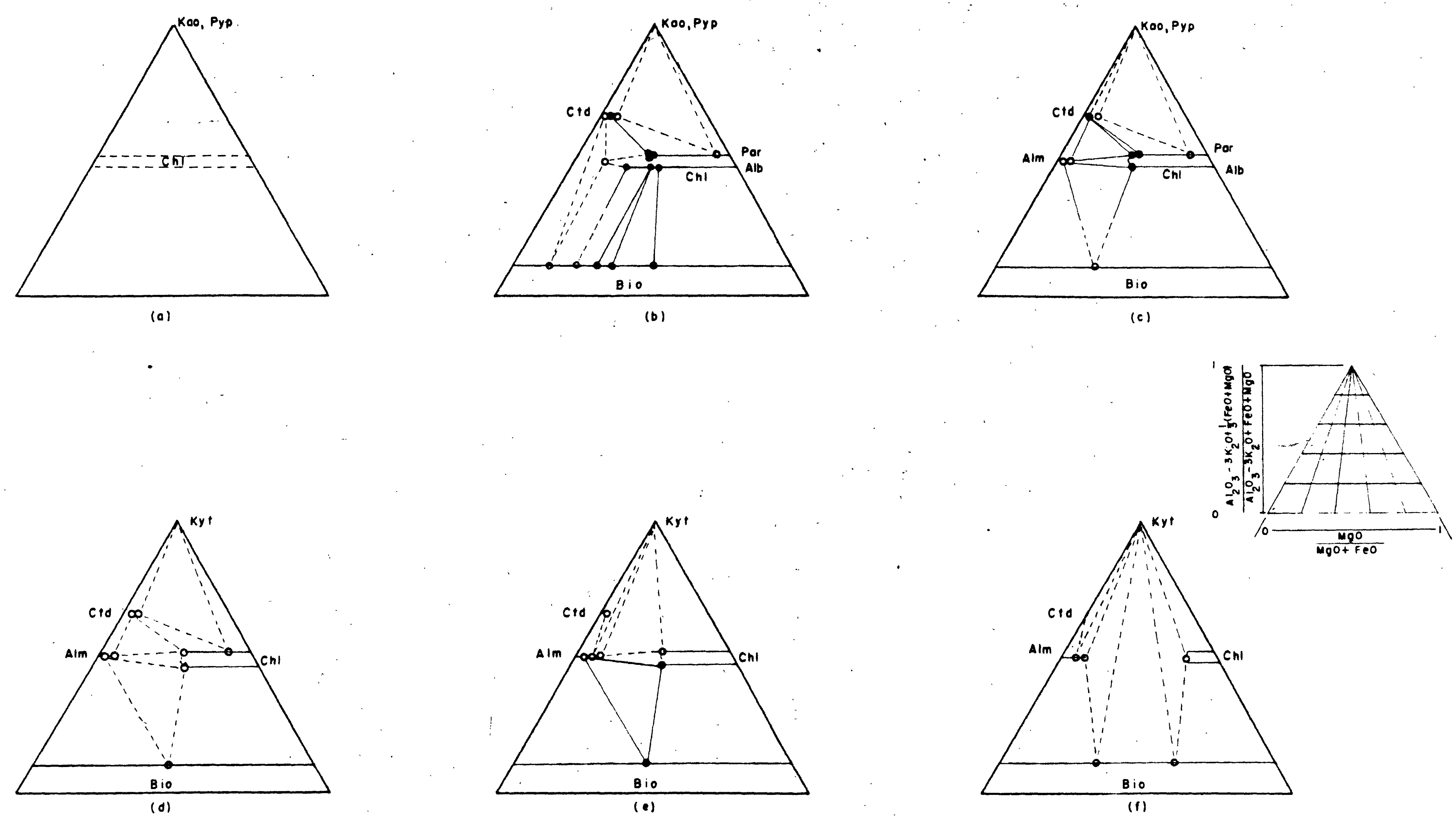

Phase Diagrams , showing mineral assemblages of quartz and muscovite schist stable at successively higher metamorphic grades in Hyde Park quadrangle. Construction of the diagrams is explained in the text and on Figure 5.

