Petrology, Structure, and Genesis of the Asbestos-Bearing Ultramafic Rocks of the Belvidere Mountain Area in Vermont

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1016

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March 3, 1849

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By A. H. CHIDESTER, A. L. ALBEE, *and* W. M. CADY

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A study of the origins of some ultramafic igneous rocks,
of their alteration products—serpentinite,
chrysotile asbestos, steatite, talc-carbonate rock,
and carbonate-quartz-rock—and of the
contact rock associations

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PETROLOGY, STRUCTURE, AND GENESIS OF THE ASBESTOS-BEARING ULTRAMAFIC ROCKS OF THE BELVIDERE MOUNTAIN AREA IN VERMONT

By A. H. Chidester, A. L. Albee, and W. M. Cady

ABSTRACT

The Belvidere Mountain area in north-central Vermont contains bodies of ultramafic rock which vary widely in size and petrographic character. The bodies within this small area show virtually the entire range of characteristics displayed within the belt of ultramafic rocks that traverses central Vermont from Massachusetts to Canada. The ultramafic rocks are emplaced in metamorphosed sedimentary and volcanic rocks characteristic of eugeosynclinal zones in the east limb of the Green Mountain anticlinorium. The hornblende isograd separates rocks of the greenschist facies in the southwest corner of the area from those of the epidote-amphibolite facies to the northwest.

Two sets of folds, their associated structural elements, and intrusive relations of the ultramafic bodies dominate the structural features of the area. Folds of an early set are closed and plastic in style, their form surfaces are bedding, and they have an associated axial-plane transverse schistosity. Folds of a later set are open and parallel in style and commonly have an axial-plane slip cleavage. Transverse schistosity, which in many places is virtually parallel to bedding, constitutes the form surfaces of the folds. In most places in northern Vermont, the effects of the early folds are not apparent in the pattern of map units, but in the Belvidere Mountain area, both the early and the late folds strongly affect the pattern.

Contacts of ultramafic bodies are predominantly parallel to, but locally crosscut the transverse schistosity; they transect, but locally are controlled by, the earlier folds. The larger ultramafic bodies are massive in the center and sheared and schistose near the margins; smaller bodies are entirely of sheared serpentinite. Layering in the massive zones is warped by the later folds. Schistose serpentinite is commonly strongly folded and locally has slip cleavage similar to the slip cleavage associated with later folds in the adjacent country rock.

The main ultramafic body in the Belvidere Mountain area is a warped lens 3,400 m (11,000 ft) long and 1,800 m (6,000 ft) wide and as much as 450 m (1,500 ft) thick. It is elongate in the direction of dip, which overall is moderate to the east. The central part is predominantly massive dunite and peridotite grading outward into massive serpentinite. Layering is prevalent in both the unaltered and the serpentinized massive rock and reflects color differences which can commonly be correlated with variations in original content of olivine, pyroxene, chromite, and magnetite. The central massive zone is surrounded by sheared serpentinite. Chrysotile asbestos occurs sparsely in cross-fiber veins in the massive dunite and serpentinite and as commercial concentrations of slip-fiber in much of the highly sheared serpentinite.

In several places in the main body, dunite and serpentinite are altered to talc-carbonate rock and carbonate-quartz rock. Commonly, these zones of alteration are grossly tabular and transect layered dunite and serpentinite. More highly quartzose rock in the central part of such zones commonly grades irregularly outward into nonquartzose talc-carbonate rock. Schist adjacent to these bodies of talcose rock is locally and irregularly altered to steatite and chlorite rock and commonly contains conspicuous segregations of sericite and albite in the outer parts of the alteration zones. In a few places, talc-carbonate rock and steatite form small masses at the margins of the body, adjacent to the contact with schist. At places, the schist is altered to blackwall chlorite rock for a few centimeters outward from the contact, and an irregular thickness of steatite intervenes between talc-carbonate rock and blackwall chlorite rock. Elsewhere at the contacts of the main body, serpentinite is separated from country rock only by a narrow zone of serpentine-chlorite rock a few centimeters thick and a zone of rodingite a few centimeters to a few decimeters thick.

Smaller ultramafic bodies contain a central core of serpentinite which is entirely schistose or which has only a small central massive zone. None of the smaller bodies is bordered by serpentine-chlorite rock and rodingite. Those emplaced in schist contain an outer shell of steatite and are bordered by thin shells of blackwall chlorite rock; those in amphibolite contain, additionally, tremolite in the outer margins of the steatite zone or contain only an outer shell of tremolite bordered by chlorite rock.

The most common theories of origin advocated in recent years for ultramafic rocks are (1) accumulation by fractional crystallization of a complex magma, and (2) variants of formation by partial fusion of the upper mantle. Both categories present serious difficulties in the way of acaccounting for field and petrologic relations of alpine-type ultramafic rocks. More recently, a theory of derivation from the upper mantle based on the concept of sea-floor spreading has been applied with remarkable success to alpine-type ultramafic rocks of the ophiolite suite. Modified to fit the situation in the Appalachian belt, the concept appears to explain the geologic relations in the Belvidere Mountain area, the range of structural and petrologic relations shown by

alpine-type ultramafic rocks, and their distribution patterns in the Appalachian belt.

On this premise, the geologic and petrologic relations of the ultramafic rocks in the Belvidere Mountain area lead to the following account of their geologic history.

At the beginning of the Paleozoic, northwestern Africa and eastern North America were joined. Mantle upwelling along a rift beneath the continental crust in the zone now occupied by the Appalachian orogen initiated the early Paleozoic eugeosyncline. Mafic lavas, formed by partial melting at depth in the upper mantle, erupted into the accumulating eugeosynclinal sediments above the rift. Mechanical interaction at the boundary between mantle and continental crust resulted in solid masses of upper mantle rocks being injected into the crust; resulting thermal changes led to their partial serpentinization. Uplift in the center of the eugeosyncline above the rift, and other factors, led to early folding of the eugeosynclinal rocks and to kneading of the ultramafic rocks through the wet eugeosynclinal pile. Progressive serpentinization, alternating with partial dehydration, occurred in response to periodic movement and shearing and facilitated tectonic transport of the solidified mass. Emplacement of the ultramafic rocks at Belvidere Mountain took place in the Early Ordovician at the end of the early period of folding, beneath a considerable cover. Minor pervasive serpentinization of dunite continued and was accompanied by fracture filling to form cross- and slip-fiber chrysotile asbestos veins, picrolite veins, and associated other veins, near the climax of the late period of folding in the Late Devonian.

Rodingite was formed by the combined effects of thermal metamorphism, desiccation by the dunite, and the addition principally of Ca from the dunite. Serpentine-chlorite rock was formed next by replacement of rodingite during falling temperatures and as the result of Mg metasomatism.

Talc-carbonate rock, carbonate-quartz rock, steatite, and rocks of related contact-rock associations were formed during regional metamorphism.; CO_2 metasomatism locally altered dunite, peridotite, and serpentinite to talc-carbonate rock and carbonate-quartz rock. The alterations in dunite and peroditite, which were accompanied by widespread loss of Mg and Si, led to extensive steatitization and chloritization of nearby country rock; alteration of serpentinite to talc-carbonate rock was essentially isochemical and independent of other changes. Contact rocks of the steatite-blackwall and the tremolite rock–chlorite rock associations were formed by metamorphic differentiation in systems virtually closed with respect to the cations.

INTRODUCTION

Ultramafic rocks in the vicinity of Belvidere Mountain, in north-central Vermont, contain the largest known reserves of chrysotile asbestos in the eastern United States. Though production is small compared with that of the largest producers in Quebec, these deposits supplied the major portion of the crysotile asbestos produced in the United States during the period 1950–60, which included the period of this investigation.

The Belvidere Mountain area encompasses about 25 km ^{2} in the towns (townships) of Eden and Lowell, Lamoille and Orleans Counties, Vt. (fig. 1; pl. 1). The dominant topographic feature is Belvidere Mountain, a prominent peak of 1,022 m (3,353 ft) elevation that marks an eastward offset of the northernmost segment of the Green Mountains. Total relief in the area is a little less than 700 m (2,300 ft). The mountain slopes are generally steep; in several places they consist of a series of nearly vertical cliffs as much as 30 m (100 ft) high. The northern and northeastern parts of the area drain into the Missisquoi River, the southern and southwestern parts into the Lamoille River.

The weather in summer is moderate, and rainfall is commonly abundant in the early part of the season. Winters are generally severe; snowfall is normally high, and temperatures as low as −20° to −40°F (−29° to −40°C) are not unusual. Most of the area is heavily timbered. A small proportion of the southeastern and southwestern parts of the area is in open pasture and cultivated fields.

The area is accessible from Hyde Park, Vt., by way of State Route 100 northeastward to Eden Mills, thence by hard-surfaced secondary road. Morrisville, the shipping point on the St. Johnsbury and Lamoille County Railroad, is about 27 km to the south. North Troy, on the Canadian Pacific Railroad, is about 30 km to the northeast.

PREVIOUS WORK

Little detailed geologic study had been done in the Belvidere Mountain area prior to the present investigation by the U.S. Geological Survey. C. H. Hitchcock (1861, p. 527) mentioned the occurrence of asbestos in the vicinity of Belvidere Mountain and in neighboring areas. Kemp (1901a, b) briefly described and called attention to the economic possibilities of the deposits. Marsters (1904) described the "serpentine belt" of Lamoille and Orleans Counties, which includes the Belvidere Mountain area, with particular reference to the occurrence and mode of origin of the asbestos deposits; and in 1905 he published an account of the geology of the Belvidere Mountain area. Bain (1932; 1936; 1942) and Keith and Bain (1932) described and discussed occurrences of ultramafic rocks and asbestos deposits in the Belvidere Mountain area in general articles concerned with the mode of origin of serpentine and of chrysotile asbestos. Bowles (1955, p. 22–23) described geologic relations and discussed the commercial aspects of the asbestos deposits. Many of the Vermont State Geologist Reports (Vermont State Geologist, 1898–[1947?]), v. 3, 4, 6, 7, 8, 12, 13, 17, 19, 20, 22, 23, 24, and 25) contain accounts

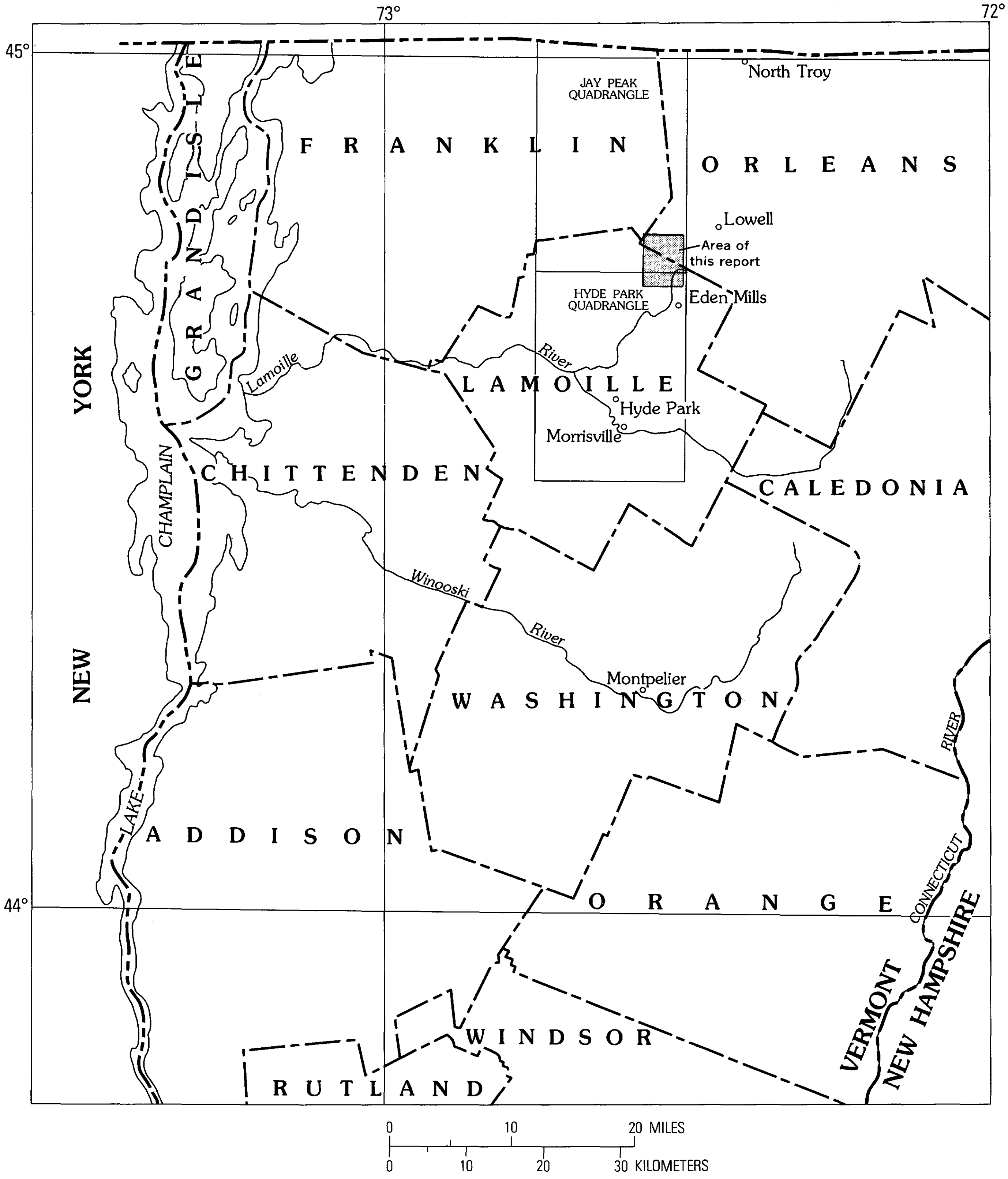

FIGURE 1.—Index map of northern Vermont, showing the location of the Belvidere Mountain area (shaded).

of historical interest concerning exploration, development, and mining activities in the Belvidere Mountain area.

PRESENT INVESTIGATIONS

The geologic investigations carried out by the Geological Survey in the Belvidere Mountain area were part of a program of mapping the regional geology and the asbestos-bearing ultramafic rocks and their associations from Belvidere Mountain northward to the Canadian border. The program was an outgrowth of a similar study in the southward extension of the belt, in which talc deposits of economic importance are associated with the ultramafic rocks. (See Chidester and others, 1951; Chidester and others, 1952a, b; Chidester, 1953; Cady, 1956; Albee, 1957; Chidester, 1962; and Cady and others, 1962.) The objectives were to determine the regional and local geologic setting, the structural controls, the metamorphic relations, and the mode of origin and geologic history of the asbestos-bearing ultramafic rocks.

The investigations in the northern belt were carried out during the period 1951–60. W. M. Cady, project chief, and A. H. Chidester were permanent members of the project staff throughout the study; A. L. Albee, from 1951 through 1954; P. L. Weiss and G. L. Koch served with the project in the summer of 1951; J. C. Ratté and F. N. Houser in the summer of 1952; and C. A. Ratté in the summer of 1953. A report on the geology of the northern belt has been published (Cady and others, 1963), and topical reports growing out of the investigations were prepared by Cady (1960, 1967, 1968, 1969).

The investigations in the Belvidere Mountain area were carried out under especially good conditions for study. Mining operations in the Lowell quarry area and in the C-area (vicinity of the C-area quarry, pl. 1) required stripping of superficial material by hydraulic and hand operations, providing clean exposures of the bedrock surface of the ultramafic rocks, the immediately adjacent country rock, and the contact zones. Successive observations during mining operations allowed us to build a composite picture of successively exposed quarry faces and levels. Drill-hole data disclosed information on the distribution of the various rock types to depths of as much as 360 m (1,200 ft) below the surface. Magnetometer surveys carried out by the mine staff provided much useful information on the location of contacts between ultramafic rocks and country rock, particularly in alluvium- and drift-covered areas in the valley of Corez Pond and on the south slopes of Belvidere Mountain (pl. 1).

Mapping of the Belvidere Mountain area was carried out at several scales and by various methods appropriate to the geology, exposure, and access. Methods included use of planetable, tape/pace/altimeter and compass, taped grids, aerial photographs, and topographic bases. Mapping procedures, credits, and source data are identified on the figures that accompany this report. Data from these various maps were then compiled at a scale of 1:6,000, using supplemental information from regional mapping to fill in some corners of the area.

Mapping of the Belvidere Mountain area was virtually completed in the period of 1951–53, but fieldwork continued intermittently through 1960 to bring maps abreast of stripping and mining operations and to gather information. The final map compilations are composites representing the optimum conditions of exposure for particular parts of the area; plates 1 and 3, in particular, do not represent the condition depicted at any given time.

Figure 2 shows the locations of geologic maps and diagrams included in the report.

ACKNOWLEDGMENTS

The owners and operators of the asbestos mine[1] at Belvidere Mountain were extremely helpful and cooperative throughout the entire investigation; without this wholehearted cooperation the investigation would have been impossible. We are greatly indebted to the staff at the mine: in particular, to Irving E. (Bill) Matthews, superintendent, and his predecessor, Michael J. Messel; James K. Gilmore, formerly chief of exploration, later chief of research; Louis Jordan, chief of exploration; David R. Nichols, geologist; and Wayne Page, chief mining engineer. We are especially indebted to Jordan and Nichols. Both spent many hours with us in discussing problems of the geology of the area, contributed in innumerable ways to our gathering of information, and aided us with plots and other information on the mine surveys and drill-hole data. Nichols participated in and greatly facilitated some of the detailed mapping in the Lowell quarry area; his analysis of magnetometer surveys and drill-hole data in the Corez Pond area and the area to the south was an important contribution to establishing

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[1] At the time of the investigations, the mine was owned and operated by the Vermont Asbestos Mines Division of the Ruberoid Company. In 1967, ownership and operation were acquired by the Industrial Products Division, GAF Corporation. In March 1975, ownership and operation were acquired by the Vermont Asbestos Group, organized by the mine employees when imposition of governmental antipollution control measures resulted in a decision by GAF Corporation to close down the mine.

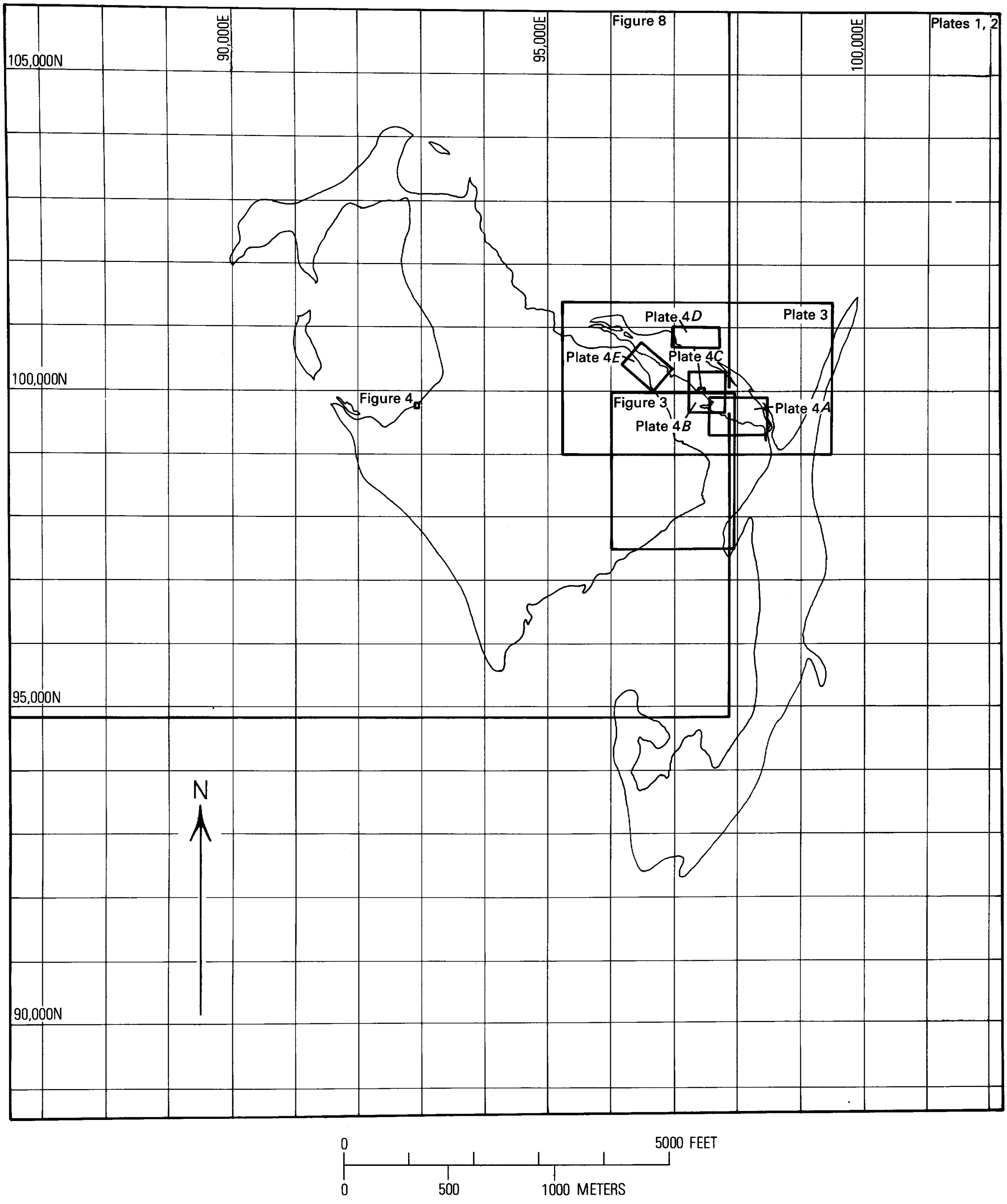


FIGURE 2.—Index to geologic maps and diagrams of the Belvidere Mountain area, Eden and Lowell, Lamoille and Orleans Counties, Vt. For each illustration, the area outlined represents the neatline of the illustration indicated. Grid is based on mine coordinate system. Outline of ultramafic rock bodies is shown.

the map pattern and to the structural interpretation of that complex area, which is largely covered by surficial deposits.

Throughout the investigation, we benefited from field conferences and discussions with many colleagues. Thomas P. Thayer twice spent several days with us in the field. Marland P. Billings, Esper S. Larsen, Jr., John C. Rabbit, Charles A. Anderson, Harold C. Bannerman, George T. Faust, James J. Norton, Allan B. Griggs, Olaf N. Rove, Walter S. White, Quentin D. Singewald, Donald E. White, and James B. Thompson participated at various times in stimulating field conferences. Clifford Frondel, Cornelius S. Hurlbut, and David Seaman generously made available the Harvard collection of minerals from the Lowell and Eden quarries.

GEOLOGIC SETTING

The Belvidere Mountain area is part of the well-known belt of ultramafic rocks within the crystalline rocks of the Appalachian Mountain system; this belt extends from Alabama to Newfoundland. Known commercial deposits of chrysotile asbestos, of which those at Belvidere Mountain are the southernmost, are limited to the northern part of the belt. The deposits in Quebec, a few miles to the north, are among the largest in the world.

The geologic map of Vermont (Doll and others, 1961) depicts the distribution of ultramafic rocks in Vermont, their stratigraphic and structural relations, and their tectonic and metamorphic setting. A report on the Missisquoi Valley area (Cady and others, 1963) describes the belt that contains the Belvidere Mountain area and discusses the local geologic setting. A study of the geology of north-central Vermont (W. M. Cady and others, unpub. data) and a regional tectonic synthesis of northwestern New England and adjacent Quebec (Cady, 1969) describe and discuss in detail the broader regional features of stratigraphy, structure, metamorphism, and tectonic relations. Here, only a brief summary of the regional relations will be given.

The Belvidere Mountain area is in a broad belt of eugeosynclinal rocks of Cambrian and Ordovician age (Cady and others, 1963, p. 7) in the east limb of the Green Mountain anticlinorium. (See inset, Tectonic map of Vermont, Doll and others, 1961.) The predominant regional structural trend is a little east of north, and planar elements (slip cleavage, schistosity, and bedding) are predominantly vertical or dip steeply east or, less commonly, west; local departures from these attitudes are principally near the noses of small folds a few decimeters in amplitude. The regional trend of linear elements (fold axes, lineations) is premodinantly a little east of north, and the plunge is gentle to moderate south or north, parallel to the Green Mountain anticlinorial axis. Locally, folds of an earlier regime are refolded by, and have linear elements that trend about at right angles to, the Green Mountain folds.

In contrast to the rest of the ultramafic belt in Vermont, structural elements in the vicinity of Belvidere Mountain and northward to about the vicinity of Jay Peak deviate conspicuously from the general regional pattern. Bedding and schistosity are steep or vertical and trend largely northwest, east, or northeast, and many of the linear elements reflect the unusual attitudes of the planar features. Boundaries between lithologic units in this area generally parallel the trends of structural elements, and reflect the pattern of folds of the earlier regime.

The rocks of Vermont may be divided into three principal metamorphic belts that trend a little east of north through the State. (See Metamorphic map of Vermont, inset *in* Doll and others, 1961. The metamorphic inset map of Vermont incorrectly shows rocks west of the Champlain thrust in the chlorite zone, whereas they are unmetamorphosed.) A narrow belt in the eastern part of the Champlain Valley east of the Champlain thrust is entirely within the chlorite zone. A broad central belt is predominantly in the biotite zone, but several small areas centered chiefly along anticlinal axes and on small plutons are in the garnet and staurolite zone. An eastern belt ranges from the garnet to the sillimanite zone in a complex pattern controlled both by structural features (domes, anticlines, synclines) and by many small to large masses of plutonic rock. In the south-central part of the State, Precambrian rocks exposed chiefly in the core of the Green Mountain anticlinorium and in several domes in the east limb of the anticlinorium were remetamorphosed compatibly with the three broad zones outlined above.

The rocks of the Belvidere Mountain area are entirely within the central metamorphic belt at the southern end of an isolated small irregularly oval area, 3 km wide and 13 km long, within which the rocks are in the garnet zone.

Bodies of ultramafic rock in the Belvidere Mountain area are representative in size and modal composition of the entire gamut of ultramafic bodies in Vermont. Small lenticular masses a few meters wide and a few tens of meters long are entirely of talc-carbonate rock and steatite. Somewhat larger masses contain a core of serpentinite and surrounding shells of talc-carbonate rock and steatite. The largest masses contain, additionally, a central core of relatively unserpentinized dunite and periodotite.

ROCKS

The Belvidere Mountain area contains metamorphosed sedimentary and volcanic rocks comprising schists, greenstone, and amphibolite of Cambrian and Ordovician age, and ultramafic igneous rocks, chiefly dunite and peridotite, whose origins date back to the origin of the mantle (Chidester and Cady, 1972), but which were emplaced in Ordovician time (Cady, 1969, p. 23–24, 108). The ultramafic igneous rocks are variously and extensively altered to serpentinite, chrysotile asbestos, amphibole asbestos, talc-carbonate rock, carbonate rock, tremolite rock, and steatite; the country rock (schist and amphibolite) adjacent to the ultramafic rock is variously altered to rodingite (calc-silicate rock), serpentine-chlorite rock, tremolite rock, blackwall chlorite rock, and steatite. The alteration products span a wide range of ages. Serpentinization of the ultramafic rocks probably started in the Cambrian or earlier, prior to the detachment of these rocks from the upper mantle, and continued during a long period of tectonic transport, emplacement, and metamorphism ending in the Devonian. The other alteration products are related to tectonic and metamorphic events that followed emplacement and that range from Ordovician to Devonian in age. The geologic relations of the rocks in the area are depicted on the map (pl. 1), and isometric fence diagram (pl. 2).

Chemical and spectrographic analyses of rocks and minerals from the Belvidere Mountain area are given in table 1. Representative modes of ultramafic rocks and contact rocks are tabulated in appropriate sections of the report.

METAMORPHOSED SEDIMENTARY AND VOLCANIC ROCKS

The metamorphosed sedimentary and volcanic rocks in the Belvidere Mountain area belong to the Hazens Notch, Belvidere Mountain, Ottauquechee, and Stowe Formations, which range in age from Cambrian(?) to Ordovician. The Hazens Notch and Belvidere Mountain Formations are divisions of the Camels Hump Group (Cady, 1956). The formations contain a wide variety of rock types, including greenstone, coarse and fine amphibolite, quartzite, graphitic and nongraphitic phyllite, quartz-muscovite-chlorite schist, and gneiss, which vary widely in the proportions of essential minerals and in texture. All the lithologic units are conformable, and they intergrade by small-scale intertonguing and by gradual change in mineral composition, both parallel and normal to the trend of the units. Each formation and many of the subunits contain several or all of the rock types in the area, but each is characterized by a particular rock type or combination of rock types.

Variations in thickness of individual map units are probably attributable in part to tectonic thinning, in part to facies changes. (See units Ꞓhg, Ꞓhc$_2$, and Ꞓbs, pls. 1, 2, and 3.) The great variation in outcrop width of some units is due principally to variations in attitude of the strata.

Most of the sedimentary and volcanic rocks are distinctly layered, but the layering varies widely in scale and character. Layers several centimeters or more thick and of distinctive lithology—such as greenstone and amphibolite, quartzite, and conglomerate—are almost certainly relict beds. Compositional layering within these beds almost always agrees in attitude with the beds and probably also is bedding. Small-scale layering in the schist and gneiss probably reflects original differences in texture or composition of successive beds or groups of beds, though considerably modified by metamorphism or tectonic action. In a few places small-scale layering in schist may be entirely of tectonic or metamorphic origin. Continuous bedding schistosity, spaced schistosity that locally transects bedding, and slip cleavage that everywhere transects bedding (see Chidester, 1962, p. 17) are present throughout these rocks. In different places, one or another is the dominant foliation. Pertinent details of these features are described and discussed for each map unit and in the section "Minor structural features."

HAZENS NOTCH FORMATION

The Hazens Notch Formation (Ꞓh, pl. 1) (Cady and others, 1963, p. 11–13) contains graphitic and nongraphitic quartz-muscovite-chlorite-albite schist, highly albitic quartz-albite-muscovite-chlorite gneiss, and thin beds of actinolitic greenstone and amphibolite. The different rock types intergrade and are intricately interbedded. Units predominantly of graphitic schist (Ꞓhc$_1$, Ꞓhc$_2$), nongraphitic schist (Ꞓhs), or albitic gneiss (Ꞓhg) are distinguished on the geologic map. Each unit contains subordinate amounts of some or all of the other rock types. In the extreme northwestern part of the map area, where exposure is poor and few observations were made, differentiation of such units in the Hazens Notch is not feasible.

The several units of the Hazens Notch crop out in rather narrow, generally parallel bands, which curve in an arcuate pattern, concave to the southeast, from the southwestern part of the area to the northeastern part. This generally simple pattern is complicated in the southwestern and northeastern parts of the area by several synclines and anticlines, which repeat some of the units and produce an intricate outcrop pattern. Probably less than 1,200 m (4,000 ft) of the upper Hazens Notch Formation occurs within the map area. Individual units that have been distinguished in plate 1 range in thickness from 0 to about 300 m (1,000 ft) and average probably about 60–120 m (200–400 ft) thick.

Each of the rock types varies rather widely in modal composition, principally in the relative proportions of essential constituents such as quartz, albite, sericite, chlorite, and graphite. Further variation is indicated by the local and relatively rare presence of garnet and biotite. Variations in the content of epidote and amphibole are due mainly to transitional relations between the schist and the greenstone and amphibolite.

GRAPHITIC AND NONGRAPHITIC SCHIST

The schist (Ꞓhc$_1$, Ꞓhc$_2$, Ꞓhs, Ꞓh) ranges in color on fresh surfaces from medium greenish gray and grayish olive green in nongraphitic types, to medium to dark gray in graphitic types; on weathered surfaces graphitic schist is commonly rust stained by the weathering of pyrite. Fine layering, produced by alternation of quartzose and micaceous layers—and also by variations in the content of graphite in the graphitic schist—is common but can rarely be traced for more than a few centimeters or a few meters. Most such layering is relict bedding, but locally, especially in the crests of some folds, a very irregular layering transverse to bedding has formed by the concentration of micaceous minerals along slip cleavage surfaces; in graphitic schist, this feature is emphasized additionally by similar concentrations of graphite. The dominant foliation is generally slip cleavage, which in highly micaceous varieties of the schist passes into a transecting spaced schistosity, transverse to bedding and not apparently related to crinkles. Continuous schistosity is commonly apparent in thin section, and locally is the dominant foliation. In some of the graphitic schist, the rock breaks along slip-cleavage surfaces in micaceous layers and along bedding-schistosity surfaces in silty layers.

The schist is composed mainly of quartz, sericite, and, in graphitic types, graphite. Chlorite is a major component of most of the nongraphitic schist and is locally abundant in the graphitic schist. Albite and biotite are commonly present, in many places as major constituents. Amphibole and epidote are present chiefly in zones transitional between schist and amphibolite. Pyrite is a common minor constituent of the graphitic schist and a rare constituent of the nongraphitic schist. Garnet is a rare constituent. The total range in composition is wide, particularly of the graphitic schist. A chemical analysis of graphitic schist is given in table 1, analysis number 19.

ALBITIC GNEISS

The albitic gneiss (Ꞓhg) is predominantly grayish olive green to greenish gray, mottled by abundant white porphyroblasts of albite. Graphitic interbeds are dark greenish gray, and mafic interbeds of actinolitic greenstone are dark yellowish green to dusky green. The gneiss is characterized by a pronounced and distinctive layering. The layering is produced by alternation of granoblastic and lepidoblastic laminae, by various changes in color and texture between both granoblastic and lepidoblastic laminae as a result of different mineral content, and by complex variations among groups of laminae. The layers are generally uniform for several tens of meters, and can commonly be traced—where exposure is adequate—for several hundred meters. The continuity, scale, lithologic relations, and the general conformity in attitude with that of the map units indicate that the layering is relict bedding.

The gneiss is composed chiefly of quartz, sericite, albite, and chlorite. Sphene, epidote, apatite, and magnetite are fairly persistent accessory minerals; garnet, biotite, carbonate, zircon, ilmenite, rutile, graphite, and sulfides are rare accessory minerals.

BELVIDERE MOUNTAIN FORMATION

The Belvidere Mountain Formation, commonly called Belvidere Mountain Amphibolite (Cady and others, 1963, p. 15–18), overlies the Hazens Notch conformably and is gradational with it. On the geologic map of Vermont (Doll and others, 1961), it is considered an upper member of the Hazens Notch, but in the context of this report it is treated as a separate formation. Throughout most of the area, the formation consists of entirely of amphibolite.

In the central part of the area, two map units, a lower one of coarse amphibolite (Ꞓbc) and an upper one of fine amphibolite containing a few thin micaceous beds (Ꞓbf), are distinguished. On the upper southwestern slopes of Belvidere Mountain, a thin lenticular unit of chlorite-sericite-quartz schist (Ꞓbs) is present between the coarse and the fine amphibolite. All three units are transitional into one another, both laterally and vertically. Elsewhere the formation is entirely fine amphibolite, except in the southwestern part of the area, where the amphibolite grades laterally into actinolitic and calcareous greenstone. The greenstone is not shown separately on the map, but it lies outside the hornblende isograd, which marks the boundary between the greenschist and the epidote-amphibolite facies (pl. 1).

The outcrop pattern of the amphibolite is intricate and varied. A flat-lying cap of amphibolite at the top of Belvidere Mountain is an erosional remnant of a facing of amphibolite that once covered the entire east slope of the mountain (fig. 16b). In the southwestern, northwestern, and east-central parts of the map-area, repetition by folding produces an irregular pattern. In several places, septa of amphibolite project into or are wholly enclosed within ultramafic rocks. Facies intertonguing between calcareous greenstone of the Belvidere Mountain and calcareous graphitic schist of the overlying Ottauquechee Formation is evident in outcrop in the southern part of the area and may account for a small part of the irregular pattern there.

The maximum thickness of the Belvidere Mountain Formation is about 300 m (1,000 ft). Along the east side of Corez Pond, the total thickness is probably less than 75 m (250 ft). The schist member is probably not more than a few meters thick at maximum.

Amphibolite, greenstone, and schist are all layered, and the layering appears to be relict bedding. The layering varies considerably in distinctness and character from place to place and from one rock types to another; it is generally best shown on weathered surfaces. It ranges in size and character from simple alternation of layers about 5 mm thick, of different composition and texture, to coarser layering caused by complex variations of packets of layers from a few centimeters to half a meter. Where the rocks are not tightly folded, the layering can be traced continuously for several hundred meters; where highly contorted, the layering is commonly discontinuous and indistinct. Nearly everywhere, the attitude of the layering conforms with that of the amphibolite as a whole.

Rocks of the Belvidere Mountain Formation are generally massive and do not have a conspicuous continuous schistosity, but the major dimensions of platy, tabular, or lathlike minerals are commonly about parallel to layering; where the proportion of these minerals is high, a continuous bedding schistosity is generally distinct. Slip cleavage is rare, but a transecting spaced schistosity is locally present in the finer grained rocks, especially in the noses of folds.

The amphibolite is composed chiefly of hornblende, epidote, and albite. Hornblende generally predominates, but locally epidote does, and in a few places albite is more abundant than either. Chlorite is a fairly persistent minor constituent and is locally a major one in the fine amphibolite. Biotite is a common but not persistent minor constituent. Garnet is a major constituent of some of the coarse amphibolite, sparse in much of the coarse, and rare or absent in the fine. Sericite is rare; it occurs only in zones transitional into rare beds of schist or into the adjacent Hazens Notch and Ottauquechee Formation. Sphene is a persistent accessory mineral; rarely, it is a major component. Rutile and ilmenite are less common and less abundant than sphene. Graphite is a distinctive and conspicuous accessory mineral of some of the amphibolite; it is more abundant and more widespread in the vicinity of the eastern contact of the Eden quarry body. Other accessory minerals are sparsely and sporadically distributed.

Each of the map units varies considerably in modal composition, but the range in chemical composition of the amphibolite is restricted generally to about that of basalt (analyses 8, 15, 31–37, table 1). Differences between coarse and fine amphibolite are no greater than variations within each rock unit, and some samples of coarse and fine amphibolite match closely in modal and chemical composition. Differences between fine amphibolite and greenstone are related to the lower grade of metamorphism in the southwestern part of the map area.

COARSE AMPHIBOLITE

The coarse amphibolite (Ꞓbc) ranges in color from dark gray or greenish black in the very coarse grained rock to medium or dark greenish gray in medium-grained varieties, and generally contains relict bedding, which is accentuated on surfaces etched by differential weathering. The most conspicuous bedding is commonly a few centimeters to half a meter thick. Thick beds consist of groups of complexly varied packets of fine beds, details of which are nearly obliterated by coarse recrystallization. Some of the amphibolite contains abundant rounded coarse grains of greenish-black hornblende dispersed in a finer greenish-gray matrix, and much of this facies shows indistinct irregular anastomosing layers of very fine light greenish-gray material. These irregular layers are generally about parallel to the bedding, but are distinctly different in character, and locally in attitude. They appear to be tiny sheared zones, possibly largely controlled by bedding.

The coarse amphibolite has a crude foliation parallel to bedding as the result of the general subparallel alinement of hornblende crystals. Slip cleavage is absent, and transverse schistosity rare.

The unsheared rock consists of a mosaic of slightly elongate hornblende crystals having interlocking sutured boundaries; the crystals range in length from 1 to 20 mm but are predominantly 5 to 10 mm. Stubby crystals of epidote, ranging from subequant anhedral grains about 1–2 mm across to subhedral laths as much as 3–5 mm long, are distributed principally between grains of hornblende but also are commonly enveloped by single crystals of hornblende. Albite forms irregular aggregates of small grains interstitial to hornblende or to hornblende and epidote and commonly contains many small inclusions of epidote; in places, albite fills tiny fractures in large crystals of garnet. Garnet as large as 5 mm in diameter is abundant in the lower part of the coarse amphibolite, decreases in abundance upward, and is sparse or absent in the upper part. Sphene is dispersed irregularly through the rock; in places it surrounds a core of rutile or a shell of rutile containing a core of ilmenite. Chlorite, or chlorite and biotite, occur as irregular patches of diversely oriented flakes that embay and invade garnet and hornblende. Graphite is concentrated at the grain boundaries and is most abundant in layers rich in albite and epidote. Apatite, carbonate, sulfides, and magnetite are distributed sparsely and erratically. The proportion of major and of some accessory minerals varies from layer to layer; this variation is well shown in outcrop.

In even the unsheared rock, some of the hornblende crystals are bent, slightly fractured, or show incipient granulation at the margins. All stages exist between virtually unsheared coarse amphibolite and highly sheared rock in which relict coarse texture is only faintly preserved. In the first stage, nearly all the coarse crystals of hornblende are bent or fractured, and the margins finely granulated. In the second stage, irregular, wavy, and branching shear zones, in which all the minerals are fractured and granulated, transect the rock. In the final stage, the rock consists of rounded grains of hornblende and epidote set in a fine granulated matrix of hornblende, epidote, and albite.

MUSCOVITE-QUARTZ-CHLORITE SCHIST

The schist member of the Belvidere Mountain Formation (Ꞓbs) is typically a silvery-green rock that has a wavy, irregular schistosity. Abundant spangles of white mica in flakes 1–2 mm across and sparsely distributed irregular black masses of fine-grained magnetite 5 to 20 cm across are distinctive features of the rock. In a few places, the rock contains ellipsoidal masses as much as 15 cm long and 8 cm thick composed almost solely of interlocking blades of actinolite 1–6 cm long and 1–5 cm wide.

Layering is not a conspicuous feature of the rock; it reflects subtle variations in units generally a centimeter or more thick. The attitude of the layering agrees with that of bedding in adjacent amphibolite. Schistosity is continuous parallel to the layering (bedding).

The schist varies considerably in modal composition principally because of transitional relations with the underlying coarse amphibolite and overlying fine amphibolite. In typical specimens, white mica, quartz, chlorite, and albite are the major minerals, in about that order of abundance. Most flakes of mica and chlorite are alined parallel, imparting a continuous schistosity, but an appreciable number of them are at a sharp angle to the schistosity. Epidote is a persistent accessory mineral and is a major component in zones transitional into amphibolite. Magnetite forms heavy concentrations of tiny grains in irregular masses of albite, which appear in hand specimen as black nodules. Ilmenite and sphene are persistent accessory minerals, and apatite, tourmaline, carbonate, rutile, biotite, and garnet occur erratically.

FINE AMPHIBOLITE

The fine amphibolite (Ꞓbf) ranges in color from greenish gray to medium bluish gray and shows both fine layering and coarse layering. The coarse layering is formed by complex variations in groups of fine layers; both are inferred to be relict bedding. The rock is too fine grained to distinguish individual minerals with the naked eye and in many places has a flinty appearance; it breaks readily along bedding surfaces into slabs as much as several centimeters thick. Partly graphitic thin zones on the southwest side of the Lowell quarry body have abundant biotite and sericite, scant quartz, are rich in albite rather than hornblende, and contain conspicuous nodular masses of epidote and hornblende 3 to 15 cm in diameter, sheathed in biotite and sericite (pl. 4*E*). Many gash veins—short, lenticular fractures alined en echelon in narrow bands—filled with albite are exposed along the southwest contact.

The fine amphibolite has a variety of textural relations. One variety consists of a mosaic of interlocking grains of hornblende, epidote, and albite, somewhat elongate parallel to the layering, and 0.1 mm to 0.3 mm in mean diameter; such rock shows no textural evidence of cataclasis. Another variety consists of rounded and broken fragments of hornblende and epidote 0.1 to 0.3 mm across, many of which show mortar structure, in a shredded and granulated matrix of hornblende, epidote, and albite grains 0.005 mm to 0.05 mm in diameter. A third variety consists of subhedral and anhedral crystals of hornblende and epidote 0.05–0.1 mm in diameter set in a matrix of interlocking grains of hornblende, epidote, and albite 0.001–0.005 mm in diameter; this rock shows no textural exidence of cataclasis.

In all the textural varieties, sphene, rutile, and ilmenite are persistent accessory minerals and are commonly distributed in runs parallel to bedding. Graphite occurs both dispersed along grain boundaries and concentrated in zones of intense granulation. Apatite, carbonate, magnetite, and sulfide are common but erratically distributed accessories. Garnet is rare or absent.

The calcitic and actinolitic greenstone, into which the fine amphibolite grades in the southwestern part of the map area, is composed of chlorite, albite, epidote, actinolite or actinolitic hornblende, and varying proportions of calcite; sphene is a persistent accessory, and apatite and sulfides are common. In other respects, the greenstone is similar to the noncataclastic fine amphibolite.

OTTAUQUECHEE FORMATION

In the Belvidere Mountain area, the Ottauquechee Formation (€o, pl. 1) (Albee, 1957; Cady, 1956; Cady and others, 1963, p. 19–22; Doll and others, 1961; Perry, 1927, p. 161, 1929, p. 27–29) consists of interbedded graphitic sericite-quartz phyllite and associated thin beds of dark-gray quartzite, light-green quartz-sericite-chlorite schist, and light-buff sericite phyllite. Thin zones of quartz-sericite-chlorite schist occur near the base of the formation, and lenticular zones of quartz-pebble and quartz-granule conglomerate are distinctive of the upper part of the formation. None of the several rock types forms large enough units or is suitably exposed to be distinguished individually on the geologic map.

The Ottauquechee is exposed widely over an irregular area in the southern part of the map area (pl. 1). The main belt extends along the southeast side of the map area in a band 500–600 m (1,500 to 2,000 ft) wide. A synclinal prong off the main belt projects northwestward along the southwest flank of Belvidere Mountain. The present thickness of the formation is probably only a little less than 500 m (1,500 ft).