Figure 7
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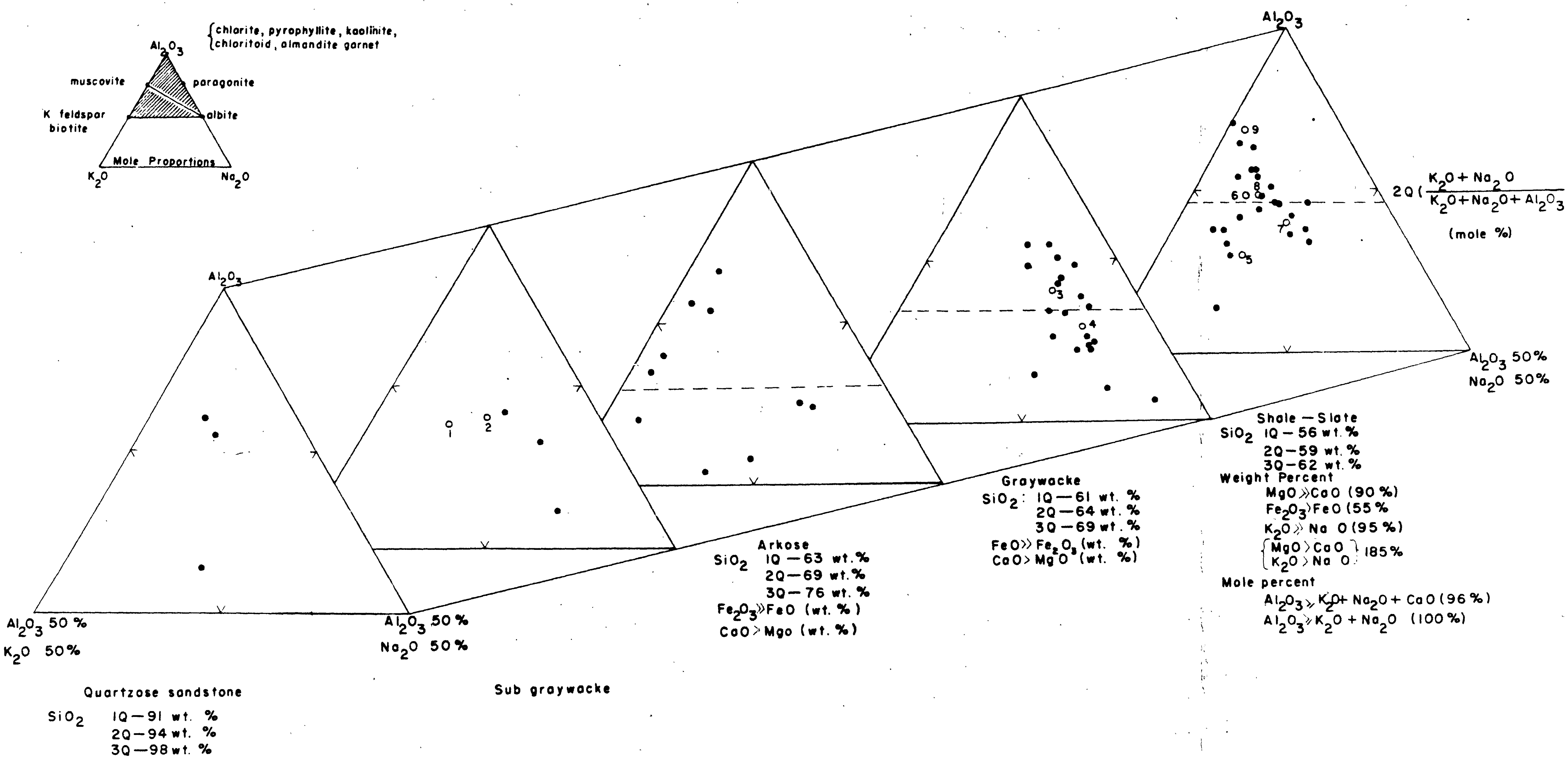
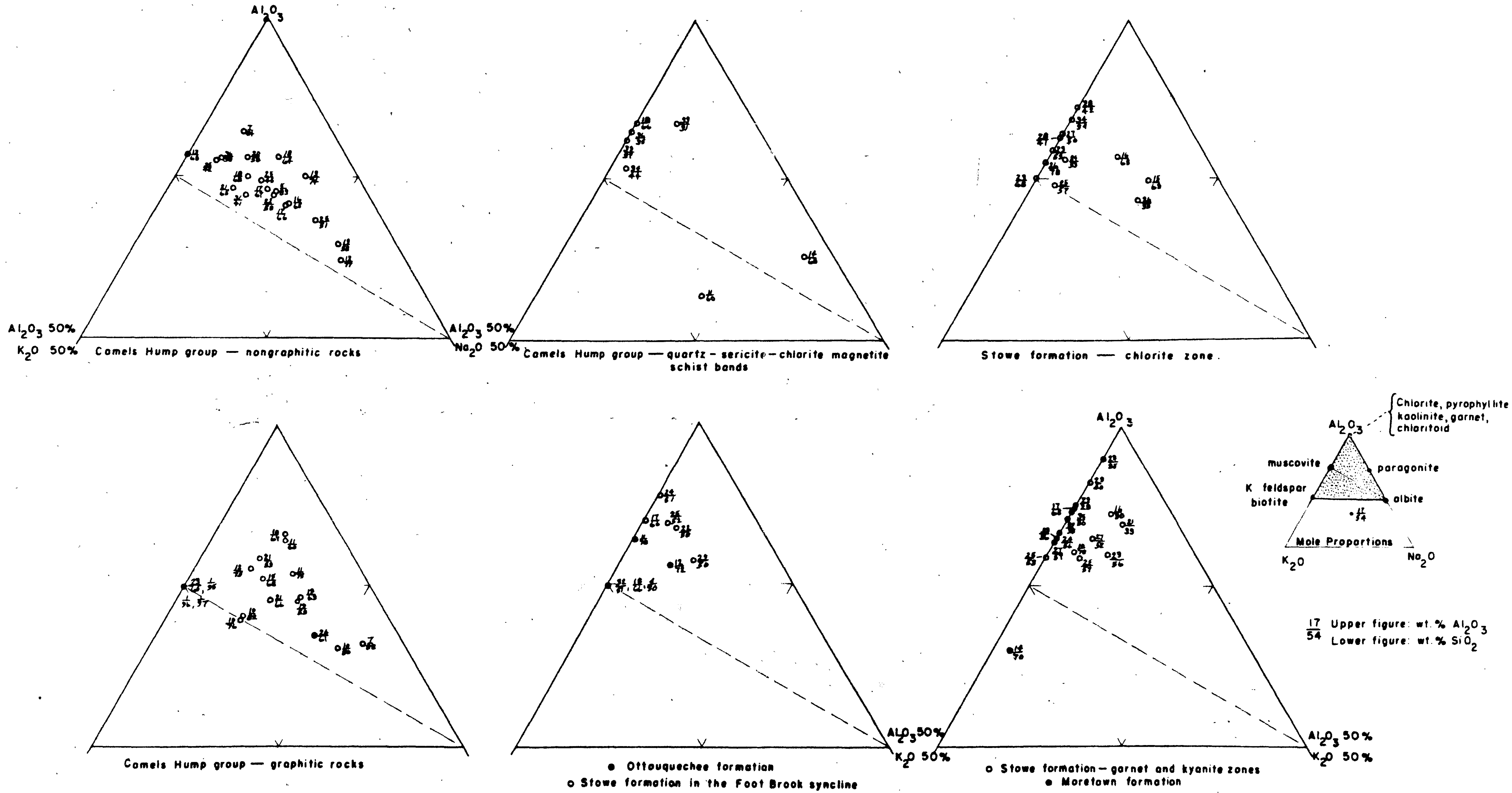


Chemical characteristics of argillaceous—arenaceous sedimentary rocks

Figure 9
57-1



Chemical characteristics of micaceous schist from Hyde Park Quadrangle

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Tabl

Number	Optic sign	Y (Beta) index of refraction	Fe_2O_3 (weight percent)	* $K_2O + Na_2O$ (weight percent)	Mole percent $\frac{FeO+MnO}{FeO+MnO+MgO+CaO}$	Mole percent $\frac{FeO+2(Fe_2O_3)+MnO}{FeO+2(Fe_2O_3)+MnO+MgO+CaO}$	Mole percent $\frac{Al_2O_3}{Al_2O_3+SiO_2}\cdot 4$	Mole percent $\frac{Al_2O_3+Fe_2O_3}{Al_2O_3+Fe_2O_3+SiO_2}\cdot 4$	Occurrence	Source
1		1.642	2.20	.00	67	68	1.46	1.52	Argillaceous slate.	Zen, 1955, p. 27 (197-2)
2		1.636	1.70	.00	64	65	1.47	1.51	Argillaceous slate.	Zen, 1955, p. 29 (212-2A)
3	-	1.628	.72	.13	50	50.5	1.35	1.37	Chlorite-epidote-calcite-albite schist.	Osberg, 1952, p. 59
4	-	1.638	2.08	.00	61	63	1.37	1.42	Quartz-albite-chlorite-muscovite schist.	Osberg, 1952, p. 49
5	-	1.627	2.43		46	48	1.41	1.47	Argillaceous schist; garnet zone.	Skehan, 1953, p. 138
6	+	1.620	2.82	.31	41	43	1.27	1.34	Greenstone.	Tilley, 1938, p. 497-511
7	+	1.622	3.88	.00	44	51	1.31	1.41	Greenstone; albite-chlorite schist.	Hutton, 1938, p. 198 (2586)
8	+	1.623	3.49	Tr	43	49	1.20	1.29	Albite-epidote-chlorite-actinolite-calcite schist.	Hutton, 1938, p. 198 (2718)
9	+	1.608	3.85		25	30	1.37	1.59	Quartz vein in phyllite.	Melon, 1938, p. 19
10	+	1.602	1.93		20	23	1.41	1.46	Quartz vein in phyllite.	Melon, 1938, p. 23
11	-	1.638	4.48		60	63	1.52	1.63	Quartz vein in phyllite.	Melon, 1938, p. 25
12	+	1.636	4.86		52	56	1.44	1.55	Quartzite.	Melon, 1938, p. 31
13	+	1.615	1.64	.03	39	40	1.21	1.26	"Black wall zone" at serpentinite contact.	Chidester, written communication (ID-50-2009)
14	±	1.626	3.4	.2	49	52	1.35	1.44	Albite dike with muscovite and sphene in metamorphic rock.	Agar and Emendorfer, 1937, p. 77
15		1.663	1.99		91	91	1.80	1.85	Cavities in bauxite.	Lyamina and Soboleva, 1937 (M.A.-8-334)
16	+	1.585	1.46	.04	17	19	1.27	1.30	Greenstone.	Simpson, 1936, p. 3
17		1.592	2.04	1.24	13	16	.96	1.03	"Black wall zone" at serpentinite contact.	Phillips and Hess, 1936, p. 340
18	+	1.613	2.76	.62	30	33	1.21	1.25	"Black wall zone" at serpentinite contact.	Phillips and Hess, 1936, p. 340
19	+	1.618	1.90		35	37	1.27	1.32	Chlorite schist associated with feldspathic amphibolite and cut by quartz-copper veins.	Orcel, 1928 (93)
20	-	1.651	1.86		74	75	1.38	1.43	Chlorite vein cutting hematite bed.	Orcel, 1928 (123)
21	-	1.667	.67	.35	91	91	1.49	1.51	In "iron bed" with hematite and magnetite.	Orcel, 1928 (128)
22		1.588	2.00		8	11	1.03	1.08	Tremolite-chlorite schist.	Orcel, 1928 (170)
23		$N_{av.}$ 1.580	0		0	0	1.00	1.00	Synthetic; pure Mg clinochlore.	Yoder, 1952, p. 576
24	+	1.581	.20		3	3.4	1.50	1.51	Associated with small serpentinite body.	Orcel, 1928 (31)
25	+	1.562	1.66		0	1	1.42	1.45	Albite pegmatite in serpentinite.	Orcel, 1928 (32)
26	+	1.581	.24	.49	1	1	1.49	1.49	Albite pegmatite in serpentinite.	Orcel, 1928 (33)
27	+	1.576	1.43		.6	1	1.36	1.38	Albite pegmatite in serpentinite.	Orcel, 1928 (34)
28	+	1.570	.45	.30	.2	.5	1.44	1.44	Albite pegmatite in serpentinite.	Orcel, 1928 (35)
29		1.593	.67		7	7.6	1.34	1.36	With corundum and spinel.	Orcel, 1928 (57)
30	+	1.588	1.04		9	10	1.40	1.41	With corundum and spinel.	Orcel, 1928 (58)
31	+	1.587	2.56		9.6	13	1.31	1.37	With corundum and spinel.	Orcel, 1928 (59)
32	+	1.594	1.45		15.4	17.4	1.41	1.44	With corundum and spinel.	Orcel, 1928 (67)
33	+	1.593	.57		17	18	1.36	1.37	With margarite.	Orcel, 1928 (68)
34	+	1.605	2.86		34.6	38	1.39	1.46	?	Orcel, 1928 (78)
35	+	1.616	2.00		38	40	1.24	1.29	In quartz veins cutting complex of aplite dikes, amphibolite gneiss and diabase dikes.	Orcel, 1928 (95)
36	-	1.649	--	1.38	94	94	1.47	1.47	?	Orcel, 1928 (132)