A spaced schistosity, uniformly steep, is distinctive of the schist and phyllite in the Ottauquechee. Commonly, the spaced schistosity is virtually parallel to relict bedding. In such places, a continuous schistosity parallel to the spaced schistosity is generally discernible. Elsewhere, in crests and troughs of folds where the spaced schistosity transects bedding, the alinement of micaceous minerals between surfaces of spaced schistosity is disturbed, and the continuous schistosity is obscured or destroyed.

The schist is characterized by small-scale layering, consisting of alternate micaceous and quartzose laminae 0.5–2 mm thick, but individual layers can rarely be traced more than a few centimeters or a few meters. The quartzite forms distinctive beds that can be traced across the extent of an outcrop; small-scale bedding in the quartzite results from minor variations in texture and in the distribution of graphite and sericite. Conglomeratic horizons in the upper part of the Ottauquechee form unmistakable beds as much as a few meters thick.

The phyllite ranges in color from light gray or buff in nongraphitic varieties to medium gray or black in graphitic types. It varies principally in the relative proportion of quartz and sericite and in the presence or absence of graphite. Disseminated fine graphite and filmy partings of sericite commonly constitute not more than 1 or 2 percent of the rock. The conglomeratic horizons differ essentially only in the presence of coarse detrital fragments. Ilmenite and sphene are persistent accessory minerals; carbonate and sulfides occur locally in small amounts. The quartzite is medium gray to dark greenish gray and is composed entirely of quartz. The quartz-sericite-chlorite schist near the base of the Ottauquechee is like the nongraphitic schist in the Hazens Notch.

STOWE FORMATION

The Stowe Formation (Os, pl. 1) Albee, 1957; Cady, 1956; Cady and others, 1963, p. 22–27) overlies the Ottauquechee conformably and intergrades with it over a distance of a few meters or a few tens of meters. Within the map-area the Stowe is composed chiefly of a distinctive grayish-green quartz-sericite-chlorite schist characterized by many lenses and lenticles of granular white quartz alined parallel to the schistosity. (See Cady, 1956; Cady and others, 1963, p. [illegible].) Thin beds of actinolitic greenstone, a few centimeters to a few meters thick, occur at several horizons in the schist.

The Stowe crops out only in a narrow strip along the southern half of the east edge of the map-area. The formation trends about north and dips steeply east. The contact with the Ottauquechee is offset to the left a few hundred meters in a couple of places by folds. The maximum thickness of Stowe exposed in the area is probably little more than 300 m (1,000 ft).

The schist is commonly not conspicuously layered. Where present, the layers cannot be traced for more than a few centimeters or a few meters. The thin units of greenstone generally are conspicuously layered parallel to the margins of the units. Spaced schistosity predominates in the schist. In many places the spaced schistosity is parallel to the discontinuous granoblastic layers; locally, where layers are preserved in the noses of folds, the schistosity cuts across the layers. Continuous schistosity is not prevalent but is locally well preserved in highly sericitic beds. Both continuous and spaced schistosity are locally crinkled and cut by a slip cleavage about parallel to the axial planes of the crinkles.

The schist is composed chiefly of quartz, sericite, and chlorite, generally in that order of abundance, though in places sericite is more abundant than quartz. Albite is sparse though not uncommon. Biotite is rare. Sphene or rutile, or both, are present as accessory minerals, and apatite, tourmaline, and sulfides are not uncommon, though very sparse. The greenstone is composed of epidote, chlorite, amphibole, and albite. Sphene is a persistent accessory and is not uncommonly a major component. Carbonate is a common minor component.

MINERALOGY AND PARAGENESIS

The mineralogy and paragenetic relations of each of the principal types of metamorphosed sedimentary or volcanic rock—quartzite, schist, gneiss, greenstone, amphibolite—are generally similar in their range of variations in all the formations. Some of the major minerals and many of the accessory minerals are uniform in optical properties, inferred composition, and textural relations; some vary only slightly and erratically. Several others, including chlorite, biotite, garnet, amphibole, and epidote, which are critical to the interpretation of the metamorphic history, vary in composition, habit, and proportion according to rock type and grade of metamorphism.

QUARTZ

Quartz is the sole or principal constituent of quartzite, of granoblastic layers in the schist and gneiss, and of the granular quartz lenses characteristic of the schist in the Stowe Formation. Isolated grains and small clusters of grains occur sparsely in micaceous layers of the schist and gneiss. Quartz is rare in the greenstone and amphibolite, where it is associated chiefly with beds of schist or forms thin discontinuous films along stratification surfaces.

Most of the quartz is unstrained, but many grains show slightly undulose extinction and abnormally biaxial interference figures. All the quartz in granular quartz lenses such as those that characterize the Stowe is unstrained. In some thin sections of the Ottauquechee, some grains contain a core of rutilated or of strained quartz, rimmed by overgrowths of clear unstrained quartz. The overgrowths are widest at the ends of elongate grains alined parallel to the layering.

ALBITE

Albite occurs sparsely in granoblastic layers of the quartzite, schist, and gneiss, as sutured grains identical with the quartz in habit. It is abundant in the gneiss as porphyroblasts 0.3 to 3 mm in diameter, both in the form of subhedral to euhedral single crystals and as aggregates of sutured grains (glomeroporphyroblasts). In the greenstone and amphibolite, albite is sparse to abundant as irregular grains interstitial to hornblende, chlorite, and epidote, and it fills gash veins in the amphibolite.

Albite of granoblastic habit in the quartzose schist, gneiss, and quartzite is mostly clear and untwinned, but an appreciable part shows polysynthetic twinning. Some of the twinned grains are conspicuously rimmed by untwinned albite. The schist member of the Belvidere Mountain Formation contains, in addition, albite of strikingly different habit: Lenses and irregular masses as much as 10 mm long, which in hand specimen appear to be black metallic masses, are composed of subequant grains of albite about 0.05 mm across, which include as much as 50 percent magnetite in euhedral and subhedral grains 0.001 to 0.003 mm across.

The porphyroblasts of albite in the gneiss are crowded with inclusions of quartz blebs, tiny grains of clinozoisite, flakes of sericite, trains of dustlike particles of graphite, and other minerals of the schist. The trains of inclusions extend from margin to margin of the porphyroblasts. Most of the trains are straight but are rotated slightly out of orientation with the layering in the schist. Only rarely is a slightly sigmoid pattern discernible. In places, the pattern of inclusions indicates that the porphyroblasts have enclosed previously formed crinkles in the schistosity.

Albite in the greenstone and amphibolite is mostly crowded with abundant inclusions of clinozoisite, whereas albite in the gash veins in the amphibolite and in the tiny veins which commonly heal fractures in garnet is essentially free of clinozoisite.

The optical properties of all the albite are uniformly $2V=75°–80°$, optic sign (+), and $\beta\approx1.732$, indicating a composition near that of pure albite, $NaAlSi_3O_8$.

WHITE MICA

White mica is a major constituent of all the schist and gneiss and is a minor accessory of the quartzite, greenstone, and amphibolite. Muscovite doubtless predominates greatly over other white micas in most of the rocks, but one or more of the minerals paragonite, pyrophyllite, margarite, and talc may be present and may rarely predominate over muscovite. In the phyllitic varieties of Ottauquechee, white mica is commonly almost the sole constituent of the lepidoblastic layers. In the rest of the schist and gneiss, white mica generally predominates over chlorite but locally is subordinate to it. Most of the white mica is sericitic in habit; only in the schist member of the Belvidere Mountain Formation is the mica consistently too coarse to be appropriately called sericite.

Flakes and shreds of sericite in schist and gneiss tend to be alined parallel to bedding, and where chlorite is also present the two are commonly intergrown. Where a transecting spaced schistosity is prominent, the parallel alinement of sericite flakes has generally been diminished and locally obliterated. In the tiny shear zones that mark the spaced schistosity, single flakes of sericite are commonly alined in the shear zones. Porphyroblasts of albite disrupt—and where abundant, almost obliterate—the alinement of sericite.

In greenstone and amphibolite, sericite occurs principally in zones transitional into schist or gneiss, where its habit is identical with that of sericite in the schist or gneiss. A small amount of sericite occurs in altered garnet in amphibolite, less commonly in altered hornblende. Such sericite is a very fine massive intermixture with chlorite and biotite.

Most of the white mica has an index of about $\beta=1.600$ and a moderate optic angle. Similar white mica in similar geologic settings elsewhere in Vermont (Chidester, 1962, p. 50–51) consists wholly of muscovite.

BIOTITE

Biotite is sparse in the metamorphosed sedimentary and volcanic rocks; it was noted only in the amphibolite of the Belvidere Mountain Formation and in the gneiss of the Hazens Notch.

In the amphibolite, biotite is confined to areas of altered garnet, where the biotite occurs as randomly oriented fine shreds and flakes intermixed with chlorite and white mica.

In the gneiss, biotite was seen only near contacts with ultramafic bodies. The biotite is patchily distributed and confined largely to lepidoblastic layers. The flakes of biotite are mostly irregular, but some are lathlike; all are diversely oriented. The patches of biotite commonly contain many irregular remnant shreds of white mica that are irregularly invaded by biotite.

All the biotite is moderately pleochroic, from yellowish brown to moderate brown, and has an optic angle of virtually zero. Little other optical data are available. Correlation of optical properties with chemical composition is not reliable, but the color and moderate pleochroism suggest a relatively magnesian biotite.

CHLORITE

Chlorite is almost ubiquitous in the metamorphosed sedimentary and volcanic rocks but varies widely in abundance and habit. It is a major constituent of all the schist, gneiss, greenstone, and fine amphibolite, and is generally a minor constituent of most of the phyllite, quartzite, and coarse amphibolite.

Chlorite occurs in the schist and gneiss principally in the lepidoblastic layers, interleaved with sericite; a large proportion of such chlorite is oriented parallel to the layering, but generally a small proportion is diversely oriented. In the granoblastic layers, and in the quartzite, chlorite is sparse and consists chiefly of diversely oriented flakes. In garnetiferous schist, diversely oriented chlorite irregularly invades much of the garnet; in some schist that contains no garnet, irregular patches of chlorite consist of diversely oriented shreddy flakes.

The relations of chlorite in the greenstone and amphibolite are more complex and varied. In the greenstone, chlorite is a major component, along with albite; where segregated in layers, the chlorite shows moderately good dimensional parallelism. In the fine amphibolite, chlorite is subordinate to hornblende, with which it is intermixed, but it is commonly a major constituent. Locally, the chlorite embays and invades the hornblende, but in most places there is no evidence of a replacement relation. Chlorite, commonly intermixed with biotite and sericite, occurs in the coarse amphibolite almost solely as aggregate pseudomorphs after garnet or as partial replacements of it. Less commonly, chlorite may partially replace hornblende.

The optical properties of the chlorite vary appreciably within most of the rock types; they differ markedly between the metamorphosed volcanic rocks (greenstone and amphibolite) and the quartzose schist, gneiss, and quartzite.

In the schist and gneiss, the indices fall generally in the range $\beta=1.628$–1.634 and pleochroism is moderate to strong in shades of green; the optic sign varies accordingly from (+) to (−), the sign of elongation from (−) to (+), and the interference colors from abnormal brown to abnormal blue or purple. Chlorite of index $\beta=1.629$ commonly shows both abnormal-brown and abnormal-blue interference colors, both (−) and (+) sign of elongation, and both (+) and (−) optic sign. These variations probably indicate small differences in ratio of $Mg:Fe^{+2}$, and possibly small differences in content of Al. The average chlorite, of about index $\beta=1.630$, is inferred (see Chidester, 1962, p. 44–46; Albee, 1962, p. 864–866) to have a composition of about

$$(Mg_{2.1}Fe^{+2}{}_{2.7})Al_{1.2}(Si_{2.8}Al_{1.2})O_{10}(OH)_8.$$

In the amphibolite, and in greenstone beds in the schist and gneiss, the range of variation is slight; optic sign (+), elongation (−), $\beta\approx1.618$, interference colors abnormal brown, pleochroism slight to moderate in light shades of green. On the basis of these optical properties, the chlorite is deduced to have approximately the composition

$$(Mg_{2.6}Fe^{+2}{}_{2.1})Al_{1.3}(Si_{2.7}Al_{1.3})O_{10}(OH)_8.$$

AMPHIBOLE

Amphibole, a major constituent of the amphibolite and a minor to major constituent of much of the greenstone, is rare in the schist and gneiss except in narrow zones adjacent to contacts with amphibolite, where it ranges from a major to a minor constituent. In the coarse amphibolite, the amphibole forms stubby anhedral crystals 5 to 25 mm long. Elsewhere it occurs as irregular blades, laths, and wispy shreds, generally less than 1 mm long.

On the basis of optical properties and chemical analyses, the chemical composition of amphibole appears to vary significantly both within and among rock types. The variations seem to correlate with coarseness of the rock, mineral associations, and rock type.

The predominant hornblende in the coarse amphibolite is represented by an analysis of a mineral separate (table 1, analysis 23) from coarse amphibolite (table 1, analysis 31) which has the following optical properties: $2V=70°$; optic sign (−); extinction angle, $\gamma\wedge c=17°$; index, $\alpha=1.650$; pleochroism, α=dusky yellow (5*Y* 6/4), β=dusky yellow green (5*GY* 5/2), γ=dusky green (5*G* 3/2). The calculated mineral formula (table 10) is:

$$(Na_{0.84}K_{0.14})_{=0.98}Ca_{1.50}[(Mg_{2.60}Fe^{+2}{}_{1.46}Mn_{0.03}Ni_{0.005}Co_{0.002}Cu_{0.003})_{=4.10}(Fe^{+3}{}_{0.39}Al_{0.78}Ti_{0.08}Cr_{0.005}Sc_{0.016})_{=1.27}]_{=5.37}[Si_{6.66}P_{0.02}Al_{1.32}]_{=8.00}O_{22.00}[O_{1.99}F_{0.01}H_{1.26}].$$

This formula has a ratio of $Mg/Mg+Fe^{+2}+Mn=0.63$, which compares closely with the ratio of about 0.62 inferred from optical properties on the basis of Foslie's (1945, fig. 1, p. 79) chart for hornblende in the epidote-amphibolite facies. Hornblende of lower index in some of the coarse amphibolite is inferred to have a ratio of $Mg/Mg+Fe^{+2}+Mn$ of about 0.76, and possibly a slightly lower content of Al.

Hornblende in the fine amphibolite ($\gamma=1.660$–1.668) has a ratio of $Mg/Mg+Fe^{+2}+Mn$ of about 0.67–0.77. A few seemingly aberrant specimens of higher index, particularly from near the base of the unit, approach the composition of the hornblende typical of the coarse amphibolite.

Amphibole in the actinolitic greenstone ($\gamma=1.650$–1.652) has a ratio of $Mg/Mg+Fe^{+2}+Mn$ near 0.8, and probably a lower content of aluminum and alkalies than does the hornblende. It corresponds in composition to actinolitic hornblende. (See Chidester, 1962, p. 57–58.)

An analyzed sample (analysis 22, table 1) of coarse amphibole from an ellipsoidal mass of amphibole in the schist unit of the Belvidere Mountain Formation is representative of amphibole of highest index in the schist and greenstone. The optical properties are: $2V=80°$; optic sign (−); extinction angle, $\gamma\wedge c=16°$; indices, $\alpha=1.619$, $\beta=1.631$, $\gamma=1.641$; pleochroism, α=grayish green (5*GY* 7/2), γ=light green (5*BG* 6/6). The calculated mineral formula, as derived from procedures outlined in the section on "Amphibole formula," is:

$$(Na_{0.40}K_{0.04})_{=0.44}Ca_{1.68}[(Mg_{4.26}Fe^{+2}_{0.47}Mn_{0.02}Ni_{0.01}\\ Co_{0.0006}No_{0.0001})_{=4.76}(Fe^{+3}_{0.20}Al_{0.35}Ti_{0.001}\\ Cr_{0.01}V_{0.003})_{=0.56}]_{=5.32}[Si_{7.78}P_{0.02}Al_{0.20}]_{=8.00}\\ O_{22.00}[_{1.96}F_{0.04}H_{1.10}].$$

The ratio of $Mg/Mg+Fe^{+2}+Mn$ in the calculated formula is 0.90, which is very close to that based upon optical properties of actinolite with a similar content of aluminum (Foslie, 1945, fig. 1, p. 79). Amphibole of lower index in the schist and greenstone is inferred to have a lower content of Al and Fe^{+2}, and to range in composition to only slightly ferroan tremolite

$$(\approx Ca_2Mg_{4.8}Fe^{+2}_{0.2}Si_8O_{22}[OH]_2).$$

EPIDOTE

Epidote is a major constituent of amphibolite and greenstone in the Belvidere Mountain Formation and is locally a major constituent of the schist member. Elsewhere, it is rare except in beds of amphibolite or greenstone in schist or gneiss. The epidote occurs in association with mafic minerals as subequant to oval grains that vary widely in size, and as numerous tiny grains associated with and included in albite. In the schist member of the Belvidere Mountain Formation, some large grains of epidote enclose cores of allanite.

On the basis of its optical properties (Winchell and Winchell, 1951, fig. 343), the epidote ranges in composition from nearly iron-free clinozoisite, $Ca_4Al_4Al_2Si_6O_{24}(OH)_2$, to pistacite ($Fe^{+3}/[Al+Fe^{+3}] \geqq 0.10$) approximately of the composition

$$Ca_4Al_4(Al_{0.86}Fe^{+3}_{1.14})Si_6O_{24}(OH)_2.$$

Expressed in terms of the ratio $Fe^{+3}/Al+Fe^{+3}$, the range in composition is from less than 0.03 to about 0.20.

In the coarse amphibolite, most of the epidote has indices of about $\beta=1.740$–1.742, corresponding to a ratio of $Fe^{+3}/(Al+Fe^{+3})$ of about 0.20. A sample of epidote separated from coarse amphibolite (table 1, analysis 24) has an $Fe^{+3}/(Al+Fe^{+3})$ ratio of 0.19 (see formula and calculation in section on "Epidote formula" and see table 11). The optical properties ($2V=80°(-)$, $\alpha=1.722$, $\beta=1.742$, $\gamma=1.76$) do not fit the Winchells' diagram well in all respects, but the ratios based on $2V$ and β and on chemical analysis are identical. A small proportion of epidote in the coarse amphibolite has indices as low as $\beta=1.728$, and a $Fe^{+3}/(Al+Fe^{+3})$ ratio of about 0.08. Fine-grained epidote associated with albite has β-index of 1.720 or lower and a ratio of less than 0.03.

In the fine amphibolite, most of the epidote, both that intermixed with hornblende and that included in albite, has indices in the range $\beta=1.712$–1.720, low birefringence, and abnormal blue and lemon-yellow interference colors, indicating a ratio of $Fe^{+3}/(Al+Fe^{+3})$ of less than 0.03. A few larger grains have indices of $\beta=1.727$–1.731, corresponding to a ratio of about 0.08.

Epidote in the greenstone and in the mafic layer in the schist and gneiss is predominantly of high birefringence (second-order blue and green interference colors) and high index ($\beta \approx 1.745$), indicating a composition near that of epidote in the coarse amphibolite, $Fe^{+3}/(Al+Fe^{+3})=0.20$.

GARNET

Garnet is confined to the Belvidere Mountain and Hazens Notch Formations. In the Belvidere Mountain, it is most abundant and most conspicuous near the base of the coarse amphibolite, where subhedral to anhedral grains as large as 1 cm constitute as much as 40 percent, but generally less than 10 percent, of the rock. The garnet decreases in size and abundance upward in the formation; more than a few tens of meters above the base of the coarse amphibolite, and throughout the fine amphibolite, garnet is sparse and generally less than 0.1 mm across. Garnet is very sparse or absent in most of the Hazens Notch but constitutes as much as 1 percent of a few thin mafic beds.

Much of the garnet in the amphibolite and most of the garnet in the schist and gneiss is rimmed, embayed, and irregularly invaded by chlorite, or pervasively altered to a mixture of very fine grained chlorite, biotite, and sericite. Some of the fractured garnets in the amphibolite are healed by tiny veinlets of albite. Many pervasively altered garnets contain veinlets of albite in a matrix of biotite, sericite, and chlorite.

Garnet in the amphibolite is reddish brown in hand specimen, pale pink in thin section, has indices of about 1.780 to 1.785, and appears isotropic. The composition is inferred to be approximately

$$Alm_{60}Pyr_{40}, \text{ or } Fe^{+2}_{1.8}Mg_{1.2}Al_2Si_3O_{12}.$$

(See Winchell, 1958, fig. 1, p. 597.) No indices were obtained from garnet in the Hazens Notch, but the garnet is similar in color to that in the Belvidere Mountain and is probably of about the same composition.

ILMENITE, RUTILE, AND SPHENE

Ilmenite, rutile, and sphene are present, singly or in combination, throughout the metamorphosed sedimentary and volcanic rocks, but the relations vary with rock type. Ilmenite and sphene, either singly or together, are prevalent in most of the schist and gneiss, where they are associated predominantly with chloritic beds. Sphene occurs chiefly as anhedral grains and aggregates of tiny grains but commonly forms subhedral to diamond-shaped crystals. Ilmenite occurs mainly as small laths and elongate irregular grains. Where the two occur together, sphene invariably rims a small core of ilmenite. In some of the Stowe schist, rutile, though very sparse, is the predominant titanium mineral, sphene is rare or absent, and ilmenite is absent.

In the amphibolite and greenstone, sphene is ubiquitous. It commonly contains a core of rutile, which rarely encloses a grain of ilmenite. The relations are, thus: ilmenite surrounded by successive shells of rutile and sphene, rutile surrounded by a shell of sphene, or sphene alone.

GRAPHITE

Graphite is most widespread and most abundant in the phyllite and schist of the Ottauquechee and Hazens Notch Formations, but it is also locally abundant in the amphibolite in the Belvidere Mountain and is a characteristic though minor constituent of much of the quartzite in the Ottauquechee.

In the schist and phyllite, graphite is associated predominantly with micaceous layers, but it also occurs abundantly in the silty layers of the phyllite. In both kinds of layers, the graphite is both concentrated at the borders of the mineral grains and disseminated through the grains as tiny inclusions. Graphite is sparse in quartzose layers coarser than silt size. Where slip cleavage and transecting schistosity are conspicuous, graphite is commonly concentrated heavily along schistosity surfaces.

In the amphibolite, graphite is commonly associated with micaceous and quartzose beds, but it also occurs abundantly in some amphibolite that has no admixed quartz and mica. Graphite generally constitutes less than 5 percent of the rock, but in a few places it forms as much as 30 percent of the schist.

The graphite is very fine and has generally a sooty appearance. Distinct flakes are rarely discernible in thin section, even under highest magnification, and the material appears to be cryptocrystalline or amorphous, but in a few thin sections, cleavage flashes were detected in the graphite. X-ray diffraction studies of several rock samples revealed only a few in which the strongest lines of graphite were detectable (F. A. Hildebrand, written commun., Oct. 13, 1955; Nov. 9, 1955; Sept. 6, 1956). These features indicate that most of the so-called graphite may be amorphous carbon, but at least a small proportion is very finely crystalline graphite. For simplicity, amorphous carbon and graphite are both referred to throughout this report as graphite.

OTHER MINERALS

Carbonate, probably chiefly calcite, occurs as scattered rare anhedral grains in the granoblastic layers of the schist and gneiss. In the greenstone and amphibolite, carbonate occurs in small veinlets and as irregular patches that embay and engulf amphibole and chlorite. Some carbonate in the calcareous greenstone appears to be pseudomorphic after amphibole. Most of the carbonate fizzes strongly in acid and appears thus to be calcite; a small proportion may be dolomite.

Apatite is common but not abundant in the schist and gneiss; it occurs sparsely and sporadically in the amphibolite. All the apatite is of very low birefringence and has an index of about $\omega = 1.633$, indicating the composition of fluorapatite, $Ca_5P_3O_{12}F$.

Rare grains of microcline, noted in the gneiss, have the same habit and associations as quartz.

Zircon was observed only in the gneiss, where it forms subrounded grains as much as 0.5 mm across.

Tourmaline is very sparse but widespread in the schist and gneiss. It ranges in habit from prismatic to subrounded and is predominantly olive or bluish green.

Allanite was noted only in the schist of the Belvidere Mountain Formation, where it forms deep yellow-brown cores in some of the larger epidote grains.

Pyrite is common throughout the schist but is associated particularly with graphitic phyllite, where it forms scattered cubes and anhedral grains.

Magnetite is associated with chloritic zones of the schist and gneiss, where it occurs sparsely in anhedral grains and octahedra as much as 1 mm across.

PETROGENESIS

Metamorphic features and relict sedimentary features modified to various degrees by metamorphism characterize the schist, gneiss, quartzite, amphibolite, and greenstone. The characteristics of the relict bedding, the compositions of the rocks, and the intergradations and facies relations among the several rock types indicate that the protoliths of the phyllite, schist, gneiss, and quartzite were graphitic and nongraphitic shale, fine-grained graywackes, siltstone, and quartzose sandstone; protoliths of the amphibolite and greenstone were mafic volcanic rocks formed largely or entirely by submarine accumulation of detrital material derived from basaltic rocks.

Most of the minerals that constitute these rocks are entirely of metamorphic origin or have been largely or entirely recrystallized during metamorphism. The principal exceptions are the larger clastic grains in the conglomeratic beds and the rare clastic grains of zircon. Despite extensive recrystallization, the distribution of minerals in the metamorphic rocks largely reflects the primary distribution of minerals in the protoliths.

PHYLLITE, SCHIST, GNEISS, AND QUARTZITE

Nearly all the quartz and probably most of the albite (as well as rare grains of microcline) in the granoblastic layers are partly of detrital origin. Except for phenoclasts in the conglomerates, the detrital outlines of most of the clastic grains are completely obliterated by recrystallization. However, scattered grains have a detrital core distinctively different from the rimming overgrowth. Among these are grains of rutilated quartz with rims of clear quartz and twinned grains of feldspar with overgrowths of untwinned feldspar. Many of the scattered small grains of quartz and some of the albite in lepidoblastic layers of the schist are similarly of detrital origin, but an appreciable proportion probably formed by metamorphic reactions. Most of the quartz lenticles in the schist probably formed by metamorphic differentiation and concretionary growth (see Eskola, 1932, p. 71–73; Ramberg, 1952, p. 91–92).

The grain size of the minerals in the granoblastic layers appears to have been little modified during metamorphism and probably reflects closely the fabric of the protolith. This is attested to by the abundance of disseminated graphite inclusions in the layers of silt-size quartz, and the relative sparsity of such inclusions in the coarser layers. The albite porphyroblasts, and perhaps many of the smaller grains of albite in the lepidoblastic layers, are almost entirely of metamorphic origin, though some may have formed on small nuclei of detrital albite; the albite probably formed by growth of larger detrital grains at the expense of small detrital grains or by metamorphic reaction of sodic clay minerals.

White mica, chlorite, and biotite are entirely of metamorphic origin. They formed principally from clay minerals, and variations in their relative proportions and composition are chiefly due to original variations in the sedimentary protoliths.

Graphite is a relict mineral of the protolith, and its distribution—most abundant in the lepidoblastic and silt-sized quartz layers—reflects closely the original distribution in the sedimentary rock. The principal change related to metamorphism is the partial recrystallization of at least some of the carbonaceous matter of the protolith to crytocrystalline graphite. In some places, the graphite is redistributed, probably mechanically, and concentrated along shear surfaces of spaced schistosity and slip cleavage. Pyrite, associated largely with graphite, probably is all of sedimentary origin, though extensively recrystallized during metamorphism.

Calcite interstitial to quartz in the granular layers is probably mainly relict calcareous cement, though completely recrystallized. Some of the calcite that appears to replace other minerals of the schist, gneiss, and phyllite may have been formed at a late stage in the metamorphism as the result of rise in the partial pressure of CO_2.

Garnet, ilmenite, rutile, sphene, epidote, and tourmaline are entirely of metamorphic origin. The cores of allanite in grains of epidote in some of the schist probably formed where rare earths were relatively abundant; the enclosing epidote formed after the rare earths were used up.

AMPHIBOLITE AND GREENSTONE

The minerals of the greenstone and amphibolite are principally or entirely of metamorphic origin. Variations in lithology reflect bedding in the protolith, but the original fabric has been extensively modified. The fineness of detail preserved varies inversely with the grain size of the rock.

Albite formed entirely by breakdown of calcian plagioclase of the protolith. Epidote is both a product of simple breakdown of the calcian plagioclase and of reaction between mafic constitutents. Amphibole, as well as garnet in the higher grade rock and chlorite in the lower grade, formed by reaction of the mafic constitutents of the protolith; ilmenite, rutile, and sphene were minor products of the reactions. Chlorite, biotite, and sericite, in aggregate pseudomorphs after garnet, are retrograde alteration products. Quartz, sericite, and biotite, in beds relatively rich in those constituents, formed from pelitic sediments intermixed with the predominant volcanic material of the protolith.

The thin films and seams of quartz along bedding surfaces show no textural evidence of detrital origin. They may be detrital layers or possibly thin chert beds, completely recrystallized, and perhaps enhanced by metamorphic differentiation, or they may have formed entirely by metamorphic differentiation.

The occurrence of graphite in particular stratigraphic zones and its abundance in some beds that contain considerable intermixed pelitic sediments indicate that the graphite is of sedimentary origin.

The textural difference between the coarse amphibolite on the one hand and the fine amphibolite and greenstone on the other depends chiefly upon grade of metamorphism. However, some of the rock mapped as fine amphibolite was originally coarser grained and was reduced by cataclasis to fine-grained rock. Some of the textural variations within each rock type are also attributable to cataclasis.

ULTRAMAFIC AND ASSOCIATED ROCKS

Ultramafic igneous rocks and rocks derived from them constitute the bedrock surface of about 10 percent of the Belvidere Mountain area and underlie another 5 percent of the area beneath mantling erosional remnants or structural infolds of amphibolite and schist (pl. 1).

The largest outcrop area of ultramafic rocks (the Eden quarry body) covers nearly the entire southeastern slope of Belvidere Mountain and extends in a relatively narrow strip around the upper slopes of the north and northwest parts of the mountain. Small isolated bodies, centered at 750 m (2,500 ft) and 810 m (2,700 ft) altitude on the southwestern flank of the mountain and at 750 m (2,500 ft) on the northeast flank, are inferred to connect with the Eden quarry body beneath the capping of amphibolite. Another large mass (the Corez Pond body) is exposed at only a few places through a deep mantle of glacial and fluvial deposits but is inferred to be present at the bedrock surface in the valley of Corez Pond. A third large body (the Lowell quarry body) is well exposed in the vicinity of the Lowell quarry. A connection at bedrock surface between the Corez Pond body and the Lowell quarry body was exposed by stripping operations. The two bodies probably are not connected with the Eden quarry body at bedrock surface, but all three are probably connected at depth beneath synclinal infolds of amphibolite.

Many small lenses, pods, and veins of ultramafic rock are isolated, at bedrock surface, in country rock near the contacts of the larger ultramafic bodies (pls. 1, 3, and 4 *D* and *E*). Probably none is connected with the main mass of ultramafic rock (the Eden quarry, Lowell quarry, and Corez Pond bodies), except that some of the clusters of tiny pods along the northeast and southwest margins of the Lowell quarry may have nebulous connections with the Eden quarry and the Lowell quarry bodies.

The main mass of ultramafic rock is a lens, tightly folded at its margins, which in overall attitude dips gently to moderately southeast. In its present configuration, the long axis of the lens trends northwest (updip) and is about 3,400 m (11,000 ft) long in plan, the intermediate axis trends northeast and is 1,800 m (6,000 ft) long, and the maximum thickness is 450 m (1,200–1,500 ft). The small isolated masses are steeply to moderately dipping pods and lenses. All the bodies are generally concordant with the bedding of the enclosing schist and amphibolite but are locally crosscutting in detail.

The main ultramafic body contains large volumes of relatively fresh igneous rock, large volumes of serpentinite, relatively minor, though appreciable, volumes of carbonate-quartz rock, talc-carbonate rock, and steatite, and negligible volumes of thin layers and irregular pods of chromite and irregular masses of chromite or chromian magnetite. The relatively fresh igneous rocks are more or less centrally located in the body. They consist chiefly of dunite and and minor amounts of peridotite. Massive serpentinite occupies an irregular and discontinuous zone around the relatively fresh dunite and peridotite, and is irregularly interlayered with it. Highly sheared schistose serpentinite occurs principally at the margins of the body but also traverses the dunite and massive serpentinite in narrow, generally irregular zones.

In many places, the dunite, peridotite, and massive serpentinite contain many veins of cross-fiber chrysotile asbestos and a few veins of dense, resinous-looking serpentine, much of which has a columnar or coarsely fibrous structure, but some of which is megascopically structureless. The schistose serpentinite commonly has short slip-fiber chrysotile asbestos distributed sparsely to abundantly along shear surfaces. In a few places along prominent and well-defined shear zones in the serpentinite, masses of very long fibers of intermixed chrysotile asbestos and fibrous calcite and brucite form spectacular but volumetrically negligible pockets of asbestos. Carbonate-quartz rock, talc-carbonate rock, and steatite (talc rock) form several irregular podlike to elongate tabular masses in the main ultramafic body. In a few places, talc-carbonate rock occurs at the margins of the body; where contacts are exposed in such places, a thin zone of steatite intervenes between the talc-carbonate rock and the contact of the ultramafic body.

The small isolated bodies of ultramafic rock consist entirely of schistose serpentinite, of a serpentinite core surrounded by successive shells of talc-carbonate rock and steatite, locally bordered by tremolite, of talc-carbonate rock surrounded by steatite (and, locally, tremolite), or entirely of steatite (and, locally, tremolite).

Country rock bordering or included in the ultramafic bodies is variously altered to contact rocks of several associations. The main ultramafic body is bordered chiefly by serpentine-chlorite rock and rodingite (diopside - garnet - epidote - vesuvianite rock). Most of the smaller ultramafic bodies and parts of the main ultramafic body are bordered by steatite and blackwall chlorite rock; a thin selvage of tremolite rock generally separates steatite and blackwall. A few of the small bodies, and the main body in a few places, are bordered by tremolite rock and chlorite rock with no intervening zone of talcose rocks. In one place, the contact-alteration zone consists of irregular and haphazardly distributed masses of talc, magnesite, quartz, albite, muscovite, and tremolite.

The relations described in the foregoing section are depicted, at various scales and in varying detail, in figures 3 and 4 and plates 1–4. In general, it was not practical to map separately varieties of ultramafic rock and contact rocks in areas of natural exposure. In the Lowell quarry area and in the C-area, dunite and the varieties of serpentinite are distinguished, where practicable, as they were exposed at the natural bedrock surface after stripping operations (pls. 3, 4*A* and *B*, and fig. 3). Contact rocks are depicted on some of the detailed maps (pl. 4*C* and *D*, and fig. 4). The relations of several of the small isolated pods are shown on plate 4*D* and *E*. Many of the exposures depicted have since been removed down to the level of the quarry floors and are now mantled with rock debris from quarry operations.

MINERALOGY OF THE SERPENTINE GROUP

The minerals of the serpentine group are varied and complex. The group has been studied intensively by many investigators, and the general scheme of classification, the structure and parameters of the unit cells, and the crystal chemistry of the several varieties have been determined.

Among the more important papers of general interest on the mineralogy and stability relations of the serpentine minerals are those by Selfridge (1936), Gruner (1937), Efremov (1939), Aruja (1945), Bowen and Tuttle (1949), Yoder (1952), Kouřimský and Šatava (1954), Kouřimský and Filčáková (1954), Roy and Roy (1954; 1955), Nagy and Faust (1956), Whittaker (1951; 1952; 1953; 1956a; 1956b; 1956c; 1957), Whittaker and Zussman (1956, 1958), Zussman and Brindley (1957), Zussman, Brindley, and Comer (1957), Künze (1956, 1958), Bates (1959), Gillery (1959), Maser, Rice, and Klug (1960), Olsen (1961), and Faust and Fahey (1962). The text by Deer and others (1962, v. 3) summarizes the mineralogy of the group, and contains an extensive list of selected references.

Three principal structural varieties of serpentine are recognized: chrysotile, lizardite, and antigorite. Chrysotile is fibrous in habit, whereas lizardite and antigorite are platy. The basic unit cell of all the minerals approximates the single-layered cell of lizardite (a=5.3 A, b=9.2 A, c=7.3 A, β=90° or $\cong$93°). However, in chrysotile the unit cell is double layered, and has a c−parameter of $\cong$14.6 A; furthermore, three subvarieties of chrysotile are distinguished—ortho-, clino-, and para-. Antigorite has a supercell that consists, essentially, of 8½ lizardite subcells, in which $a \cong 43.5$ A. A fourth variety, six-layer orthoserpentine, has a supercell in which six layers are stacked along the c−axis (c=43.9 A); it reportedly occurs in both fibrous and massive form (Deer and others, 1962, p. 173).

Lizardite and six-layer orthoserpentine are elongated parallel to a; antigorite is elongated parallel to b. Orthochrysotile and clinochrysotile have a parallel to the fiber axis; parachrysotile has b parallel to the fiber axis.

The ideal structural formula of all the varieties of serpentine except antigorite is that of lizardite, $Mg_6Si_4O_{10}(OH)_8$, or an appropriate multiple (two for chrysotile, six for six-layer orthoserpentine). Because of distortion in the antigorite structure, antigorite departs slightly from the simple formula of 8½ times the formula of lizardite; it is reported as $Mg_{48}Si_{34}O_{85}(OH)_{62}$. (See note.)

NOTE.—Faust and Fahey (1962) record slightly different structural formulas derived by Zussman (1954) and by Kunze (1956, 1958). The recorded difference in Mg-content appears to be due to an arithmetical or typographical error in Kunze's paper (1956, p. 105). "$Mg_{2.8125}$" should read "$Mg_{2.8235}$." The difference in content of (OH) appears to result from different conclusions concerning the (OH) ions that would be omitted from the unit cell: Zussman concluded that 2(OH) would be eliminated from the tetrahedral sheet and 3(OH) from the octahedral sheet; Kunze, 6(OH) from the octahedral sheet. Inversion of the polar sheet at the point of inflection, postulated by both authors, would appear to make Kunze's omission of 6(OH) ions the more likely. These considerations reconcile the two formulas exactly.

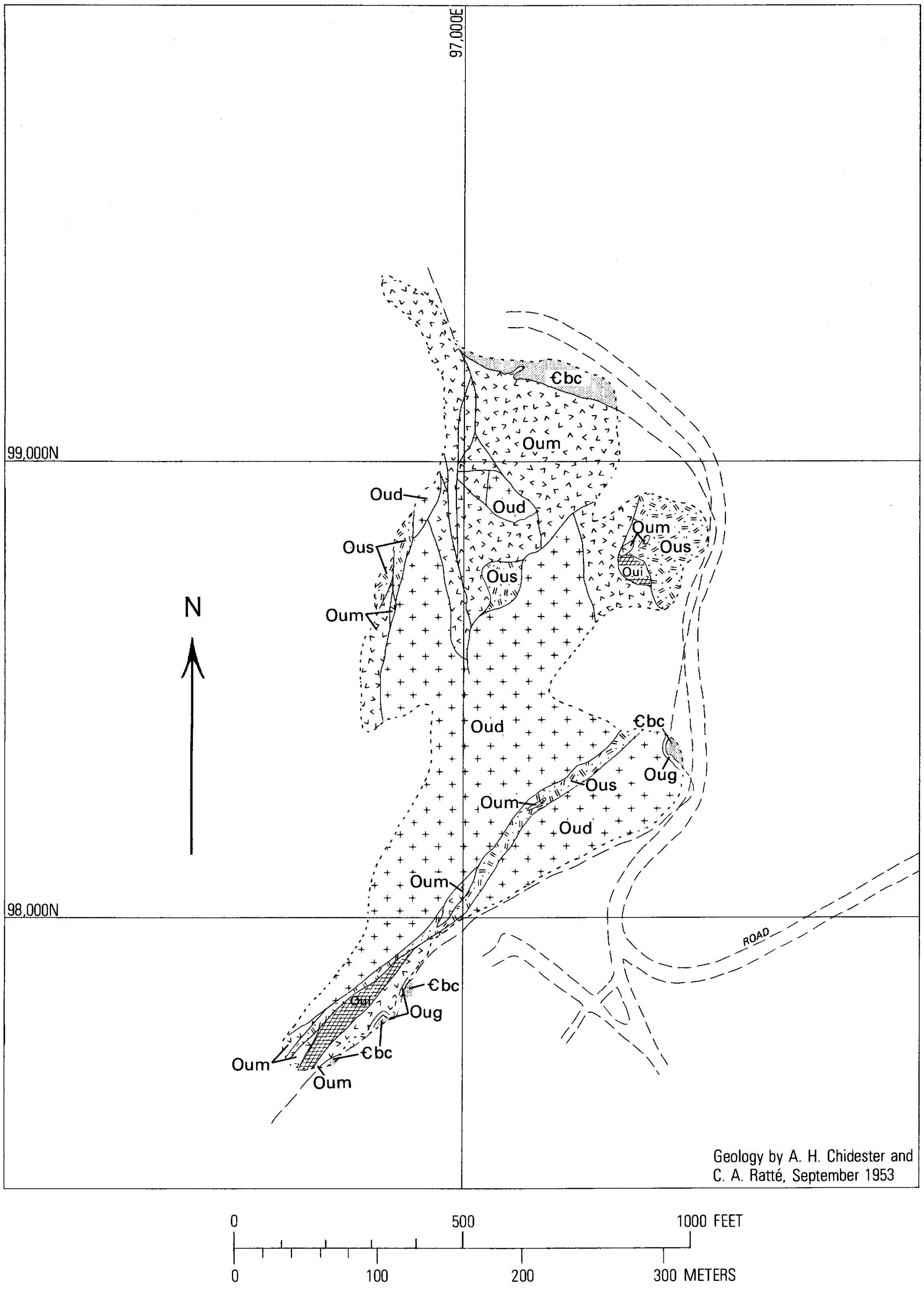
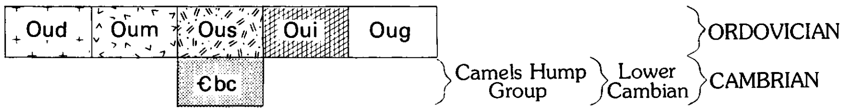


FIGURE 3.—Geologic map of part of the C-area, Eden and Lowell, Vt.

CORRELATION OF MAP UNITS



DESCRIPTION OF MAP UNITS

Oud	DUNITE—Slightly to moderately serpentinized. Layering, which reflects variations in content of pyroxene and opaque minerals, is ubiquitous; layers range in thickness from 1 in. (2.5 cm) to as much as 2 ft. (60 cm)
Oum	MASSIVE SERPENTINITE— Layering obscure. Irregular fractures several inches to a few feet apart impart a blocky structure
Ous	SCHISTOSE SERPENTINITE—Composed entirely of small chips and lenticular fragments of serpentinite. In tabular bodies the chips and lenticular fragments are alined parallel to the shear contacts. In irregular bodies the orientation of fragments is variable
Oui	SCHISTOSE SERPENTINITE—Contains abundant lenticular and irregular masses of tremolite-chlorite rock which has a highly contorted schistosity and in which scattered nodules of coarse actinolite are distinctive
Oug	GRAPHITIC SCHISTOSE SERPENTINITE—The Ordovician age designation refers to the age of emplacement of the intrusive ultramafic rocks, not to the age of the parent igneous rocks or the metamorphic derivatives (see text subheading discussions under "Tectonic and petrogenic synthesis")
Ꞓbc	BELVIDERE MOUNTAIN FORMATION—Coarse amphibolite
————	Contact or shear surface—Solid where exposed, dashed where covered. All contacts between ultramafic rocks are shear surfaces; the contact between ultramafic rock and amphibolite generally is moderately sheared
- - - - - -	Limit of exposure

FIGURE 3.—Continued.