* $K_2O + Na_2O$ in weight percent is given where reported unless the analysis was corrected by the author, in which case the corrected analysis was used. Where a chlorite analysis contains much alkali, the sample probably contained impurities.

Table 10. Chlorite analyses and optical data.

	Hitchcock and others (1861)	Richardson and others (1902, 1906, 1919, 1919b, 1924, 1927)	Perry (1929) Whittle (1894)	Currier and Jahns (1941)	Doll (1945)	Present usage
Middle Ordovician to Devonian(?)	Calciferous-mica schist				Meetinghouse slate (1945)	Meetinghouse slate
					Gile Mountain schist (1945)	Gile Mountain formation
		Waits River limestone phase (1907) interbedded with Randolph phyllite phase (1925)	Waits River limestone and Randolph phyllite	Waits River formation	Standing Pond amphibolite Waits River formation	Standing Pond amphibolite Waits River formation
	Clay slate	Memphremagog slate (1907)		Northfield slate (1941)		Northfield formation
	Talcose schist	"Irasburg conglomerate"		Shaw Mountain formation (1941)		Shaw Mountain formation
Middle Ordovician(?)		Missisquoi group (1919, 1923)	Missisquoi group	Cram Hill formation (1941)		Cram Hill formation Moretown formation (Cady, 1956)
Cambrian to Middle Ordovician		Bethel schist (1925)	Bethel schist			Stowe formation (Cady, 1956)
	clay slate clay slate		Ottauquechee phyllite and quartzite (1929)			Ottauquechee formation
	Talcose schist		Pinney Hollow schist (1929)			Camels Hump group (Cady, 1956) Various subdivisions in different areas shown in table 15
	gneiss		Upper quartzite Middle quartzite, dolomite, and conglomerate Albite mica schist Lower quartzite and dolomite "Older Cambrian rocks"			
			Mendon series (1894)			
Precambrian			Mt. Holly series (1894)			Mt. Holly "series" or complex"

Table 14. Development of the rock units of Eastern Vermont sequence. Date in parenthesis after names still in use gives date of introduction.