For purposes of comparison, it is convenient to write the mineral formulas of all the structural varieties in terms of the structural formula of lizardite. Allowing for (1) the substitution of Fe^{+2} (and generally small amounts of other divalent ions) in octahedral position; (2) of Al and Fe^{+3} (and very small amounts of other trivalent ions) in direct substitution in octahedral position, which results in vacancies; and (3) of Al in coupled substitution in octahedral and tetrahedral position, the formula for chrysotile (all varieties), lizardite, and six-layer orthoserpentine, then becomes

$$\left[(\mathrm{Mg},\ \mathrm{Fe}^{+2})_{\left(6-\frac{3x}{2}-y\right)}(\mathrm{Al},\ \mathrm{Fe}^{+3})_{(x+y)}\right]^{\mathrm{VI}}_{=\left(6-\frac{x}{2}\right)}\left[\mathrm{Si}_{(4-y)}\mathrm{Al}_{y}\right]^{\mathrm{IV}}_{=4}\mathrm{O}_{10}(\mathrm{OH})_{8},$$

where x is equal to (Al, Fe^{+3}) that substitutes directly in octahedral position, and y is equal to Al that is in coupled substitution in octahedral and tetrahedral position (Chidester, 1962, p. 46). Similarly, the formula for antigorite becomes

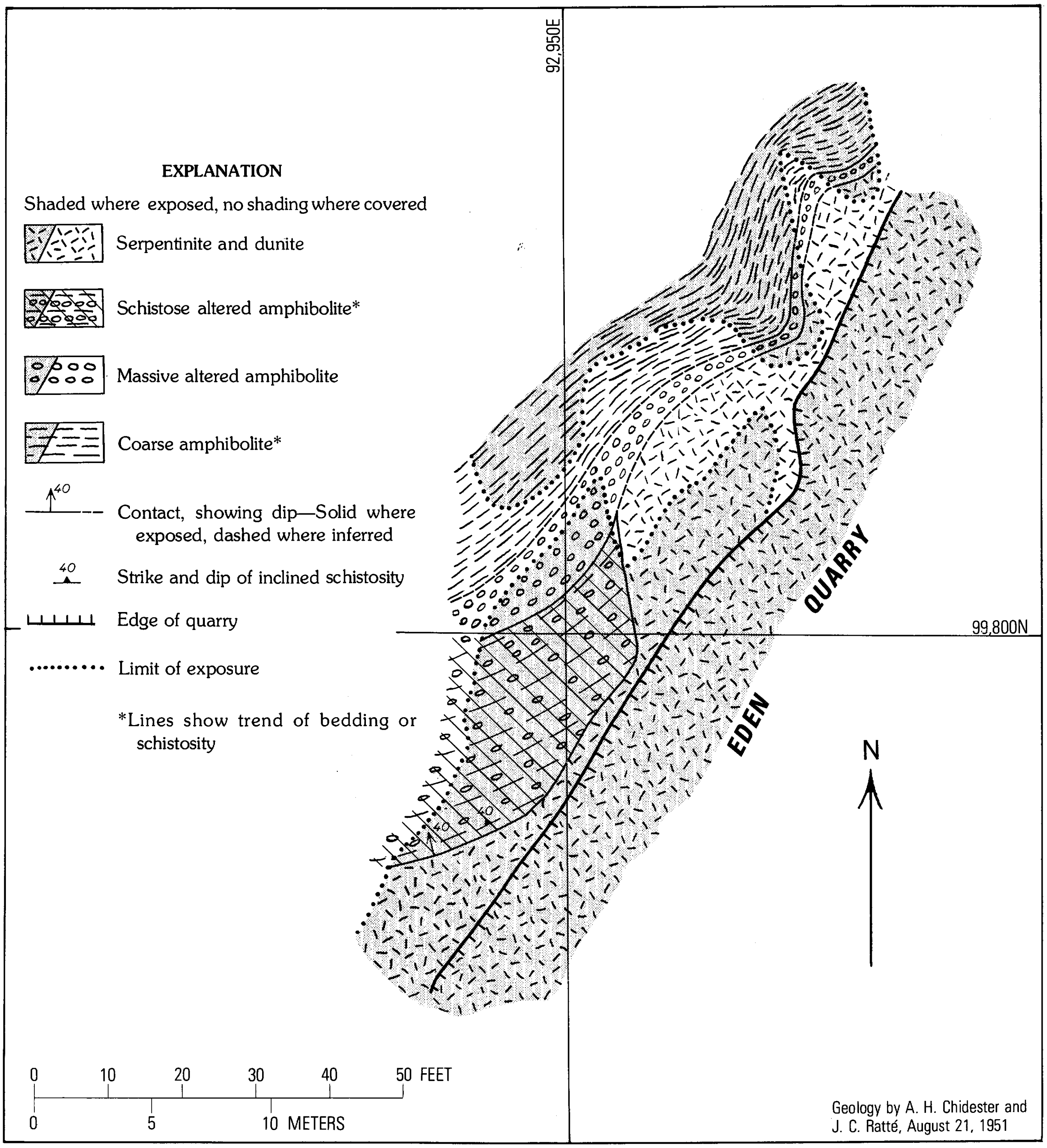


FIGURE 4.—Geologic map of the contact between ultramafic rock and amphibolite, Eden quarry, Eden, Vt.

$$\left[(\mathrm{Mg},\ \mathrm{Fe}^{+2})_{\left(5.65-\frac{3x}{2}-y\right)}(\mathrm{Al},\ \mathrm{Fe}^{+3})_{(x+y)}\right]^{\mathrm{VI}}_{=\left(5.65-\frac{x}{2}\right)}\left[\mathrm{Si}_{(4-y)}\mathrm{Al}_{y}\right]^{\mathrm{IV}}_{=4}\mathrm{O}_{10}(\mathrm{OH})_{7.29}.$$

In practice, the formula of lizardite suffices as a general formula for the serpentine group, and the Roman numerals designating the coordination position are dropped. Furthermore, it is not generally possible to determine precisely the role of the different types of substitution on the basis of present chemical analysis and sampling techniques, and it is sufficient to indicate simply that the trivalent ions in tetrahedral position are not necessarily equal to those in octahedral position. The simplified formula thus becomes

$$[(Mg, Fe^{+2})_{(6-x)}(Al, Fe^{+3})_y]\ [Si_{(4-z)}Al_z]\ O_{10}(OH)_8,$$

where, theoretically, x is equal to 3/2 times the (Al, Fe^{+3}) in direct substitution in octahedral position plus the Al in octahedral position that is in coupled substitution with Al in tetrahedral position; y is equal to the Al in octahedral position both in direct substitution and in coupled subsitution with Al in tetrahedral position; and z is equal to Al in tetrahedral position in coupled substitution with Al in octahedral position. The precise formulas in the preceding paragraph should make these relations clear. (See note.)

NOTE.—In addition to the compositional variations discussed, it seems likely that the entrance of Al into six- and four-coordination in antigorite might cause structural modifications that would result in changes in the structural formula of antigorite: The replacement of Mg and Si by Al would lead to better fit between the "brucite" layer and the Si_2O_5-layer, thus decreasing the curvature of the unit cell, and reducing the likelihood of "coincidence deficiency" (Kunze, 1956, p. 82) in the unit cell. In such a case, the structural formula of antigorite would change toward that of lizardite. Perhaps such a change would take place suddenly, at a critical point with respect to the content of Al, in any particular unit cell. For macroscopic samples of antigorite, however, the compositions might be expected to vary statistically through a considerable part of the range between that of antigorite and that of lizardite. When the Al content is great enough to cause a good fit between the "brucite" layers and the Si_2O_5 layers, thereby reducing the curvature of the unit cell sufficiently to allow accommodtion of an additional 6(OH) and 3 Mg per unit cell, probably the lizardite structure, rather than the antigorite structure, would be stable. The lizardite structure differs from the postulated structure of antigorite that contains a high content of Al only in that the polar layers are not inverted at the midpoint of each supercell (8½ unit cells of lizardite).

Unfortunately, no general distinction based on optical properties can be made between varieties of serpentine that have been distinguished by X-ray, DTA, and electron microscope investigations. In a particular geologic environment, however, or perhaps only within the limits of a particular locality, a distinction on the basis of optical properties can be made between some of the varieties of serpentine. (Chidester, 1962, p. 46–48). Even then, it is generally possible to distinguish only between chrysotile on the one hand and the platy varieties of serpentine (antigorite, lizardite, and six-layer orthoserpentine) on the other. And where two or more varieties are intimately intergrown, the distinction by means of optical properties may be uncertain or impossible.

Thus, in the Belvidere Mountain area, chrysotile can generally be distinguished by its asbestiform habit and low indices relative to platy serpentine. The apparent optic axial angle in sections cut normal to the fiber axis is $2V=0°$; in oblique sections, the angle increases as obliquity increases. The apparent optic sign is (+); the sign of elongation, (+). Measured indices are $n=1.532–1.545$, $N=1.550–1.562$; the measured value of N is probably near true γ, which vibrates parallel to the fiber axes, but the value of n is intermediate between α and β. (See note.)

NOTE.—It seems probable that, fundamentally, the optic sign of chrysotile is negative, the same as that of other varieties of serpentine. The apparent positive optic sign, and the apparent optic angle, which varies with the obliquity of the section with respect to the fiber axis, appear to result from the cylindrical structure of the lattice. In sections of orthochrysotile and clinochrysotile cut normal to the fiber axes, γ vibrates parallel to the microscope tube, and all the transmitted light, vibrating in the plane of the section, ranges in index from α to β. Thus, the index observed is an average of the two; the mineral appears to be uniaxial and the optic sign, (+). In sections cut moderately oblique to the fiber axes, light vibrating parallel to the microscope tube is a nearly fixed large value of γ'. Light vibrating parallel to the long axis of the elliptical section ranges in index from a smaller γ' to α', so that the index observed is an average near β. Light vibrating parallel to the short axis of the elliptical section ranges in index from α to β, so that the index observed is a small value of α'. As a consequence, the observed interference figure is biaxial, $2V$ is moderate, and the optic sign is (+). As obliquity increases, the apparent optical axial angle increases until, when the fiber axes are parallel to the section, the interference figure is like that of a flash figure of an uniaxial positive mineral.

The same line of argument would appear to hold for parachrysotile, but with this difference: If $\beta \approx \gamma$, the apparent optic angle, apparent optic sign, and sign of elongation would be the same as for orthochrysotile and clinochrysotile, but the apparent birefringence would be appreciably lower. If $\alpha \approx \beta$, the relations of the apparent optic angle would be the same, but the optic sign and sign of elongation might be negative; the apparent birefringence would also be low.

On the other hand, the platy varieties of serpentine (chiefly antigorite) have the following optical properties: $2V$, small to moderate; optic sign, (−); sign of elongation, (+); birefringence colors, abnormal blue; indices, $\alpha=1.560–1.565$, $\beta \approx \gamma = 1.565–1.573$.

These observations form the basis of data presented in table 2, which relates approximately the structural parameters, habit, and optical properties of the serpentine minerals.

Many varietal names that designate chiefly types of serpentine defined on the basis of habit and association have been discredited; only antigorite, lizardite, six-layer orthoserpentine, and the three types of chrysotile are now recognized as valid species. In this report, "picrolite" is retained, in addition to the species names listed above; it is used in a purely descriptive sense for nonasbestiform vein serpentine that is dense and of generally columnar or coarsely fibrous but locally massive habit.

IGNEOUS ROCKS, SERPENTINITE, AND VEINS

Ultramafic rocks that have more than about one-third of the essential primary minerals (olivine and pyroxene) remaining retain generally the appearance of the primary igneous rock and are classed in this report as dunite or peridotite. More highly serpentinized rock is classed as serpentinite. The classification as dunite or peridotite is based on the original content of pyroxene. Rocks containing less than 5 percent pyroxene or recognizable pseudomorphs after pyroxene are classed as dunite; those containing more than 5 percent, as peridotite.

The transition from dunite to peridotite is commonly abrupt; layers that contain 5–10 percent pyroxene are in sharp contact with layers that contain little or no pyroxene. Less commonly, the transition is gradational over several centimeters. Both the dunite and peridotite grade into massive serpentinite. Normal to the layering, the transition is fairly sharp, the gradation from little or moderately serpentinized rock to highly serpentinized rock taking place within a fraction of a centimeter. Parallel to the layering, the transition is commonly very gradual, taking place over several meters.

Chromitite forms sharply bounded layers in the dunite and massive serpentinite, and irregular pods in schistose serpentinite. Several types of veins crosscut layering in the dunite, peridotite, and massive serpentinite; are parallel to the irregular schistosity of the schistose serpentinite; or are in transecting shear zones. Most of the veins are of serpentine, carbonate, and associated minerals, but magnetite also forms distinctive veins.

Table 3 contains selected modes of ultramafic igneous rocks and their serpentinized derivatives.

DUNITE AND PERIDOTITE

Dunite crops out extensively on the southeastern slopes of Belvidere Mountain and was well exposed by stripping operations in the Lowell and C-area quarries. Areas predominantly of dunite are delineated on the large-scale maps of the Lowell quarry (fig. 5) and the C-area (fig. 7). Minor amounts of

TABLE 2.—*Unit-cell dimensions, habit, and optical properties of the structural varieties of serpentine minerals*

	Chrysotile			Antigorite	Lizardite	Six-layer orthoserpentine
	Ortho	Clino	Para			
			Unit-cell dimensions			
a	5.34 A	5.34 A	5.34 A	43.5 A	5.31 A	5.32 A
b	9.2 A	9.2 A	9.2 A	9.2 A	9.2 A	9.23 A
c	14.63 A	14.65 A	14.63 A	7.26 A	7.31 A	43.59 A
β	--	93° 16′	--	91° 23′	--	--
			Optical properties			
$2V$[1]	0°-large	0°-large	0°-large	Small to moderate	Small to moderate (?)	?
Optic sign[1]	(+)	(+)	(+) or (−)?	(−)	(−)	(−)?
Sign of elongation[1]	(+)	(+)	(+) or (−)?	(+)	(+)	(+)
α-index[2]	1.532–1.545	1.532–1.545	?	1.560–1.565	?	?
β-index	?	?	?	$\approx\gamma$	?	?
γ-index	1.550–1.562	1.550–1.562	?	1.565–1.573	1.573	1.575(?)
Optical orientation	$\alpha=c$ $\beta=b$ $\gamma=a$	$\alpha=c$ $\beta=b$ $\gamma=a$	$\alpha=c$ $\beta=b$ $\gamma=a$	$\alpha=c$ $\beta=b$ $\gamma=a$	$\alpha=c$ $\beta=b$ $\gamma=a$	?
Extinction	Parallel	Parallel	Parallel	Parallel	Parallel	?
Abnormal birefringence colors	None	None	None	Blue	Blue	
			Habit			
	Fibrous; a= fiber axis	Fibrous; a= fiber axis	Fibrous; b= fiber axis	Platy to columnar; elongated parallel to b.	Platy to columnar; elongated parallel to a.	Structureless to platy; elongated parallel to a.

[1] See text discussion on optic axial angle, optic sign, and sign of elongation, of chrysotile.

[2] Actually an α'-index in the case of chrysotile.

TABLE 3.—*Selected modes of dunite and serpentinite*

[Tr, trace]

Specimen No.	Protolith, (based on content of pseudomorphosed pyroxene: dunite <5 percent peridotite >5 percent)	Olivine	Pyroxene	Chromite	Magnetite	Antigorite or lizardite	Chrysotile	Chlorite	Brucite	Anthophyllite	Sulfides	Carbonate	Talc	Graphite
A. Dunite and massive serpentinite														
105	Dunite	16	Tr	--	2	80	--	1	--	1	--	--	--	--
110	do	33.3	--	Tr	3.8	59.3	--	1.6	--	2.0	--	--	--	--
AR-42	do	--	--	--	7	83	--	1	8	--	Tr	1	--	--
VAE-1	do	90	--	Tr	<1	10	<1	<1	--	--	--	--	--	--
AR-13b	do	63.0	--	--	1.1	34.9	--	.7	.3	--	Tr	--	--	--
AR-13f	do	76.4	--	--	4.0	19.6	--	--	--	--	--	--	--	--
AR-73	do	--	--	--	5.6	84.6	--	4.1	5.7	--	--	--	--	--
AV-15	do	1.3	--	--	4.9	83.7	3.4	4.0	2.7	--	--	--	--	--
AV-29	Peridotite	20	Tr	--	6	68	--	1	--	5	--	--	--	--
AV-99	Dunite	--	--	--	5.1	78.2	--	7.3	9.4	--	--	--	--	--
AV-129	Peridotite	Tr	Tr	--	10	78	--	2	Tr	10	--	Tr	--	--
AV-248	Dunite	85.9	--	Tr	1.0	12.7	--	Tr	.4	--	--	--	--	--
A-BM-53-1	do	--	--	--	5	76	--	--	--	--	--	--	--	19
A-BM-53-12	do	--	--	--	5	95	--	--	--	--	--	--	--	--
A-BM-57-3	do	60	--	--	2	37	--	1	Tr	--	--	--	--	--
B. Schistose serpentinite														
AV-74	Dunite	--	--	--	3.2	86.6	6.4	3.8	--	--	--	--	--	--
AV-81	do	--	--	--	4.0	80.2	13.2	1.3	1.2	--	--	0.1	--	--
AV-209	do	--	--	--	1	91	1	--	--	--	--	7	--	--
A-BM-30	do	--	--	1	--	99	--	--	--	--	Tr	--	Tr	--
A-BM-53-20	do	--	--	--	Tr	94	--	--	--	--	--	--	1	5
A-BM-53-43	do	--	--	--	Tr	100	--	--	--	--	--	--	--	--
EQ-4-909	do	--	--	--	--	60	--	2	--	--	--	1	--	37
CP-2-126	do	--	--	--	7	90	--	Tr	3	--	--	--	--	--

peridotite are interlayered with the dunite but nowhere form mappable units. Dunite is abundant in the ultramafic rocks on the steep slopes under and south of the abandoned aerial tramway line on the east side of Belvidere Mountain (fig. 3) but is not mapped separately.

The dunite and peridotite are generally massive and range from grayish yellow green to dark greenish gray. They are predominantly fine grained and break unevenly along minutely grainy surfaces characterized by cleavage flashes of olivine and tiny lustrous silken threads and patches of serpentine. Olivine is the only essential mineral of the dunite, and some of the dunite is virtually free of pyroxene or its pseudomorphs; however, most of the dunite contained originally from a small fraction of 1 percent to as much as 5 percent pyroxene. The peridotite contains pyroxene as an essential constituent in addition to olivine. The original pyroxene content of the peridotite ranged from 5 percent to about 10 percent. Chromite, magnetite, and sparse sulfides and sulfarsenides are common primary accessory minerals of the dunite and peridotite. Serpentine and magnetite are ubiquitous secondary minerals, and brucite, chlorite, carbonate, and anthophyllite are common ones in both rocks. Both rocks vary widely in degree of serpentinization, but all contain more than 10 percent serpentine. Representative modes of dunite and peridotite are given in table 3*A*.

On weathered surfaces long exposed to the air, dunite and peridotite characteristically range from buff or yellowish gray to moderate reddish brown, but weathered surfaces from which mantling vegetation has been recently removed are commonly pale yellowish green to almost white.

Layering is ubiquitous in the dunite and peridotite. Most of the layers are 1–25 cm thick, and few are as thick as a meter. Some of the thinner layers consist of groups of uniform layers a few millimeters thick. Individual layers are distinguished by differences in texture and color; the layering reflects chiefly original differences in texture and content of accessory minerals but also marks differences in the pattern and degree of serpentinization. The distribution of pyroxene pseudomorphs especially contributes to the layering; the content of pyroxene is commonly significantly different in adjacent layers, and the alinement of the platy crystals parallel to the layering reinforces the layered effect. On weathered surfaces, differences in color and texture are accentuated by differences in weathered color of olivine, pyroxene, serpentine, and brucite, by the variously knotty or pitted weathered surface of pyroxenic rock, and by the increased contrast between silicate and metallic minerals.

Much of the dunite and peridotite has a crude to pronounced foliation parallel to the layering. The foliation is due to the parallel alinement of platy minerals, particularly pyroxene, to the tendency of the rock to break along surfaces of contact between layers of contrasting composition, and, in a few specimens, to a discernible parallel alinement of crystallographic planes in the olivine grains.

CHROMITITE

Concentrations of chromite (chromitite) are rare in the Belvidere Mountain area and were noted only in the main body of ultramafic rock. A layer of massive chromitite about 30 cm thick is exposed for a length of about 3 meters at 102,950 N., 90,940 E. (pl. 1) on the west peak of Belvidere Mountain. Several thin layers of chromitite 1–10 cm thick were exposed by stripping operations near 98,260 N., 97,160 E. (fig. 3 and pl. 1) in the C-area. These layered bodies are conformable with the layering in the enclosing dunite and massive serpentinite. Where the layered dunite is folded, the chromitite veins are folded in conformity. A small irregular pod of massive chromitite is exposed southwest of Corez Pond, at 92,700 N., 96,900 E. (pl. 1). The contacts are not exposed, and the pod is surrounded by talc-carbonate rock; so the structural relations are not known. Veins of pale-green coarsely foliated talc border and vein the chromitite pod.

The tabular bodies of chromitite range in fabric from massive throughout to finely layered. The fine layering is as little as 1 mm thick and consists of alternate laminae predominantly of chromite and olivine or serpentine. The pod of chromitite is entirely massive.

A sample (A–BM–29) of the 30-cm-thick layer of chromite from the west peak of Belvidere Mountain has a modal composition of: chromite, 66 percent; olivine and serpentine, 34 percent. The modal composition of a sample (A–BM–53–116) from a tabular body 10 cm thick, in the C-area, is: chromite, 55 percent; magnetite (probable chromian), 13 percent; olivine and serpentine, 32 percent.

SERPENTINITE

Serpentinite has a variety of textural features and mineral associations, but can be divided into two textural end types: massive serpentinite and highly sheared and schistose serpentinite (hereafter referred to simply as "schistose serpentinite"). Serpentinite intermediate in character is common.

Massive serpentinite is well exposed near the west peak and on the southeast slopes of Belvidere Mountain, where it is interlayered with dunite and peridotite, and in the Eden, C-area, and Lowell quarries, where it is marginal to and interlayered with dunite and peridotite (pl. 1). Schistose serpentinite is well exposed in the three quarries, where it is principally adjacent to the contacts of the ultramafic bodies but also occurs in irregular zones that transect both dunite and massive serpentinite. Schistose serpentinite is also well exposed in the many small satellitic bodies around the main body of ultramafic rock. Moderately sheared serpentinite forms discrete masses in the Lowell and C-area quarries. It is mapped generally with the schistose serpentinite, but plate 4*C* illustrates the variation in textural pattern of moderately sheared serpentinite. The general distribution patterns of serpentinite are depicted in figure 3 and plates 3 and 4.

Massive serpentinite is medium greenish or bluish gray to very dusky blue green or greenish black. It retains the gross textural features of the dunite or peridotite but breaks with a hackly rather than grainy fracture. The layering present in the dunite has been retained with little change except in the color. In several places, the massive serpentinite contains layers composed of diversely oriented flakes as much as 2 mm across; these layers are commonly not more than a few centimeters thick. Mineralogically, the massive serpentinite is like the partly serpentinized dunite and peridotite except for the relative proportions of the minerals. Relict olivine is commonly present, and remnants of pyroxene and anthophyllite are not uncommon, but in many places these minerals are entirely altered to serpentine. Magnetite is more abundant than in the dunite and peridotite, particularly as very fine grains and dustlike particles. Patches and veins of brucite and halos of chlorite around grains of chromite are common. Chromite, the larger grains of magnetite, and the sulfides and sulfarsenides are in the same relations as in the dunite and peridotite. Some of the massive serpentinite is slightly to highly graphitic, particularly in the flaky layers.

Schistose serpentinite is similar in color to the massive serpentinite, but it retains almost none of the textural features of the igneous rock. It consists of irregular polyhedral units of unsheared serpentinite bounded by thin zones of highly sheared and schistose serpentinite. The unsheared polyhedral units range in size from tiny chips to masses several meters across; the schistose zones in which macroscopic unsheared chips are virtually absent are commonly a small fraction of a centimeter thick, but may be as thick as a few centimeters. The schistose serpentinite consists essentially of platy serpentinite; in most specimens, fibrous serpentine is absent from the groundmass, and in none does it constitute more than a few percent of it. Magnetite is the most abundant accessory mineral. Carbonate, brucite, chlorite, chromite, and sulfides and sulfarsenides are common but sparse accessories; talc is rare but locally abundant. Locally, graphite is sparse to very abundant in the schistose serpentinite, especially along the eastern contact of the Eden quarry body.

Both the massive and schistose serpentinite weather to a very pale green, very pale buff, or white chalky surface. Commonly, smooth surfaces are marked by a reticulate pattern of finely etched lines. Remnants of olivine in the massive serpentinite and patches and flakes of brucite in both the massive and the schistose serpentinite weather brown to buff and show up against the white-weathered serpentine as distinctive specks and patches.

SERPENTINE VEINS

Serpentine veins in the ultramafic rocks have a wide variety of structural and textural relations and mineral associations, but they fall into two basic categories: asbestos veins and picrolite veins. Asbestos veins are of two habits: cross-fiber veins, in which the fiber axes are at a large angle to the vein walls, and slip-fiber veins, in which the fibers are nearly parallel to the vein walls. Picrolite veins, in which the serpentine mineral is dense rather than asbestiform, are usually coarsely fibrous or columnar but are locally massive. Some veins are composite and contain both picrolite and asbestos. Microscopic veins, which are observable only in thin section, or whose textural features are discernible only under the microscope, can generally be classified in one or another of these categories, but some have features not displayed by macro-veins.

In dunite and peridotite, serpentinized zones are conspicuous adjacent to both cross-fiber and picrolite veins. In massive serpentinite, these zones are generally indistinct, and in schistose serpentinite none were observed along the rare tiny cross-fiber veins.

Crosscutting relations between different kinds of serpentine veins are scarce, indicating that the veins are generally of the same age. In a few places, veins of slip-column picrolite crosscut veins of cross-fiber asbestos. Veins of cross-fiber asbestos were not seen to crosscut veins of picrolite, except in the limited sense that the two kinds of serpentine crosscut each other in composite veins, but such a relationship may exist.

CROSS-FIBER ASBESTOS

Veins of cross-fiber asbestos are the most conspicuous serpentine veins and are the source of the most prized specimens of chrysotile asbestos, but they are quantitatively of relatively minor importance. Cross-fiber veins are composed chiefly of asbestiform fibrous serpentine (chrysotile). The principal accessory mineral is magnetite concentrated sparsely to heavily along partings and at the margins of the veins, and locally intergrown with the fiber. Brucite, crudely fibrous in habit, is sparsely intergrown with the asbestos in some veins, and patches of calcite occupy irregular volumes in a few veins.

On fresh surfaces of compact unopened fiber, the asbestos is lustrous grayish green, dusky yellowish green, or bronzy yellowish brown. Rarely, it shows faint and delicate variations in color in laminae parallel to the vein walls. Such "color lamination" is more common and more pronounced in some of the asbestos deposits in Quebec (Riordan, 1955, p. 67–81, fig. 4). When teased and separated, the fibers are silky and flexible and almost white. Weathered surfaces are pale green, and weathered fiber generally rather stiff and brittle.

Veins range in thickness from less than a millimeter to as much as 5 cm; most are probably 0.5–5 mm thick, and veins thicker than 3 cm are rare. In tabular dimensions the veins range from a fraction of a centimeter to several tens of meters. The distribution pattern varies greatly, from closely spaced subparallel veins of relatively uniform thickness ("ribbon veins") to diversely oriented, widely spaced veins that vary greatly in thickness. In places, the veins form conjugate sets, but attitudes of the conjugate sets vary markedly from place to place; in other places no systematic orientation of the veins is discernible. Not uncommonly, some of the veins form conjugate sets, whereas others conform to no apparent system. In many places the spacing of the veins varies with their thickness: thin veins tend to be closely spaced, thicker ones, widely spaced.

Cross-fiber veins show considerable diversity in the configuration of their walls, the relations of central partings or seams of magnetite, the distribution pattern of inclusions of serpentine, the orientation of the asbestos fibers with respect to the vein walls, and the relations between veins at their intersections. Figure 5 illustrates some of the configurations and relations shown by cross-fiber veins.

The smaller veins, those a millimeter to several centimeters long, have mostly the form of thin doubly convex lenses. Larger veins are commonly tabular or sheetlike; they are of relatively uniform thickness throughout most of their extent, and wedge out or lens out to knife edges at their margins. In some places, the wedging out is rather abrupt, in others, gently tapering. Some grossly tabular veins consist,

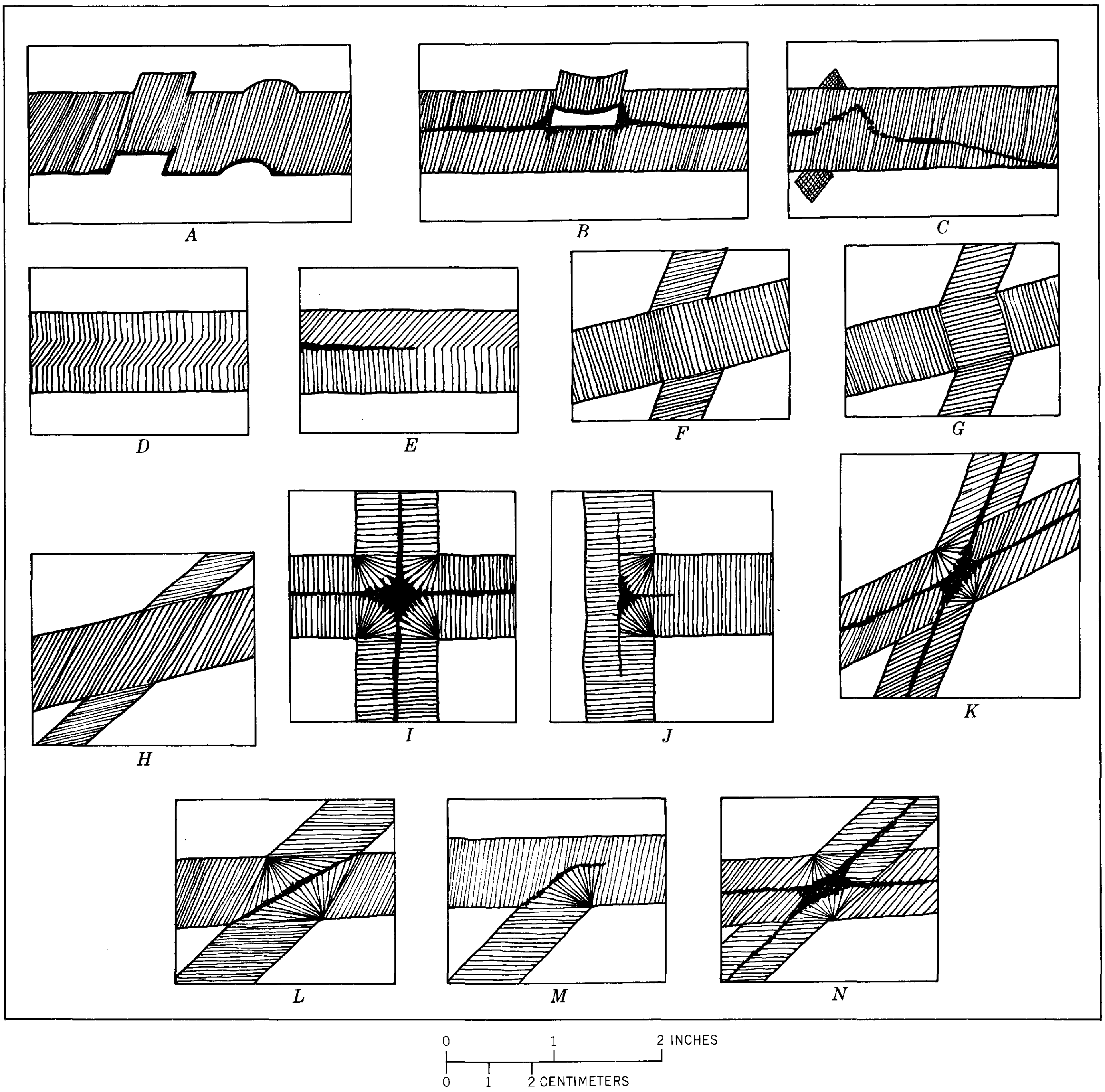


FIGURE 5.—Sketches illustrating some of the geometric relations displayed by cross-fiber chrysotile asbestos veins. White areas represent serpentine, closely spaced fine lines represent chrysotile asbestos fibers, irregular black areas represent magnetite, and cross-hatched pattern represents chromite. *A*, single-fiber vein, showing matching irregularities at opposite ends of fibers; magnetite concentrated along one wall. *B*, double-fiber vein, with magnetite concentrated along central parting, chip of serpentinite displaced from wall. *C*, double-fiber vein, magnetite concentrated along irregular crenellated parting. *D*, single-fiber vein displaying open chevron pattern of fibers. *E*, double-fiber vein with magnetite along central parting merging into single-fiber vein marked by sharp flexure of the fibers along a plane that coincides with that of the central parting. *F–N*, geometric relations of intersecting and merging veins.

in whole or in part, of a series of en echelon gash veins. Most veins are nearly planar or only gently curved, but some are more irregular and depart considerably from planar form.

Nearly all the veins have generally matching opposite walls, and irregularities commonly correspond in minute detail. Inclusions of massive serpentine within cross-fiber veins are commonly seen to fit into cusps and irregular embayments in the adjacent wall. In a few places, matching fragments of fractured grains of chromite or of pseudomorphosed pyroxene appear on opposite walls of a vein. Almost without exception, matching irregularities, fragmented crystals, and inclusions and embayments are at opposite ends of bundles of fibers. Irregularities unequivocally indicative of replacement are rare, and the part for which origin by replacement is indicated constitutes only a small proportion of the total vein width.

Many cross-fiber veins contain one or more partings of thin seams of magnetite—along which may be alined thin chips of serpentinite—that separate the vein into two or more layers of fibers. In others, single fibers extend the full width of the vein. On the basis of partings, it is a common practice to divide cross-fiber veins into single-fiber and double- or multiple-fiber veins (Cooke, 1937, p. 91–99).

Many single-fiber veins have almost no magnetite associated with them, or only inconspicuous concentrations at either or both walls. Others have moderately heavy concentrations of magnetite impregnating the wallrock bordering the veins. Double- and multiple-fiber veins generally have similar concentrations of magnetite at either or both walls, in addition to the central parting seams, but in some veins the margins are virtually free of magnetite.

In some veins, the central parting or partings are fairly regular and occupy about the same relative position throughout the veins. In others, the partings are irregular and cut from near one side of the vein to near the other, or cut back and forth. In some veins, the partings are grossly smooth and continuous; in others, they are strongly crenellated and discontinuous, the discontinuous segments capping bundles of fibers. In detail, even the smooth continuous partings are slightly crenellated. The partings are commonly of fairly uniform thickness, but many pinch and swell; some widen out to small irregular concentrations of magnetite—locally intergrown with the asbestos and of fibrous or columnar habit—that may in places occupy nearly the whole thickness of the vein. The distribution pattern of chips of serpentinite in the veins is similar to that of crenellated segments of magnetite partings.

Single-fiber veins appear to be associated more commonly with massive serpentinite, and double- or multiple-fiber veins, with dunite, but each occurs in both types of rock. Veins of both types in dunite commonly have marginal concentrations of magnetite at either or both walls, whereas those in massive serpentinite have none, or only inconspicuous marginal concentrations.

In most cross-fiber veins, the asbestos is oriented so that its fiber axes are about normal to the vein walls. In few are fibers more oblique than 45°, but in some, the fibers approach the obliquity of slip-fiber. Even where the fibers display marked obliquity, evidence of shearing movement between fiber veins and vein wall is extremely rare or nonexistent. In both the single- and multiple-fiber veins, the fibers generally are straight and are uniformly oriented. In a relatively small proportion of the single-fiber veins, the fibers are flexed into open chevron-type folds along one or more planes parallel to the vein walls. In a few double-fiber veins, the fibers on opposite sides of a parting diverge in orientation by as much as 30°; commonly, such veins pass laterally into single-fiber veins that have a plane of flexure that is coplanar with the parting.

The pattern of intersecting cross-fiber veins is almost infinitely varied, but fundamentally the intersections are of two types, crosscutting and merging. All crosscutting veins are some variation of a basic X-form, whereas merging veins may have a Y, T, or X form. Most intersections are merging. Simple crosscutting intersections are rare, but complex intersections, which have both merging and crosscutting relations, are not uncommon.

In simple crosscutting veins, one vein can be traced distinctly through another. The attitude of the fibers in each vein is virtually uniform at and on either side of the intersection. When crosscutting veins intersect obliquely, one or another of the veins is invariably offset forming an open Z-pattern. Which vein is offset depends solely on the geometric relations of the veins and the fiber orientation, rather than on age relations. Figure 5*F*, *G*, and *H* illustrate some of the possible relations.

In merging intersection, the fibers of the intersecting veins merge by gradual and systematic change in attitude, approximately radially about the edges of intersection of the vein walls (fig. 5 *I-N*). Where intersecting veins are about perpendicular to one another, the fibers commonly are radial to the four (or two, in T-shaped intersections) edges of intersection of the vein walls (fig. 5 *I* and *J*). Where veins intersect in a small dihedral angle, the fibers commonly are radial only about the two (or one, in slant-T intersections) edges at the apices of the obtuse dihedral angles of the walls (fig. 5 *K* and *N*). Commonly, partings of magnetite and chips of serpentinite separate groups of fibers that have different radial foci. In predominantly single-fiber veins, such partings are generally thin and extend only for short distances from the junctions, or join the apices of opposite acute dihedral angles of the vein walls (fig. 5 *J, L,* and *M*). In double-fiber veins, the partings in the junction are commonly irregular and contain abundant magnetite, chips of serpentinite, and disordered bundles of short asbestos fiber (fig. 5 *I, K,* and *N*).

Complex intersections were observed only in multiple-fiber veins and may be confined to them. Features characteristic of merging intersections predominate. The possible relations are extremely varied, but commonly the layers of asbestos adjacent to the vein wall in the obtuse angles merge, and the layers adjacent to the vein wall in the acute angles partly or entirely crosscut each other and the merging layers (fig. 5 *N*).

For additional descriptions and illustrations of relations shown by asbestos veins, refer to Cooke (1937, p. 91–98) and Riordon (1955, p. 68–75).

SLIP-FIBER ASBESTOS

The term "slip-fiber asbestos" is applied to asbestos of two distinctively different habits and modes of occurrence: (1) thin seams of short fibers distributed along closely spaced irregular surfaces of schistosity in schistose serpentinite and along widely spaced minor shear surfaces in dunite and massive serpentinite, and (2) very long fibers in pockets along shear zones or faults that are fairly regular in attitude, tens or hundreds of meters in extent, and a centimeter to a meter wide. The proportion of slip-fiber asbestos in the schistose serpentinite is not apparent except upon close examination by an experienced observer, but such slip fiber constitutes most of the asbestos in the Belvidere Mountain area. The pockets of very long slip fiber in faults and shear zones are spectacular occurrences, but the amount of such material at Belvidere Mountain is negligible.

Slip-fiber asbestos associated with schistose serpentinite is short, generally much less than a centimeter. The fiber veins are thin, and most of the fibers appear to be nearly parallel to the vein walls. In detail, however, the fibers make a very small acute angle with the vein walls, and have a shingled—or, perhaps more accurately, a thatched—arrangement. Over small areas of a surface covered with slip fiber the fibers are essentially parallel to each other, but the orientation varies slightly from place to place. The overall pattern of such a surface forms an irregular mosaic of groups of fibers that have the same general orientation, but are slightly divergent from one another. The boundaries between divergent groups commonly coincide with irregularities in the schistosity surfaces. Most of the fiber is firmly attached to both walls; some is in sheared contact, but the shearing may have occurred during quarrying operations.

In color, flexibility, and silkiness, the slip fiber in schistose serpentine, and also that in dunite and massive serpentinite, appears to be similar to the cross fiber, and the mineralogy does not differ significantly (see preceding section on "Cross-fiber asbestos").

Pockets of slip fiber in faults and shear zones contain fibers as much as 1 m long. All such slip fiber is intergrown with abundant fibrous calcite, which commonly is more abundant than chrysotile asbestos and is as delicately fibrous. The fibrous calcite is stiff and brittle, however, whereas the asbestos is flexible, soft, and silky. The fibers are parallel to the shear zones; no shingled or thatched arrangement is discernible, and the fiber ends are unattached to either wall.

PICROLITE

Veins of picrolite have the same general distribution and size range as veins of cross-fiber and slip-fiber asbestos, but are less abundant than either and are more variable in structure and mineral associations. Nearly all the picrolite has a crudely fibrous or columnar structure, but some is megascopically structureless and massive.

The picrolite ranges from pale olive through dark greenish yellow to dusky green; it is microcrystalline, dense, and has a waxy luster. Coarsely fibrous to crudely columnar picrolite has a splintery fracture; dense and massive picrolite has a conchoidal fracture. The picrolite veins commonly contain small to large amounts of brucite. The brucite is commonly fibrous or flamboyant and is intergrown with the picrolite. Fibrous calcite and clots and disseminated grains of magnetite are abundant in a few veins.

In most veins associated with dunite, peridotite, and massive serpentinite, the columnar structure is about perpendicular to the vein walls, as the fibers are in cross-fiber asbestos. In veins in schistose serpentinite, the columnar structure is nearly parallel to the vein walls, as the fibers are in slip-fiber veins. "Cross column" picrolite veins are mostly less than 2 cm thick, rarely as thick as 10 cm, and the columns are of comparable length. "Slip column" picrolite veins are seldom more than 5 cm thick, but the columns are commonly 15 cm to more than 30 cm long.

The walls of cross-column picrolite veins commonly have matching irregularities at opposite ends of columns, but irregular penetration of the walls by the vein minerals is rather common, though relatively minor. Some picrolite veins of the cross-column type are indistinctly laminated parallel to the vein walls. The laminae reflect chiefly differences in color, but also slight differences in orientation of the coarse fibers and columns in successive laminae (Riordan, 1955, p. 67–81, figs. 3 and 5). Few if any simple unlaminated veins contain partings of magnetite, or significant amounts of serpentinite inclusions. On the other hand, composite veins that contain both cross-fiber asbestos and picrolite commonly have partings associated with the cross-fiber asbestos, and laminated picrolite veins commonly contain thin discontinuous layers of magnetite parallel to the color lamination.

COMPOSITE VEINS

Most composite veins contain both cross-fiber asbestos and cross-column picrolite, but some contain cross-fiber asbestos and slip-column picrolite; perhaps other composite types also exist. Commonly picrolite and asbestos form two or more distinct layers parallel to vein walls. The transition between layers is very sharp, and the boundary grossly regular but minutely irregular in detail. In some composite veins, there is a lateral transition, gradational over a short distance, from a vein composed entirely of cross-fiber asbestos to a vein composed entirely of picrolite. A few composite veins consist of irregular intergrowths of cross-column picrolite, wispy bundles of wavy cross-fiber asbestos, and fibrous brucite.

In the relations of the vein walls, mineralogy, and other features, the composite veins combine the features of the simple component veins.

MICROSCOPIC VEINS

Many of the microscopic veins represent in miniature one or another of all the categories of macroveins, but some have features apparently unlike those of any of the macro-veins. In one type, regular blades of platy serpentine, many times longer than they are wide, are alined parallel to the veins. Under low magnification they appear fibrous, but under very high magnification the apparently fibrous structure is seen to be a perfect cleavage parallel to the length of the blades. Some of the veins of platy serpentine are composite with brucite veins. Typically, they consist of a central vein of brucite about 0.001 mm thick, bordered on either side by fibrous brucite veins about 0.005 mm thick, the fibers at a large angle to the vein; the brucite veins are succeeded outward by veins of platy serpentine about 0.01 mm thick. A second type of micro-vein consists, in thin section, of irregular elliptical particles of serpentine, probably spindle shaped, elongate parallel to the veins. In routine examination, the spindle-shaped particles appear to be structureless, but careful examination under very high magnification indicates that they probably have a very fine fibrous structure about perpendicular to the spindle axes. A third type, observed in only a few thin sections, has an irregular fibrous structure at varied angles to the vein walls; the serpentine commonly is distinctly banded parallel to the walls and has a marked yellowish tinge. The anomalous appearance of this rare third type of vein is due to the orientation of the fiber axes at a relatively large angle to the plane of the thin section.

Most of the micro-veins merge at their intersections, but a few are crosscutting. In one section, yellowish veins of the third type, which were just described, are cut by micro-veins of colorless cross-fiber asbestos.

Most of the micro-veins contain only serpentine, but some contain tiny grains and dustlike particles of magnetite, sparse flakes of brucite, and irregular patches of carbonate.

SERPENTINIZED ZONES AT THE MARGINS OF VEINS

Marginal zones of alteration adjacent to serpentine veins are associated only with cross-fiber asbestos veins and cross-column picrolite veins. Marginal alteration zones in dunite and peridotite are conspicuous features, readily apparent even from a considerable distance (fig. 6). In massive serpentinite they are generally faint because of the decreased contrast in color; thus, the distinctness of the marginal zones varies directly with the proportion of unaltered olivine in the rock.

The alteration zones consist essentially of serpentine, brucite, fine-grained or dusty particles of magnetite, and tiny relict grains of olivine and chromite; variations in relative proportions, texture, and dis-

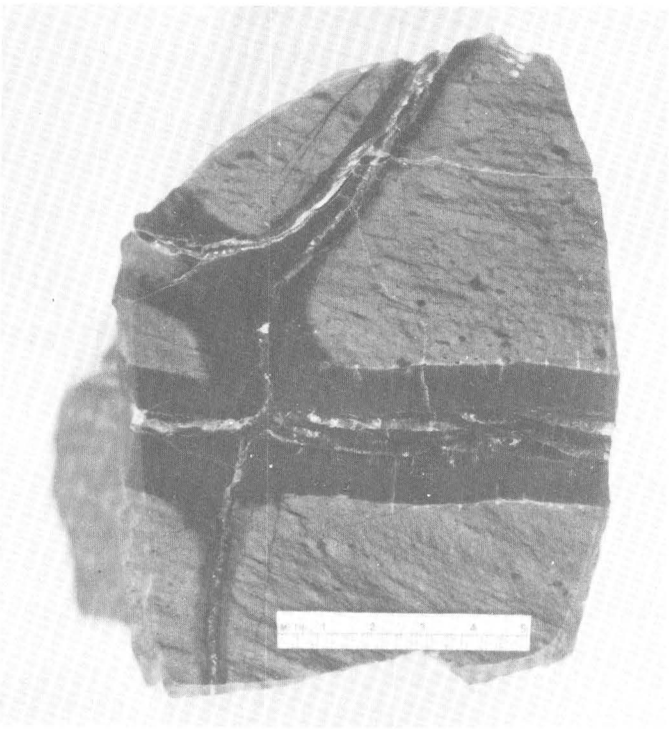


FIGURE 6.—Photograph of cross-fiber vein in dunite, showing marginal alteration zones.

tribution impart a varied but distinct layering to the zones. The outer margins are generally marked by a thin layer that is 1 mm or less wide; its color is grayish yellow green. Not uncommonly there are several such layers, of different shades of yellow green and increasingly less distinct inward toward the walls of the asbestos or picrolite veins. The inner layers weather white, very pale green, or very pale buff; the outer layers, various shades of buff and light brown.

Typically, immediately next to the vein the alteration zone consists of irregular shreds and chips of fibrous serpentine in a subordinate matrix of nearly isotropic serpentine of uncertain but probably platy habit. A few millimeters away from the vein scattered flakes and blades of platy serpentine appear. Outward toward the dunite, the proportion of fibrous serpentine and isotropic serpentine decreases, and the proportion of distinctly platy serpentine increases. Brucite occurs throughout the entire zone, commonly as disseminated flakes, and it tends to be more abundant in the outer part of the zone. Magnetite occurs as scattered grains and dispersed dustlike particles; its distribution is commonly irregular, but in many marginal zones it is concentrated more heavily in a narrow zone immediately next to the vein, or between the asbestos and the vein wall. Very sparse relict grains of olivine occur chiefly in the outer parts of the alteration zones. Chromite is sparsely distributed as in the unaltered dunite.

Some alteration zones consist almost entirely, except for abundant grains of magnetite, of diversely oriented shreds and flakes of fibrous serpentine and almost no isotropic or platy serpentine. Others contain many veins and disseminated flakes of brucite. Layering near the outer margins of some of the zones reflects chiefly different concentrations of brucite in the different layers. A few of the marginal zones are strikingly asymmetric in the distribution of minerals. In some of them, the marginal zone on one side of the vein shows the typical relations just described, whereas the zone on the opposite side of the vein consists almost entirely of brucite and only sparse bladed serpentine, fibrous serpentine, and magnetite. In others, the distribution of brucite is seemingly haphazard, and the relations of the serpentine minerals are irregular.

The boundaries of marginal alteration zones are smooth and sharp in overall appearance. In detail, the transition from the altered zone to dunite is gradational over a distance of a fraction of a millimeter, and the boundary is minutely irregular. Where the alteration zone crosscuts layering in the dunite, tiny tongues from the alteration zone penetrate as much as 1–2 mm along some of the layers. In some places, tiny fractures about normal to the marginal zone boundaries extend a few millimeters into the dunite and serpentinite on either side of the boundary and commonly mark sawtoothed irregularities in the boundary.

Geometrical relations of the marginal alteration zones are varied, but they conform in aggregate to a general pattern. Most of the marginal zones are roughly symmetrical with respect to the vein throughout most of its extent and are of about uniform width along it, and the outer margins of the zones conform grossly with the vein walls. At vein intersections, the marginal zones are commonly somewhat thicker and more irregular. Not uncommonly, the marginal zones thicken or thin locally in gently tapering fashion along a vein of uniform thickness; in some places the marginal zones maintain a generally uniform thickness along a vein that varies appreciably in thickness. Along some veins, the marginal zone on one side is somewhat thicker than the zone on the other side, and not uncommonly the relation is reversed at different places along the vein. In a few places, veins as much as 5 mm thick have no marginal alteration zones, even in dunite, and some less than 1 mm thick have marginal zones several tens of times as thick.

Though in a few places the ratio of vein thickness to total thickness of vein plus bordering alteration zones varies widely from the average, the ratio is fairly constant for the great majority of veins. For most of the cross-fiber asbestos vein in the Belvidere Mountain area, the ratio ranges from about 1:4 to about 1:10, and the average is probably near 1:6 or 1:7. No significant difference between veins in fresh dunite and in relatively completely serpentinized massive serpentinite was detected, but the data are insufficient to be conclusive. Few reliable observations are available on picrolite veins, but the ratio appears to be somewhat larger, perhaps near 1:3 or 1:4. These ratios are similar to those obtained by Cooke (1937, p. 103–110) in a careful study of asbestos deposits in Quebec.

AMPHIBOLE ASBESTOS

A small mass of amphibole asbestos, consisting of fibers more than 30 cm long intergrown with fibrous calcite, was found in quarrying operations in a shear zone at or near the northeast contact of the Lowell quarry body; none was observed in place. The fiber is similar to masses of slip-fiber chrysotile, except that it is more brittle. Mineralogical examination by G. T. Faust (written commun., 1951) disclosed the asbestos to be a member of the tremolite-actinolite series.

OTHER VEINS

Less abundant and generally less conspicuous than the serpentine veins are veins of magnetite, carbonate, brucite, and magnetite-brucite-carbonate. Of these, veins mainly of magnetite are the most abundant.

Veins of magnetite are of two types: moderate to heavy concentrations of disseminated fine-grained magnetite in generally regular veinform masses in massive serpentine or dunite; and irregular pods and lenses of massive, coarse-grained chromian magnetite in schistose serpentinite. The veinform masses of fine-grained magnetite generally are clearly controlled by layering or by crosscutting fractures, but most such magnetite is heavily disseminated in, and bears a replacement relation to, the adjoining serpentinite. In many places, seams of magnetite extend for considerable distances beyond the lateral limits of asbestos veins, along the same fractures. Most of the veins are fairly uniform and less than a centimeter or so thick, but some widen out locally to form irregular masses several centimeters across which are broadly gradational into the serpentinite.

Pods and lenses of massive chromian magnetite are very rare and no more than a few centimeters thick at Belvidere Mountain, but a mass more than 3 m thick and 30 m long is exposed about 3 km north-northeast of Troy village (Cady and others, 1963, p. 71, pl. 1). This pod is very irregular, and the massive magnetite is everywhere in sheared contact with the enclosing schistose serpentinite. The magnetite rock in the pod consists of closely packed anhedral grains of magnetite 1–5 mm across in a matrix of bladed serpentinite. Ilmenite forms narrow rims around scattered grains of sphene that have cores of rutile and tiny veinlets along cleavage surfaces in magnetite. Calcite occurs as sparsely scattered grains and tiny veinlets. Magnetite constitutes 65–75 percent of the rock; serpentine and carbonate, 35–20 percent; and ilmenite, rutile, and sphene, <1–5 percent.

Brucite and calcite and mixtures of one or both with magnetite commonly form scattered tiny veinlets in the dunite, peridotite, and serpentinite. Commonly the veinlets are fracture controlled. Brucite veins of this type associated with serpentine veins have been described under "Microscopic veins." Some veins are of uniform width, have sharp boundaries, and show little evidence of replacement. Some of the regular veins widen out locally to irregular volumes showing predominantly features characteristic of replacement origin. Many veins are entirely or predominantly irregular. Commonly, the veins having features characteristic of replacement contain narrow central fissurelike zones entirely of magnetite. In others, the distribution of minerals is fairly uniform or entirely irregular.

These veins range in composition from entirely calcite or brucite to various intermixtures of magnetite, brucite, and calcite. In many, the textural relations of the three minerals are not diagnostic, but commonly the calcite replaces brucite.

MINERALOGY AND PARAGENESIS

Dunite, peridotite, and chromitite contain primary minerals, which are the earliest formed minerals in the decipherable history of the rock—that is, those that are not recognizably derived from preexisting minerals. These three rocks also contain secondary or metamorphic minerals, derived from the primary minerals or introduced into a host rock during the metamorphic history. Serpentinite is constituted entirely of secondary minerals or contains sparse relict primary minerals of the dunite and periodotite. Serpentine veins and their marginal alteration zones, and brucite-calcite-magnetite veins are composed entirely of secondary minerals. Magnetite veins are largely secondary, though some of the veins parallel to layering in the dunite and peridotite, and some of the irregular pods may be entirely or largely primary.

The primary minerals are olivine, pyroxene, chromite, magnetite (chiefly or entirely chromian), sulfides, and sulfarsenides. They vary little in composition, habit, and association, or vary systematically. The secondary minerals are varieties of serpentine, anthophyllite, brucite, chlorite, carbonate, and magnetite (nonchromian). They vary widely in textural relations and mineral association.

The primary minerals are anhedral to subhedral grains that form a granular mosaic texture that is characteristic of many plutonic igneous rocks; the secondary minerals form irregular interlocking mosaics, felted aggregates, and diverse fibrous textures that are characteristic of metamorphic minerals; the secondary minerals also show characteristic zonal relations to one another and to primary minerals that are indicative of genetic sequence. The minerals in veins and marginal alteration zones commonly have a zonal distribution. Chromite grains and associated rimming minerals show distinctive zoning. A translucent core of chromite is surrounded by an opaque rim of chromite or chromian magnetite and magnetite, which in turn is rimmed irregularly by chlorite. In many places the chlorite rim is bordered by fresh olivine; more commonly, the zoned assemblage is surrounded by antigorite. In some such associations, zoning in the antigorite adjacent to the chlorite is indicated by a zonal pattern in the abnormal interference colors. In a few slides, grains of opaque chromite in massive serpentinite have partial rims of olivine tightly stuck to the chromite. Such grains have no rim of chlorite adjacent to the olivine, but where the olivine rim is absent the chromite is separated from the enclosing serpentine by a rim of chlorite.

OLIVINE

In the least serpentinized dunite and peridotite, olivine forms a mosaic of subequant grains predominantly 0.5 to 2 mm across and of irregularly polygonal outline in thin section. A few crystals of olivine are much larger. One rock specimen contained a single crystal more than 3 cm long and 2 cm thick, which enclosed several small subhedral crystals of olivine of different crystallographic orientation. All the olivine grains are highly fractured, but commonly many or most show uniform extinction. In every thin section, a few grains show undulatory extinction, and a few grains are twinned; in many thin sections, most of the grains show undulatory extinction, and many are twinned. The large single crystal shows pronounced undulatory extinction and locally a vague mosaic pattern of areas that extinguish at slightly different positions.

In even the freshest rocks, most of the olivine grains are slightly altered at their boundaries to bladed serpentine and are locally rimmed by tiny veins of chrysotile. Serpentine invades many grains along fractures and cleavage directions, particularly (010) cleavage. As serpentinization increases, the control by morphologic and crystallographic features is less evident. In massive serpentinite, olivine remains only as small, very irregular relict grains. Marginal zones of such relict grains are somewhat cloudy. Under very high magnification, these cloudy zones are seen to contain many closely spaced hairline fractures, commonly only 0.002 mm apart, along which the olivine is serpentinized. Many of the tiny fractures have a six-rayed stellate pattern.

The olivine shows several distinctive variations in habit. Inclusion-free grains of olivine are grayish yellow green (5*GY* 7/2)[2] in hand specimen, and colorless in thin section. Most of the grains show only an irregular fracture, but many show good (010) and poor (001) and (021) cleavage. A few contain conspicuous planes of tiny inclusions, chiefly of unidentifiable nonopaque grains but also of magnetite, along crystallographic directions (021), (001), and (101). In a somewhat similar manner, olivine is intergrown in a few places with skeletal crystals of magnetite so that (021) of the olivine is parallel to the dominant direction of skeletal growth of the magnetite. The result of this relation is that the olivine has apparently inclined extinction, and such intergrowths are easily mistaken for pyroxene grains in which magnetite is concentrated along cleavage directions.

The olivine is generally uniform in optical properties and shows no systematic variations within the ultramafic body. The optic axial angle, $2V$, is about 85°–90°, the optic sign is (+), $\beta = 1.664–1.671$, and the birefringence about 0.038–0.042.

Although no analyses of olivine separated from the dunite were obtained, two analyses of the freshest available dunite (analyses 1 and 45, table 1; specimen numbers AR–13f, AV–248, table 3*A*), comparatively free of accessory primary minerals, yield reliable mineral formulas of olivine, based on cal-

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[2] Color designations are based on the Rock-color chart of the National Research Council (Goddard, 1948).

culated modes. The calculated formulas for the olivine, corrected for contaminating spinellids, serpentine, and brucite, and reduced to ideal values, are:

AR–13f:

$$Mg_{1.96}Fe^{+2}_{0.04}Si_{1.00}O_{4.00},$$

AV–248:

$$Mg_{1.95}Fe^{+2}_{0.05}Si_{1.00}O_{4.00}.$$

The ratio of Fe^{+2}/Mg in the calculated formulas is 0.02–0.025. The ratio of Fe^{+2}/Mg, based on the optical properties (AR–13f: $2V=88°$ (+), $\beta=1.66$; AV–248: $2V=88°$ (+), $\alpha=1.644$, $\beta=1.664$, $\gamma=1.683$) (Deer and others, 1962, v. 1, fig. 11), is about 0.05 and 0.06. For AR–13f and AV–248, respectively, the ratio of total Fe/Mg in the whole-rock analyses is 0.07 and 0.06, and of Fe^{+2}/Mg, 0.036 and 0.046 (table 1, analyses 1 and 45). Inasmuch as an appreciable proportion of the Fe is contained in magnetite and other spinellids and in serpentine, it is evident that the ratio of Fe^{+2}/Mg in the olivine must be appreciably lower than in the whole rock. Consequently, the calculated analyses are believed to indicate more reliably than the optical data the proportion of Fe in the olivine, though they may err slightly on the low side because of possibly slightly high estimates of the proportion of magnetite in the calculated modes.

PYROXENE

Though pyroxene originally constituted as much as 5 percent of the dunite and 5–10 percent of the peridotite, it is now largely altered to anthophyllite(?), antigorite, brucite, and magnetite, which are pseudomorphic after the pyroxene in gross outline or retain features of the pyroxene cleavage. Unaltered pyroxene occurs only very sparsely as scattered relict cores of partly altered grains.

The pseudomorphs after pyroxene are generally tabular and range in outline in thin section from irregularly elongate to elliptical or rectangular; most are 2–4 mm long and 1–2 mm wide. Some contain blebs and irregular stringers of olivine which appear to be relicts of inclusions in the pyroxene and of tiny veinlets of olivine that healed fractures in the pyroxene.

Unaltered pyroxene is so sparse that it was not practicable to determine indices of refraction. Other optical properties, where determinable, were uniformly: $2V=60°$, optic sign (+), extinction angle $\gamma\wedge c=45°$, nonpleochroic, and colorless. On the basis of these optical properties, the pyroxene is inferred to be diopside or augite that contains an appreciable content of Fe^{+2}.

CHROMITE AND MAGNETITE

Chromite, the sole essential constituent of the chromitite, is also a common constituent of the dunite, peridotite, and serpentinite but rarely constitutes as much as 1 percent of the rock. Magnetite, the sole essential constitutent of magnetite veins, is commonly a minor constituent of the chromitite and is a ubiquitous but generally minor constituent of the dunite, peridotite, and serpentinite; it is also present in most of the serpentine and other veins, commonly as a minor constituent but locally in proportions greater than 50 percent.

In the chromitite layers and pods, chromite occurs as subequant to slightly elongate anhedral grains as much as 5 mm across, commonly in a closely packed granular mosaic. The chromite grains are commonly highly fractured and the fractures filled with platy serpentine. In some of the chromitite bodies, chromite is the sole spinellid; in others, the chromite grains are partially rimmed by, veined by, and intergrown with varying but subordinate amounts of magnetite.

The chromite varies appreciably in physical and optical properties. Hand specimens are black and have a dull metallic luster; the streak ranges from brown to brownish gray. The powdered chromite ranges, in reaction to a hand magnet, from nonmagnetic to weakly magnetic. In polished specimens from the west peak of Belvidere Mountain and from the C-area quarry (A–BM–29 and A–BM–53–116; see section on "Chromitite" under "Igneous rocks, serpentinite, and veins" in the section on "Ultramafic and associated rocks"), the chromite has the following optical properties, as observed with an ore microscope:

A–BM–29: gray in reflected light, having red internal reflections under crossed nicols; isotropic, having black cross on white field under conoscopic observation using crossed nicols; dispersion of the reflection rotation (Cameron, 1961, p. 115) is DR_r $+r>v$; not scratched readily with a needle.

A–BM–53–116: light gray in reflected light, no internal reflections detected, isotropic, black cross under conoscopic observation using crossed nicols; dispersion of the reflection rotation not detectable; not scratched readily with a needle. Magnetite in A–BM–53–116 is silver white in reflected light, isotropic, and does not show detectable dispersion of the reflection rotation.

In the magnetite veins, serpentine veins, and brucite-calcite-magnetite veins, magnetite forms irregular grains and aggregates 0.1–1 mm across; in addition, magnetite forms fibrous intergrowths with asbestos in many of the cross-fiber asbestos veins. In thin section, observed by reflected light, the magnetite is light steel gray.

In dunite, peridotite, and serpentinite, spinellids occur chiefly as grains that fall into three size ranges of distinctive habit, associations, and properties. The largest size range consists of euhedral to anhedral grains 1–2 mm across, having grossly regular boundaries. Most have translucent cores that are dark reddish brown in transmitted light and brownish black in reflected light. Surrounding the translucent core, and bordering fractures in it, is a shell of opaque spinellid of slightly higher reflectance. In massive serpentinite, and in moderately to extensively serpentinized dunite and peridotite, many of the grains contain a narrow outer rim composed of an aggregate of tiny grains that are steel gray in reflected light. The intermediate size group consists of anhedral opaque grains predominantly 0.1 mm across. These grains have generally an appearance in reflected light similiar to that of the inner opaque shells surrounding the translucent cores of the larger grains, but some are brighter and grayer. Many contain narrow rims of aggregates of tiny grains having steel-gray reflectance. The smallest size group consists of opaque dustlike particles and tiny grains and irregular clusters of grains generally less than 0.05 mm across.

Other magnetite is varied in habit but uniformly steel gray in reflected light. Concentrations of magnetite form tabular aggregates along relict cleavage directions in pseudomorphs after pyroxene and scaly aggregates of tiny grains 0.1–1 mm in diameter along schistosity surfaces in serpentinite. Skeletal crystals in the form of incomplete reticulate networks of irregular laths and elongate particles of magnetite are locally intergrown with olivine.

Though precise data on the variations in composition of the spinellids are not available, several chemical analyses of chromite and magnetite, together with observed variations in optical properties of the spinellids, provide general information on their composition.

Chromite in the massive chromitite layers and pods ranges in composition from near that of the chromite end member of the spinel series to compositions considerably higher in iron and lower in chromium. A chemical analysis of massive chromite (sample A–BM–29, analysis 41, table 1) yields a calculated formula composition, corrected for serpentine and olivine, of

$$(\mathrm{Fe}^{+2}{}_{0.49}\mathrm{Mg}_{0.43}\mathrm{Mn}_{0.05})\quad(\mathrm{Cr}_{1.56}\mathrm{Fe}^{+2}{}_{0.20}\mathrm{Al}_{0.25})\mathrm{O}_4.$$

The atomic ratio of Cr/Fe is about 2.3. A partial chemical analysis of material from a small chromite-magnetite vein (sample A–BM–53–116, analysis 42, table 1) indicates an average atomic ratio of Cr/Fe of about 0.8. About one-fourth of the material was magnetite, so for the chromite mixture, the ratio of Cr/Fe is near 1.1. The total range in composition of the chromite in the layers and pods is therefore at least as great as is indicated by these analyses, possibly greater. Disseminated chromite in the dunite, peridotite, and serpentinite has a range in optical properties comparable with that of the analyzed specimens and probably has a similar range in composition. The translucent reddish-brown grains probably approach A–BM–29 in composition. The inner opaque rims and the larger opaque grains probably have an Cr/Fe ratio as low as or lower than that of A–BM–53–116, and appreciably less Mg than A–BM–29.

All the magnetite appears to contain small amounts of chromium, aluminum, and magnesium, and lesser amounts of titanium and manganese. An analysis of magnetic concentrate from the tramp-iron magnet (sample VT–24, analysis 20, table 1), which consisted chiefly of magnetite but contained a small proportion of admixed chromite, is represented by the following formula:

$$(\mathrm{Fe}^{+2}{}_{0.88}\mathrm{Mg}_{0.15})\ (\mathrm{Fe}^{+3}{}_{1.86}\mathrm{Al}_{0.06}\mathrm{Cr}_{0.04})\ \mathrm{O}_4;$$

an analysis of scaly aggregates of magnetite (sample A–BM–56–6, analysis 47, table 1) from schistosity surfaces in serpentinite is represented by the following formula:

$$(\mathrm{Fe}^{+2}{}_{0.90}\mathrm{Mg}_{0.00}\,\mathrm{Mn}_{0.01}\mathrm{Ti}_{0.01})(\mathrm{Fe}^{+3}{}_{1.84}\mathrm{Cr}_{0.09}\mathrm{Al}_{0.09})\mathrm{O}_4.$$

SULFIDES AND SULFARSENIDES

Sulfides and sulfarsenides occur sparsely in the dunite, peridotite, and serpentinite. They occur chiefly as isolated elongate blebs and irregular grains, but some are intergrown with magnetite.

The sulfides are predominantly pale brassy yellow in reflected light. Most appear to be pyrite (FeS_2), but some may be pyrrhotite ($\mathrm{Fe}_{1-x}\mathrm{S}$).

The sulfarsenides are silvery gray to grayish blue in reflected light. Specific identification of the minerals was not made; elsewhere in Vermont, Clemmer and Cooke (1936, p. 12) identified gersdorffite in the ultramafic rocks. Probably, therefore, gersdorffite, and perhaps arsenopyrite, are the principal sulfarsenides in the ultramafic rocks.

AMPHIBOLE

Two varieties of amphibole occur in the ultramafic rocks: anthophyllite as an alteration product of pyroxene, and amphibole asbestos in slip-fiber veins. Anthophyllite is common but sparse in the more pyroxenic dunite, peridotite, and massive serpentinite. In a few thin sections, many of the grains of anthophyllite surround relict cores of pyroxene; most grains, however, contain no relict pyroxene but preserve the form and cleavage of pyroxene grains which they have replaced. Many grains of anthophyllite are slightly to extensively altered to serpentine.

The anthophyllite is fibrous in habit. It is commonly cloudy and composed of an aggregate of small grains, so that optical properties are difficult to determine. The color in thin section is pale clove brown. The optic axial angle, $2V$, is about 90°; indices are about 1.625–1.632; and extinction is parallel to the fiber direction. Positive identification by optical methods is uncertain, but the mineral probably is a magnesian anthophyllite. On the basis of Rabbitt's (1948, fig. 4, p. 295) chart, it contains about 5 percent combined $FeO+Fe_2O_3+TiO_2+MnO$, which corresponds to a formula composition in which the ratio of Mg/Fe^{+2} is on the order of 6.

The amphibole asbestos was identified as tremolite-actinolite by George T. Faust (written commun., 1951).

SERPENTINE

Serpentine constitutes an appreciable proportion of even the least serpentinized dunite and peridotite and is the principal constitutent of the massive serpentinite, the schistose serpentinite, and the serpentine veins. The habit and associations of the serpentine are many and varied. Fundamentally, individual particles are either fibrous or platy, but each kind has a wide variety of textural relations and habits.

All the serpentine is moderate yellow green (5*GY* 6/2) to dusky yellow green (5*GY* 7/2) in hand specimen and colorless in thin section. Massive material commonly is darker because of abundant finely disseminated magnetite. Asbestiform serpentine is pale green to almost white when the fibers are separated and fluffed.

On the basis of optical properties determined by the authors, and DTA and X-ray studies made by G. T. Faust (written commun., 1961), all or nearly all the platy serpentine in the dunite, peridotite, and serpentinite—that is, exclusive of serpentine veins and adjacent alteration zones in the dunite and massive serpentinite—is identified as antigorite, and all the fibrous serpentine, as chrysotile. Some nearly isotropic serpentine apparently of platy habit (so-called serpophite) may be six-layer orthoserpentine. Some serpentine identified as antigorite appears crudely fibrous, but under highest magnification it proves to have a fundamentally platy habit; the apparently fibrous structure results from an excellent cleavage in very regular laths or flakes, or from the wispy habit. All the serpentine in the asbestos veins and the picrolite veins is chrysotile. The serpentine in the zones of marginal alteration bordering the serpentine veins is predominantly a mixture of lizardite and chrysotile, and, locally, minor antigorite.

Platy serpentine is the most abundant type of serpentine mineral; it ranges in particle shape from bladed, lathlike, or flaky to subequant grains having very irregular boundaries, and in size from blades as much as 1–2 mm long to grains, flakes, and shreds as small as 0.005 mm in maximum dimension. In the relatively unserpentinized dunite and peridotite, large blades of platy serpentine commonly occur at and parallel to grain boundaries, and along cleavage and other crystallographic directions of the olivine. Where the olivine has a lattice-oriented fabric, the blades of platy serpentine are commonly parallel. Fine-grained platy serpentine commonly invades the olivine irregularly along grain boundaries and fractures. Platy serpentine that has replaced pyroxene and anthophyllite generally has a parallel or rectilinear arrangement of bladed particles.

Large blades of platy serpentine commonly form diversely oriented, parallel, or reticulate patterns in a groundmass of fine-grained platy serpentine. In places, the coarse blades form flamboyant sheaves. Patches of platy serpentine pseudomorphic after pyroxene or anthophyllite have a distinctive parallel or reticulate pattern. In some samples, coarse blades of platy serpentine form a network that encloses subequant patches of fine-grained platy serpentine, or of very low birefringent serpentine that is mostly of platy habit, but that may be in part composed of diversely oriented submicroscopic fibers. The network of large blades has the same pattern as the grain boundaries of olivine in the dunite.

Fibrous serpentine is generally markedly subordinate to platy serpentine but locally constitutes as much as 30–40 percent of the matrix serpentine—

that is, exclusive of macroscopic asbestos veins. In one thin section, all the matrix serpentine appeared to be fibrous. Much of the disseminated fibrous serpentine occurs in microscopic tabular to lenticular veinlets of cross-fiber asbestos, but in many thin sections an appreciable proportion forms wispy bundles and sheaves of parallel fiber not obviously veinform. All have a silky luster and parallel arrangement of the fibers. In many or most thin sections, the apparently nonveinform patches may result from the thin section just grazing the surface of tiny lenses of cross-fiber asbestos. That this is commonly the case is suggested by the observation that many of the isolated sheaves and bundles of fibrous serpentine give biaxial positive interference figures with small to moderate optic axial angles; such would be the result where the thin section is nearly parallel to the vein wall, at a large angle to the fiber axes. (See section on "Mineralogy of the serpentine group" under "Ultramafic and associated rocks.") However, in the samples very high in, or entirely of fibrous serpentine, the fibers cannot be veinform. Some patches of very low birefringent serpentine on which an interference figure cannot be obtained may be composed of diversely oriented, microscopically indistinguishable fibers, but most are of platy serpentine.

Admixtures of fibrous serpentine modify the textural patterns of platy serpentine in varying degree. The fibrous serpentine is distributed in tiny tabular veinlets that reflect the control of grain boundaries, fractures, and cleavage; in tiny lenticular veinlets; and in wispy sheaves and bundles. One thin section in which the matrix serpentine is entirely fibrous consists of an irregular interlocking mosaic of subequant wispy patches of parallel fibers. The apparent birefringence of the patches ranges from yellow to dark gray, depending upon the orientation of the fibers with respect to the thin section. Bundles whose fiber axes are at a large angle to the thin section have much the same appearance as very fine granular antigorite.

Fibrous serpentine in asbestos veins forms fine silky fibers that have uniformly parallel arrangement. In picrolite veins, it forms bundles and sheaves of slightly disarrayed and commonly less distinct fibers; the bundles and sheaves are nearly parallel and impart a coarsely fibrous or columnar structure.

In the massive serpentinite, the platy and the fibrous serpentine combine in an almost infinite variety of textures, but each is characterized by various combinations of a few basic textures.

Small blades, laths, and elongate shreds generally form a felted mass of diversely oriented particles; not uncommonly, a few of the blades and laths form a rectilinear pattern in a felted matrix. Coarsely flaky particles of serpentine range from diversely oriented to subparallel, and in thin section present a variable, mosaic to schistose texture. Under crossed nicols, subequant granular particles of platy serpentine have a texture like that of galvanized metal. In particular specimens, either the felted texture or the "galvanized" texture may predominate almost to the exclusion of the other, or both may characterize the rock in varying degree. Textures characteristic of flaky particles commonly characterize an entire sample, but some samples contain small admixtures of flaky serpentine dispersed throughout bladed or subequant particles.

ANTIGORITE

All the antigorite is generally similar in optical properties but varies appreciably within rather well-defined limits. Abnormal blue interference colors are characteristic. They range from very pale blue (5*B* 8/2) to moderate blue (5*B* 5/6); most are pale blue (near 5*B* 6/2). Antigorite next to halos of chlorite that rims grains of chromite is deeper blue than the surrounding antigorite. The optic axial angle, $2V$, ranges from 0° to 65°, but is predominantly less than 20°. The small apparent optic axial angle may result largely from superposition of layers having random orientation about the c-axis. The optic sign is $(-)$, and the sign of elongation $(+)$. The β index of refraction ranges from 1.562 to 1.573 but is predominantly in the range 1.567–1.570. The α and γ indices can seldom be measured reliably, but minimum α' and maximum γ' indices are seldom more than 0.003 smaller and larger than the β index. The apparent birefringence ranges from 0.002 to 0.009 and is generally about 0.003–0.005.

Analyses of pure antigorite were not obtained, but analyses were made of several samples of rock consisting principally or almost entirely of antigorite. By correcting for contaminating minerals whose compositions and proportions in the samples are fairly accurately known, the approximate compositions of the antigorite can be calculated. The calculated formula compositions of the antigorite in these samples are tabulated below.

Formula composition of antigorite calculated from chemical analyses of serpentinite

[Corrected for estimated contaminants of magnetite, chlorite, and carbonate]

AV–81m (VT–4), analysis 3, table 1:

$$(\mathrm{Mg}_{5.6}\mathrm{Fe}^{+2}{}_{0.04}\mathrm{Fe}^{+3}{}_{0.1})\ (\mathrm{Al}_{0.1}\mathrm{Si}_{3.9})\,\mathrm{O}_{9.9}\,(\mathrm{OH})_{7.6})$$

AR–73 (VT–6), analysis 4, table 1:

$$(\mathrm{Mg}_{5.3}\mathrm{Fe}^{+2}{}_{0.02}\mathrm{Fe}^{+3}{}_{0.2})\ (\mathrm{Al}_{0.2}\mathrm{Si}_{3.8})\,\mathrm{O}_{9.9}\,(\mathrm{OH})_{7.2}$$

AV–127 (VT–8), analysis 5, table 1:

$$(\mathrm{Mg}_{5.0}\mathrm{Fe}^{+2}{}_{0.4})\ (\mathrm{Al}_{0.1}\mathrm{Si}_{3.9})\,\mathrm{O}_{9.8}\,(\mathrm{OH})_{7.2}$$

AV–209 (VT–16), analysis 12, table 1:

$$(\mathrm{Mg}_{5.2}\mathrm{Fe}^{+2}{}_{0.4})\ (\mathrm{Al}_{0.1}\mathrm{Si}_{3.9})\,\mathrm{O}_{10.0}\,(\mathrm{OH})_{7.2}$$

A–BM–53–12, analysis 21, table 1.

$$(\mathrm{Mg}_{4.8}\mathrm{Fe}^{+2}{}_{0.3})\ (\mathrm{Al}_{0.1}\mathrm{Si}_{3.9})\,\mathrm{O}_{10.2}\,(\mathrm{OH})_{6.3}$$

A–BM–53–1, analysis 30, table 1:

$$(\mathrm{Mg}_{5.2}\mathrm{Fe}^{+2}{}_{0.1}\mathrm{Fe}^{+3}{}_{0.04})\ (\mathrm{Al}_{0.1}\mathrm{Si}_{3.9})\,\mathrm{O}_{9.7}\,(\mathrm{OH})_{7.3}$$

AV–127, analysis 55, table 1:

$$(\mathrm{Mg}_{5.0}\mathrm{Fe}^{+2}{}_{0.4}\mathrm{Fe}^{+3}{}_{0.05})\ (\mathrm{Al}_{0.1}\mathrm{Si}_{3.9})\,\mathrm{O}_{10.0}\,(\mathrm{OH})_{7.2}$$

CHRYSOTILE

Structural varieties of chrysotile cannot be distinguished from one another by optical properties. In the DTA and X-ray studies, clinochrysotile was specifically identified in several samples; in other samples containing chrysotile, the specific structural type of chrysotile was not identified. Clinochrysotile predominates greatly over orthochrysotile in worldwide occurrence. Orthochrysotile was not identified from deposits in Vermont, but specimens from Thetford, Quebec, contain as much as 7 percent orthochrysotile (Whittaker, 1956b, p. 862; Whittaker and Zussman, 1956, p. 114–115). Therefore, orthochrysotile is probably present in small amounts at Belvidere Mountain, but subordinate to clinochrysotile.

All the chrysotile is colorless in thin section, and most shows normal first-order yellow birefringence colors. Some groundmass chrysotile in rock composed largely or entirely of chrysotile has low apparent birefringence that has a faint abnormal blue tinge. The anomalously low birefringence and abnormal color may be caused by superposition of wispy layers of fibers having different orientations.

Optical properties of the chrysotile are distinctive, but they vary appreciably in different specimens. The variable optic axial angle, depending upon the orientation of the thin section with respect to the fiber axes, and the positive optic sign, are distinctive of chrysotile (see section on "Mineralogy of the serpentine group" under "Ultramafic and associated rocks"). Another distinctive optical property is the low index relative to that of associated antigorite. The greater index of refraction, N, probably is generally near true γ; it ranges from 1.538 to 1.562, but is predominantly in the range 1.545–1.555 and is always less than the lowest index of associated antigorite. The smaller index of refraction, n, is of less fundamental significance, because it represents some average between α and β. Measured values range from 1.518 to 1.555 and are predominantly in the range 1.538–1.548. The highest measured birefringence was 0.020, but generally it is in the range 0.010–0.015. Extinction is parallel, and the sign of elongation (+) is parallel to the fiber axes.

Two chemical analyses of chrysotile fiber were obtained, one of milled fiber (sample A–BM–57–1, analysis 50, table 1), and one of long silky fiber intergrown with fibrous calcite (sample AR–55, analysis 56, table 1). All rock particles and nearly all the magnetite were removed from the material in analysis 50 (A–BM–57–1); so the analysis is of virtually pure chrysotile. In analysis 56 (AR–55), the removal of all the magnetite from the material left a mixture of almost pure calcite (80 percent) and chrysotile (20 percent). A reliable formula composition of the chrysotile was therefore obtained by subtracting from the analysis amounts Ca, Mg, and Mn equal to the content of CO_2. The calculated formula compositions are:

A–BM–57–1:

$$(\mathrm{Mg}_{6.02}\mathrm{Fe}^{+2}{}_{0.04}\mathrm{Fe}^{+3}{}_{0.13})\ (\mathrm{Si}_{3.87}\mathrm{Al}_{0.12}\mathrm{Fe}^{+3}{}_{0.01})\,\mathrm{O}_{9.80}\,(\mathrm{OH})_{8.77}.$$

AR–55:

$$(\mathrm{Mg}_{6.12}\mathrm{Fe}^{+2}{}_{0.07}\mathrm{Fe}^{+3}{}_{0.10})\ (\mathrm{Si}_{3.89}\mathrm{Al}_{0.09}\mathrm{Fe}^{+3}{}_{0.02})\,\mathrm{O}_{9.64}\,(\mathrm{OH})_{9.15}.$$

Analyses of picrolite that is composed essentially of chrysotile yield the following calculated formula compositions:

AR–52 (VT–15), analysis 11, table 1:

$$(\mathrm{Mg}_{6.6}\mathrm{Fe}^{+2}{}_{0.02}\mathrm{Fe}^{+3}{}_{0.2}\mathrm{Al}_{0.2})\ (\mathrm{Si}_{3.6}\mathrm{Al}_{0.4}\mathrm{Fe}^{+3}{}_{0.1})\,\mathrm{O}_{11.0}\,(\mathrm{OH})_{8.0}.$$

AR–64, analysis 54, table 1:

$$(\mathrm{Mg}_{5.58}\mathrm{Fe}^{+2}{}_{0.04}\mathrm{Fe}^{+3}{}_{0.05}\mathrm{Al}_{0.01})\ (\mathrm{Si}_{3.94}\mathrm{Al}_{0.06})\,\mathrm{O}_{10.0}\,(\mathrm{OH})_{7.4}.$$

AR–3x, analysis 10, table 1:

$$(\mathrm{Mg}_{5.85}\mathrm{Fe}^{+2}{}_{0.01}\mathrm{Fe}^{+3}{}_{0.05}(\mathrm{Si}_{3.94}\mathrm{Al}_{0.02}\mathrm{Fe}^{+3}{}_{0.04})\,\mathrm{O}_{9.8}\,(\mathrm{OH})_{8.1}.$$

Analysis 10 agrees reasonably closely with the ideal formula; analysis 54, somewhat less so. Sample AR–52, analysis 11, contained appreciable but undetermined amounts of brucite and magnesite, which accounts for the high content of Mg. Despite their limitations, the analyses do indicate a low content of Fe^{+2}, Fe^{+3}, and Al in the chrysotile in picrolite veins.

LIZARDITE

Lizardite is generally indistinguishable from antigorite in habit and optical properties. None was distinguished optically, but it was identified by X-ray and DTA studies in the zones of marginal alteration (George Faust, written commun., 1961). The platy serpentine in these zones of marginal alteration, which is the lizardite, has the general appearance and optical properties of antigorite.

Analyzed samples (table 1, analyses 2 [AR–13f, VT–2] and 52 [A–BM–57–3]) of rock from alteration zones bordering chrysotile veins consisted almost entirely of lizardite and minor amounts of chrysotile (identified on the basis of DTA and X-ray analysis), brucite, carbonate, and magnetite. Most of the magnetite was removed from sample AR–13f (VT–2) before analysis, leaving as significant contaminants only brucite and magnesite. Sample A–BM–57–3 contained microscopically determined amounts of brucite, magnesite, and magnetite. Thus, for each analysis, the formula composition of the serpentine mixture could be determined with considerable accuracy. The calculated formula compositions, therefore, represent rather closely the composition of the lizardite. They are:

AR–13f (VT–2):

$$(Mg_{5.82}Fe^{+2}{}_{0.02}Ni_{0.02}Fe^{+3}{}_{0.10})\,(Si_{3.86}Al_{0.08}Fe^{3+}{}_{0.06})\,O_{9.71}(AH)_{8.16}.$$

A–BM–57–3:

$$(Mg_{5.74}Fe^{+2}{}_{0.01}Ni_{0.03}Fe^{+3}{}_{0.21})\,(Si_{3.79}Al_{0.16}Fe^{+3}{}_{0.05})\,O_{10.04}(OH)_{7.90}.$$

SIX-LAYER ORTHOSERPENTINE

No six-layer orthoserpentine was positively identified in the ultramafic rocks of the Belvidere Mountain area. The nearly isotropic serpentine locally abundant in the dunite and massive serpentine and tentatively identified as possibly six-layer orthoserpentine has indices similar to those of antigorite, negligible birefringence, and no discernible structural habit. No information is available on the chemical composition.

CHLORITE

Chlorite is a common accessory mineral of the ultramafic igneous rocks and serpentinite. It generally constitutes less than 2 percent, and rarely as much as 5 percent of the rock. Most of the chlorite is associated with chromite and the larger grains of magnetite, forming complete or partial coronas which separate the grains from surrounding antigorite. In some of the schistose serpentinite, the chlorite occurs in isolated flakes or irregular patches of flakes adjacent to the spinellid grains, but not in distinctive coronas. In a few thin sections, chlorite is interlayered with antigorite in pseudomorphs after pyroxene.

All the chlorite has similar optical properties: $2V=0°–20°$, optic sign (+), sign of elongation (−), abnormal brown to pale-green and yellow-orange interference colors, $\beta=1.565–1.587$, birefringence very low. On the basis of these optical properties, the chlorite is inferred to range in formula composition from about

$$(Mg_{5.1}Fe^{+2}{}_{0.2}Al_{0.7})\ (Si_{3.3}Al_{0.7})O_{10}(OH)_2$$

to

$$(Mg_{4.2}Fe^{+2}{}_{0.8}Al_{1.0})\ (Si_{3.0}Al_{1.0})O_{10}(OH)_2.$$

(See Chidester, 1962, p. 44–46, fig. 14.)

BRUCITE

Brucite is a common constituent of the ultramafic rocks and associated veins. It is generally a minor component of the slightly serpentinized dunite and peridotite and locally a major component of the serpentinite, in which it occurs as disseminated flakes and irregular aggregates of flakes. It is locally intergrown sparsely with chrysotile in cross-fiber asbestos veins and is a major component of many veins of columnar picrolite. Brucite is an essential constituent of the marginal zones of alteration that border cross-fiber asbestos and cross-column picrolite veins and is commonly associated with magnetite and calcite in veinform aggregates.

The brucite ranges in habit from flaky to fibrous. Observed optical properties are: uniaxial, optic sign (+), sign of elongation (−), $\omega=1.566–1.573$, birefringence$\approx$0.11. These optical properties indicate a composition near that of the end member, $Mg(OH)_2$, having at most only a small content of Fe^{+2}.

CARBONATES

Magnesite, calcite, and dolomite are almost ubiquitous accessory constituents of the serpentinized ultramafic igneous rocks and of the several kinds of veins, and not uncommonly are major components. Generally only one species is present in a single hand specimen. Where more than one is present, the different species are generally distinguished by different habits or associations.

Magnesite, the most abundant carbonate mineral in the ultramafic rocks, is generally sparse or absent in the least serpentinized dunite and peridotite but constitutes as much as 20 percent of massive and schistose serpentinite. The magnesite occurs in the massive serpentinite and partly serpentinized dunite principally as disseminated grains about 1–2 mm across, less commonly as aggregates of small grains. In the schistose serpentinite, it commonly forms small irregular masses and veins and locally is segregated in thin lenticles parallel to the schistosity. Measured indices range from $\omega = 1.700$ to $\omega = 1.705$, indicating a composition near pure magnesite, $MgCO_3$.

Calcite is associated principally with veins, in which it varies widely in abundance and habit; it occurs only sparsely in the serpentinite, in the form of dispersed grains and small lenticles. In cross-fiber asbestos veins, calcite is locally intergrown in fibrous or feathery form with the chrysotile or forms irregular patches within the vein. In cross-column and massive picrolite veins, calcite forms irregular, commonly flamboyant, intergrowths with the chrysotile. In masses of long slip-fiber chrysotile asbestos, finely fibrous calcite—megascopically distinguishable from the chrysotile asbestos only by its brittleness—commonly forms as much as 50 percent or more of the asbestiform mass. Granular calcite is commonly associated with brucite or calcite, or both, in small veins in the dunite, peridotite, and massive serpentinite.

Calcite in all the associations shows only a small range in optical properties, from $\omega = 1.658$ to $\omega = 1.660$. These properties indicate a composition near that of pure calcite, $CaCO_3$.

Dolomite occurs sparsely in small veins in both the massive and the schistose serpentinite. Measured indices of $\omega = 1.684$ indicate the composition of a somewhat magnesian dolomite.

GRAPHITE

Graphite occurs sparsely to abundantly in a small proportion of the serpentinite. It is confined to a few thin coarsely flaky layers in the massive serpentinite, to coarsely fibrous picrolite veins in such layers, and to relatively minor, but appreciable, volumes of schistose serpentinite. The graphite is disseminated sparsely in irregular cloudy patches in the flakes and shreds of serpentine and is concentrated moderately to heavily at the margins of flakes of serpentine and along slip cleavage surfaces and tiny shear zones in the schistose serpentinite. Thin sections of graphitic massive serpentinite contain 1–19 percent graphite; thin sections of schistose serpentinite, 5–37 percent. Specimens of massive serpentinite that were analyzed chemically for graphite (analyses 25, 26, and 30, table 1) contained 1.4–14.4 percent.

All the graphite is very fine grained and may be partly amorphous. X-ray powder photographs of graphitic serpentinite show only the strongest lines of graphite (Fred A. Hildebrand, written commun., Oct. 13, 1955; Nov. 9, 1955). Some of it, however, particularly that concentrated along grain boundaries and in shear zones, shows flashes off crystal faces and is silvery gray in reflected light.

STEATITE, TALC-CARBONATE ROCK, AND CARBONATE-QUARTZ ROCK

Steatite, talc-carbonate rock, and carbonate-quartz rock are associated with the ultramafic bodies in distinctive structural and petrologic relations. The three rock types are completely intergradational, but the compositional range and the association of rock types is related to structural setting. Talc-carbonate rock and steatite form concentric shells at the margins of ultramafic bodies, and quartz is very sparse or absent in such bodies. Talc-carbonate rock and carbonate-quartz rock are associated within ultramafic bodies as irregular or lenticular masses whose distribution suggests control by faults or shear zones. In the Eden quarry ultramafic body, the masses labeled "Outc" near grid line 94,000 E. (pl. 1) are predominantly of variably quartzose talc-carbonate rock in which no zonal distribution pattern is discernible. Near the south end of the Corez Pond body, carbonate-quartz rock and talc-carbonate rock show a zonal relation; talc-carbonate rock is marginal to a central mass of carbonate-quartz rock. Relations with nearby serpentinite are not exposed, but talc-carbonate rock appears everywhere to intervene between serpentinite and carbonate-quartz rock.

Grains of chromite and magnetite are sparse relict igneous minerals in all three rock types. Table 4 contains selected modes of the end-member rocks and of some transitional types.

STEATITE

Steatite forms thin shells a few centimeters to a few meters thick at the outer margins of the small lenses of ultramafic rocks northeast of the Lowell quarry body, west of the Eden quarry body, and near the southeast corner of the Belvidere Mountain area (pl. 1). In the stripped area centered at 100,900 N., 97,200 E. (figs. 5, 11), along the northeast contact of the Lowell quarry body, thin seams of steatite that

TABLE 4.—*Selected modes of steatite, talc-carbonate rock, and carbonate-quartz rock*

[Tr, trace]

Specimen No.	Talc	Carbonate minerals	Quartz	Tremolite	Chlorite	Sphene	Magnetite	Sulfide minerals	Graphite
				Steatite					
A–BM–126	97	--	--	3	--	--	Tr	--	--
A–BM–53–3	99	--	--	--	1	--	--	--	--
T2–6–652	70	--	--	30	<1	--	--	--	--
T2–8–1204	96	--	--	--	--	--	<1	--	4
				Talc-carbonate rock					
AV–212	88	10	--	--	--	--	Tr	2	--
EQ–1–175	60	40	--	--	--	--	--	Tr	--
EQ–5–94	87	10	--	--	2	1	--	--	--
T–32–568	60	40	--	--	Tr	--	Tr	--	--
T–32–664	60	40	--	--	--	--	<1	--	--
				Carbonate-quartz rock					
A–BM–108	45	45	9	--	--	--	1	--	--
A–Ch–56–535c	1	70	29	--	Tr	--	--	--	--

swell locally into pods as much as 2 m thick extend along bedding surfaces for tens of meters beyond the ends of serpentinite lenses.