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Table 15

	Ludlow quadrangle	Rutland, Woodstock, and Hanover quadrangles	Rochester, Randolph, and Strafford quadrangles	Lincoln Mountain, Barre, and East Barre quadrangles	Camels Hump, Montpelier, and Plainfield quadrangles	Mt. Mansfield, Hyde Park, Hardwick, and St. Johnsbury quadrangle	Jay Peak, Irasburg, and Memphremagog quadrangles	Eastern Townships of Quebec
	Thompson (1950, 1952)	Brace (1953), Thompson (1952), Chang (1950), Lyons (1955)	Osberg (1952), Thompson (1952), Doll (1945), White and Jahns (1950)	Cady, Albee, and Murphy, White and Jahns (1950)	Cady (1956)	Albee; Dennis (personal communication)	Albee, Cady, Chidester, Doll (1951)	Cooke (1951), Ambrose (1942), Clark (1934)
Middle Ordovician to Devonian(?)	Gile Mountain formation	Gile Mountain formation	Gile Mountain formation	Gile Mountain formation	Gile Mountain formation	Gile Mountain formation	Westmore formation	St. Francis group
	Standing Pond amphibolite	Standing Pond amphibolite	Standing Pond amphibolite	Standing Pond amphibolite	? ?	Standing Pond amphibolite	?	(Tomifobia slate)
	Waits River formation	Waits River formation	Waits River formation	Waits River formation	Waits River formation	Waits River formation	Barton River formation Irasburg conglomerate member Ayers Cliff formation	(Glenbrooke group in synclines separated from St. Francis group)
	Northfield formation	Northfield formation	Northfield formation	Northfield formation	Northfield formation	Northfield formation	Northfield slate	
	Shaw Mountain formation	Shaw Mountain formation	Shaw Mountain formation	Shaw Mountain formation	Shaw Mountain formation	Shaw Mountain formation	Shaw Mountain formation	Sherbrooke group (Peaseley Pond (agl)
Middle Ordovician(?)	Cram Hill formation Moretown formation	Barnard gneiss Missisquoi group	Missisquoi group	Cram Hill formation Moretown formation	Moretown formation	Cram Hill and Moretown formations	Cram Hill and Moretown formations	Beauceville group (Magog slate)
		"Bethel formation"				Umbrella Hill formation		?
Cambrian to Middle Ordovician	Stowe formation		Stowe formation	Stowe formation	Stowe formation	Stowe formation	Stowe fm	
	Ottauquechee formation	Ottauquechee formation	Ottauquechee formation	Ottauquechee formation	Ottauquechee formation	Ottauquechee formation	Ottauquechee formation	
						Belvidere Mtn. amphibolite		?
	Pinney Hollow formation	Pinney Hollow formation	Pinney Hollow formation					
			Granville formation					
	Plymouth member	Plymouth member	Battell member	Camels Hump group	Camels Hump group	Camels Hump group	Camels Hump group	Sutton Mountain schists
	Hoosac (Grahamville formation)	Grahamville formation	Monastery formation					
	Tyson formation	Tyson formation	Tyson member		Not exposed	Not exposed	Not exposed	Not exposed
	Saltash formation	Saltash formation	"Mt. Holly complex"					
Precambrian	Mt. Holly series	Mt. Holly series	Mt. Holly complex	Mt. Holly complex				

Table 15. Correlation of rock units of the Eastern Vermont Sequence from south to north.