Steatite is relatively rare in the main ultramafic body. It is exposed at the contacts only near 102,000 N., 94,300 E., and 99,000 N., 92,000 E. (pl. 1). Several diamond-drill holes in the vicinity of these exposures, and at the south end of the Corez Pond body, penetrated as much as 1 m of steatite at the lower contact between the ultramafic body and the country rock.

The steatite is uniformly fine grained, commonly schistose but locally massive, and medium gray to light greenish gray on fresh surfaces. Weathered surfaces are pale buff to light brownish buff. In some places, steatite in the outer part of the zone contains streaks and knots of minerals and subtle variations in color and texture that impart a faint layering similar in pattern to that in the adjacent country rock.

The steatite consists of a fine-grained aggregate consisting of shreds and flakes of talc predominantly 0.01 to 0.1 mm long but locally as long as 0.3 mm. In the schistose steatite, narrow shear zones 0.001 to 0.01 mm wide, within which the particles are uniformly alined, are separated by zones 0.005 to 0.1 mm wide, in which the particles range from diversely oriented to uniformly parallel to the shear zones.

The steatite is composed essentially of talc, but chlorite and dusty magnetite are sparse accessory constituents throughout the rock. Magnesite occurs in the transition zone between steatite and talc-carbonate rock, and tremolite or abundant chlorite occurs in the zone transitional into tremolite rock or blackwall chlorite rock. Relict grains of the igneous minerals chromite and magnetite are distinctive of the inner and major part of the steatite zone. Relict grains of sphene, apatite, and, rarely, other minerals of the country rock are distinctive of a narrow outer part of the steatite zone, particularly where relict layering is preserved.

TALC-CARBONATE ROCK

Talc-carbonate rock constitutes the central core, or forms an envelope from a few centimeters to several tens of meters thick about a central core of serpentinite, in the small ultramafic lenses northeast of the Lowell quarry body (pls. 3 and 4*D*), west of the Eden quarry body, and in the southeast part of the map area (pl. 1). Talc-carbonate rock also forms large irregular masses in the Eden quarry body and the Corez Pond body.

The talc-carbonate rock consists of a fine-grained groundmass of talc throughout which are dispersed abundant single crystals and aggregates of magnesite. The porphyroblasts of magnesite range in size from less than 1 mm to as large as 3 cm and impart a faint mottling to the rock, which varies from medium gray to light greenish gray. Weathered surfaces are rusty reddish brown. The rock is predominantly massive but locally has a crude spaced schistosity. Some of the massive talc-carbonate rock has faint layering resembling that of the dunite and massive serpentinite.

The composition of the talc-carbonate rock varies widely, on a small scale throughout the rock body and on a large scale in transition zones into serpentinite, carbonate-quartz rock, and steatite. Outside the transition zones, representative samples contain about 60 percent talc and 40 percent magnesite.

Relict igneous minerals and dusty magnetite commonly constitute less than 1 percent of the rock, and chlorite is locally present in small amounts.

CARBONATE-QUARTZ ROCK

Carbonate-quartz rock forms an appreciable mass only at the south end of the Corez Pond body, near the center of a larger mass of talc-carbonate rock. Quartz occurs erratically in small to moderate amounts in the irregular bodies of talc-carbonate rock in the Eden quarry body and in the small pod of ultramafic rock near 91,500 N., 101,000 E. (pl. 1).

The carbonate-quartz rock is medium to light gray or greenish gray and generally retains the gross appearance of the serpentinite from which it was derived. Quartz forms a mosaic of anhedral grains 0.02 to 0.2 mm across, throughout which anhedral grains, rhombs, and aggregates of magnesite 0.1 to 2 mm across are irregularly interspersed. In the typical specimen, magnesite constitutes about 70 percent of the rock and quartz, 30 percent. Grains of chromite and magnetite occur as sparse relict igneous minerals. Small amounts of dusty magnetite and shreds of talc are common accessories.

MINERALOGY AND PARAGENESIS

Steatite, talc-carbonate rock, and carbonate-quartz rock are composed chiefly of talc, magnesite, and quartz that have characteristic felted and interlocking mosaic textures. Relict minerals of the preexisting rocks are volumetrically negligible but are diagnostic of the metamorphic origin of the steatite, talc-carbonate rock, and carbonate-quartz rock. Granular magnetite, chromite, pyrite, and sulfarsenide are relict minerals inherited from the igneous rocks. Dustlike particles of magnetite, chlorite associated with magnetite grains, and serpentine are relict minerals of the serpentinite. Chlorite and sphene in the outer part of the steatite zone are relicts of incomplete alteration of the blackwall chlorite rock. Graphite is a relict mineral of the graphitic schist, amphibolite, and serpentinite.

TALC

Talc occurs in shreds and flakes generally less than 0.5 mm long. The optical properties are generally uniform. The optic axial angle, $2V$, is small to moderate. The β-index of refraction varies through a small range, from about 1.588 to 1.591. Maximum interference colors indicate a birefringence of about 0.045.

Chemical analyses of talc similar in optical properties and geologic occurrence elsewhere in Vermont indicate a composition of about

$$(Mg_{5.6}Fe^{+2}_{0.3}Al_{0.1})(Si_{7.9}Al_{0.1})O_{20}(OH)_4$$

(see Chidester, 1962, p. 79–80). The talc in the Belvidere Mountain area is inferred to have about the same composition.

CARBONATE

Only magnesite was noted in the talc-carbonate rock and carbonate-quartz rock, but calcite and dolomite in small amounts may possibly exist as relict minerals of the serpentinite. The magnesite occurs as anhedral to rhombic grains as much as 1 mm across and aggregates as large as 3 cm.

The ω-index of refraction ranges from 1.700 to 1.720. Analyzed carbonate minerals that have a similar range in index from ultramafic rocks elsewhere in Vermont range in composition from about pure magnesite, $MgCO_3$, to about

$$(Ca_{0.01}Mg_{0.86}Fe^{+2}_{0.13})CO_3.$$

QUARTZ

Quartz in the carbonate-quartz rock is uniformly fine grained and uniformly shows normal extinction. In rock containing little or no talc, inclusions are relatively rare except for magnesite grains enclosed in aggregates of quartz grains. In rock containing moderate to abundant talc, quartz grains and aggregates commonly enclose numerous shreds of talc as well.

OTHER MINERALS

Relict grains of chromite, magnetite, and sulfides and sulfarsenides differ little from those in the dunite, peridotite, and serpentinite (see section on "Chromite and magnetite" under "Mineralogy and paragenesis" in the section on "Igneous rocks, serpentinite, and veins") and are inferred to have changed little or not at all in composition. Locally, grains of magnetite and chromite appear to have been replaced slightly by carbonate.

Chlorite is a sparse relict of serpentinite or occurs in steatite transitional into blackwall. The optical properties and inferred composition are like those of chlorite in serpentinite (see section on "Chlorite" under "Mineralogy and paragenesis" in the section on "Igneous rocks, serpentinite, and veins") or in the blackwall (see section on "Chlorite" under "Mineralogy and paragenesis" in the section on "Contact rocks").

Graphite persists as a relict mineral only rarely in thin shear zones and along schistosity surfaces. Sphene is a rare relict mineral inherited from schist or amphibolite.

CONTACT ROCKS

Each of the three main associations of contact rocks—rodingite and serpentinite-chlorite rock, steatite and blackwall chlorite rock, and tremolite rock and chlorite rock—conforms generally to broad regular patterns of interrelations and mineral assemblages but varies considerably in detailed relations, in the proportions of major minerals, and in the identity and proportions of accessory minerals (see beginning discussion in section on "Ultramafic and associated rocks").

As the ultramafic bodies are generally conformable with bedding schistosity in the enclosing country rocks, the boundaries of the contact rocks also generally conform grossly with the bedding schistosity. Where the ultramafic bodies crosscut the bedding and where the configuration of the contact rocks is controlled by joints and shear zones, the boundaries of the contact rocks are grossly crosscutting; where the boundaries of a contact rock are irregular and the amplitude of the irregularity is appreciably greater than the thickness of bedding in the adjacent country rock, the contacts are crosscutting in detail. At such places, wherever layering can be traced from one rock type to another, there is no discernible difference in thickness of layers in the different rock types, nor is there discontinuity in the bedding surfaces at the contacts between rock types.

Table 5 contains selected representative modes of contact rocks. Table 6 contains modes of selected suites of specimens across contacts between ultramafic rocks and several kinds of country rock; it is intended to illustrate the variable and complex compositional relations of contact rocks.

THE RODINGITE AND SERPENTINE-CHLORITE ROCK ASSOCIATION

Rocks of the rodingite and serpentine-chlorite rock association border serpentinite at contacts with the country rocks and at the margins of inclusions throughout almost the entire Lowell quarry body, in much of the Eden quarry body, and in probably a smaller proportion of the Corez Pond body. The association occurs both where the adjacent country rock is schist and gneiss, such as at the northeast margin of the Lowell quarry body, and where it is amphibolite, which borders much of the main ultramafic body elsewhere. The serpentine-chlorite rock zone is immediately adjacent to the serpentinite; the rodingite zone is adjacent to the unaltered country rock.

TABLE 5.—*Selected modes of contact rocks*

Specimen No.	Diopside	Vesuvianite	Garnet	Epidote or clinozoisite	Prehnite	Serpentine	Chlorite	Tremolite	Talc	Biotite	Carbonate minerals	Apatite	Rutile	Sphene	Ilmenite	Magnetite	Sulfide minerals	Graphite
A. Rodingite																		
TPT–Vt–10	29.2	--	21.6	25.0	--	--	19.6	--	--	--	--	--	--	4.3	0.3	--	--	--
AR–18	--	50	30	--	--	--	20	--	--	--	Tr	Tr	--	--	--	--	--	--
AR–20	31.0	4.6	3.3	49.5	--	--	7.1	--	--	--	0.3	--	--	4.2	--	--	--	--
AV–5a	10	4	70	--	--	--	15	--	--	--	--	--	--	1	--	--	--	--
AV–5c	40	Tr	33	24	--	--		--	--	--	--	--	--	2	--	--	1	--
AV–71	65	--	30	5	--	--	--	--	--	--	--	--	--	Tr	--	--	Tr	--
AV–90	89.4	--	3.3	.3	--	--	5.4	--	--	1.1	--	--	--	Tr	--	--	.4	--
AV–131b	41.7	--	10.4	32.4	2.0	--	--	--	--	8.2	5.1	--	--	.2	--	--	--	--
AV–242	3.8	31.2	45.4	--	--	--	19.6	--	--	--	--	--	--	--	--	--	--	--
VAE–3	35	40	25	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--
B. Serpentine-chlorite rock																		
TPT–Vt–11	--	--	--	--	--	23.7	67.4	--	--	--	--	--	--	--	8.9	--	--	--
AR–20	--	--	--	--	--	51.1	45.5	--	--	--	--	--	--	--	3.4	--	--	--
AV–5a	--	--	--	--	--	60	30	--	--	--	--	2	--	--	8	--	--	--
AV–90	--	--	--	--	--	48.8	50.3	--	--	--	--	--	--	--	.9	--	--	--
AV–238	4.0	--	--	--	94.1	.3	--	--	--	--	--	--	--	--	1.6	--	--	--
AV–241	1.1	--	--	--	90.3	3.9	--	--	--	--	--	--	--	--	4.7	--	--	--
C. Blackwall chlorite rock and tremolite-chlorite rock																		
A–BM–82	--	--	--	--	--	--	97	--	--	--	--	--	1	--	--	--	--	2
A–BM–109	--	--	--	--	--	--	97	--	--	--	--	--	3	--	--	--	--	--
A–BM–53–41	--	--	--	3	--	--	88	--	--	--	--	--	--	9	--	--	--	--
EQ–1–740	--	--	--	1	--	--	89	--	--	--	--	Tr	--	10	--	--	--	--
T–67–1004c	--	--	--	<1	--	--	97	--	--	--	--	--	--	1	--	2	--	--
AV–206	--	--	--	--	--	--	73.7	7.4	--	8.6	--	.4	--	9.9	--	--	--	--
A–BM–140	--	--	--	--	--	--	25	70	5	--	--	--	--	--	--	--	--	--
A–BM–53–57	--	--	--	--	--	--	20	30	--	--	--	--	--	--	--	--	--	50
AV–204	--	--	--	--	--	--	11.9	88.1	--	--	--	--	--	--	--	--	--	--
AV–205	--	--	--	--	--	--	6.3	93.6	--	--	--	--	--	--	--	--	0.1	--
AV–203	--	--	--	--	--	--	5.0	94.8	--	--	--	--	--	--	--	--	.2	--
A–BM–130	--	--	--	--	--	--	1	99	--	--	--	--	--	--	--	--	--	--

Both the rodingite and the serpentine-chlorite rock show variations related to structure or to lithology of the adjacent country rock or to both. The rodingite is uniformly about a meter thick where the adjacent country rock is schist or gneiss in conformable contact with the ultramafic body. Where the adjacent country rock at conformable contacts is amphibolite, the rodingite is commonly only a few centimeters or decimeters thick. Serpentine-chlorite rock, on the other hand, shows no significant variations in thickness related to lithology, ranging generally from 2 to 8 cm thick at conformable contacts. The rodingite is conspicuously thicker—in several places as much as 3 m—in both amphibolite and schist or gneiss where the contact of the ultramafite is strongly crosscutting and where septa and inclusions of country rock in the ultramafic body are highly sheared and fractured. The serpentine-chlorite rock varies less markedly in thickness but is noticeably thicker at some structural irregularities, particularly where thin tongues of serpentinite penetrate along bedding planes of country rock at crosscutting contacts; such composite zones are as much as 1 m thick. Along joints that intersect the ultramafic contact, the serpentine-chlorite rock zone tapers from 5 cm thick to zero. In addition to such variations in thickness, the rodingite shows conspicuous variations in texture and grain size related to joints and shear zones.

Contacts reflect the intrusive or replacement relations of the associated rocks. The contact between serpentine-chlorite rock and serpentinite is generally regular and sharp but is commonly difficult to pinpoint because of the similar appearances of the two rock types. In many places, the contact is a shear surface; elsewhere, the junction of the two rocks is tight. The contact between serpentine-chlorite rock and rodingite is very irregular and is gradational over a very short distance—commonly less than 1 mm (see fig. 7). The contact between rodingite and country rock is variable. In most places, the gradation takes place over a distance of 2 cm or less. In some places, particularly in the coarse amphibolite, the rodingite grades irregularly into unaltered country rock over a distance of several centimeters to several decimeters; not uncommonly, such gradational zones contain an unusually large proportion of chlorite.

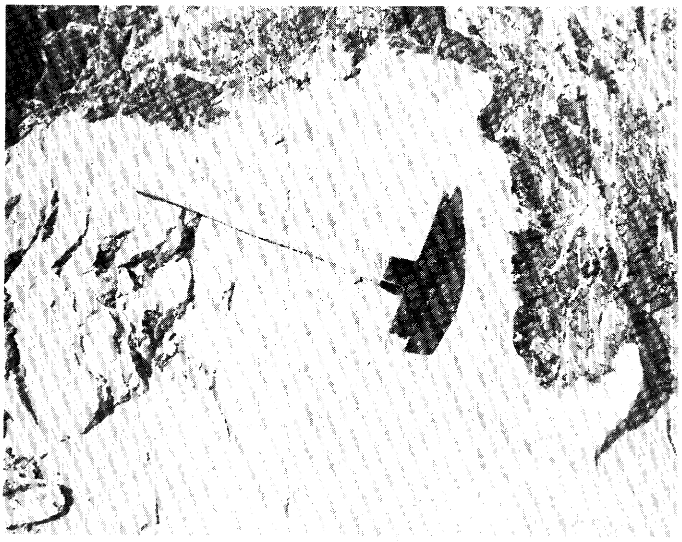

FIGURE 7.—Photograph of contact between serpentine-chlorite rock and rodingite, northeast side of Lowell quarry. The rodingite, under the hammer, is pale buff, the serpentine-chlorite rock dark bluish green.

In a few places, clear-cut zonal relations of contact rocks are not apparent. At the contact between serpentinite and coarse amphibolite at the west edge of the Eden quarry, no well-defined serpentine-chlorite rock and rodingite zones border the serpentinite. (fig. 4). Rather, a single zone, variable in width, consists of variable proportions of diopside, chlorite, and relict minerals of the amphibolite. Along the southwestern margin of the Lowell quarry body, serpentine-chlorite rock projects in narrow fingers beyond the rodingite zone along many of the numerous joints and small faults that intersect the contact of the ultramafic body; in a few places, the serpentine-chlorite rock is bordered at the outer margin by amphibolite, without an intervening zone of rodingite.

Irregular veins of coarse chlorite, calcite, and magnetite are associated with the rodingite and serpentine-chlorite rock in a few places at the eastern edge of the Eden quarry body and at the southwestern end of the Lowell quarry body. Most such veins are in the rodingite and serpentine-chlorite rock but some penetrate into the serpentinite for a few centimeters. The veins appear to be largely fracture controlled, but in detail the boundaries are commonly slightly irregular and locally very irregular.

RODINGITE

Rodingite occurs in two distinctive textural varieties, each in a distinctive structural setting. That in structurally simple settings is uniformly fine grained and of dense flinty appearance. That in and immediately adjacent to highly fractured and sheared zones is coarser grained and more varied in appearance. At the margins of the highly fractured zones, the coarse rodingite grades irregularly into

TABLE 6.—*Modes of selected suites of specimens*

[Tr,

Specimen No.		Distance (cm) into the ultramafic body (−), or into the contact rock and country rock (+), measured normal to the contact	Serpentine	Chromite	Magnetite	Talc	Tremolite	Chlorite
A. Contact of the rodingite and serpentine-chlorite rock type in graphitic schist of the Hazens Notch								
AV-5a-1	Serpentine-chlorite rock	+6	60	---	8	---	---	30
AV-5a-2	Rodingite	+7	---	---	---	---	---	15
AV-5b	do	+9	---	---	---	---	---	3
AV-5c	do	+12	---	---	---	---	---	---
AV-5d	do	+21	---	---	---	---	---	Tr
AV-5e	do	+28	---	---	---	---	---	3
AV-6	do	+152	---	---	---	---	---	---
B. Contact of the rodingite and serpentine-chlorite rock type in graphitic schist of the Hazens Notch formation								
AV-131a	Rodingite	+9	---	---	---	---	---	---
AV-131b	do	+91	---	---	---	---	---	---
AV-131c	Transition rodingite-schist	+167	---	---	3.7	---	---	---
AV-131d	Hornblendic graphitic schist	+229	---	---	---	---	---	4.2
C. Contact of rodingite and serpentine-chlorite rock type in coarse amphibolite of the Belvidere Mountain								
A-BM-117a	Rodingite	+3	---	---	Tr	---	---	8
A-BM-117b	do	+9	---	---	---	---	---	95
A-BM-117c	do	+24	---	---	---	---	---	73
A-BM-117d	do	+30	---	---	---	---	---	22
A-BM-118	Amphibolite	+122	---	---	---	---	---	6
D. Contact of the rodingite and serpentine-chlorite rock type in fine amphibolite of the Belvidere Mountain [Serpentine-chlorite rock containing irregular lenses								
AV-243	Serpentinite	−1	93.8	---	6.0	---	---	---
AV-242	Rodingite	+6	---	---	---	---	---	19.6
AV-241	Serpentine-chlorite rock	+15	90.3	---	4.7	---	---	3.9
AV-240	Serpentinized rodingite	+30	88.3	---	.9	---	---	.3
AV-239	Rodingite	+40	---	---	7.2	---	---	8.8
AV-238	Serpentinized rodingite	+49	94.1	---	1.6	---	---	---
AV-237A	Rodingite	+63	---	---	10	---	---	33
AV-237B	do	+64	---	---	Tr	---	---	24
AV-237C	do	+66	---	---	Tr	---	---	6
E. Contact intermediate between the steatite and blackwall type and the tremolite and chlorite rock type [Drill hole inclined about 70° to bedding schistosity. Last set of								
T2-6-584	Serpentinite	-	98	---	2	---	Tr	---
T2-6-652.5	Tremolitic steatite	-	---	Tr	---	70	30	Tr
T2-6-654	Tremolite-chlorite rock	Contact	---	---	---	Tr	60	40
T2-6-655	Tremolite rock	-	---	---	---	---	98	2
T2-6-656	Schist	-	---	---	Tr	---	---	8
T2-6-658	do	-	---	---	---	---	---	---
T2-6-692	do	-	---	---	1	---	---	13
F. Contact intermediate between the steatite and blackwall type and the tremolite and chlorite rock type [Drill hole inclined about 60° to bedding schistosity. Last set of								
T2-8-1195	Serpentinite	-	83	---	2	---	---	---
T2-8-1202	Graphitic serpentinite	-	83	1	6	---	---	---
T2-8-1204	Graphitic steatite	-	---	---	Tr	96	---	---
T2-8-1205	Graphitic tremolite rock	Contact	---	---	---	---	98	---
T2-8-1206	Graphitic schist	-	---	---	---	---	2	---
G. Contact of the tremolite and chlorite rock type in coarse amphibolite of the Belvidere								
AV-202	Tremolite rock	Contact	---	---	Tr	---	100	Tr
AV-203	Tremolite-chlorite rock	+21	---	---	---	---	94.8	5.0
AV-204	do	+40	---	---	---	---	88.1	11.9
AV-205	do	+46	---	---	---	---	93.6	6.3
AV-206	Chlorite-tremolite rock	+82	---	---	---	---	7.4	73.7
AV-207	Amphibolite	+326	---	---	---	---	---	8.9
AV-208	do	+570	---	---	---	---	---	7.9
H. Contact of the tremolite and chlorite rock type in coarse amphibolite of the								
A-BM-53-45	Serpentinite	−122	96	---	3	---	---	Tr
A-BM-53-44	do	−24	97	---	3	---	---	---
A-BM-53-43	do	−3	100	---	Tr	---	---	---
A-BM-53-42	Tremolite rock	+18	---	---	---	---	95	5
A-BM-53-41	Chlorite rock	+46	---	---	---	---	---	88
A-BM-53-40	Amphibolite	+152	---	---	---	---	---	4
A-BM-53-39	do	+244	---	---	---	---	---	2

---

[1] Veins.

across the contacts of ultramafic bodies
trace]

Diopside	Vesuvianite	Garnet	Prehnite	Carbonate minerals	Epidote/ clinozoisite	Hornblende	Plagioclase	Quartz	Muscovite	Biotite	Apatite	Sphene	Sulfide minerals	Graphite
formation ($€hc_2$) near the center of the northeast contact of the Lowell quarry body—Contact regular														
---	---	---	---	---	---	---	---	---	---	---	2	---	---	---
10	4	70	---	---	---	---	---	---	---	---	---	1	---	---
55	---	10	---	---	30	---	---	---	---	---	---	2	---	---
40	Tr	33	---	---	24	---	---	---	---	---	---	2	1	---
40	---	Tr	---	---	53	5	---	---	---	---	Tr	2	---	---
30	---	2	1	---	60	2	---	---	---	---	---	2	---	---
25	---	65	---	---	3	---	---	---	---	---	---	---	---	7
($€hc_2$) near the center of the northeast contact of the Lowell quarry body—Contact very irregular														
39.6	---	5.0	(1)	---	53.6	---	---	---	---	---	---	---	---	---
39.7	---	10.4	2.0	5.1	32.4	---	---	---	---	8.2	---	0.2	---	---
---	---	---	---	---	58.7	20.0	---	14.1	---	---	---	3.5	---	---
---	---	---	---	Tr	6.1	2.8	1.2	16.7	36.0	---	---	---	0.4	32.5
Formation (€bc) at the upper contact of the Eden quarry body, west side of the Eden quarry														
92	---	---	---	---	---	---	---	---	---	---	---	---	---	---
3	---	---	---	---	---	---	---	---	---	---	---	2	---	---
10	---	---	---	---	15	---	---	---	---	---	---	2	---	---
17	---	---	---	---	33	26	---	---	---	---	---	2	---	---
---	---	---	---	---	34	36	22	---	---	---	---	2	---	---
Formation (€bf)) at the southeast end of the southwest contact of the Lowell quarry body of rodingite grades sharply into massive rodingite at +60 cm]														
---	---	---	---	0.2	---	---	---	---	---	---	---	---	---	---
3.8	31.2	45.4	---	---	---	---	---	---	---	---	---	---	---	---
.6	---	---	---	---	0.5	---	---	---	---	---	---	Tr	---	---
6.0	---	---	---	---	3.5	---	---	---	---	---	---	1.0	---	---
78.0	---	---	---	---	4.0	---	---	---	---	---	---	2.0	---	---
2.0	---	---	---	.3	2.0	---	---	---	---	---	---	Tr	---	---
---	20	46	---	1	---	---	---	---	---	---	Tr	Tr	---	---
38	---	24	---	4	1.0	---	---	---	---	---	---	Tr	---	---
48	---	---	---	2	42	---	---	---	---	---	---	2	---	---
in graphitic schist of the Hazens Notch Formation ($€hc_2$) beneath the Eden quarry body, drill-hole T2–6 digits in specimen number indicates footage from collar of drill hole]														
---	---	---	---	Tr	---	---	---	---	---	---	---	---	---	---
---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
---	---	---	---	---	53	---	---	---	15	23	---	1	---	---
---	---	---	---	---	1	---	30	10	12	38	---	3	---	5
---	---	---	---	---	4	---	9	25	45	---	---	3	---	---
in graphitic schist of the Hazens Notch Formation ($€hc_2$) beneath the Eden quarry body,drill-hole T2–8 digits in specimen number indicates footage from collar of drill hole]														
---	---	---	---	15	---	---	---	---	---	---	---	---	---	---
---	---	---	---	---	---	---	---	---	---	---	---	---	---	10
---	---	---	---	---	---	---	---	---	---	---	---	---	---	4
---	---	---	---	---	---	---	---	---	---	---	---	---	---	2
---	---	Tr	---	---	10	---	54	---	10	20	---	1	---	3
Mountain Formation (€bc) near the southeast end of the northeast contact of the Lowell quarry body														
---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
---	---	---	---	---	---	---	---	---	---	---	---	---	0.2	---
---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
---	---	---	---	---	---	---	---	---	---	---	---	---	.1	---
---	---	---	---	8.6	---		---	---	---	---	0.4	9.9	---	---
---	---	---	---	---	3.2	76.5	---	---	---	---	.1	4.4	---	---
---	---	---	---	---	6.4	70.4	8.2	---	---	---	---	6.8	.3	---
Belvidere Mountain Formation (€bc) in septum at northwest end of the Lowell quarry body														
---	---	---	---	1	---	---	---	---	---	---	---	---	---	---
---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
---	---	---	---	---	3	---	---	---	---	---	---	9	---	---
---	---	---	---	---	22	64	5	---	---	---	---	5	---	---
---	---	---	---	---	20	58	16	---	---	---	---	4	---	---

the dense flinty rodingite, commonly within a few centimeters.

FINE-GRAINED RODINGITE

The fine-grained, dense, flinty rodingite is light bluish gray, light greenish gray, or pale buff. Layering, on the scale of that in the bordering schist, gneiss, or amphibolite, is typical of most of the rock and is marked by faint differences in color that are accentuated by weathering. Though generally rather uniform in megascopic appearance, the rock shows considerable variation in mineralogy and in microfabric.

Significant variations in mineralogy are predominantly on a scale larger than that which can be seen in a hand specimen, but some thin sections show significant variations within the area of the section, and a few show different assemblages of minerals in adjacent layers. The principal assemblages observed are:

diopside-clinozoisite
diopside-clinozoisite-garnet
diopside-clinozoisite-garnet-chlorite
diopside-clinozoisite-garnet-chlorite-vesuvianite
diopside-clinozoisite-garnet-vesuvianite
diopside-clinozoisite-garnet-prehnite
diopside-clinozoisite-chlorite
diopside-clinozoisite-prehnite
diopside-garnet-vesuvianite
diopsite-garnet-vesuvianite-chlorite
diopside-chlorite-(magnetite)
clinozoisite-garnet-chlorite
clinozoisite-chlorite
garnet-chlorite-vesuvianite-(magnetite)

In addition to the major minerals, sphene is an almost ubiquitous accessory mineral, and apatite, rutile, magnetite-ilmenite, and calcite are common accessories. Quartz, hornblende, sulfides, graphite, biotite, and plagioclase are common and locally abundant relict minerals of the country rock; microcline is very rare.

The variations in mineralogy conform generally to two patterns. One type of variation is related to position in the rodingite zone and to local structural controls. Diopside and clinozoisite predominate at the inner margin, clinozoisite and chlorite at the outer margin; garnet and vesuvianite, though irregular in distribution and absent from much of the rodingite, are most abundant at the inner margin of the zone and adjacent to coarse rodingite associated with fractures and shear zones. The second type of variation is unrelated to position in the zone, but reflects original differences in composition of the country rock from which the rodingite was derived. This type of variation is exemplified by the different assemblages of minerals in different relict beds in a single specimen.

Textures of the fine-grained rodingite may be grouped into only a few basic types, which in slight variations and different combinations contribute considerable textural diversity to the rock. Much of the rock is a fine-grained mixture of constituents resolvable only under high magnification. Some consists of an equigranular aggregate of subequant serrated grains that are readily resolvable under the microscope. A third type consists of small subequant grains of one or more mineral species, interstitial to which other mineral species form irregular patches, shreds, and wispy flakes. In a few places, groundmass material having one of the above basic textures contains conspicuously larger, diversely oriented, tabular crystals of diopside. Some of the rodingite derived from amphibolite consists of a fine-grained groundmass throughout which are scattered larger, rounded grains 0.1 to 0.5 mm across. Most of the rounded grains are zoned clinozoisite; a few are diopside. All these textural varieties locally contain small veinform or irregular masses of distinctly coarser grained minerals, some of them flamboyant in habit.

For the most part, layering in the rodingite is parallel to bedding, and where inequidimensional minerals are present, they are commonly alined subparallel to the layering. Not uncommonly, several of the basic textures are shown by different layers in a single specimen. In a few places where bedding is tightly folded, quartz is concentrated in the troughs and crests of tiny folds, and inequidimensional grains of dark minerals arranged tangentially to layering are concentrated along the limbs; the resulting layering is at a large angle to bedding.

COARSE-GRAINED RODINGITE

The coarse-grained rodingite is variable in color and texture and is generally massive. It ranges in color from light gray to various shades of green and brown, or is mottled in those shades. In texture, it ranges from medium fine-grained granular aggregates of anhedral and subhedral grains to very coarse aggregates of anhedral and subhedral crystals as much as 8 cm in maximum dimension, many of which show excellent crystal forms. Some joint surfaces are covered with drusy aggregates of one or more minerals. In several places, large crystals form spectacular "pockets" from which interstitial calcite has been dissolved to give a vuggy appearance. Such pockets are the source of many prized mineral specimens.

The coarse rodingite contains the same minerals and has a variety of mineral assemblages similar to those of the fine-grained rodingite, but the variations are more irregular and on a larger scale, so that some masses of hand-specimen size consist almost entirely of only two or three minerals. In addition, calcite is an almost ubiquitous major constituent, and zoisite is locally abundant. The surfaces of a few open joints are spotted with botryoidal masses or radial aggregates of calcite and siderite. The principal assemblages observed in rock masses of hand specimen size are:

diopside-vesuvianite-garnet-calcite
diopside-vesuvianite-garnet-clinozoisite-calcite
diopside-garnet-clinozoisite-calcite
diopside-clinozoisite-calcite
diopside-calcite
garnet-clinozoisite-calcite
garnet-calcite
clinozoisite-zoisite.

SERPENTINE-CHLORITE ROCK

The serpentine-chlorite rock is fine grained, massive, dark greenish gray to greenish black, and breaks with a hackly to conchoidal fracture. Though uniform in megascopic appearance, the serpentine-chlorite rock varies considerably in microscopic texture, and has a consistent pattern of variation in mineralogy.

Serpentine and chlorite are the sole essential minerals of the rock, but magnetite and ilmenite are ubiquitous accessory minerals and are commonly major components. The proportions of serpentine and chlorite are variable, and the variations are somewhat erratic; but, in general, serpentine predominates over chlorite at the inner margin of the zone, next to the serpentinite, and decreases in abundance outward toward the rodingite zone. Near the contact with rodingite, chlorite commonly predominates over serpentine. Relict minerals from rodingite are common accessories in the zone of transition from serpentine-chlorite rock to rodingite, and the titanium minerals show consistent paragenetic relations. At sharp contacts, favorably situated single grains consist of sphene on the rodingite side and ilmenite on the serpentine-chlorite rock side.

The textural pattern of the serpentine-chlorite rock ranges from a felted aggregate of interlocking flakes, shreds, and blades of chlorite and serpentine to a mosaic of irregularly polygonal and elliptical patches of nearly isotropic serpentine containing disseminated flakes and blades of chlorite and serpentine. In some places, the elliptical and polygonal patches of serpentine contain almost no chlorite, but the interstitial areas are predominantly chlorite. In the transition zone from rodingite, the interstitial areas of chlorite commonly contain remnant grains of diopside. Anhedral grains of magnetite and ilmenite are scattered throughout the rock.

CHLORITE-CALCITE-MAGNETITE VEINS

Chlorite-calcite-magnetite veins vary from tabular to irregular in shape, from a millimeter to a decimeter thick, and from a decimeter to more than 3 m long. The contacts of the veins with the enclosing rocks are sharp but are commonly irregular in detail. Some of the veins are distinctly zoned, chlorite or chlorite and magnetite predominating at the margins of the veins and calcite predominating in the central part. In such zoned veins, the chlorite flakes and books at the outer margins are commonly alined normal to the contacts of the veins. Elsewhere, the distribution of the minerals is variable, and their orientation apparently random.

The minerals of the veins are coarse. Chlorite and magnetite form spectacular euhedral to subhedral crystals. Calcite is mainly in anhedral aggregates interstitial to the chlorite and magnetite. Small knots of epidote are scattered sparsely and unevenly through the veins in a few places.

THE STEATITE AND BLACKWALL CHLORITE ROCK ASSOCIATION

Rocks of the steatite and blackwall chlorite rock association border all the ultramafic bodies where they contain talcose rocks at the margins. These sites include most of the small lenses of ultramafic rocks (see section on "Steatite, talc-carbonate rock, and carbonate-quartz rock" under "Ultramafic and associated rocks"), the central parts of the northeastern and the western contacts of the Eden quarry body, and probably much of the southern end of the Corez Pond body. The zones are generally most conspicuous and most fully developed in the schist and more obscure and narrower in the amphibolite.

All the rocks of the association are intergradational, and they have interdependent distribution relations. The steatite zone varies in width from a centimeter to as much as several decimeters; the blackwall is typically 10 cm wide where the steatite zone is 30 cm or more but is appreciably thinner elsewhere. A narrow zone of tremolite rock, commonly about 5 cm wide, intervenes between the steatite and blackwall in many places but is absent in the less mafic varieties of schist. Where the adjacent country rock is albitic schist or gneiss, a zone as much as 8 cm wide at the outer margin of the blackwall contains a relatively high proportion of albite porphyroblasts (the albite porphyroblast zone).

The tremolite rock is grayish green and composed of interlaced to radial aggregates of tremolite needles as much as 3 cm long, intermixed with variable proportions of talc and chlorite. Sphene, epidote, and magnetite are sparse and rare accessories.

Blackwall is typically dark greenish gray and fine grained. It consists essentially of chlorite but commonly contains accessory sphene, ilmenite, rutile, and apatite; rarely, relict grains of quartz and albite occur. The blackwall commonly has poor to good schistosity. In thin section, the schistosity is seen to result from the partial to almost complete parallel alinement of the chlorite flakes (continuous schistosity), or from the presence of closely spaced tiny shear zones within which chlorite flakes are alined and between which the chlorite flakes range from alined to diversely oriented (spaced schistosity). In some places, both types of schistosity are present and are parallel.

Albite porphyroblast rock is variegated in grayish green and light gray, the overall appearance depending chiefly on the proportion of albite porphyroblasts. Except for its more visible content of albite, the rock ranges in character and mineralogy from albitic schist to blackwall, with both of which it intergrades.

Muscovite-albite-quartz-(chlorite) rock forms irregular masses near the western contact of the main ultramafic body at 99,800 N., 91,600 E. (pl. 1). The muscovite is compact and fine grained and is irregularly intermixed with mosaic aggregates of albite. Quartz and chlorite occur in moderate to small amounts. The rock replaces adjacent amphibolite, but its relations to nearby steatite and talc-carbonate rock are hidden by a covered interval of about 1 m. This muscovite-albite rock may occupy the albite porphyroblast zone and may be simply an unusual and extraordinarily large variant mass.

THE TREMOLITE ROCK AND CHLORITE ROCK ASSOCIATION

Contact rocks of the tremolite and chlorite rock association occur in only a few places: along the northeast contact of the Lowell quarry body near 99,910 N., 98,270 E., and at the margins of the small serpentinite masses in the amphibolite tongue between the Lowell quarry and Eden quarry bodies, near 100,500 N., 96,600 E. (pl. 3). In all places, the country rock is amphibolite.

Tremolite rock is medium gray to pale green, fine grained, and massive. A very irregular and inconspicuous spaced schistosity is locally discernible. Needles of tremolite are rarely large enough to be distinguished readily without a hand lens. Tremolite is the sole essential constituent; chlorite is common and is increasingly abundant away from the ultramafic body; magnetite and pyrite occur sparsely and erratically.

Chlorite rock is dark greenish gray and fine grained. A crude spaced cleavage is commonly sufficently well developed that the rock breaks into irregular slabs. In addition to chlorite, tremolite is a common constituent; sphene, carbonate, apatite, and other relict minerals from the amphibolite occur unevenly in the chlorite rock and are abundant in the transition into amphibolite.

The tremolite and chlorite contact rocks border serpentinite without intervening zones of talc-carbonate rock and steatite. Tremolite rock adjacent to the serpentinite is transitional by gradual increase in content of chlorite and decrease of tremolite into chlorite rock adjacent to the country rock. The contact between serpentinite and tremolite rock is sharp. The width of the tremolite rock and chlorite rock zones, and of the transitional zone between, ranges from a few centimeters to a few decimeters and varies with the size of the associated ultramafic body. The transition from chlorite rock to amphibolite is irregular and variable, but generally takes place within a few centimeters.

MINERALOGY AND PARAGENESIS

The minerals of the contact rocks differ widely in grain size, textural relations, and variability of composition, but all conform to a general pattern of paragenetic relations. Thus, minerals of the serpentine-chlorite rock replace minerals of the rodingite and locally those of the amphibolite; minerals of the rodingite replace those of the schist, gneiss, and amphibolite. Similarly, minerals of the steatite replace minerals of the blackwall chlorite rock, and minerals of the blackwall replace those of the schist, gneiss, and amphibolite. Furthermore, in the rodingite, some of the minerals have an orderly sequential relation: garnet and vesuvianite replace diopside and epidote, and chlorite commonly replaces all four of them. Garnet and vesuvianite commonly are euhedral in contact with calcite; vesuvianite is euhedral in contact with garnet. All embay diopside irregularly, generally having scalloped boundaries convex toward the diopside, and all locally contain island remnants of diopside and of relict minerals of the country rock. Garnet, diopside, epidote, and vesuvianite commonly have ragged boundaries with chlorite, and in places form tiny island remnants in patches of chlorite.

DIOPSIDE

Diopside is an almost ubiquitous constituent of the rodingite, is generally a major constituent, and is commonly the most abundant mineral. In coarse rodingite, the diopside ranges in size and form from anhedral grains a few millimeters to several centimeters across, to prismatic crystals as much as 3 cm long. Diopside in dense flinty rodingite is all fine grained and ranges in habit from subequant anhedral grains, wispy blades, and irregular poikiloblastic grains to subequant subhedral grains and well-formed tabular crystals. The poikiloblastic grains enclose quartz, epidote, and other relict minerals from the country rocks. In places, fine-grained diopside irregularly rims relict grains of coarse hornblende, and invades these grains along fractures and cleavages. Diopside in small veins and irregular coarser grained patches in fine rodingite is commonly bladed or flamboyant in habit.

The diopside varies in color from moderate yellow green to grayish olive. Euhedral crystals are generally short prisms terminated by basal pinacoids; {100}, {010}, and {111} faces are subdued. A few crystals are tabular, parallel to {010}. Prismatic cleavage is generally distinct.

The diopside varies appreciably in optical properties and inferred chemical composition. The optic axial angle, $2V$, varies from 60° to 70°, the optic sign is (+), the extinction angle is nearly 45°, and the β-index of refraction varies from 1.671 to 1.705. These data indicate a range in composition, in terms of the diopside and hedenbergite end members, from He_0 to about He_{50}, or from $CaMgSi_2O_6$ to about

$$Ca\,(Mg_{0.5}Fe^{+2}_{0.5})Si_2O_6.$$

CLINOZOISITE

Clinozoisite is nearly as prevalent as diopside in the rodingite and not uncommonly is the most abundant mineral. In fine-grained rodingite, clinozoisite occurs as tiny anhedral subequant grains in an equigranular mosaic, as larger rounded grains in a fine-grained groundmass, and as subhedral crystals in patchy and veinform aggregates. In coarse rodingite, clinozoisite forms striated bladed and anhedral crystals as large as 3 cm across.

The clinozoisite is dusky yellow green, light olive gray, or olive gray, and well-formed crystals are strongly striated. In thin section, it appears colorless and commonly has two good cleavages about at right angles. The fine-grained material is mostly untwinned and shows uniform extinction. Coarse-grained material commonly is strongly twinned. Most of the large rounded grains set in a fine groundmass are zoned, the inner part of the grain commonly showing undulatory extinction and the outer part showing uniform extinction and slightly different optic orientation.

The optical properties vary within a relatively small range. The optic axial angle, $2V$, is large, and the optic sign varies from (+) to (−). Abnormal blue and lemon-yellow interference colors are distinctive. The β-index of refraction ranges from 1.712 to 1.728 but is chiefly in the range 1.717–1.723. These properties indicate a range in composition from close to that of iron-free epidote, $Ca_4Al_4Al_2Si_6O_{24}(OH)_2$, to perhaps as much as

$$Ca_4Al_4(Al_1Fe^{+3}_1)Si_6O_{24}(OH)_2,$$

and a predominant composition of about

$$Ca_4Al_4(Al_{1.3}Fe^{+3}_{0.7})Si_6O_{24}(OH)_2.$$

ZOISITE

Zoisite was noted only in coarse rodingite, where locally it forms coarse radial aggregates of bladed white crystals. The zoisite contains two optically distinct phases in intimate association. One phase has $2V=0°–50°$; optic sign (+); dispersion, $r<v$ strong; abnormal blue interference colors; indices, $\alpha=1.703$, $\beta=1.704$, $\gamma=1.708$; optic orientation, $a=\gamma$, $b=\beta$, $c=\alpha$. Zoisite having such optical properties is commonly correlated with iron-free or α-zoisite. The second phase has $2V=0°–35°$; optic sign (+); dispersion, $r>v$ distinct; abnormal brown interference colors; indices $\alpha=1.701$, $\beta=1.701$, $\gamma=1.706$; optic orientation, $a=\gamma$, $b=\alpha$, $c=\beta$. Zoisite having such properties is commonly correlated with ferrian or β-zoisite. The slight difference in optical properties suggests that the difference in composition between the two phases is also slight. The "β-zoisite" has only slightly higher indices than an analyzed sample of zoisite listed by Deer, Howie, and Zussman (1962, v. 1, p. 188, table 32, analysis 3), in which the ratio $Fe^{+3}/(Al+Fe^{+3})=0.033$. The "$\alpha$-zoisite" and "$\beta$-zoisite" from Belvidere Mountain are inferred to have comparable $Fe^{+3}/(Al+Fe^{+3})$ ratios, or formula compositions near

$$Ca_4Al_4(Al_{1.8}Fe^{+3}_{0.2})Si_6O_{24}(OH)_2.$$

GARNET

Garnet is a common and generally abundant constituent of both fine and coarse rodingite. It forms euhedral to subhedral crystals as much as 1 cm across in coarse rodingite, and small anhedral grains in fine rodingite. Garnet varies widely in color and optical properties in both types of rock and shows a variety of textural relations in fine rodingite.

In fine rodingite, garnet forms massive aggregates of small grains and scattered clusters of grains. Patches of massive garnet have scalloped borders that are convex toward the diopside, and the garnet commonly encloses irregular shreds of diopside. Small veinlets of garnet or garnet and vesuvianite are common. Some such veinlets are complex; they consist of a central straight-sided vein about 0.05 mm wide of isotropic or nearly isotropic garnet, bordered on either side by narrower zones of birefringent garnet whose contact with diopside is very irregular—as if the central vein had formed by fissure filling, the outer zones by replacement.

The garnet has a wide range of colors, chiefly shades of green, brown, and red. Most of the garnet is slightly to distinctly birefringent. The index ranges from 1.740 to >1.789. In general, euhedral garnet in coarse rodingite and garnet in straight-sided veinlets are of low index, and garnet in the veinlets is isotropic; anhedral fine-grained garnet, whose relations suggest a replacement origin, is of relatively high index and is distinctly birefringent. However, there are many exceptions to these general relations; in particular, many of the large euhedral crystals of garnet are distinctly birefringent and of relatively high index.

Microprobe analyses of garnets from rodingite, mostly coarse grained and euhedral to subhedral but representative of a wide range of colors and indices of refraction, indicate that the garnets are of two limited compositional ranges (gro:and:alm:sp): (1) 86–91:14–4:4–1:0–4, and (2) 45–66:55–32:0–2:0. Representative formula compositions for each range are:

(1) $[Ca_{2.86}Mn_{0.02}Fe^{+2}_{0.12}][Al_{1.92}Fe^{+3}_{0.07}Ti_{0.01}]Si_{2.99}O_{12}$, and

(2) $[Ca_{2.99}Mn_{0.02}][Al_{0.94}Fe^{+3}_{1.04}]Si_{3.0}O_{12}$.

Data are meager, but no correlation between composition and physical and optical properties is apparent.

VESUVIANITE

Vesuvianite, though less common and abundant than diopside or garnet, is a fairly common and locally abundant constituent of the rodingite. It ranges in habit and grain size from anhedral fine grains to prismatic crystals as much as 3 cm in diameter and 10 cm long, some of them singly terminated by pyramids and basal pinacoids, and Belvidere Mountain has long been highly prized for vesuvianite by mineral collectors. Massive vesuvianite commonly consists of anhedral grains; isolated grains are commonly subhedral. As opposed to garnet, vesuvianite generally is euhedral.

The vesuvianite ranges in color chiefly from grayish yellow green to moderate yellow green. The optical properties are generally uniform; the ω-index is chiefly in the range 1.718 to 1.720. One specimen (AV–237) has an index $\omega=1.730$. The optic axial angle range is moderate (40°–50°), the optic sign (−), the birefringence low, and interference colors of abnormal violet and yellowish brown are characteristic.

On the basis of the light-green color (Deer and others, 1962, v. 1, p. 118) the vesuvianite is inferred to contain chiefly ferric, and little ferrous, iron. The indices are near those of a light-green vesuvianite listed by Deer and others (1962, v. 1, p. 116, table 21, analysis 5). Neither the exact mineral formula nor the correlation of optical properties with chemical composition are adequately known, but on the basis of the data available, it appears that the average composition of the vesuvianite, in terms of the most widely accepted ideal formula, is approximately $Ca_{10}(Mg_{1.0}Fe^{+3}_{0.8}Fe^{+2}_{0.2})Al_4[Si_2O_7]_2[SiO_4]_5(OH,F)_4$.

PREHNITE

Prehnite forms distinctive fine drusy aggregates on a few joint surfaces; it is rarely a sparse constituent of fine-grained rodingite, where it forms anhedral patches interstitial to diopside and clinozoisite. The optical properties are somewhat variable but show no consistent differences between drusy prehnite and that in the groundmass of fine-grained rodingite. The optic axial angle is $2V=70°–75°$; the optic sign (+); indices, $\alpha=1.614–1.620$, $\beta=1.623–1.634$, $\gamma=1.642–1.643$; extinction is parallel. The properties indicate a composition in which the ratio of $Fe^{+3}/(Al+Fe^{+3})$ is in the range 0.02–0.10. The average formula composition is inferred to be about

$$Ca_2(Al_{1.9}Fe^{+3}_{0.1})Si_3O_{10}(OH)_2.$$

CALCITE

Calcite is a common and abundant constituent of coarse rodingite, where it forms anhedral grains interstitial to other minerals. It is a sparse constituent of fine rodingite, of blackwall, and of tremolite rock. The ω=index is uniformly 1.658 to 1.660, indicating a composition near that of pure calcite, $CaCO_3$.

MAGNETITE

Magnetite is a relatively uncommon and generally sparse constituent of rodingite, blackwall chlorite rock, and tremolite rock; it is a ubiquitous and commonly abundant constituent of serpentine-chlorite rock and of calcite-chlorite-magnetite veins associated with rodingite and serpentine-chlorite rock.

Magnetite in rodingite, blackwall chlorite rock, and tremolite rock is anhedral, subequant to irregular, and has generally a distribution pattern similar to that of magnetite in adjacent country rock. Magnetite in serpentine-chlorite rock is varied in habit, ranging in grain size from 0.015 to 1.5 mm and in form from anhedral to euhedral. Some thin sections contain both scattered large anhedral grains and abundant small euhedral to subhedral grains. In all sections, the proportion of magnetite in serpentine-chlorite rock is higher than in adjacent rodingite. Magnetite in the calcite-chlorite-magnetite veins forms crystals as much as 1 cm across which commonly show octahedral and dodecahedral forms.

No information is available on possible variations in composition of magnetite in the several contact rocks. An analysis of a sample from a calcite-chlorite-magnetite vein (table 1, analysis 46) gave a calculated formula composition of

$$(Fe^{+2}_{0.91}Mg_{0.01}Mn_{0.01}Ti_{0.04}V_{0.01})\\(Fe^{+3}_{1.91}Al_{0.05}Cr_{0.02})O_4.$$

CHLORITE

Chlorite is the sole essential constituent of blackwall chlorite rock, an essential and major constituent of serpentine-chlorite rock and calcite-chlorite-magnetite veins, and a commonly abundant constituent of rodingite and tremolite rock. The habit and textural relations of the chlorite are varied. It occurs as blades, flakes, and shreds in blackwall, serpentine-chlorite rock, rodingite, and tremolite rock. It forms distinctive books of platy crystals in large calcite-chlorite-magnetite veins, as well as in varied tiny veinlets in the rodingite. In the large veins, these chlorite books are commonly hexagonal and are as large as 3 cm across. The chlorite is intimately intermixed, in apparent equilibrium relation, with serpentine in serpentine-chlorite rock; it invades and replaces garnet, vesuvianite, diopside, and clinozoisite in rodingite; it replaces tremolite in tremolite rock; and it replaces relict minerals of the country rock in the blackwall.

Chlorite in contact rocks is almost all of relatively low index, positive optic sign, negative elongation, and abnormal brown interference colors. It shows some minor but distinctive variations among different contact rocks.

Chlorite in the calcite-chlorite-magnetite veins shows little variation in optical properties. An analyzed sample (A–BM–56–4b, table 1, analysis 49) showed $2V=30°$, optic sign (+), elongation (−), abnormal brownish-gray interference colors, $\alpha \approx \beta = 1.576$, $\gamma = 1.580$. The formula composition calculated from the chemical analysis (Chidester, 1962, p. 132–138) is

$$(Mg_{4.80}Fe^{+2}_{0.11}Mn_{0.005})(Al_{0.86}Fe^{+3}_{0.06}Cr_{0.01}Ti_{0.001})\\(Si_{3.06}Al_{0.94})O_{9.43}(OH_{8.85}F_{0.003}).$$

This composition fits rather closely with that based upon the optical correlations of Winchell and Winchell (1951, fig. 267; see Chidester, 1962, fig. 13, p. 45).

Measured indices of chlorite in blackwall range from $\beta=1.594$ to $\beta=1.613$ and are predominantly near $\beta=1.600$. Chlorite of index $\beta=1.600$ has a composition of about

$$(Mg_{3.5}Fe^{+2}_{1.3}Al_{1.2})Si_{2.7}Al_{1.3})O_{10}(OH)_8.$$

Observations elsewhere in Vermont (Chidester, 1962, p. 66–69) suggest that the total range in composition of chlorite in blackwall may be appreciably greater than is indicated by the relatively few optical determinations made on blackwall chlorite at Belvidere Mountain, and that the variation is systematic within the zone, the ratio of $Mg/Mg+Fe^{+2}$ increasing sharply near a contact with steatite and decreasing sharply near a contact with schist (Chidester, 1962, fig. 17, p. 67).

Chlorite in serpentine-chlorite rock has a similar range in optical properties; variations appear not to be systematic with respect to position in the zone. Some of the chlorite is nearly isotropic, consists predominantly of abnormal brown (optically positive) flakes intermixed with a relatively small proportion of abnormal blue (optically negative) flakes, and has indices of about $\beta=1.575$. Such chlorite, on the basis of Winchell and Winchell's data (Chidester, 1962, fig. 13, p. 45) is inferred to have a composition of about

$$(Mg_{5.0}Fe^{+2}_{0.3}Al_{0.7})(Si_{3.3}Al_{0.7})O_{10}(OH)_8.$$

In index, much of the chlorite reaches a maximum of $\beta=1.62$, and in birefringence reaches $\gamma-\alpha=0.005$. The corresponding limiting range in composition is inferred to be about

$$(Mg_{2.6}Fe^{+2}_{2.1}Al_{1.3})(Si_{2.7}Al_{1.3})O_{10}(OH)_8.$$

Chemical analyses were obtained of three samples of serpentine-chlorite rock (TPT–Vt–11(VT–11), AV–78 (VT–13), and AR–110 (VT–20), analyses 7, 9, and 16, table 1). The distribution of minerals in serpentine-chlorite rock is erratic, so that accurate modes of the material analyzed are not available. It is estimated, however, that the material of analysis 7 consists predominantly of chlorite and serpentine in the proportion of about 25/75; that of analysis 9, about 50/50; and that of 16, about 60/40. Calculated formula values of the serpentine-chlorite mixture in each show R^{+3}/Si ratios of 2/3, 1.4/3.3, and 1.3/3.4, and $Mg/Mg+Fe^{+2}$ ratios of 0.88, 0.99, and 0.98. As the admixed serpentine has a much lower content of R^{+3} than the chlorite, these ratios set lower limits for the content of R^{+3} in the chlorite. Most of the chlorite must therefore have a composition near that indicated for chlorite of β–index$=1.62$ (see above), a ratio of R^{+3}/Si of about 1.

Chlorite in tremolite rock has about the same range in optical properties as that in blackwall but varies erratically. Measured indices of chlorite in rodingite range from $\beta=1.585$ to $\beta=1.621$, corresponding to a range in composition of

$$(Mg_{4.2}Fe^{+2}_{0.7}Al_{1.1})(Si_{2.9}Al_{1.1})O_{10}(OH)_8 \text{ to } (Mg_{2.5}Fe^{+2}_{2.2}Al_{1.3})(Si_{2.7}Al_{1.3})O_{10}(OH)_8.$$

Chlorite in analyzed samples of tremolite-chlorite rock (sample AV–205 (VT–17), analysis 13, table 1) and chlorite-tremolite rock (sample AV–206 (VT–18), analysis 14, table 1) is inferred to have a composition of

$$(Mg_{3.6}Fe^{+2}_{1.3})(Si_{2.9}Al_{1.1})O_{10}(OH)_8$$

on the basis of optical properties. Cross calculations based on modal compositions were carried out on the two analyses (see Chidester, 1962, p. 132–138); these calculations yielded a calculated formula of

$$(Mg_{3.5}Fe^{+2}_{1.2}Mn_{0.02})(Fe^{+3}_{0.1}Al_{1.05})(Si_{2.5}Al_{1.15}Ti_{0.3}P_{0.05})O_{10}(OH)_{8.2},$$

very close to that based on optical properties.

Chlorite in the single and unusual occurrence of massive fine-grained chlorite in an irregular veinlike mass between serpentinite and rodingite is nearly isotropic, has a $\beta=$index of 1.632, and shows both abnormal pale-pink and pale-blue interference colors. The composition is inferred to be about

$$(Mg_{1.8}Fe^{+2}_{2.8}Al_{1.4})(Si_{2.6}Al_{1.4})O_{10}(OH)_8.$$

SERPENTINE

Serpentine in the contact rocks occurs only in serpentine-chlorite rock, where it is everywhere a major component but varies in proportion from as little as 25 percent to as much as 95 percent. The serpentine has a variety of textural features. Most occurs as irregular flakes and blades, intimately intermixed with chlorite or segregated rather irregularly into layers predominantly of serpentine. Isotropic patches of macroscopically structureless serpentine appear to be pseudomorphic after garnet and vesuvianite. Rarely, tiny veinlets of fibrous serpentine occur next to the contact with the ultramafic body.

The serpentine varies from colorless or very pale green to yellowish brown in thin section, and from virtually isotropic to distinctly abnormal blue or light gray under crossed nicols. Measured indices of refraction range from $\beta=1.548$ to $\beta=1.573$, and birefringence from nearly zero to 0.009. The sign of elongation is uniformly positive. The optic sign of all platy serpentine is negative; the apparent optic sign of fibrous serpentine is positive.

X-ray and DTA studies by G. T. Faust (written commun., 1961) of two specimens of serpentine-chlorite rock indicate the presence of lizardite and chrysotile, in addition to chlorite. Chemical analyses of the serpentine-chlorite rock (TPT–Vt–11 (VT–11), AV–78 (VT–13), and AR–110 (VT–20), analyses 7, 9, and 16, table 1) cannot yield accurate formula compositions for the serpentine minerals, but the average ratio of $Mg/Mg+Fe^{+2}$ for the serpentine minerals must be greater than that of the mixture in each analysis (see preceding section on "Chlorite" under this section on "Mineralogy and paragenesis"). Thus the average ratio of $Mg/Mg+Fe^{+2}$ must be near 0.99 or higher. No useful information on the content of Al and Fe^{+3} in the serpentine minerals can be deduced. Though the content must be low (most Al and Fe^{+3} are required for the chlorite), limitations of the data allow for a range of variations considerably greater than possible critical differences between different serpentine minerals.

TREMOLITE

With the exception of relict hornblende in rodingite and blackwall chlorite rock, tremolite is the sole amphibole in contact rocks. Tremolite forms a thin zone between steatite and blackwall and in a few places occupies the position of the entire steatite zone. The tremolite varies from finely fibrous to blades as much as 1 mm wide and 5 mm long. The blades are commonly bent, broken, or folded; not uncommonly, laths of tremolite tail out into finely fibrous bundles of asbestiform tremolite.

The optical properties vary only moderately: $2V=75°–85°$, the optic sign $(-)$, the extinction angle $\gamma \wedge c=15°–20°$, and the γ-index of refraction ranges from 1.625 to 1.636 but in most places is near 1.632. These optical properties correspond to a range in composition from pure tremolite, $Ca_2Mg_5Si_8O_{22}(OH)_2$, to actinolite of formula composition

$$Ca_2(Mg_{4.5}Fe^{+2}_{0.5})Si_8O_{22}(OH)_2;$$

most is about

$$Ca_2(Mg_{4.7}Fe^{+2}_{0.3})Si_8O_{22}(OH)_2$$

(Foslie, 1945, fig. 1).

A chemical analysis of tremolite-chlorite rock in the position of the steatite zone (sample AV–205 (VT–17), analysis 13, table 1), the tremolite of which has a γ-index of 1.632, yields the following formula composition (after correcting for about 12 percent chlorite of β-index$=1.599$, and very small amounts pyrite and calcite):

$$Ca_{1.9}(Mg_{4.9}Fe^{+2}_{0.2})Si_{8.0}O_{22.0}[O_{0.4}(OH)_{1.6}].$$

ILMENITE, RUTILE, AND SPHENE

All the contact rocks generally contain ilmenite, rutile, and sphene, singly or in combination, locally as major constituents. In the different contact rocks one or another of the three minerals commonly predominates, not uncommonly to the almost complete exclusion of the others. In some of the contact associations, the titanium minerals vary regularly between rock types, and in some of the individual contact rocks they show regular variations with respect to position in the zone, or regular sequential relations with one another.

Ilmenite is a ubiquitous and commonly a major constituent of the serpentine-chlorite rock. It is rare in the rodingite and uncommon in the blackwall, in both of which it occurs only as relict grains in association with rutile or sphene, or both. The ilmenite ranges in habit from tiny laths to irregular grains.

Rutile is relatively uncommon in the contact rocks. It is most abundant in the blackwall; there it forms prismatic crystals, or anhedral aggregates associated with either ilmenite or sphene or both. Rutile occurs rarely in the rodingite, and always in association with ilmenite and sphene.

Sphene is the most common and abundant mineral in the contact rocks. It is nearly ubiquitous in the blackwall and rodingite; it is common in tremolite rock, in steatite derived from schist or amphibolite, and in miscellaneous irregular types of contact rocks. Only serpentine-chlorite rock is devoid of sphene. The sphene ranges in habit from irregular aggregates of tiny grains to diamond-shaped single crystals.

The relations of ilmenite, rutile, and sphene are regular for each association but differ between associations. In the rodingite, sphene generally occurs alone, but in some places, particularly in the outer parts of the zone, some of the sphene contains a core of ilmenite or a shell of ilmenite around a core of rutile. At the contact between rodingite and serpentine-chlorite rock, elongate grains that project across the contact consist of sphene on the rodingite side of the contact and ilmenite on the serpentine-chlorite rock side. In the blackwall, the overall relation among the three minerals is that of a core of ilmenite surrounded by a shell of rutile and an outer rim of sphene; any one or a pair of the three may be absent locally (Chidester, 1962, p. 69.)

OTHER MINERALS

Talc is a minor constituent of the transition zone between steatite and blackwall chlorite and is the essential major constituent of the steatite, the outermost few millimeters of which belong to the contact-rock association. The properties are virtually the same as those of talc derived from ultramafic rocks.

White mica, quartz, and albite occur principally in the outer margins of the rodingite and the blackwall zones, as relict minerals of the adjacent country rock. The three minerals are associated in the irregular masses of muscovite-albite-quartz rock which bear a replacement relation to the adjacent amphibolite near 99,800 N., 91,600 E. (see section on "The steatite and blackwall chlorite rock association" under "Contact rocks" in the section on "Ultramafic and associated rocks"). The mica occurs as compact fine-grained masses; the quartz and albite as irregular anhedral grains. The mica is identified as muscovite on the basis of X-ray pattern and optical properties of $2V=30°$, optic sign $(-)$, $\gamma=1.599$, $\gamma-\alpha=0.034$.

Apatite, hornblende, biotite, sulfides, and graphite, all are relict minerals of the country rocks from which the contact rocks were derived. All are similar in habit and composition to their equivalents in the unaltered country rocks.

PETROGENESIS

Igneous and metamorphic features and geologic relations of the ultramafic rocks attest to an igneous origin of the dunite, peridotite, and chromitite, to later alteration of much of the igneous rock to serpentinite, carbonate-quartz rock, serpentine and magnetite veins, talc-carbonate rock, and steatite, and to a regular sequential relation among them. Mineralogical and geological relations and textural features of the contact rocks attest to their origin by metasomatic replacement of the country rock and point to systematic genetic and chronologic relations between the contact rock and the ultramafic rocks.

Direct evidence of relative ages based on field relations between the various ultramafic rocks and the contact rocks demonstrates only that the serpentinite is younger than the primary igneous minerals, that the serpentine veins are younger than the consolidation of the ultramafic igneous rocks, and that the several talcose rocks are younger than the serpentinite. Inasmuch as contact rocks of the rodingite and serpentine-chlorite rock association, on the one hand, and those of the steatite and blackwall association or the tremolite and chlorite rock association, on the other, are mutually exclusive in occurrence, their field relations are ambiguous. Elsewhere in Vermont, steatitized ultramafic bodies contain talc pseudomorphic after asbestos (Chidester, 1962, p. 76–77), which demonstrates that the steatite is younger than the asbestos.

Serpentine veins appear to be generally younger than the pervasive serpentinization because the veins, which are unsheared, were formed late in the tectonic history, whereas pervasive serpentinization was early. Pervasive serpentinization and the formation of serpentine veins may have overlapped in time to a significant extent. Vein formation took place in a relatively short interval late in the tectonic history of the area, whereas pervasive serpentinization probably took place over a much longer interval that began before final emplacement of the ultramafic bodies. Very probably, pervasive serpentinization extended into the period of vein formation (see following section on "Serpentinite" under this section on "Petrogenesis").

IGNEOUS ROCKS

An intrusive igneous origin for the dunite and peridodite is indicated by the locally crosscutting relations of the ultramafic body, the mosaic texture of olivine in dunite and peridotite and of chromite in chromitite, and the phenocrystic habit of the pyroxene and chromite. Layering of the igneous rocks is inferred to be, in part, at least, the result of crystal accumulation, but it may have been modified and enhanced tectonically by flowage during emplacement of the mass. Some of the thinner layering may be entirely of tectonic origin. Modification of the original igneous texture by cataclasis is indicated by the prevalence of undulatory extinction and glide twinning in part of the olivine throughout the dunite and peridotite, the vague mosaic extinction in large crystals of olivine, and the parallel dimensional and crystallographic orientation of the olivine in some of the dunite.

Olivine, pyroxene, and chromite formed early and more or less contemporaneously. The inclusion of olivine grains in some single crystals of pyroxene indicates that pyroxene was in part later than olivine. As crystallization proceeded, translucent reddish-brown chromite gave way to opaque chromian magnetite. The chromian magnetite formed as rims surrounding cores of translucent chromite, as dispersed discrete grains, as concentrations of small grains interstitial to olivine grains, as vein fillings in fractured chromite grains, and as thin discrete layers in dunite and peridotite. Pyrite and sulfarsenides formed at about the same time as magnetite, with which they are locally intergrown.

Pyroxene is pseudomorphically replaced by anthophyllite, which is in turn replaced by serpentine. Alteration of pyroxene to anthophyllite probably took place in an early postmagmatic stage, during cooling of dunite and peridotite.

Other minerals in the igneous rocks are related to serpentinization, and will be discussed in the following section.

SERPENTINITE

Many features demonstrate that serpentinite was formed by alteration of dunite and peridotite. For massive serpentinite, these include a gradational relation with dunite and peridodite, preservation of relict layering, aggregate pseudomorphs after olivine, replacement of olivine by serpentinite, and the presence of relict olivine, pyroxene, chromite, and magnetite. For schistose serpentinite, the principal evidence is a gradational relation to massive serpentinite and a prevalence of relict chromite grains.

Textural features of massive serpentinite indicate that serpentinization was effected without perceptible change in volume. There is no evidence in the schistose serpentinite bearing on volume change during serpentinization. Shear polyhedrons formed during tectonic transport are evidence of early serpentinization, perhaps beginning in the upper mantle. (See section on "Emplacement and structural history.")

Antigorite, the predominant mineral in the serpentinite, formed chiefly by replacement of olivine and only in minor amounts by replacement of pyroxene or anthophyllite. Chrysotile, which is relatively minor in the groundmass serpentine, formed chiefly in tiny fractures in and at the margins of grains of olivine but locally appears to have replaced whole grains or groups of grains. Possible six-layer orthoserpentine, which occurs in the centers of mesh-structure serpentine, formed by replacement of olivine.

Chromite and the larger grains of magnetite are relict igneous minerals. Magnetite in the form of dispersed fine dusty particles, scattered small grains, concentrations of tiny grains along relict cleavage surfaces of altered pyroxene, and aggregates of tiny grains that rim many of the chromite grains and the larger magnetite grains was formed during serpentinization.

A small amount of chlorite was formed by the alteration of pyroxene or anthophyllite to antigorite and chlorite during serpentinization, but most of the chlorite was formed by interaction between chromite or chromian magnetite and the surrounding antigorite, during and immediately following serpentinization. During the reaction between the spinellids and antigorite, aluminum was lost from the spinellids, and entered the chlorite and locally the adjacent antigorite to form a thin halo of aluminian antigorite. The loss of aluminum from chromite and magnetite grains effected a change in composition of the outer shell of the grains, but the resulting reaction zone cannot generally be distinguished from the zones of opaque chromian magnetite of igneous origin. Perhaps the thin opaque borders of fractures in translucent chromite grains are chiefly or entirely such reaction zones.

Brucite in the groundmass of the serpentinite formed entirely by replacement of olivine, to which its relation is clearly shown in the partly serpentinized dunite.

Magnesite, dispersed in the groundmass serpentine and as small lenticles, is a product of serpentinization of the olivine. Dispersed grains of calcite are probably all associated with late veins. Talc, which occurs only rarely in the schistose serpentinite and near the margins of the ultramafic bodies, is probably all later than the serpentinization. (See following section on "Steatite, talc-carbonate rock, and carbonate-quartz rock" under "Petrogenesis" in the section on "Ultramafic and associated rocks.")

The origin of the graphite in serpentinite is uncertain. Concentration of graphite near the margins of the ultramafic body probably occurred during serpentinization or steatitization, or both, and redistribution along schistosity and slip cleavage surfaces was probably accomplished mechanically during folding. Graphite was not seen in fresh dunite and peridotite. Trofimov (1940, p. 34–35) reported carbon-bearing peridotite in Siberia that contains traces to as much as 0.38 percent carbon; at Belvidere Mountain it is possible that such graphite—which would be very difficult to detect microscopically in small amounts—was concentrated at the margins during serpentinization. Concentrations of graphite at the outer margin of the blackwall chlorite rock zone of steatitized ultramafic bodies elsewhere in Vermont have been interpreted as being formed by outward displacement from graphitic schist during the blackwall-steatite reaction (Chidester, 1962, p. 123, 127). Thayer (1966, p. 697–700) suggested that the reduction of carbon dioxide by hydrogen produced by decomposition of water during serpentinization may result in the formation of graphite. Hydrogen might also be formed in a similar manner during steatitization. In either case, reaction of such hydrogen with carbon dioxide mobilized during regional metamorphism would be expected at the margins of the ultramafic bodies. Such a reaction would explain well the distribution and apparent age relations of the graphite.

SERPENTINE VEINS

That irregularities in opposite walls of the serpentine veins match even down to fine details indicates that the veins have formed chiefly by fracture filling. Only locally has an appreciable proportion of a vein formed by replacement; elsewhere, replacement was negligible. In both the cross-fiber asbestos veins and the cross-column picrolite veins, the observation that matching irregularities in opposite vein walls are at opposite ends of a group of fibers indicates that the axes of the fibers and columns mark the relative direction of movement of the walls during opening of the fissure. Though the relations are generally obscure for the so-called slip fiber in the schistose serpentinite, the orientation and length of the fibers probably also mark the direction and amount of movement on opposite walls of much of the slip-fiber veins.

Riordon (1955) concluded, for asbestos deposits in Quebec, that the delicate layering parallel to the walls in many of the picrolite and cross-fiber veins, and the interrelations of the veins, indicate that all the serpentine veins were injected into open fissures as "collodial serpentine," that the colloidal serpentine later recrystallized to picrolite, and that the picrolite subsequently recrystallized to cross-fiber asbestos.

We propose, and favor, an alternative explanation for these features.

We suggest, on the basis of the relations previously described (in the section on "Serpentine veins" under "Igneous rocks, serpentinite, and veins" in the section on "Ultramafic and associated rocks") that the thin delicate layering in the veins records discrete increments in the opening of the fissures occupied by the veins. Differences in color of different layers probably reflect very slight differences in composition of successive layers; slight differences in orientation of the fibers in successive layers record slight differences in direction of relative movement of the vein walls during successive increments of opening. We further suggest that when the rate of opening of the fissure equaled or only minutely exceeded the rate of influx of material into the fissure, the conditions led to an almost perfect orderly growth of parallel tubular fibers of chrysotile. When the rate of opening exceeded the rate of addition of material to the fissure by a discrete but small amount, the fibers of chrysotile grew in a slightly disarrayed but subparallel fashion to produce coarsely fibrous to columnar picrolite. When the rate of opening of the fissure was so rapid that an appreciable void resulted, the chrysotile fibers formed in a sufficiently disarrayed manner so that the picrolite appears massive, without megascopically discernible fibrous or columnar structure. The simple veins of chrysotile asbestos record a history in which the rate of opening of a fissure uniformly almost equaled the rate of growth of the fiber; the simple veins of picrolite record a history in which the rate of opening uniformly exceeded the rate of growth. The composite veins record a complex history in which the rate of opening at times about equaled, and at times exceeded, the rate of growth of the vein material. Veins of unlayered asbestos and picrolite record a history in which the rate of opening was steady, or in which there was no significant change, in successive impulses, in the relative direction of movement of the walls and in composition of the vein material.

Concentrations of magnetite along partings and at margins of serpentine veins probably consist chiefly of magnetite formed during fissure filling, but some may consist partly of original seams of magnetite along which fissures opened in the rock. In single-fiber veins, separation during opening of a fissure occurred repeatedly at the junction of fiber and magnetite seam or fiber and vein wall. In multiple-fiber veins, separation occurred successively at opposite sides of a magnetite seam or at opposite vein walls. In veins with strongly crenellated partings, separation took place irregularly along the vein at one side or another of the partings in successive impulses of vein opening. Chips of serpentinite along partings mark sites where fractures extended locally into wallrock, rather than uniformly along a junction of wallrock and magnetite seam or fiber.

The relations of fibers at vein intersections indicate relative ages of fissures. Merging relations of the fiber indicate that intersecting fissures opened and vein material formed contemporaneously. Simple crosscutting relations indicate that one fissure opened and its vein material formed entirely before the other fissure opened. Complex intersections indicate that fissures and vein materials are in part contemporaneous and in part sequential.

Pockets of long slip fiber that are intergrown with abundant fibrous calcite in shear zones are difficult to explain. Perhaps they formed under early static conditions, by slow growth of patches of fiber that had previously been sheared into parallelism with the enclosing shear zones.

The fibrous habit of much of the calcite and brucite, which occur in generally minor amounts in many of the serpentine veins, suggests that they formed largely contemporaneously with the serpentine, but calcite of granular habit appears locally to replace chrysotile asbestos.

Marginal zones of alteration adjacent to serpentine veins are inferred to be contemporaneous with and genetically related to the veins. The zones were formed by replacement of olivine by chrysotile, lizardite, and brucite in dunite and massive serpentinite bordering the veins. Layering in marginal zones probably records successive outward waves of alteration corresponding to successive episodes of vein filling.

The general relation that the width of the marginal zone of alteration is proportional to the width of the adjacent vein suggests that the vein material (except water) was derived largely from immediately adjacent wallrock. Many local departures from this general relation indicate that in places significant amounts of material moved along the veins. The relative amounts involved are illustrated in table 7, which compares the composition of equal volumes of cross-fiber chrysotile asbestos and of serpentinite from serpentinized zones bordering asbestos veins with that of dunite, moderately serpentinized dunite, and massive serpentinite. This comparison is in terms of the modified standard cell described by Chidester (1962, p. 95–97).

Table 7 shows that gains and losses in alteration zones are primarily dependent upon the degree of serpentinization of host rock. Minor differences in composition of host rocks account for other variations, particularly of minor constituents. For veins in only slightly serpentinized dunite (table 7, section *A*), the principal changes are loss of Mg and Si from, and introduction of OH into, the border zones. For veins of average ratio of width of border zone to width of vein, the proportional loss of Si and Mg from the border zones considerably exceeds the proportional amount of those constituents introduced into chrysotile veins. The anomalous gain in Al and Fe^{+3} in the border zone, indicated in table 7, section *A*, results from the atypical character of the analyzed sample of dunite (AV–248). The sample was selected for its freshness and freedom from pyroxene and magnetite and is therefore lower than average in its content of iron and aluminum. Sample AV–13f (table 7, section *B*) doubtless represents more closely the average content of aluminum and iron in the ultramafic rocks.

These considerations indicate that during formation of veins in relatively fresh dunite and peridotite, the supply of constituents from host rock adjacent to vein walls was more than adequate to form the veins and that appreciable Si, Mg, Fe, Al, and Ca must have migrated outward beyond the domain of the veins. This migration could account for many magnetite veins in the ultramafic rocks and possibly for calcite-chlorite-magnetite veins in rodingite and serpentine-chlorite rock.

In extensively serpentinized host rocks, the loss of Si and Mg (table 7, sections *B* and *C*) from the walls of the adjacent host rock is inadequate (for the average ratio of width of serpentinized border zone to width of asbestos vein) to supply the amounts in the veins. Clearly then, in most instances, those constituents of the serpentine veins cannot be exclusively, or even largely, of local derivation. They must have been supplied from more distant sources by the serpentinization process. This conclusion strongly suggests that pervasive serpentinization of the dunite and peridotite continued through the period of formation of serpentine veins.

OTHER VEINS

Thin veins and seams of magnetite are fracture controlled and are formed largely by replacement. The common association with asbestos veins and the invariable association with serpentinized zones tie them to formation of serpentine veins. Many veinform masses parallel to layering may be partly of magmatic origin, but most or all were modified and enlarged during serpentinization.

Because of their rarity and the lack of diagnostic features in the Belvidere Mountain area, irregular pods of chromian magnetite, such as the large pod exposed in the town (township) of Troy (fig. 1), are of uncertain mode of origin. They may be of magmatic origin and modified by intense local shearing, or they may have formed by replacement along shear zones.

Veins consisting of varied proportions of magnetite, brucite, and calcite formed partly by fracture filling, partly by replacement. The association of many such veins with, and their local replacement relations to, serpentine veins indicates that they are genetically related to and were emplaced during and late in the stage of formation of serpentine veins (see preceding section on "Serpentine veins").

Though precise relations are unknown, amphibole asbestos probably formed late during serpentinization by reaction of material in the shear zone with magnesium-bearing solutions from the ultramafic body.

STEATITE, TALC-CARBONATE ROCK, AND CARBONATE-QUARTZ ROCK

Geologic relations of the rocks and paragenetic relations of the minerals demonstrate that talc-carbonate rock and carbonate-quartz rock formed by replacement of ultramafic igneous rocks and serpentinite; steatite formed chiefly by replacement of serpentinite but partly by replacement of country rock. The absence of a serpentinized zone between some talc-carbonate rock and enclosing dunite in the Eden quarry body, and the zoned replacement relation of talc-carbonate rock to serpentinite elsewhere (see Chidester, 1962), demonstrate that steatitization is later than and unrelated to serpentinization.

Relict grains of chromite and chromian magnetite throughout talc-carbonate rock and carbonate-quartz rock indicate that the rocks were derived from ultramafic igneous rocks. The gradational relations

TABLE 7.—*Chemical composition of equal volumes of cross-fiber vein chrysotile, marginal alteration zones,*

[Chemical composition is in terms of the content of the modified standard cell (MSC) (Chidester, 1962, p. 95, 132–137); gains and losses are deter-
of the vein. The product is the proportional amount of each constituent gained or

Specimen No.	Zone	Width of zone	Si	Al	Fe^{3+}	Fe^{2+}	Mg	Ca
						A.	**Host rock, slightly**	
AV–248	Dunite	>1 m	41.6	0.19	1.4	3.5	76.3	0.09
A–BM–57–3	Serpentinized border zone	32 mm	34.1	1.5	2.7	.41	56.1	.02
A–BM–57–1	Chrysotile vein	4 mm	32.6	1.0	1.2	.37	51.0	.20
	Proportional gain or loss in the serpentinized border zone		−240	+42	+41.6	−99	−646	−2
	Proportional amount of constituent introduced into the chrysotile vein		130	4	5	2	204	1
						B.	**Host rock, moderately**	
AR–13f (VT–1)	Dunite	>1 m	38.3	1.3	2.2	2.5	69.3	0.05
AR–13f (VT–2)	Serpentinized border zone	28 mm	35.8	.83	1.6	.19	57.1	.00
AR–55	Chrysotile vein	4 mm	34.4	.8	1.0	.58	54.2	.00
	Proportional gain or loss in the serpentinized border zone		−70	−13	−17	−65	−342	−1
	Proportional amount of constituent introduced into the chrysotile vein		138	3	4	2	217	0
					C.	**Host rock**	**highly**	**serpentinized**
AR–73 (VT–6)	Dunite	>1 m	34.9	1.9	3.7	1.2	54.6	0.07
A–BM–57–3	Serpentinized border zone	80 mm	34.1	1.5	2.7	.41	56.1	.02
AR–55	Chrysotile vein	8 mm	34.4	.8	1.0	.58	54.2	.00
	Proportional gain or loss in the serpentinized border zone		−64	−32	−80	−63	+120	−4
	Proportional amount of constituent introduced into the chrysotile vein		275	6	8	5	434	0

[1] The total cations per MSC is, by definition, equal to $100 \times F_c$, and is here rounded to one decimal place. Because of rounding procedures, the total
tion by a number as large as 80),

between talc-carbonate rock and serpentinite and between talc-carbonate rock and dunite and peridotite indicate that talc-carbonate rock replaced serpentinite, dunite, and peridotite directly. Relations of carbonate-quartz rock to dunite or peridotite and serpentinite are not directly ascertainable because of exposure conditions. However, the zonal distribution and gradational relations of carbonate-quartz rock and talc-carbonate at the south end of the Corez Pond body, where carbonate-quartz rock forms a small central mass within a larger mass of talc-carbonate rock, suggest that the carbonate-quartz rock and talc-carbonate sequentially replaced serpentinite or dunite, or both.

The steatitization process is discussed in detail in an earlier report (Chidester, 1962) based on studies of extensive, well-exposed talc deposits elsewhere in Vermont. That report shows that the alteration of serpentinite to talc-carbonate rock is essentially isochemical with respect to the principal cations and involved chiefly the introduction of CO_2 and loss of (OH). Steatite formed as the result of metamorphic reaction between serpentinite and adjacent country rock, chiefly by outward migration of Mg and (OH) from the serpentinite and inward migration of Si from the schist. The two processes were essentially independent, though virtually contemporaneous. These conclusions are borne out in the Belvidere Mountain area, and similar analysis is not repeated here. However, the relations of talc-carbonate rock and carbonate-quartz rock to each other and to serpentinite and dunite provide additional information about the carbonatization of ultramafic bodies that warrants brief comment.

Calculations based on the assumption that carbonatization was isochemical with respect to the principal cations (see Chidester, 1962, p. 121–122) yield the approximate modal compositions for talc-carbonate rock and carbonate-quartz rock derived from dunite and serpentinite that are shown in table 8.

Comparison of the chemical composition of modified standard cells of rocks of the modal compositions shown in the table with that of dunite and serpentinite from which the talc-carbonate rock was derived indicates that only for the alteration of serpentinite to talc-carbonate rock can isochemical alteration take place without change in volume—that is, that the *and host dunite and serpentinite; and proportional gains and losses of constituents during vein formation*

mined by multiplying the width of the zone or vein by the difference between the cell content of the host rock and the zone, or by the cell content lost per zone, or introduced into a vein; dashed line, not determined. Data from table 1]

Na	K	Ti	P	Cr	Ni	Co	Mn	C^{4+}	Total cations [1]	H	F	Cl	O
serpentinized dunite													
0.00	0.00	0.00	0.00	0.21	0.24	0.02	0.14	0.10	123.7	23.9	---	0.01	178.2
.02	.01	.01	.00	.26	.23	.01	.08	.26	95.7	79.7	.00	.02	172.1
.02	.02	.01	.01	.12	.16	.01	.05	.53	87.2	73.8	.00	.03	158.4
+.6	+.3	+.3	0	+2	−.3	−.3	−2	+5	−896	+1786	0	+.3	−195
.1	.1	.04	.04	.5	.6	.04	.2	2	349	295	0	.1	634
serpentinized dunite													
0.00	0.00	0.01	0.02	---	---	---	0.18	0.20	113.9	47.6	---	0.05	178.1
.00	.00	.00	.00	.13	.18	.01	.08	.30	96.2	81.1	.00	.03	174.1
.00	.00	.06	.00	.10	.10	.00	.00	.00	91.3	81.0	.00	.05	167.2
0	0	−.3	−.6	---	---	---	−3	+3	−496	+938	---	−.6	−112
0	0	.3	0	.4	.4	0	0	0	365	324	0	.2	669
dunite (massive serpentinite)													
0.00	0.00	0.01	0.03	---	---	---	0.08	0.18	96.7	72.3	---	---	170.8
.02	.01	.01	.00	.26	.23	.01	.08	.26	95.7	79.7	.00	.02	172.1
.00	.00	.06	.00	.10	.10	.00	.00	.00	91.3	81.0	.00	.05	167.2
+2	+1	0	−2	---	---	---	0	+6	−80	+592	---	---	+140
0	0	.5	0	.8	.8	0	0	0	730	648	0	.4	1338

does not necessarily equal the sum of the cations shown to the left of the total. In calculations of gains or losses from a zone (involving multiplica- the discrepancy may be fairly large.

TABLE 8.—*Calculated modes of talc-carbonate rock and carbonate-quartz rock formed by isochemical alteration of dunite and serpentinite, and consequent percentage increase in volume (=percentage loss of Mg plus Si in isovolumetric alteration)*

	Talc-carbonate rock		Carbonate-quartz rock	
	From dunite	From serpentinite	From dunite	From serpentinite
Talc	50	60	--	--
Magnesite	50	40	70	65
Quartz	--	--	30	35
Percentage volume increase (=percentage loss of Mg plus Si)	35	0	45	15

composition of the modified standard cell for each rock is the same with respect to the principal cations. For dunite, equal-volume alteration to talc-carbonate rock of the modal composition indicated in table 8 requires a loss, for both Si and Mg, of about 35 percent of the amount present in the parent rock; equal-volume alteration of dunite to carbonate-quartz rock requires a loss of about 45 percent. On the same basis, the alteration of serpentinite to carbonate-quartz rock requires a loss of about 15 percent for Si and Mg; the alteration to talc-carbonate rock, as stated above, is both isochemical and isovolumetric.

Another way of stating the above relations is that isochemical alteration during carbonatization would require volume changes on the same order as those cited for changes in composition. The geologic and petrographic evidence available at Belvidere Mountain and elsewhere in Vermont (Chidester, 1962, p. 121; Jahns, 1967, p. 157) supports the conclusion that carbonatization was essentially isovolumetric. Furthermore, the irregular, abnormally wide, and mineralogically unusual zones of contact alteration observed in several places adjacent to talc-carbonate and carbonate-quartz rock masses in dunite suggest that appreciable material was displaced outward from the dunite during carbonatization. On the basis of the range in modal compositions observed, and on petrological considerations, it is inferred that the proportion of Mg to Si that was displaced varies considerably, but on the average is near their ratio in parent rock. Proportionately larger percentages of minor constituents, particularly Al and Fe^{+3}, were displaced; at the ultimate stage of carbonatization (carbonate-quartz rock), nearly all the Al and Fe^{+3} present in parent dunite or serpentinite was displaced.

These considerations suggest that, on the average, talc-carbonate rock and carbonate-quartz rock derived from serpentinite should differ characteristically from those rocks derived from dunite. Conditions of exposure at Belvidere Mountain do not permit quantitative evaluation of this conclusion, but the modes in table 4, section *B* and *C* lend support to it.

The association of carbonate-quartz rock and quartzose talc-carbonate rock with faults or shear zones in the Eden quarry body, and the zonal distribution of carbonate-quartz rock and talc-carbonate rock with relation to serpentinite in the Corez Pond body, suggest that carbonate-quartz rock is localized near channels that provided free access of CO_2, and where the partial pressure of CO_2 was consequently high. This relation, together with the petrologic considerations that have been discussed, indicate that carbonate-quartz rock is the end product of carbonatization of dunite and serpentinite. The normal succession, where the partial pressure and supply of CO_2 was sufficiently high, was the formation of talc-carbonate rock followed by the formation of carbonate-quartz rock. Local variations in the supply or partial pressure of CO_2 could account for local variations in the paragenetic sequence of minerals.

The presence of relict chromite in much of the steatite shows that the steatite was derived largely from ultramafic rock, but the presence of relict sphene in the outer part of the steatite zone indicates that the outer few millimeters or centimeters of steatite is derived from schist or amphibolite. The gradational relations of steatite and talc-carbonate rock in most places, and of steatite and serpentinite in only a few, indicate that in most places formation of steatite followed a preceding stage of formation of talc-carbonate rock, and only rarely was steatite formed directly by replacement of serpentinite. The invariable gradation outward of steatite into blackwall chlorite rock—in some places including an intervening narrow zone rich in tremolite—shows that steatitization of country rock followed an advance wave of chloritization and, locally, tremolitization.

CONTACT ROCKS

The diversity of contact rocks is related both to major differences in composition of the country rocks and to the several stages in the history of the ultramafic rocks.

Inasmuch as the rodingite and serpentine-chlorite rock association exists within a few meters of the steatite and blackwall association, and both associations are in the same kind of country rock (Ꞓhc, pl. *4D*), the two would seem to have formed under different physical conditions and are inferred therefore to be of different ages. Because rodingite and serpentine-chlorite rock occur only adjacent to serpentinite, at contacts where talcose rocks are absent, this association seems to be genetically related to serpentinization or to some older stage. Blackwall chlorite rock, however, occurs only at contacts where the ultramafic bodies are marginally altered to talcose rocks. Therefore, this association seems to be genetically related to steatitization. (See Chidester, 1962, p. 89–93, for a detailed discussion of the age relations of serpentinite to steatite and blackwall.) It follows, from the conclusion that serpentinization is older than and unrelated to steatitization, that the rodingite and serpentine-chlorite rock association is older than and unrelated to the steatite and blackwall association. The association of rodingite and serpentine-chlorite rock with ultramafic bodies that contain large volumes of dunite suggests further that the rodingite and serpentine-chlorite rock may be entirely or partly related to the stage of emplacement of the ultramafic bodies.

Neither the age of the tremolite rock and chlorite rock association relative to the other contact associations nor its genetic relation to the ultramafic rocks can be determined with complete assurance on the basis of geologic relations. The absence of talcose rocks between tremolite rock and serpentinite would seem to rule out a genetic relation to talcose rocks. On the other hand, the similarity of the tremolite rock and chlorite rock association—except for the absence of the talcose rocks—to the steatite and blackwall association that contains a tremolite zone is striking. The most significant fact is that the tremolite rock and chlorite rock association is confined to amphibolite. This association is probably a genetic and temporal equivalent of the steatite and blackwall association, and probably forms, because of compositional limitations, where country rock bordering ultramafic rock is amphibolite.

In each association, the zoned arrangement of the contact rocks peripheral to the ultramafic bodies, and the replacement relation the rock of a zone bears to that of the next zone outward, demonstrates a sequential relation marked by successive mineralogical changes that proceeded outward from the ultramafic bodies. This sequential relation requires that the outer part of each zone be younger than the adjacent part of the next zone outward, but permits an entire zone to be younger than the outward adjacent zone. In the first case, the zones would be essentially contemporaneous, and probably closely related genetically; in the second case, the zones might be of very different ages and perhaps unrelated in origin.

The several rocks of the steatite and blackwall association (steatite, tremolite, blackwall chlorite rock, albite porphyroblast rock) are contemporaneous and are genetically related. This conclusion is based upon the fact that nowhere do rocks of an inner zone cut out the rocks of an outer zone if the composition of the country rock is favorable for the development of all the zones. The conclusion is supported by petrologic arguments which are essentially the same as those presented by Chidester (1962, p. 91–93) for deposits elsewhere in Vermont.

The rocks of the tremolite rock and chlorite rock association are inferred also to be of contemporaneous origin on the basis of their inferred similarity and genetic equivalency to the rocks of the steatite and blackwall association. Observed relations in the relatively scarce exposures of this association support this conclusion.

Though all are older than the steatitization, the rocks of the rodingite and serpentine-chlorite rock association show intersecting relations which indicate that, rather than being essentially contemporaneous, some of the rocks were separated in time of formation by a significant but probably small interval. The cutting out of rodingite by serpentine-chlorite rock at some places next to amphibolite, and the projection of serpentine-chlorite rock beyond the outer limits of rodingite zones along joints that intersect the zones, strongly suggest that the serpentine-chlorite rock is entirely younger than the rodingite. The intersection of rodingite by calcite-chlorite-magnetite veins shows that the veins are younger than the rodingite, but allows them to be the same age as serpentine-chlorite rock. The apparent absence of a serpentine-chlorite rock zone at the west side of the Eden quarry may be due to locally diffuse and inconspicuous development of the serpentine-chlorite rock zone. On the basis of the association of rodingite and serpentine-chlorite rock with serpentinite and dunite, and of the observed age relations of the rodingite and serpentine-chlorite rock, logical conclusions are that the rodingite may be related to the stage of emplacement of ultramafic rocks that were appreciably hotter than the surrounding country rocks, and that the serpentine-chlorite rock and perhaps also the calcite-chlorite-magnetite veins are probably wholly related to serpentinization.

Layering in contact rocks, which is observable commonly in rodingite and locally in serpentine-chlorite rock, blackwall, and steatite, is entirely inherited from the protoliths, though considerably modified by metasomatism. The distribution and kind of relict minerals, the similarity in appearance between layering in contact rocks and that in adjacent country rock, and the existence locally of layers that can be traced across one or more contact zones into adjacent country rock, all attest to the relict character of the layering. Where contacts are crosscutting and layers can be traced from one zone to another, absence of discontinuity or of systematic distortion of layers at contact between zones indicates that replacement was on a volume-for-volume basis (see Chidester, 1962, p. 93–94).

Chemical changes that took place in the blackwall and steatite reaction are discussed in detail in an earlier report (Chidester, 1962, p. 122–124) and are not repeated here. Table 9 assembles analyses of suites of specimens of the rodingite and serpentine-chlorite rock association and of the tremolite rock and chlorite rock association. The analyses are presented in terms of composition of equal volumes of rock (the modified standard cell) so that chemical changes can be deduced for reactions that took place on a volume-for-volume basis. These changes are considered in the appropriate sections ("Rodingite," "Serpentine-chlorite rock," and "Rocks of the tremolite rock and chlorite rock association") below.

RODINGITE

Rodingite formed by replacement and reconstitution of country rock bordering ultramafic bodies. That the rodingite formed indiscriminately from schist, gneiss, and amphibolite is evident from its transitional relations to these rocks and from paragenetic relations of the minerals. The striking development of rodingite in schist and gneiss along the northeastern contact of the Lowell quarry body, and the relatively inconspicuous development in amphibolite at a nearby contact, indicate that in comparable structural settings the composition of the country rock exercises major control in the rodingite reaction.

The paragenetic relations of the minerals distinguish two stages in the formation of the rodingite. The first stage is characterized by alteration of mafic minerals to diopside, clinozoisite, and predominantly birefringent garnet; the second stage is characterized by replacement of diopside, clinozoisite, and other minerals by isotropic garnet and vesuvianite. The two stages were probably not appreciably separated in time; the relations require only that stage one preceded stage two at any particular point. The distribution of the minerals indicates that the two

TABLE 9.—*Chemical composition and calculated gains or losses of equal volumes of rocks of contact sequences*

[Chemical composition is in terms of the contents of the modified standard cell (MSC). (Chidester, 1962, p. 95, 132–137). Dashed line, not determined. Analyses from table 1.]

Specimen No.	Rock type	Zonal position	Si	Al	Fe^{3+}	Fe^{2+}	Mg	Ca	Na	K	Ti	P	Mn	C^{4+}	Total cations[1] in MSC	H	O
A. Rocks of the rodingite and serpentine-chlorite rock association adjacent to amphibolite																	
AV-78(VT-13)	Serpentine-chlorite rock.	Adjacent to intrusive contact	31.1	11.1	1.6	0.38	50.7	0.34	0.00	0.00	0.56	0.13	0.29	0.14	96.4	74.9	172.2
TPT-Vt-11(VT-11)	do	Adjacent to rodingite	27.2	15.9	2.3	5.6	40.4	.25	.00	.00	.65	.09	.30	.16	93.0	70.8	165.7
TPT-Vt-10(VT-10)	Rodingite	Center of rodingite	47.0	18.4	1.9	10.5	13.6	23.3	.43	.00	.96	.17	.21	.08	116.3	23.2	186.5
AV-78(VT-12)	Fine amphibolite	Country rock	55.6	20.4	2.0	4.8	6.1	10.9	9.6	.40	.70	.14	.23	.07	110.9	6.1	176.7
A-BM-53-140	do	do	52.3	18.7	1.7	7.6	12.7	13.2	4.2	.20	.79	.08	.19	.01	111.8	17.7	181.9
A-BM-53-98	Coarse amphibolite	do	48.4	17.1	4.0	7.6	17.4	13.0	4.0	.34	.8	.15	.22	<.07	113.2	15.8	178.9
A-BM-53-139	do	do	47.2	20.1	3.4	10.1	15.4	12.6	4.5	.84	.96	.01	.27	.01	115.7	20.4	183.2
		Gain (+) or loss (−) per MSC in alteration of:															
		1. VT-12 to VT-10	−8.6	−2.0	−.1	+5.7	+7.5	+12.4	−9.2	−.40	+.26	+.03	−.02	+.01	+5.4	+17.1	+9.8
		2. A-BM-53-140 to VT-10	−5.3	−.3	+.2	+2.9	+.9	+10.1	−3.8	−.20	+.17	+.09	+.02	+.07	+4.5	+5.5	+4.6
		3. A-BM-53-98 to VT-10	−1.4	+1.3	−2.1	+2.9	−3.8	+10.3	−3.6	−.34	+.16	+.02	−.01	--	+3.1	+7.4	+7.6
		4. A-BM-53-139 to VT-10	−.2	−1.7	−.5	+.4	−1.8	+10.7	−4.1	−.84	0	+.16	−.06	+.07	+.6	+2.8	+3.3
		5. VT-10 to VT-11	−19.8	−2.5	+.4	−4.9	+26.8	−23.1	−.43	0	−.31	−.08	+.09	+.08	−23.3	+47.6	−20.8
		6. VT-11 to VT-13	+3.9	−4.8	−.7	−5.2	+10.3	+.09	0	0	−.09	+.04	−.01	−.02	+3.4	+4.1	+6.5
B. Rocks of the rodingite and serpentine-chlorite rock association adjacent to schist																	
AR-110(VT-20)	Serpentine-chlorite rock.	Adjacent to rodingite	30.8	9.7	1.5	0.75	49.8	1.1	0.00	0.00	0.62	0.16	0.29	1.2	96.0	75.9	175.6
AR-110(VT-21)	Rodingite	Adjacent to serpentine-chlorite rock.	55.0	20.1	2.7	2.9	8.2	20.9	.10	.02	.70	.20	.23	<.07	111.2	14.5	185.8
AR-110(VT-22)	do	8 cm from serpentine-chlorite rock.	60.8	18.2	2.3	3.3	5.4	14.6	.27	.64	.59	.19	.18	<.07	106.7	6.3	181.2
AR-111(VT-23)	Graphitic schist	Country rock	51.1	19.7	1.3	5.2	4.8	5.2	3.1	4.5	.68	.25	.12	<.07	96.0	18.7	163.8
		Gain (+) or loss (−) per MSC in alteration of:															
		1. VT-23 to VT-22	+9.7	−1.5	+1.0	−1.9	+.6	+9.4	−2.8	−3.9	−.09	−.06	+.06	--	+10.7	−12.4	+17.4
		2. VT-22 to VT-21	−5.8	+1.9	+.4	−.4	+2.8	+6.3	−.17	−.62	+.11	+.01	+.05	--	+4.5	+8.2	+4.6
		3. VT-21 to VT-20	−25.0	−10.4	−1.2	−2.2	+41.6	−19.8	−.10	−.02	−.08	−.04	+.06	>1.1	−15.2	+61.4	−10.2
C. Rocks of the tremolite rock and chlorite rock association																	
AV-209(VT-16)	Serpentinite	Adjacent to tremolite rock	37.2	1.2	1.0	3.9	50.1	0.80	0.00	0.00	0.01	0.01	0.06	1.13	95.5	68.0	169.0
AV-205(VT-17)	Tremolite-chlorite rock.	Adjacent to serpentinite	55.1	3.0	.4	3.4	37.0	12.5	.04	.00	.01	.01	.09	.15	111.9	21.3	179.5
AV-206(VT-18)	Chlorite-tremolite rock	Adjacent to unaltered amphibolite.	31.9	15.6	.8	9.0	33.2	5.8	.02	.00	2.0	.33	.15	2.4	101.3	61.0	176.9
AV-208(VT-19)	Coarse amphibolite	45 cm from chlorite-tremolite rock.	44.5	17.2	1.1	11.2	21.6	7.8	3.2	.13	.94	.10	.26	<.08	107.9	34.9	177
		Gain (+) or loss (−) per MSC in alteration of:															
		1. VT-16 to VT-17	+17.9	+1.8	−.6	−.5	−13.1	+11.7	+.04	.00	.00	.00	+.03	−.98	+16.4	−46.7	+10
		2. VT-19 to VT-18	−12.6	−1.6	−.3	−2.2	+11.6	−2.0	−3.2	−.13	+1.1	+.23	−.11	+2.4	−6.6	+26.1	−0.9

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[1] The total cations per modified standard cell (MSC) is, by definition, equal to $100 \times F_c$, and is here rounded to one decimal place. Because of rounding procedures, the total does not necessarily equal the sum of the cations shown to the left of the total.

stages of alteration proceeded outward from the contacts of the ultramafic bodies or—particularly in the case of the coarse rodingite—from fractures or fracture zones adjacent to the bodies.

Minerals of coarse rodingite—garnet, vesuvianite, clinozoisite, zoisite, diopside, calcite, and locally, prehnite—formed largely by fracture filling, and the two stages distinguishable in the fine rodingite are not evident except locally at the margins of fissures.

The fabric of fine rodingite is considerably different from that of the protolith. Alteration to diopside resulted generally in a reduction in grain size with respect to the protolith. Development of garnet and vesuvianite imparted a massive character to the rock. Clinozoisite, most relict minerals, and the titanium minerals retain generally the textural features and distribution pattern of the protolith.

The first stage in the development of rodingite was marked by mineralogical changes and little or only moderate change in composition. Diopside formed almost solely by alteration of mafic minerals, chiefly and perhaps entirely hornblende; the relation is shown by the incomplete alteration of hornblende to diopside along cleavages and at margins of grains, at many places in the transition zone between rodingite and country rock. Much, probably most, clinozoisite is a relict of the protoliths, but most was appreciably modified in composition (the content of iron reduced) during alteration. Only cores of the larger rounded grains appear to retain their original somewhat more ferriferous composition.

The second stage was marked by greater compositional changes. Garnet and vesuvianite formed by replacement of a variety of minerals of contrasting composition—diopside, clinozoisite, albite, and other relict minerals of the country rock. Scalloped boundaries of the garnet and vesuvianite, convex toward the rest of the minerals, and inclusion within them of diopside, clinozoisite, and other minerals, are evidence of the mode of origin.

The abundance locally of island remnants of garnet, diopside, epidote, and vesuvianite in areas of chlorite, and the persistence with which chlorite invades and embays both those minerals and the minerals of the country rock, indicate the replacement origin of much of the chlorite. On the other hand, the occurrence in a few places of chlorite in contact with diopside, clinozoisite, garnet, and vesuvianite, but not embaying or invading them, suggests the compatibility and contemporaneous origin of the minerals concerned. There is no direct evidence of the age of chlorite that replaces the country rock minerals, relative to chlorite that replaces the minerals of the rodingite. The simplest interpretation appears to be that the chlorite in apparent equilibrium with diopside and other minerals of the rodingite, and the chlorite that replaces the minerals of the country rock, are contemporaneous; chlorite that replaces the minerals of the rodingite is distinctly later.

In the development of rodingite, the titanium minerals appear to follow the progressive alteration of rutile to ilmenite to sphene, for sphene is greatly predominant, rutile is present only as a core surrounded successively by shells of ilmenite and sphene, or ilmenite forms the core alone. Some sphene, particularly euhedral grains, possibly originated in the alteration of slightly titaniferous hornblende to less titaniferous diopside. If so, the proportion is small, inasmuch as protolithic amphibolite contains about the same amount of titanium minerals as rodingite.

Prehnite mostly fills tiny fractures in fine rodingite. In rare instances where it is intermixed with other groundmass minerals, prehnite presumably formed by replacement, possibly of albite.

Other minerals in rodingite are relicts of the protoliths.

Increase in content of Ca is the dominant chemical change in the alteration of both amphibolite and schist to rodingite; the content of Na and K also decreased consistently (table 9, sections *A* and *B*). Other changes show no significant pattern. Inferences concerning changes in content of (OH) are unreliable because of widespread retrograde alterations (particularly chloritization) that affect both rodingite and amphibolite.

Assemblages in the rodingite are of apparently higher metamorphic grade than those of the schist and amphibolite, but the narrowness of the rodingite zone indicates that factors leading to formation of rodingite were limited to a narrow zone at margins of the larger ultramafic bodies. The possible controlling factors in the narrow zone include increased temperature, decreased partial pressure of water, and influx of Ca. Increased temperature in a narrow zone surrounding the ultramafic body may have resulted from the ultramafic rocks being hotter than surrounding country rock at the time of emplacement. A lowered partial pressure of water immediately adjacent to the ultramafic body may be attributed to the desiccating effect of dunite in the larger bodies. Influx of Ca may have resulted from the outward migration of Ca from the ultramafic rocks as a result of serpentinization (see table 7, section *A–D*).

SERPENTINE-CHLORITE ROCK

Serpentine-chlorite rock formed almost entirely by replacement of rodingite. Locally, along small faults that intersect the rodingite zone and where serpentine-chlorite rock transects the rodingite zone, serpentine-chlorite rock replaces amphibolite directly, but such places are relatively rare, and the amount of serpentine-chlorite rock so formed is small. These conclusions are evident from relations of the rocks, relict minerals, and preservation of a distinctive pattern of former grains and patches of garnet and vesuvianite at many places in the serpentine-chlorite rock.

Almost the sole reaction in formation of serpentine-chlorite rock is wholesale conversion of the minerals of rodingite—or, in rare instances, amphibolite—to a mixture of chlorite and serpentine. The only exception is a change involving the titanium minerals sphene, rutile, and ilmenite.

The predominance of chlorite in the outer part (next to rodingite) of the serpentine-chlorite rock zone and of serpentine in the inner part (next to the ultramafic body) indicates that at all stages during development of the zone chlorite predominated over serpentine near the interface with rodingite, and serpentine over chlorite near the interface with serpentinite. The initial stage at any particular point was thus marked by alteration of minerals of the rodingite to chlorite and minor serpentine. As the zone widened, more and more of the chlorite must have been converted to serpentine, so that the end product was a zone composed at the inner margin predominantly of serpentine and of only minor chlorite, and at the outer border, predominantly of chlorite and of only minor serpentine.

During formation of serpentine-chlorite rock, both sphene and rutile altered to ilmenite, for both are absent, whereas ilmenite is present. The relation of single titanium-mineral grains at the contact between rodingite and serpentine-chlorite rock (see section on "Serpentine chlorite rock" under "Contact rocks" in the section on "Ultramafic and associated rocks"), which consist of sphene on the rodingite side and of ilmenite on the serpentine-chlorite side of the contact, further supports this conclusion. An increase in total content of titanium minerals in serpentine-chlorite rock as compared with rodingite requires a source outside the titanium minerals. Diopside, garnet, and vesuvianite are possible sources of both the required iron and titanium.

Table 9, sections *A* and *B*, show that the dominant and only consistent chemical change was a severalfold increase in content of Mg. Other changes vary with composition of the country rock and reflect chiefly decreases in content of Al, Fe^{+3}, Ca, Na, and K imposed by the limiting compositional ranges of chlorite and serpentine.

CHLORITE-CALCITE-MAGNETITE VEINS

Veins of chlorite, calcite, and magnetite associated with rodingite and serpentine-chlorite rock are younger than rodingite, which they crosscut. They are also at least partly younger than serpentine-chlorite rock, for locally they crosscut it, but they are virtually contemporaneous with the serpentine-chlorite rock.

The minerals of the veins formed largely by fissure filling, as is shown by the euhedral form of much of the chlorite and magnetite, the common orientation of columnar calcite and elongated books of chlorite normal to the vein walls, and the geometry of the veins—locally wedge form, generally smooth walls, matching irregularities in opposite walls (though rare). The generally small scale of irregularities indicative of replacement implies that replacement was of minor importance in formation of the veins.

These relations, together with petrologic considerations based on interrelations between serpentine veins, pervasively serpentinized dunite, rodingite, and serpentine-chlorite rock (see section on "Serpentine veins" under "Petrogenesis" in the section on "Ultramafic and associated rocks" and in the same section see "Serpentine-chlorite rock" under "Contact rocks"), suggest that chlorite-calcite-magnetite veins formed essentially contemporaneously with serpentine veins. These relations also suggest that the constituents were derived from excess cations (chiefly Mg, Fe, Ca, Al, Si) displaced from wallrock adjacent to the serpentine veins and from that part of the pervasive serpentinization of dunite that accompanied formation of the serpentine veins.

ROCKS OF THE STEATITE AND BLACKWALL CHLORITE ROCK ASSOCIATION

The geologic relations, relict minerals, and textural features indicate that blackwall and steatite are contemporaneous and that blackwall formed by replacement of the schist bordering the serpentinite, whereas the steatite formed chiefly by replacement of serpentinite or talc-carbonate rock but partly by replacement of blackwall. Where the schist contained sufficient calcic and mafic minerals, a tremolite zone formed between the steatite and blackwall; where it was albitic, an albite porphyroblast rock zone formed at the outer margin of the blackwall zone. By analogy, it seems likely that a relatively high content of muscovite in the country rock may have led to heavy concentration of muscovite at the outer part of the blackwall zone.

The textural relations of the minerals indicate that albite porphyroblast rock has replaced the schist, blackwall has replaced the albite porphyroblast rock, and steatite has replaced the blackwall. Such relations clearly indicate sequential development of the rocks. These relations and the contemporaneity of the zones indicate that the zones formed simultaneously by outwardly migrating reactions, so that the outer part of each zone is younger than the inner part of the next zone outward.

Tremolite rock in the blackwall-steatite association replaces both blackwall and steatite at either margin. It therefore widened at the expense of both and differs from other rocks in the association in that at its outer margin it is younger than the inner part of the blackwall and at its inner margin it is younger than the outer part of the steatite. The zone as a whole, however, is the same age as the others.

Chlorite, the sole essential constituent of the blackwall, formed by replacement of all the minerals of the schist. Albite formed in the adjacent albite porphyroblast zone by enlargement of preexisting crystals of albite and by growth of new crystals; concentrations of muscovite in the same relative position probably formed similarly. Tremolite grew by replacement both of chlorite in the blackwall and of talc in the steatite. Talc in the steatite zone formed chiefly by replacement of serpentine, in some places directly but in most places with an intermediate stage of formation of talc-carbonate rock. Where tremolite rock is absent, talc also replaced chlorite in the blackwall zone for a short distance beyond the original contact of the ultramafic rock.

In the blackwall, albite porphyroblast rock, and tremolite rock zones the titanium minerals underwent a progressive alteration from ilmenite→rutile→sphene, illustrated at many places by cores of ilmenite surrounded successively by shells of rutile and sphene and by the great preponderance of sphene over ilmenite and rutile in general. The decrease in content of sphene in the alteration of blackwall to steatite suggests that in the alteration, appreciable titanium was incorporated in talc.

ROCKS OF THE TREMOLITE AND CHLORITE ROCK ASSOCIATION

Petrogenic relations of the tremolite and chlorite rock association are somewhat doubtful, partly because the rocks are unevenly distributed and erratically exposed. Chlorite rock replaces amphibolite bordering serpentinite, and tremolite replaces the chlorite rock. Tremolite may replace serpentinite, but evidence was not seen. The tremolite and chlorite rock association is inferred to be equivalent to the steatite and blackwall association in an amphibolite environment, and, by analogy, tremolite rock is inferred to be the same general age as chlorite rock.

OTHER ASSOCIATIONS

Irregular alteration zones in country rock adjacent to masses of talc-carbonate rock and carbonate-quartz rock in dunite superficially resemble the blackwall assemblage but are probably related to carbonatization rather than to steatite-blackwall alteration. Because exposure is inadequate, precise relations are not known. However, the general association of such alteration zones with talc-carbonate and carbonate-quartz masses in dunite, and petrologic considerations as discussed before (in section on "Steatite, talc-carbonate rock, and carbonate-quartz rock" under "Petrogenesis" in section on "Ultramafic and associated rocks") suggest that the alteration of country rock in these zones resulted from influx of material displaced from dunite when the dunite was altered to talc-carbonate rock and carbonate-quartz rock. The influx of constituents (principally Mg, Si, Al, and Fe) resulted in outward displacement of constituents of the schist, forming crudely zoned but irregularly distributed concentrations of talc, chlorite, sericite, and albite.

STRUCTURE

The structure of the Belvidere Mountain area is dominated by folds and by intrusive bodies of ultramafic rock, whose combined effect imparts a complicated sinuous pattern to the lithic units (pls. 1, 2, and 3). Faults play a comparatively inconspicuous role, being apparent chiefly in minor offsets in contacts of ultramafic bodies. In addition to major structures, a wide variety of minor structural features—layering, foliations, planar features, warped surfaces, and lineations—impart distinctive characteristics to the rocks and bear consistent and significant relations to each other and to major structures.

MAJOR STRUCTURAL FEATURES

On the basis of major structural trends (that is, those evinced by map patterns), the area is divided into three parts, each having characteristic rock associations. In the northeastern part of the area, which consists chiefly of metamorphosed sedimentary and volcanic rocks, structural trends are predominantly about east. In the southwestern part, which also consists chiefly of metamorphosed sedimentary and volcanic rocks, the trends are about north. In the zone between these two parts, which consists chiefly of intrusive ultramafic rocks, structural features of the intrusive bodies are prominent, and structural trends are diverse (pl. 1).

The bodies of ultramafic rock are grossly conformable with the country rock formations and with intervening contact rocks. The configurations of the bodies vary in complexity with size. The smaller bodies show chiefly intrusive structures and are simple lenses. The large bodies show both intrusive structures and two sets of folds and are complexly warped.

Two sets of major folds emerge from the pattern of diverse and contrasting structural trends. In the northeastern part of the area, the two sets of folds are markedly divergent, and east- to southeast-trending folds predominate greatly over southwest-trending ones. Southwestward, both sets warp around more and more into the south, and the two sets become less and less divergent, so that in the southwesternmost part of the area they are nearly parallel and trend about north.

The character and style of the two sets of major folds (hereafter referred to simply as "southeast-trending" and "southwest-trending") and the nature of the relation between the two sets are not known with certainty, because conditions of exposure are inadequate to determine precisely the map pattern of the rock units. Differences in style are not striking, but the southeast-trending folds generally are compressed and attenuated, and the southwest-trending ones are somewhat more open.

Faults and shear zones in the area commonly do not intersect major folds, and so the relations of the one to the other cannot be determined unequivocally. Both faults and shear zones are unaffected by folding, and both intersect and offset contacts of intrusive bodies.

ULTRAMAFIC INTRUSIVE BODIES

Most of the major structural features of the ultramafic bodies, such as shape, layering, inclusions and septa, and schistose zones characterized by shear polyhedrons, are independent of the adjacent country rocks, but folds and shear zones are shared with the country rocks. In general, features of the ultramafic bodies are intersected by one or more sets of the folds and shear zones.

SIZE AND SHAPE

The ultramafic intrusive bodies are lenses for which the ratio of strike length to thickness appears to be roughly on the order of 5:1. Three-dimensional relations can be determined directly only for the main ultramafic body. In it, the intermediate axis is about parallel to the strike of the principal plane; the major axis is about downdip and roughly twice the length of the intermediate axis. The geometric relations of the other ultramafic bodies are inferred to be generally similar.

The main ultramafic body is bordered on the southwest and northeast by southeast-trending synclines; between them, the now eroded upper contact of the body forms a broad anticlinal arch trending southeast about coincident with the major axis of the lens. Small sharp warps in the lens marked by infolds of Belvidere Mountain Formation (Ꞓbf)—one just southwest of the Lowell quarry, a second near 103,500 N., 93,000 E. on the northeast flank of Belvidere Mountain, and a third between the main peak and the west peak of Belvidere Mountain (pl. 1)—are produced by synclinal folds of the southeast-trending set. The folds that bring the main ultramafic body to the surface in an anticline in the vicinity of Corez Pond, and probably the broad syncline that warps the body below a cap of amphibolite at the crest of Belvidere Mountain, belong to the southwest-trending set. In overall attitude, the main body strikes about northeast and dips gently to moderately southeast.

CONTACT RELATIONS

Intrusive contacts of the small ultramafic bodies are almost entirely conformable; those of the main ultramafic body are generally grossly conformable with the adjacent map units of country rock, but they are slightly to moderately crosscutting in many places. These contact relations are shown at various scales in figures 3 and 4, and plates 1–4. Both the strictly conformable and the crosscutting relations are best shown in the vicinity of the Lowell quarry, in the area of the tenuous connection between the Lowell quarry body and the Corez Pond body, and at several places along both the northeast and the southwest contacts. Abundant isolated inclusions and projecting septa of wallrock in the serpentinite further demonstrate the prevalence of slightly crosscutting relations in the Lowell quarry body.

INCLUSIONS AND SEPTA

Inclusions and septa of wallrock are exposed at many places in the Lowell quarry body and at a few places in the less well exposed Eden quarry body. A large projecting septum of coarse amphibolite separates two clawlike prongs of ultramafic rock at the northwest end of the Lowell quarry body near 101,000 N., 96,000 E. (pls. 1 and 3). Lenses of coarse amphibolite extend from this septum southeastward in the ultramafic body in an irregular and discontinuous pattern toward another septum that projects from the northeast wall near the southeasern end of the quarry. Near 99,900 N., 98,300 E. (pl. 3, this septum becomes glued tight to the schist, but about 100 m farther southeast it is separated from the schist and is breached by a thin tongue of serpentinite that projects from the Corez Pond body and connects with the Lowell quarry body. A large slab of schist and gneiss projects conspicuously from the northeast contact near the center of the Lowell quarry (pl. 3). Several small septa of amphibolite are exposed near the southeast contact of the Eden quarry body (pl. 1).

Small pendulous or lenticular masses of amphibolite, completely isolated within the serpentinite or joined to wallrock by only a narrow neck are a distinctive feature of the southwest contact of the Lowell quarry body between 100,000 N., 97,400 E., and 99,800 N., 97,600 E. (pl. 3). The contacts of these masses are conformable with bedding schistosity except in narrow necked-off parts of a mass and at faults, and are conformable with thin zones of serpentine-chlorite rock and rodingite that border the masses. Some of the masses may be nearly cigar-shaped in plan, but most or all are probably elongated vertically, so that their shape is roughly like that of an airplane wing (pl. 4*B* and *C*). These features are discussed further in the section "Folds in ultramafic rocks."

LAYERING

The overall pattern of layering in the ultramafic rocks is known only sketchily for most of the ultramafic bodies because of both insufficient outcrop and insufficient detailed study outside the active quarry areas. Scattered observations on the eastern slope of Belvidere Mountain and detailed mapping in the C-area (pl. 1) suggest that the layering there is very roughly conformable with the intrusive contacts. Layering in the dunite near the center of the Lowell quarry is markedly unconformable with the contacts of the body. In this large mass of dunite, the layering has a mosaic pattern in which the units are irregular fault blocks a few tens to a few hundred feet across (pl. 3).

SHEARING

The outcrop pattern of shearing in the ultramafic bodies ranges from essentially linear tabular shear zones (fig. 3 and pl. 3) to very irregular schistose serpentinite masses characterized by shear polyhedrons (pls. 3 and 4*A*). In places, tabular shear zones intersect and can be traced through irregular zones of schistose serpentinite. Elsewhere, the shear zones can be traced only a short distance into the irregular schistose areas, where they lose their identity (fig. 3 and pl. 4*A*, *B*, and *C*).

Some of the shear zones are about parallel to the northeast-trending set in the adjacent wallrock, and some—such as those in the dunite mass near the center of the Lowell quarry—have regularities of pattern throughout areas as much as a few hundred feet across. Many are diversely oriented, however, and no general pattern is discernible for the area as a whole. The irregular zones of sheared serpentinite are generally peripheral to the bodies. In many places, these general relations are obscured by irregularities.

SOUTHEAST-TRENDING MAJOR FOLDS

Several folds of the southeast-trending set are distinguishable in the area. Most conspicuous is an east-trending syncline that parallels the eastern half of the northern border. Southwestward, successive synclines have increasingly southeast trends (compare the synclines northeast and southwest of the Eden quarry body). A syncline west of Scofield Ledges, near the 89,000 E. grid line (pl. 1), whose trough is marked by a long-narrow, north-trending belt of fine amphibolite, probably also belongs to this set of folds, but it cannot be distinguished with certainty from the southwest-trending set. These four narrow synclines are separated by appreciably wider anticlinal tracts. Other, smaller folds in the southwestern part of the area may also belong to this set. All folds of the set tend to converge in plan toward the northwest corner of the area.

Folds of the southeast-trending set are nearly isoclinal; axial planes are about vertical. The gross map pattern requires the fold axes to plunge gently to the east, southeast, or south. Except near the main ultramafic body, second-order features in the fold pattern are inferred from generally inadequate outcrop data. Most of those shown in figure 3 are largely interpretive, but none show mirror symmetry with respect to the axial planes of the major folds. Rather, the pattern is the same in the region of the axial plane and in both limbs of the major fold.

SOUTHWEST-TRENDING MAJOR FOLDS

Folds of the southwest-trending set are subordinate to those of the other set. The principal ones are the synclinal and anticlinal folds at the southeast margin of the Eden quarry body, responsible for the complex map pattern in the vicinity west of Corez Pond (pl. 1); the folds in the contact between the Ottauquechee and Stowe along the southeastern margin of the outcrop area of the Ottauquechee; the broad synclinal fold in the amphibolite that caps Belvidere Mountain; and the folds that impart a zigzag pattern to the northeastern contact of the Eden quarry body and to the nearby formations to the northeast.

The folds are moderately open in style and have steep axial planes. The plunges of the axes are variable, depending largely upon relations to the southeast-trending folds and to the contacts of the main ultramafic body, but the predominant plunge is gently to moderately south or southwest. Representation of the detailed configuration of the folds is to a considerable extent interpretive, though appreciably less so than in the case of the southeast-trending folds, but it is reasonably clear that the secondary fold patterns have mirror symmetry with respect to the axial planes of the major folds.

FAULTS

Observed faults, minor in their extent and structural effect, are confined virtually to the vicinity of the ultramafic bodies, but such location may be largely due to conditions of exposure. A small fault at the northeast margin of the Eden quarry body just west of the Lowell quarry (pl. 3), exposed for a map distance of about 100 m and inferred to extend along the contact for an additional 400 m, probably has a displacement of not more than a few tens of meters. The fault trends northwestward and dips moderately to the southwest.

Narrow shear zones a few tens of meters to more than a hundred meters long are abundant in the vicinity of the ultramafic bodies (pl. 3). Most of them belong to a set that trends a little east of north and is nearly vertical. Others are diversely oriented, commonly irregular, and dip moderately to steeply. Though all types occur in both the ultramafic bodies and in the country rock, the set that trends east of north predominates in the country rock, the diversely oriented ones predominate in the ultramafic bodies.

The linear belts of talc-carbonate rock that trend northward within the Eden quarry body near the 94,000 E. mine coordinate may mark fault zones, but there is no direct evidence to so identify them.

MINOR STRUCTURAL FEATURES

Minor structural features (which are represented on maps by a structural symbol) in the area comprise a wide variety of foliations, structural surfaces, zones of distributed shearing, folds, and lineations. For these, the petrographic aspects have been described under the section on "Rocks." In the present section, the geometric relations of the structural elements are described.

The terminology is for the most part that commonly used, but a statement of usage for some terms may be desirable (see Chidester, 1962, p. 17). *Bedding schistosity* is everywhere parallel to bedding, even in the noses of small isoclinal folds; it is generally, perhaps always, a *continuous schistosity,* in which all platy minerals are uniformly parallel. *Transverse* or *transecting schistosity* transects bedding (and bedding schistosity) in suitable structural situations, such as in the limbs of folds that diverge sufficiently from the axial planes, and in the noses of isoclinal folds. Transverse schistosity is always a *spaced schistosity* which consists of discretely spaced surfaces of flexure, fracture, or shear that mark discontinuities in the fabric of the rock. *Slip cleavage* is a transverse schistosity associated with clearly identifiable flexures or crinkles in an older schistosity. *Fracture cleavage* consists of closely spaced fractures without significant alinement of minerals along them. Under field conditions, transverse schistosity is not always clearly and readily identifiable as either slip cleavage or fracture cleavage; indeed, it may appear megascopically to be a continuous schistosity. Microscopically, however, it is always identifiable as a spaced schistosity.

In much of the ultramafic rock, schistosity cannot be classified on the basis of intersecting relations and must be classified simply as continuous or as spaced schistosity.

The associations and intersecting relations of the minor structural features are systematic, and their spatial patterns are of differing complexity (homogeneity and symmetry) at the scale of normal megascopic observations and for the field of the entire area. In general, the simpler the spatial pattern of a structural element, the larger the minimum domain for which the element is penetrative (for example, the wider the spacing of a regularly spaced schistosity) and the fewer the structural elements that intersect it. From these relations, it follows that the simpler the spatial pattern of a structural element, and the larger the minimum domain for which the element can be considered to be penetrative, the younger the element.

On the basis of these relations, many of the classes of structural features can be divided into two or more sets of different age, such as older and younger folds and related structural elements, and several ages of schistosity.

As a general rule, minor structures in the country rock cannot be related directly to those in the ultramafic rocks, including those that involve the contacts, because of the lack of suitable elements in common. Similarities in style and pattern, and other relations, permit indirect correlation. Because they are not directly related, however, the structural features in the two kinds of rocks are best treated separately.

BEDDING

Bedding commonly can be traced for as much as several tens of meters or the entire extent of an outcrop in quartzite, in highly quartzose schist and gneiss, and in much of the amphibolite. In the less quartzose thinly laminated schists, the bedding is similarly distinct and traceable in favorable structural settings, but generally it is traceable for only a few centimeters or is virtually indistinguishable, particularly in the highly graphitic schist and in the chloritic schist characteristics of the Stowe. Individual bedding laminations appear to be parallel to larger bedding units in most places but diverge slightly where facies relations are evident, as in the area depicted on plate 4*D*.

The attitude of bedding varies greatly, from virtually planar through simply folded to intricately contorted. Individual measurements of bedding attitudes throughout the area range virtually through 360° in strike and vary from horizontal to vertical in dip. In the northeastern part of the area, strikes deviate only moderately from east, and dips are moderate southward. In the southwestern part of the area, strikes of bedding are predominantly a little either side of north; dips are steeply either side of vertical. The central wedge-shaped area has more diverse bedding patterns and no predominant trend.

BEDDING SCHISTOSITY

Bedding schistosity is best developed in rocks that have distinctive micaceous layers, but it has commonly been obscured or obliterated by intersecting structural features. The spatial pattern of bedding schistosity is, of course, identical with that of bedding.

LAYERING IN ULTRAMAFIC ROCKS

Layers from a centimeter to as much as half a meter thick are conspicuous in most of the dunite and periodotite and in much of the massive serpentinite. The layering results from differences in color, texture, and mineralogy.

The layering is nearly planar or only moderately warped in most places, but at several places in the C-area it is folded in simple patterns on a small scale. At the many small faults that offset the layering by a few centimeters, the layering generally differs appreciably in attitude on either side of the fault. Layering is commonly transected at a high angle by zones of schistose serpentinite, but the relations are variable, and in places the boundaries of schistose zones are parallel to layering.

Observed attitudes of layering are diverse and conform to no simple pattern.

DIMENSIONAL ORIENTATION OF PRIMARY MINERALS IN DUNITE AND PERIDOTITE

Most of the primary minerals in the ultramafic igneous rocks are about equidimensional; therefore, dimensional parallelism of minerals in these rocks is seldom noticeable. In places, pseudomorphs after pyroxene and rare relict grains of pyroxene are of tabular habit; such tabular grains are generally alined roughly parallel to the layering.

Preferred crystallographic orientation of olivine in dunite is shown in a few thin sections by crude mass parallel extinction. The olivine is commonly alined with (010) cleavage parallel to preferentially serpentinized layers, which locally have a parallel continuous schistosity. The macroscopic relations of this foliation in the dunite and peridotite are not known, but these few microscopic observations suggest that the foliation may be parallel to layering.

SHEAR POLYHEDRONS

Shear polyhedrons (Chidester and others, 1951, p. 7) consist ideally of closely packed polyhedral units of relatively unsheared massive serpentinite surrounded by generally thin zones of highly sheared schistose serpentinite. The polyhedral units commonly present an irregular pattern, and the schistosity in the enclosing thin sheared zones conforms with the outlines of the units. In many places, the polyhedrons are distinctly triaxial, their principal planes are crudely alined with the zone of shearing, and the schistosity tails off at the ends of polyhedrons tangentially to their principal planes. These relations are complicated and obscured in many places by the intersection of distributed shears related to small faults.

FOLDS

Minor folds are prevalent in both the metamorphosed sedimentary and volcanic rocks and the ultramafic rocks. In both, the folds can be grouped into several sets on the basis of style and interrelations, and, in both, the form surfaces (McIntyre and Weiss, 1957, p. 149; Weiss, 1959, p. 16) that define the folds can be divided into types that differ in kind, particularly in tectonic relations.

FOLDS IN THE METAMORPHOSED SEDIMENTARY AND VOLCANIC ROCKS

Minor folds in the metasedimentary and metavolcanic rocks may be classified principally into two groups on the basis of style and intersecting relations: an older set of tightly compressed plastic folds, and a younger set of more open brittle-style folds which refold the older folds where the geometric relations are appropriate. Both these sets of folds range in amplitude from a few centimeters to as much as several meters and involve generally a considerable number of beds. A third type of fold is consistently on the order of a centimeter in amplitude and is shown only by the pattern of quartzose layers. These folds are plastic in style, and the limbs are tightly compressed. The distinctness of the fold form varies widely, so that in places the feature is distinguishable only as quartz rodding.

In general, the axes of the older folds and of the folded quartz lenses are at a large angle to the axes of nearby younger folds. Some of the older folds are complex in form, and their structural elements vary markely in attitude. The younger folds, though also variable, are more regular and are uniformly simple in form.

Form surfaces that define older minor folds and isoclinally folded quartz lenses are invariably bedding or bedding schistosity, whatever the scale of observation. Where older folds and folded quartz lenses are absent, bedding and bedding schistosity also serve as the form surfaces of the younger folds; where older folds are present, the form surfaces of the younger folds consist of transverse schistosity and axial surfaces of the older folds.

Older folds.—Minor folds of the older set are sparse in the area. An excellent example of one, folded by a later fold, is exposed near 103,500 N., 94,300 E. (pl. 1). The limbs are isoclinal, the axial plane dips moderately south, and the fold axis plunges gently west overall, but all these elements are folded about a younger southwest-trending axis. A few other minor folds having isoclinal limbs and anomalous axial trends, but having no evident relation to younger folds, are probably of the older set; the fold symbolized near 100,940 N., 96,580 E. is probably an example (pl. 3). The distinctive features of these folds appear to be a plastic style, virtually isoclinal limbs, nearly east-trending axes, and—depending upon position with relation to younger folds—locally warped and variable limbs, axial planes, and axes.

Folded quartz lenses.—Isoclinally folded quartz lenses are relatively rare in the area and occur chiefly in the highly schistose rocks. The folded quartz layers are greatly thickened in the noses of the folds and thinned or pinched out on the limbs, so that the form is that of a flattened cylinder folded about axes parallel to that of the cylinder. In the more highly deformed lenses, the limbs are nearly obliterated, and the thickened noses coalesce so as to be nearly indistinguishable from quartz rods.

The attitude of the rodding varies, but most rods are oriented about down the dip of bedding schistosity.

Younger folds.—Younger minor folds are abundant in the metasedimentary and metavolcanic rocks, but they vary in expression and abundance according to rock type. They range in style from moderately tight, straight limbed, sharp crested folds of generally small amplitude in the micaceous schist, to broad open warps in the more massive rocks such as the amphibolite.

The younger folds vary appreciably in attitude, and the variations show a crude pattern with respect to areal position. Axial planes are predominantly steep; they range in strike from northeast or east in the northeastern part of the area, to north or northeast in the southwestern part. The plunge of axes ranges from steeply southward to gently north or south. The folds have dextral, neutral, and sinistral patterns, all associated with both northerly and southerly plunges, but dextral patterns associated with southerly plunges predominate. Folds near ultramafic bodies deviate from the general trend.

FOLDS IN ULTRAMAFIC ROCKS

Minor folds are common in the ultramafic rocks. The folds vary widely in form and orientation: from isoclinal or attenuated and plastic to open and brittle in style, and from steep to gentle and through a wide range of trends in the attitudes of axes and axial planes. They are defined by form surfaces of two types: (1) surfaces that bear no genetic relation to folding, such as layering in the ultramafic rocks and intrusive contacts; and (2) foliation that is demonstrably related to folding, such as schistosity and cleavage that is axial plane to folds in the contacts of the ultramafic bodies (fig. 3, pls. 1, 2, 3, and 4*A*, *B*, and *C*).

The relation between the style and pattern of a fold and the kind of form surface that defines it is not simple. The sparse folds in layering range from nearly isoclinal to open and are variable in attitude and pattern. Folds in the contacts between serpentinite and amphibolite are mostly plastic in style, and the limbs are isoclinal or greatly attenuated. Some, however, are fairly open. Those in contacts between serpentinite and gneiss or schist are all open and tend toward a more brittle style. Most of the folds in the contacts have moderate to steep axes and a dextral fold pattern in plan, but a few are sinistral or neutral; near 101,000 N., 95,000 E. (pl. 1), a regular alternation of groups of dextral and sinistral small-scale folds that have gentle to moderate plunges results in a generally neutral pattern. Folds in the schistosity of the serpentinite are open and have fairly sharp troughs and crests. Some, such as those at 101,000 N., 95,000 E., are similar in style and attitude to nearby folds in the country rock. Others, such as those at 100,000 N., 97,430 E. (pl. 4*C*), have attitudes that diverge greatly from those of folds in the country rock.

TRANSVERSE SCHISTOSITY

Schistosity that transects or is transverse to bedding in the metamorphosed sedimentary and volcanic rocks is varied in appearance and in relations to bedding and folding. Two classes are distinguished on the basis of these relations: an *older transverse schistosity,* which bears an axial-plane relation to the older folds and which is virtually parallel to bedding, except in the noses of the older folds into which the schistosity abuts; and a *younger transverse schistosity,* which bears an axial-plane relation to the younger folds and is nearly everywhere divergent from bedding.

OLDER SCHISTOSITY

The older transverse schistosity consists entirely of generally parallel shear surfaces so closely spaced that they are resolvable with difficulty or are only resolvable microscopically. These shear surfaces are parallel to bedding and bedding schistosity throughout most or the entire extent of observation, but wherever the relations to older folds are seen, the shear surfaces are axial plane to the folds and butt into them.

Transverse schistosity generally parallel to bedding is seen chiefly in the gneiss and the more quartzitic varieties of schist and is subordinate to continuous bedding schistosity. Within the field of most outcrops, the pattern of this older transverse schistosity is virtually identical with that of bedding schistosity; it is only in much larger fields that the two diverge significantly. Consequently, measurements of attitudes of bedding schistosity nearly always also serve to record attitudes of the older transverse schistosity; a separate symbol for the older schistosity is almost never recorded.

YOUNGER SCHISTOSITY

The younger transverse schistosity occurs in distinctively different habits that correlate with lithology and that intergrade in conformance with changes in lithology. In granular rocks (quartzite and micaceous quartzite), the younger schistosity is a *fracture cleavage;* in quartzose mica schists, schistosity is a *slip cleavage.* In highly graphitic and phyllitic schists, the younger schistosity is generally classifiable as slip cleavage only in thin section.

All types of younger transverse schistosity conform to the same general spatial pattern and have about the same range of variation in attitude as do the younger folds. The schistosity is predominantly steep and varies significantly in strike only on a fairly large scale and in a simple manner. In the northeastern part of the area, it strikes chiefly northeast or east; in the southwestern part, it strikes north or northeast.

Fracture cleavage.—Fracture cleavage is characteristic of granular rocks such as quartzite and micaceous quartzite and is generally subordinate to bedding schistosity and to older transverse schistosity. The cleavage consists of parallel or subparallel fractures spaced a few millimeters apart, along which differential movement is very small or indiscernible. Micaceous minerals in the rock are not alined parallel to the fractures. Fracture cleavage is generally most conspicuous in troughs and crests of folds; commonly the cleavage surfaces are not strictly parallel but fan symmetrically on either side of the axial plane.

Slip cleavage.—Slip cleavage is a spaced schistosity that bears a regular relation to crinkles in an older schistosity and consists of discretely spaced surfaces of parting or incipient parting subparallel to the limbs of the crinkles and about parallel to their axial planes. It is distinctive of much of the schist and gneiss in the area. (See White, 1949, for a detailed discussion of slip cleavage.) It varies in expression from predominant over bedding schistosity to subordinate to it. The micaceous minerals of the rock are not generally alined parallel to the slip cleavage, though the shingled arrangement of micas in the attenuated limbs of crinkles commonly results in near parallelism. Where the attenuated limbs have actually been disrupted and sheared, some of the mica flakes are commonly parallel within the shear zones.

Slip cleavage in graphitic and phyllitic schist is commonly not readily distinguished in the field. In such rocks, the transverse schistosity consists of very closely spaced shear zones, commonly of such fine texture that they cannot be detected without the use of a microscope. The transverse schistosity grades imperceptibly into slip cleavage. For the most part, the micaceous minerals are not parallel to the schistosity, but mica flakes are locally alined within individual shear zones; in some places such parallel flakes are abundant. Bedding and bedding schistosity in such rocks are discontinuous; segments can rarely be traced more than a few centimeters and in places form only shredded fragments. Where the noses of folds are discernible, the transverse schistosity is parallel to the axial planes of the folds.

UNCLASSIFIED SCHISTOSITY

When the relations of schistosity to bedding are unknown, the schistosity is here termed "unclassified." Commonly, the relations are unknown because bedding was originally poorly developed or has been largely obliterated during metamorphism and deformation, or the conditions of exposure were poor. Probably most unclassified schistosity is equivalent to divergent schistosity in which bedding has been almost entirely obliterated. A small proportion may be equivalent to the older transverse schistosity. Most of the structural attitudes recorded for unclassified schistosity are markely divergent from bedding and of the same pattern as the younger transecting schistosity.

CLEAVAGE IN ULTRAMAFIC ROCKS

Cleavage in the ultramafic rocks is confined chiefly to serpentinite. Locally, talc-carbonate rock, carbonate-quartz rock, and steatite have similar cleavage, but they are generally poorly exposed. Dunite and peridotite are chiefly massive and unfoliated.

The cleavage is of three distinctive types: (1) paper-thin schistosity characteristic of the thin schistose zones that enclose the massive shear polyhedrons; (2) subparallel irregular fractures or shear zones spaced as much as several centimeters apart; (3) sets of closely spaced fractures that intersect in a rhombic pattern, bounded by or grading into faults or shear zones of small displacement. In addition, relict-bedding schistosity is preserved in a few places in steatite derived from schist.

The paper-thin schistosity characteristic of the thin zones surrounding shear polyhedrons is a fine-textured spaced schistosity. Between shear surfaces, the serpentine particles are common largely alined almost parallel to the shear zones, but rarely so well as to impart a good continuous schistosity. In the talcose and carbonatian rocks only the gross pattern of the shear zones remains, and the particles of talc are generally diversely oriented. The pattern of the schistosity conforms with that of the surfaces of the shear polyhedrons; although very irregular in detail, the overall trend is in most places roughly parallel to the principal plane of the lenticular bodies.

The widely spaced irregular parallel fractures and thin shear zones range in spacing from a fraction of a centimeter to several centimeters. This type of cleavage is distinctive of moderately sheared serpentinite containing shear polyhedrons and of many of the massive shear polyhedrons themselves. The attitude of this widely spaced cleavage is highly variable, and the cleavage is locally folded. At the southwestern contact of the Lowell quarry body in the vicinity of 100,000 N., 97,420 E. (pl. 4*C*), the cleavage is axial plane to bulbous tightly compressed folds.

Closely spaced fractures that intersect in a rhombic pattern are of two principal types: (1) sharply bounded tabular zones a few centimeters to a meter or more thick that transect massive serpentinite and, locally, dunite or peridotite; and (2) branching distributive shear zones that transect schistose serpentinite characterized by shear polyhedrons and that grade into faults or tabular shear zones.

The relations of the rhombic cleavage are distinct where it forms sharply defined tabular zones. Both sets of intersecting fractures are steep, and they intersect in dihedral angles of 30° and 150° to form crude rhombic prisms generally less than a centimeter on the rhombic edge. The prismatic axis is about down the dip, and the long diagonal of the rhombic section is parallel to the margins, of the zone. The strike of the fractures varies widely in different zones.

Within zones of schistose serpentinite characterized by shear polyhedrons, rhombic cleavage is generally indistinct because of the interference of the irregular schistose zones that surround the shear polyhedrons.

The relations between the three types of cleavage distinctive of the ultramafic rocks seem fairly simple, though not entirely beyond dispute. The schistosity surrounding shear polyhedrons intergrades with spaced cleavage formed by parallel fractures and thin shear zones. The rhombic cleavage integrades with faults and shear zones of minor offset and intersects both the schistosity that surrounds the shear polyhedrons and the widely spaced cleavage. Along the southwest margin of the Lowell quarry body (pl. 4*C*), the widely spaced cleavage is axial plane to bulbous folds in the contact of the ultramafic body, and the minor faults and tabular zones of rhombic cleavage are axial plane to folds in the widely spaced cleavage.

LINEATIONS

In addition to the axes of folds, significant linear features in the area include quartz rodding, crinkles, striations, and intersections of cleavage and bedding schistosity. Lineation resulting from mineral elongation is virtually confined to the asbestos and picrolite veins. This rather special kind of lineation is discussed under both sections on "Serpentine veins."

Quartz rodding is at least partly, and probably entirely, equivalent to the isoclinally folded quartz lenses described in the section on "Folds in the metamorphosed sedimentary and volcanic rocks" under "Minor structural features" and is similar in attitude. The rods have the form of irregular elliptical cylinders and are generally strongly striated parallel to the cylindrical axis. They are commonly about 1 cm by 3 cm in cross section and as much as 60 cm long.

Lineation marking the intersection of cleavage and bedding or bedding schistosity takes several forms. Most commonly, for example, where the cleavage is slip cleavage or transecting schistosity, the intersection is marked by a crinkle, which is generally a millimeter or less and rarely as much as several millimeters in wavelength. Where fracture cleavage intersects bedding, the lineation is marked by a distinct fracture line. The strong striations distinctive of quartz rods and isoclinally folded quartz lenses are the imprint of transecting schistosity on the quartz beds, and therefore are really lineations resulting from intersection of bedding and cleavage.

In a few places where a conspicuous crinkle lineation marks the intersection of slip cleavage and bedding schistosity, another set of crinkles is oriented at a large angle to the first set. Commonly, crinkles of the second set are smaller than those of the first. Intersecting relations between crinkles of the two sets are generally rather indistinct and ambiguous; where clearcut, the crinkles associated with the axial-plane slip cleavage are intersected and offset by crinkles of the other set.

Quartz rodding and striations on the quartz rods and folded quartz lenses have the same spatial pattern as the older fold axes. Lineation that marks the intersection of bedding schistosity and younger transecting schistosity (fracture cleavage, slip cleavage, or divergent schistosity) is strictly parallel to the axes of the younger folds. Crinkles that make a large angle with those associated with the transecting schistosity are about parallel to the older fold axes in most places; however, the intersecting relations indicate that they are about the same age as the crinkles parallel to the younger folds and so are unrelated to the older folds.

JOINTS

Joints occur erratically in the area and are most conspicuous in the more massive varieties of both country rock and ultramafic rocks. Perhaps because of better conditions of exposure in the ultramafic rocks, joints seem more abundant there than in the country rock.

Observations on joints in the country rock are scattered and confined almost entirely to the amphibolite; the joints show no discernible systematic pattern and no apparent significant relation to other structural features. In the gneiss at the northeast contact of the Lowell quarry body, a regular widely spaced set of joints was exposed by stripping operations during quarrying; they are parallel to a faint slip cleavage which is axial plane to the younger folds (pl. 4*D*). These "joints" probably formed during quarrying operations by fracturing of the rock along planes of weakness determined by the slip cleavage.

Joints in the ultramafic rocks are virtually confined to the dunite, peridotite, and massive serpentinite. Their apparent absence in the talcose and carbonatian rocks probably results from conditions of exposure, for joints in talc-carbonate rock are locally conspicuous in ultramafic bodies elsewhere in Vermont (Chidester, 1962, p. 24). The joints range in individual configuration from nearly planar to very irregular; in their interrelations, they range from clearly recognizable systems of two or three sets, through clearly defined sets with no apparent systematic relation between sets, to apparently randomly oriented fractures. Even among conspicuous sets and systems of joints, no overall pattern is evident. Particular spatial patterns prevail throughout areas several tens of meters in extent, but the relations between such fields are varied.

Throughout the main ultramafic body, many of the joints contain veins of cross-fiber chrysotile asbestos. Commonly, all the joints of a system contain such cross-fiber veins, but in some places, joints of one set contain veins, whereas those of a second set, which is virtually complementary to the first, contain no veins. Where several sets of joints are clearly not members of a single system, it is common for one to contain cross-fiber veins and for the others to be barren.

Joints in the massive serpentinite and dunite of the Eden quarry body in the vicinity of the bodies of talcose rocks near 99,000 N., 92,000 E., and 100,000 N., 94,000 E. (pl. 1), are commonly the centers of irregular thin tabular zones of talc-carbonate rock. In many such places, the talc-carbonate alteration is confined to a single set of joints, and other sets of joints are barren of talc. Elsewhere in the main body, talc is almost entirely absent from all joints, except for a single occurrence a few centimeters thick and a few meters in extent along a joint in the Lowell quarry body near 99,850 N., 98,320 E. (pl. 3).

FAULTS

Observed faults in the area are virtually confined to the vicinity of the main ultramafic body, doubtless largely as the result of better conditions of exposure in the quarry areas (fig. 3, and pls. 1–4).

Nearly all of these faults are steep to moderate in dip, and range in displacement from only a few centimeters (so that distinguishing them from joints becomes largely academic) to a few decimeters. They range in length from only a few decimeters to a few hundred meters. In most, the zone of shearing is not more than 5 or 6 cm thick and is sharply bounded by rather smooth surfaces of unsheared rock. Slickensides and gouges on the fault surfaces are mostly downdip. Most of the faults are entirely within the ultramafic bodies or intersect the contacts and extend only several tens or a hundred meters or so into the country rock; some, however, are confined to the country rock. Two predominant attitudes define two fairly apparent sets of faults, one trending northeast and one trending northwest. A few faults, most of which are very irregular in plan, are diverse in orientation and do not seem to fit either set.

The fault at the northeast margin of the Eden quarry body, near 100,200 N., 96,600 E., (pls. 1 and 3), has been excellently exposed by stripping operations throughout a length of about 100 m and appears to be of greater magnitude than other faults in the area. A conspicuous gouge zone ranges from 15 to 25 cm in thickness, and the minimum displacement appears to be a few tens of meters (pl. 3, cross-section *A–A′*). The fault dips moderately southwest. Despite its greater magnitude, this fault is not greatly different from those of the northwest-trending set of smaller faults, and it probably belongs to that set.

Nearly all the faults that show evidence of relative movement sense are reverse faults, having only a small but commonly appreciable strike-slip component. In many of the faults the movement sense is indeterminate.

Within the country rock zone, along all the faults that intersect the contacts of the Lowell quarry and Eden quarry bodies, the rock bordering the faults and the gouge and fragmented rock within the fault zones are altered for a few millimeters or centimeters to calc-silicate minerals and to serpentine-chlorite rock (see section on "The rodingite and serpentine-chlorite rock association" under "Contact rocks" in section on "Ultramafic and associated rocks.")

Within the ultramafic bodies, the faults are barren of calc-silicate minerals. Many are virtually barren of chrysotile asbestos, but some contain slip fiber asbestos and picrolite.

TECTONIC AND PETROGENIC SYNTHESIS

The Belvidere Mountain area records a history of sedimentation, volcanic activity, and emplacement of ultramafic rocks in an early Paleozoic eugeosyncline, followed by a complex history of folding and metamorphism which extended into the Devonian. Most of this history is imprinted in the rocks of the area and has been documented in appropriate sections of this report, but establishment of some of the chronology and of the broader correlations depends upon regional studies which have been cited and summarized. The ultimate origins of the ultramafic rocks are largely inferential and encompass broad problems of petrology, global tectonics, and earth history, which are highly speculative and which allow for wide latitude in interpretation. Consequently, this aspect of the history has long been involved in intense controversy. (See Chidester, 1962, p. 87–88; Jahns, 1967, p. 155–156; Thayer, 1960, 1967, 1969, p. 511–515; Wyllie, 1967, 1969; Chidester and Cady, 1972.)

ORIGIN OF THE ULTRAMAFIC ROCKS

Two alternative modes of origin have generally been advocated for alpine-type ultramafic rocks: (1) accumulation by crystal fractionation of a complex (for example, basaltic) magma and (2) derivation from the upper mantle by one or more of several, commonly vaguely specified, processes. In either mode of origin, emplacement as a crystalline mass is generally recognized. Both models imply a close genetic relation between the ultramafic rocks and mafic volcanic rocks interbedded in a eugeosynclinal pile. In the first model (fractionation of a complex magma) the ultramafic rocks are regarded as crystal cumulates and the basalts are regarded as fluid fraction. In the simplest form of the second model (partial melting of upper mantle material), the ultramafic rocks are regarded as refractory residue and the basalts, as fluid filtrate. Many variations on this theme are possible. In earlier reports (Chidester, 1962, 1968; Cady and others, 1963; Cady 1969, p. 24–25), we have adhered to these conventional theories as alternatives for the origin of the ultramafic rocks of Vermont. Both theories, however, have serious drawbacks to their acceptance, such as implied structural and petrologic features consequent upon a particular theory of origin, which do not fit observed field relations (for example, the absence of associated granitic rocks), and the need to invoke special conditions to explain the wide variety of relations shown by alpine-type ultramafic rocks.

A theory of origin of ultramafic rocks based upon the concept of sea-floor spreading proposed by Hess (1962; Dietz, 1961) has been applied with considerable success to full ophiolite complexes in such widely separated areas as Papua-New Guinea (Davies, 1968), Oman (Reinhardt, 1969), Greece (Moores, 1969), Cyprus (Moores and Vine, 1971), and California (Bailey and others, 1970). However, the application of the concept to the alpine-type ultramafic rocks of Vermont, which are exclusively *intrusive* into fairly continuous sequences of eugeosynclinal sedimentary and volcanic rocks, requires modification of the model developed for the full ophiolite assemblages, which includes sheeted diabase, pillow lava, and chert. We (Chidester and Cady, 1972) have recently presented such a model specifically to reconcile the mode of origin of the alpine-type ultramafic rocks of the Appalachians with the concepts of sea-floor spreading. The salient features of the model are summarized as follows:

In late Precambrian time, mantle upwelling began along a sinuous belt that now marks the site of the Appalachian orogen. This sinuous belt was 300 to 800 km northwest of the line along which Africa and North America were later to separate by continental drift, beginning in the Mesozoic. At this early stage, however, the two were parts of one continent, and the floor of the incipient Appalachian eugeosyncline was thick sialic crust. Outward movement of upper mantle rocks on either side of a developing rift beneath the continental crust led to distention and necking of the continental crust and downwarping of the surface, thereby initiating the Appalachian eugeosyncline. Basaltic lava, formed by partial melting of material deeper in the upper mantle, erupted along the rift. As the subcontinental rift continued to form, and as the accumulation of sediments in the eugeosyncline increased, basaltic lava erupted within the eugeosyncline as dikes and sills, surface flows, pyroclastic layers, and volcanic piles, which were continually reduced by erosion and incorporated as detrital volcanic beds into the sedimentary sequence. As a result of cooling and hydration during upwelling, the peridotite of the upper mantle beneath the rift became partly serpentinized. As the upper mantle moved outward away from the rift beneath the continental crust, interactions at the surface between crust and mantle produced geanticlinal tracts and projections and irregular bulges of rocks of mantle composition into the crust. As activity continued along the rift zone and as the eugeosyncline formed, many masses of such material, varying widely in size, intruded the continental crust and eugeosynclinal pile. Coarse layering that reflects complex variations of primary minerals in dunite and peridotite has the earmarks of magmatic crystallization and accumulation. Such layering is inferred to be a relict from the upper mantle.

EMPLACEMENT AND STRUCTURAL HISTORY

The ultramafic rocks of the Belvidere Mountain area are representative of ultramafites in the northern Appalachian belt, which were emplaced at various times in the early and middle Paleozoic, during a long interval in which sets of early folds, plastic and similar in style and associated with an axial-plane schistosity, were successively formed (Cady, 1969, p. 44, 47, 108, 113). Some of the early folds are clearly Ordovician in age (Cady, 1969, p. 40–41; Stevens and others, 1969) ; some are dated as Devonian (Jahns, 1967, p. 138–141). The ultramafites were emplaced prior to a set of late folds, open and generally parallel in style and having an axial-plane cleavage, that form the Green Mountain anticlinorium. The late folds are Middle and Late Devonian (Jahns, 1967, p. 141–144; Cady, 1969, p. 75–80). (see section on "Folds in ultramafic rocks"; see also Chidester, 1968, p. 348–350.) In neighboring Quebec, ultramafic bodies are truncated by a Middle Ordovician unconformity (Cady, 1969, p. 23–24); emplacement of the ultramafites in the Belvidere Mountain area probably also took place in Early Ordovician.

Plastic deformation of the rocks in the eugeosynclinal pile began early in the geosynclinal history and continued episodically throughout it. In this process, bodies of ultramafic rock were nipped off the upper mantle. Stresses were mainly subhorizontal and were triggered collectively, and in varying degree, by local uplift in the eugeosyncline (perhaps principally over the subcontinental rift), and by compaction-diagenesis-metamorphism in the wet eugeosynclinal sediments (Cady, 1969, p. 42, 143–145.) These stresses produced tightly compressed plastic-style folds, and kneaded and squeezed previously emplaced bodies of ultramafic rock along bedding planes and developing structural surfaces in the sedimentary and interbedded volcanic rocks. As the ultramafic rocks were transported to successively higher positions, they passed through newly deposited and newly folded rocks.

Serpentinization, already begun as pervasive but partial alteration of the dunite and peridotite when the upper mantle rocks were carried upward into a lower temperature regime, continued and was assisted by contact with wet sediments through which the bodies of ultramafic rock were kneaded upward. Transport was intermittent; episodes of movement were separated by periods of immobility. Sharing of the mass during transport subsequently facilitated the entry of water and consequent further serpentinization during periods of immobility. Such incremental serpentinization further increased the mobility of the mass in subsequent episodes of movement, during which dehydration of the serpentine under shear stress at high pressure and temperature (Raleigh and Paterson 1965, p. 3965; Reicker and Rooney 1966, p. 196), was perhaps also effective. A postulated sequence of events implies repeated episodes of serpentinization, shearing, and dehydration in alternating stages of movement and immobility during tectonic transport, and increasingly extensive serpentinization accompanied by increased penetration of shearing. No unequivocal petrographic evidence was observed to confirm such a history, but as Jahns (1967, p. 156) has stated, disseminated flakes of talc in massive serpentinite may be relicts of such dehydration. Serpentinization during the stage of tectonic transport produced chiefly highly sheared serpentinite at the margins of the body and in sheared zones that transect the central parts. As the process continued, pervasive serpentinization extended into the more massive zones, becoming complete in the outer parts of such zones and decreasing generally but irregularly inward. In the central parts of the larger bodies (such as the main body at Belividere Mountain), partial or only slight serpentinization resulted, and large parts of the olivine escaped alteration.

In contrast to coarse layering (see section on "Origin of the ultramafic rocks" under "Tectonic and petrogenic synthesis"), fine layering in dunite, peridotite, and massive serpentinite reflects simple textural and mineralogical alternations and probably was formed by flowage of the crystalline mass, particularly in the early stages of transport when entire masses were subjected to simple deformational stresses.

Complex shearing patterns, which formed in the serpentinite as it was kneaded through the country rock, led to the formation of shear polyhedrons in the marginal zones of the ultramafic bodies. Unsheared units of serpentinite, ranging in size from small chips to masses several or many feet in diameter, were milled and rotated as they moved past one another, producing a matrix of finely comminuted and sheared serpentinite that forms shells around and fills interstices between polyhedrons. The mosaic pattern produced by diverse orientations of relict layering in adjacent larger polyhedral units results from rotation of the blocks relative to one another. In the central parts of the larger bodies of ultramafic rock, large masses of only slightly serpentinized dunite and peridotite were carried along relatively intact, mechanically buffered by yielding in the highly sheared marginal zones.

After the emplacement of the ultramafic bodies at their present sites, the late set of folds, open in style and divergent from the early set in attitude, was imposed on the country rock and the ultramafic bodies (Jahns, 1967, p. 141–144). Slight movement of the ultramafic rocks relative to the enclosing country rock may have occurred locally during the late episode of folding. Such movement is suggested by the local absence of rodingite at the margins of the main Belvidere Mountain body. (See section on "Metamorphic history" under "Tectonic and petrogenic synthesis.")

Figure 8 illustrates the evolution of the structural pattern of the area and the emplacement of the ultramafic rocks. The three diagrams depict successive stages in the geologic history, as if at each stage the rocks were exposed at the present surface.

In favorable structural situations, such as at Belvidere Mountain, the late folding strongly warped tabular bodies of ultramafic rocks, bending layers and schistosity in conformity with late folds in the country rocks and producing a slip cleavage in folded schistose serpentinite identical in style with that in the adjacent country rock. Where intrusive contacts partly cut across early folds and partly conform to and were controlled by bedding surfaces warped by the early folds, the structural pattern of the contacts reflects part of the pattern of the early folds and the entire pattern of the late folds. Joint systems in the ultramafic rocks, which are best developed in the relatively fresh massive dunite, are inferred to be related to the late folds. The pattern of joints, though locally consistent, varies irregularly from place to place in a body because of irregularities in the internal stress field induced by inhomogeneities and irregularities in the ultramafic bodies.

In unfavorable structural situations, such as in podlike bodies emplaced in simple homoclinal sections, the effects of the late stage of folding are inconspicuous in the ultramafic bodies.

Near the climax of the late period of folding, joints in the massive ultramafic rock and fracture surfaces in sheared serpentinite opened along zones under tensional stress.

METAMORPHIC HISTORY

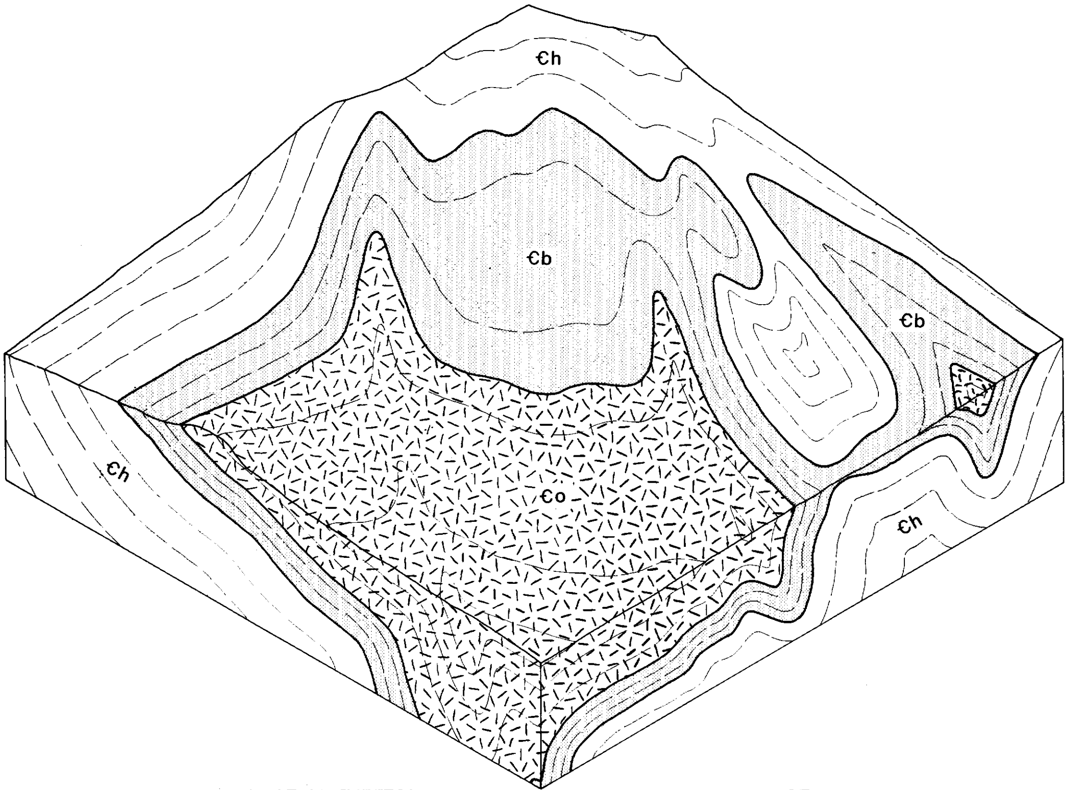
Interpretation of the early metamorphic history of the ultramafic rocks depends to a considerable extent upon inferences regarding their history before emplacement, which remains highly speculative. In particular, conclusions concerning two stages of the early history—origin of the primary ultramafic rocks and emplacement of the ultramafic bodies—are critical to interpretation of the metamorphic history. The following outline of events and their relations to one another is based upon the above discussions of concepts of origin along a subcontinental rift and solid emplacement by tectonic transport.

Serpentinization is inferred to have begun early in the stage of mantle upwelling along the rift beneath the eugeosyncline, in accordance with the concepts developed by Hess (1962) for sea-floor spreading at an oceanic rift. Extensive further serpentinization occurred during tectonic transport of the ultramafic masses, as an integral part of the mechanism of emplacement. Probably most of the serpentinization in each body was accomplished by the time the bodies were finally emplaced.

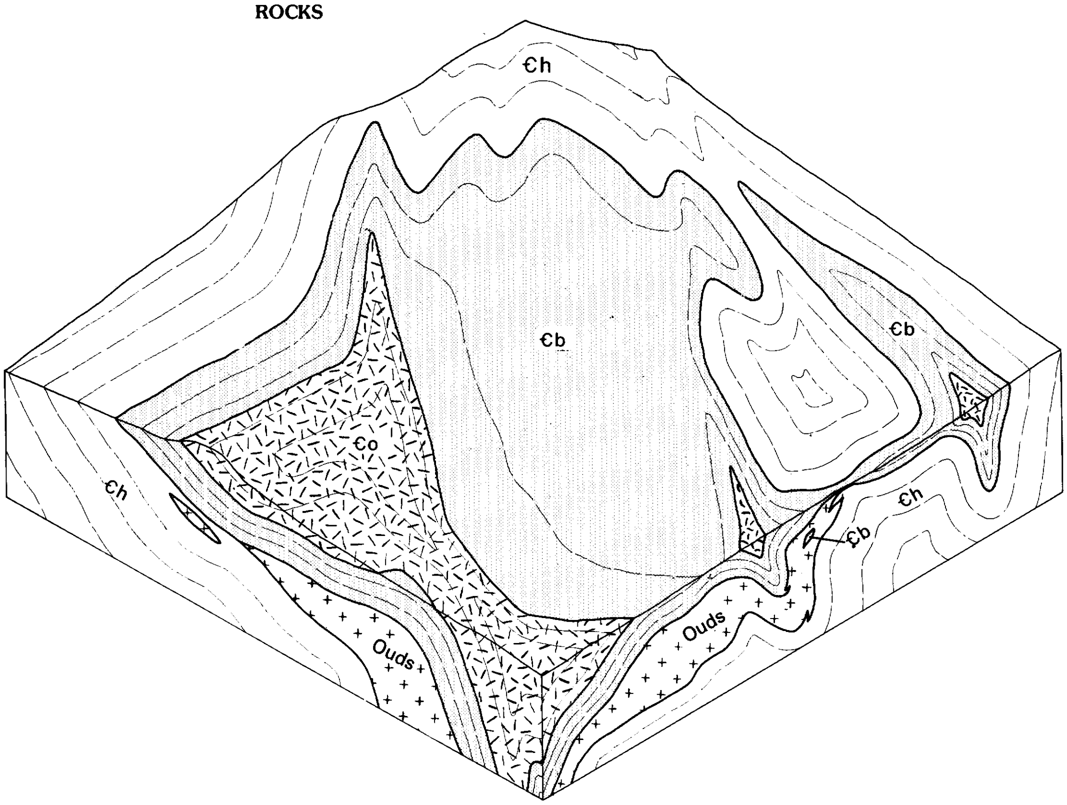
After the emplacement of the ultramafic rock, initial metamorphic changes were confined to systems not greatly larger than the ultramafic bodies (except for the sources of water). These changes included further pervasive serpentinization on a small scale, formation of serpentine and related veins in and adjacent to the ultramafic bodies, and the formation of narrow contact metamorphic and metasomatic aureoles adjacent to the larger ultramafic bodies. After formation of these features, metasomatic changes related to regional metamorphism occurred. In part, these changes involved extensive migration of CO_2 and resulted in the local alteration of dunite and serpentinite to talc, magnesite, quartz, and tremolite, in distinctive patterns of assemblages. Concomitantly, related alterations in the adjacent country rock produced distinctive contact-rock associations.

The thermal history of a mass of upper mantle rock intruded into a eugeosynclinal pile would depend upon a complex interrelation among many variables; primarily, however, the body's temperature and its difference from ambient temperature would depend on its size and on the rate and distance of transport after detachment from the mantle. Initially the body of rock would be at a higher temperature than that of the surrounding rocks. As it was kneaded through the pile, it would lose heat to its surrounding and gradually approach the ambient temperature. In terms of simple conductive loss, larger less far traveled bodies would be hotter than smaller farther traveled bodies at any particular time after detachment from the mantle.

Though all the ultramafic rocks in the Belvidere Mountain area were emplaced at considerable depth and about simultaneously, thermal considerations

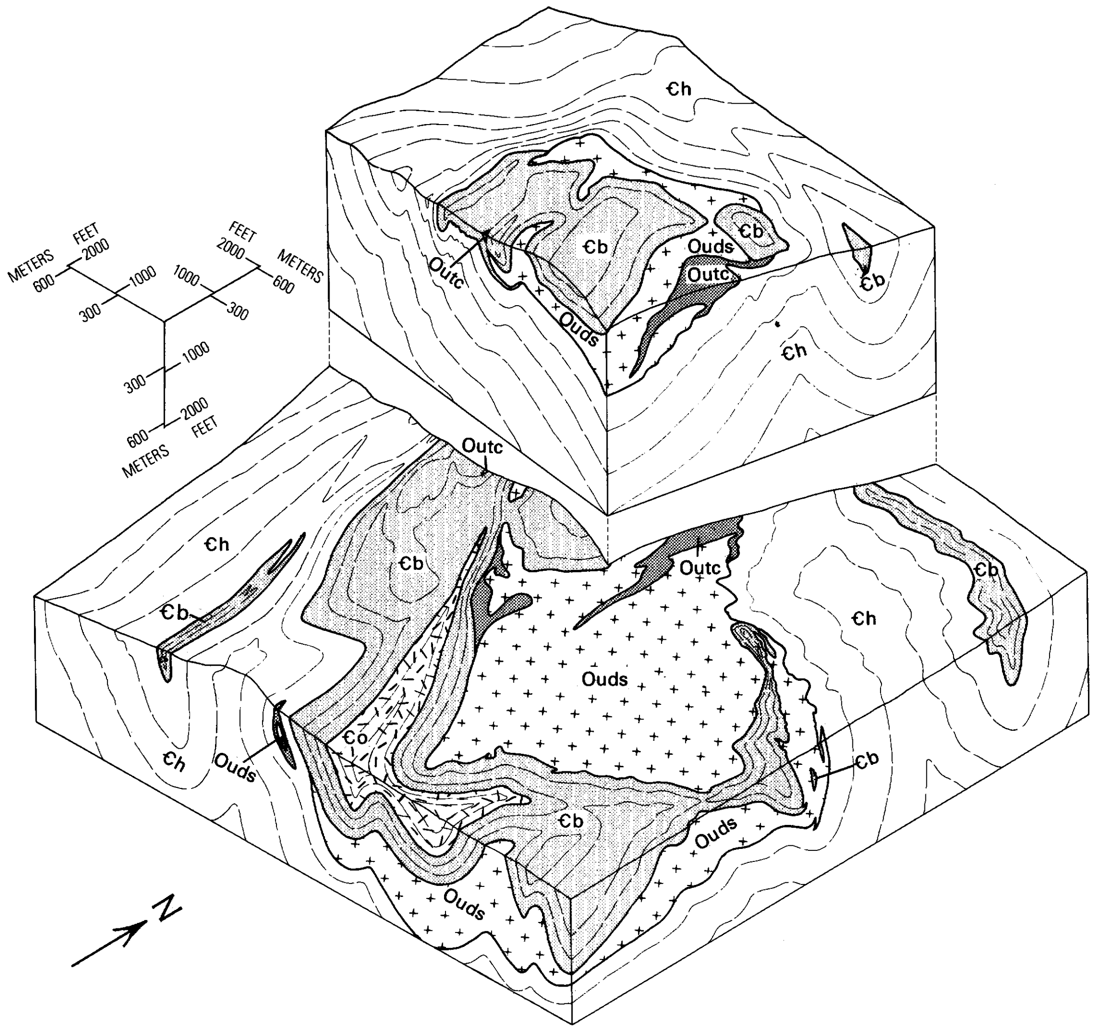


A. AT AN INTERMEDIATE STAGE OF THE FIRST PERIOD OF FOLDING, PRIOR TO EMPLACEMENT OF THE ULTRAMAFIC ROCKS



B. AT THE END OF THE FIRST PERIOD OF FOLDING, FOLLOWING EMPLACEMENT OF THE ULTRAMAFIC ROCKS

FIGURE 8.—Isometric block diagrams illustrating critical stages in the structural history and emplacement of ultramafic rocks in the Bedvidere Mountain area. Rock units are shown at each stage as if they were exposed at the present land surface. *A*. At an intermediate stage of the first period of folding, prior to emplacement of the ultramafic rocks. *B*. At the end of the first period of folding, following emplacement of the ultramafic rocks. *C*. At the end of the second period of folding, and following metamorphism. Note the bodies of talc-carbonate and carbonate-quartz rock, absent in diagram *B*.



C. AT THE END OF THE SECOND PERIOD OF FOLDING, AND FOLLOWING METAMORPHISM. NOTE THE BODIES OF TALC-CARBONATE AND CARBONATE-QUARTZ ROCK, ABSENT IN DIAGRAM *B*

EXPLANATION

Outc	TALC-CARBONATE ROCK AND CARBONATE-QUARTZ ROCK (ORDOVICIAN)[1]
+ Ouds +	DUNITE AND SERPENTINITE (ORDOVICIAN)[1]
Ꞓo	OTTAUQUECHEE FORMATION (UPPER AND MIDDLE CAMBRIAN)
Ꞓb	BELVIDERE MOUNTAIN FORMATION (LOWER CAMBRIAN)
Ꞓh	HAZENS NOTCH FORMATION (CAMBRIAN(?))

[1] The Ordovician age designation refers to the age of emplacement of the intrusive ultramafic rocks, not to the age of the parent igneous rocks or the metamorphic derivatives (see text subheading discussions under "Tectonic and petrogenic synthesis").

FIGURE 8.—Continued.

and the mechanism of emplacement advocated here suggest that the larger bodies differed significantly in temperature and mineralogy from the smaller bodies at the time of emplacement. The larger bodies, such as the main ultramafic body at Belvidere Mountain, contained large masses of unaltered dunite in the central part of the body, and their temperatures were appreciably, though not greatly, above the temperature of the surrounding rocks. The smaller bodies were completely serpentinized when emplaced and were virtually at the temperature of the enclosing rocks.

Under these circumstances, the temperature of the country rock at the margins of the large ultramafic bodies would be increased, whereas that at the margins of the small bodies would not. Serpentinization of dunite would continue so long as water was available, producing such effects as partial desiccation and the addition of Ca and Mg (see section on "Rodingite" under "Contact rocks" in the section on "Petrogenesis" under "Ultramafic and associated rocks" and see tables 7 and 9). These conditions combined to produce narrow contact-metamorphic and metasomatic aureoles of rodingite around the larger ultramafic bodies that contained considerable dunite, but no discernible effects at the margins of the smaller completely serpentinized bodies.

Limited serpentinization of dunite in the larger bodies extended into the period of late folding. As temperatures fell in the contact zones, serpentinization of dunite and consequent efflux of Mg led to replacement of rodingite by serpentine-chlorite rock in a narrow zone next to the contact of the ultramafic body. Near the climax of the late folding, the dunite and serpentinite opened up along joints, fractures, and shear surfaces under tensional stress. As such openings formed, they were filled with chrysotile to form cross- and slip-fiber asbestos veins and picrolite veins. The character of the serpentine veins depended primarily upon local mechanical relations and less on physico-chemical variations. Because the stress field in the ultramafic bodies varied erratically owing to irregularities and inhomogeneities in the bodies, the direction and rate of opening along fractures and shear surfaces varied markedly from place to place and resulted in a complex distribution pattern of the various kinds of serpentine veins. Simultaneously with the formation of the serpentine veins, constituents released during their formation formed segregations of magnetite in the serpentine veins or in nearby fractures in the ultramafic bodies and contributed to the formation of chlorite-calcite-magnetite veins in the bordering country rock.

During regional metamorphism, which culminated during and extended somewhat beyond the climax of the late stage of folding, dunite and serpentinite were locally altered to talc-carbonate rock and carbonate-quartz rock. The distribution of these rocks at the margins of the ultramafic bodies, or in linear zones that appear to be controlled by faults or shear zones, indicates that the processes were controlled by access of CO_2 into the ultramafic rocks. The general absence of such rocks adjacent to contacts where rodingite is present suggests that the rodingite acted as an effective barrier to the access of CO_2, or that the structural conditions that led to preservation of the rodingite also prevented access of CO_2. The distribution of carbonate-quartz rock with respect to talc-carbonate rock suggests that highly quartzose zones are centered along channelways that provided ready access of CO_2, and indicates that formation both of talc-carbonate rock and carbonate-quartz rock was a single process whose end product was carbonate-quartz rock. Where dunite and peridotite were extensively altered to carbonate-quartz rock and talc-carbonate rock, expulsion of large amounts of Mg and Si led to extensive and irregular steatitization, chloritization, and related metasomatic replacement of the adjacent country rock.

In the smaller ultramafic bodies that have no rodingite border, and in the larger bodies where rodingite is locally absent, serpentinite and adjacent siliceous schist reacted to form steatite (chiefly by replacement of serpentinite) and blackwall chlorite rock (entirely by replacement of schist). Where the country rock bordering the serpentinite is amphibolite, tremolite was formed additionally, or in place of talc. Where rodingite borders serpentinite, it acted as an effective barrier to metamorphic interaction between serpentinite and schist—perhaps in part because, as an effective barrier to ingress of CO_2, it prevented the triggering action of CO_2 metasomatism (Chidester, 1962, p. 120; Cady and others, 1963, p. 47; Cady, 1969, p. 155).

CALCULATIONS

Throughout this report, chemical changes in metamorphic and metasomatic processes inferred to have taken place at constant volume are deduced by comparison of the chemical content of equal volumes ("modified standard cells") of rocks (Chidester, 1962, p. 95). Chemical analyses in table 1 are presented both in terms of weight percentages and in terms of cation percentages, and cell factors (F_c)

are included so that the content of the modified standard cell for each analysis can be computed readily by multiplying the cation percentage for each component by the cell factor.

The method of deriving the modified standard cell, the methods of computation, and related problems and methods of calculating mineral formulas from chemical analyses have been described in an earlier report (Chidester, 1962, p. 94–97, 129–205).

In the present report, all serpentine formulas are calculated simply on the basis of $(\mathrm{Si}+\frac{1}{2}R^{+3})=4$, because the methods of sampling and analysis do not warrant further refinements in calculation. (See text discussion in section "Mineralogy of the serpentine group" under "Ultramafic and associated rocks"; Bates, 1959.)

In analyses of rocks consisting essentially of serpentine and minor brucite, the content of brucite is calculated from the relations:

1. $N_s+N_B=C$,

and

2. $\frac{3}{5}N_s=\mathrm{Si}+\frac{1}{2}R^{+3}$,

where N_s = equivalent molecular percentage of serpentine; N_B = equivalent molecular percentage of brucite; C = sum of equivalent molecular percentages of serpentine and brucite; Si = cation percentage of silica; and R^{+3} = sum of cation percentages of trivalent elements (essentially $\mathrm{Al}+\mathrm{Fe}^{+3}$).

Mineral formulas of amphibole and epidote were calculated from analyses of mineral separates, which contained small but significant amounts of impurities, by solving simultaneous equations relating the two analyses, yielding calculated analyses of the pure minerals. The methods of calculation are described below.

ANALYSES OF PURE AMPHIBOLE AND PURE EPIDOTE

1. Determine the mode of each analyzed sample (from thin section, slide mounts of sized grains of the material analyzed, or otherwise) and recalculate to the basis of amphibole + epidote = 100.

	Amphibole sample (containing minor impurities)	*Epidote sample (containing minor impurities)*
Amphibole	M	P
Epidote	N	Q

where $M+N=100$, and $P+Q=100$.

2. Recalculate each analysis of the impure material into equivalent molecular (= cation) percent. On the basis of the approximation that cation percentage = volume percentage, correct the analysis for impurities of fixed composition (sphene, rutile, albite, and the like) by subtracting appropriate amounts of Ca, Ti, Si, Al, and so forth. The balance, recalculated to the basis of the sum of the cations = 100, represents, for each analysis, analyses of mixtures solely of amphibole and epidote.

3. From the modes recalculated to the basis of amphibole + epidote = 100 (1, above), and the calculated analyses of mixtures consisting solely of amphibole and epidote (2, above), the following simultaneous equations, valid for each constituent in the analysis, can be derived:

$$(1)\quad mC_{\mathrm{a}}+nC_{\mathrm{e}}=C_{\mathrm{ae}}$$

$$(2)\quad pC_{\mathrm{a}}+qC_{\mathrm{e}}=C_{\mathrm{ea}}$$

where m, n, p, q are equal to $\frac{M}{100}$, $\frac{N}{100}$, $\frac{P}{100}$, $\frac{Q}{100}$. C_a represents the cation percentage of each constituent in the analysis of pure amphibole; C_{e} represents the cation percentage of each constituent in the analysis of pure epidote; C_{ae} represents the cation percentage of each constituent in the mixture of amphibole and epidote; and C_{ea} represents the cation percentage of each constituent in the mixture of epidote and amphibole.

4. Solve (1) and (2) simultaneously:

Multiple (1) by q:

$$(3)\quad qmC_{\mathrm{a}}+qnC_{\mathrm{e}}=qC_{\mathrm{ae}}.$$

Multiply (2) by n:

$$(4)\quad pnC_{\mathrm{a}}+qnC_{\mathrm{e}}=nC_{\mathrm{ea}}.$$

Subtract (4) from (3) and solve for C_{a}:

$$qmC_{\mathrm{a}}-pnC_{\mathrm{a}}=qC_{\mathrm{ae}}-nC_{\mathrm{ea}}.$$

$$(5)\quad C_{\mathrm{a}}=\frac{q}{qm-pn}C_{\mathrm{ae}}-\frac{n}{qm-pn}C_{\mathrm{ea}}.$$

Substitute (5) in (2) and solve for C_{e}:

$$(6)\quad C_{\mathrm{e}}=\frac{m}{qm-pn}C_{\mathrm{ea}}-\frac{p}{qm-pn}C_{\mathrm{ae}}.$$

5. For each component, substitute appropriate values of m, n, p and q, and of C_{ea} and C_{ae} in equation (5) to calculate the analysis of pure amphibole, and in equation (6) to calculate the analysis of pure epidote. The resulting calculated analyses are in terms of cation percentages.

6. For example, in table 1, analyses 23 and 24, of concentrates of amphibole and epidote, respectively:

$m=0.9900$ $p=0.0152$
$n=0.0100$ $q=0.9848$

Substituting these values in formulas (5) and (6), get:

TABLE 10.—*Calculation of amphibole formula*

[Sample No. A-BM-53-139a, Lab. No. 143380a. Analyses from table 1. 0, looked for and not found]

	Molecular (cation) present								
	Analyzed material	Impurities[1] (except epidote)			C'_{ae} (amphibole plus epidote)	$C_{ae} = \frac{100\,C'_{ae}}{\Sigma C'_{ae}}$	C_{ea} (from table 11)	C_a (pure amphibole) $=1.0103\,C_{ae} - .0103\,C_{ea}$	Formula Nos. (O+OH =24)
		Sphene	Rutile	Albite					
SiO_2	41.96	0.04	--	0.03	41.89	41.97	36.97	42.02	6.66
$AlO_{3/2}$	13.40	--	--	.01	13.39	13.42	29.57	13.25	2.10
$FeO_{3/2}$	2.52	--	--	--	2.52	2.53	6.89	2.49	.39
FeO	9.08	--	--	--	9.08	✓ 9.10	.74	9.19	1.46
MgO	16.24	--	--	--	16.24	16.27	.71	16.43	2.60
CaO	9.64	.04	--	--	9.60	9.62	24.05	9.47	1.50
$NaO_{1/2}$	5.26	--	--	.01	5.25	5.26	.60	5.31	.84
$KO_{1/2}$	.88	--	--	--	.88	.88	.10	.89	.14
$HO_{1/2}$	7.86	--	--	--	7.86	7.88	3.65	7.92	1.26
TiO_2	.58	.04	.03	--	.51	.51	.00	.51	.08
$PO_{5/2}$	.14	--	--	--	.14	.14	.14	.14	.02
F	.06	--	--	--	.06	.06	.03	.06	.01
$CrO_{3/2}$	.03	--	--	--	.03	.03	.08	.03	.005
NiO	.03	--	--	--	.03	.03	.002	.03	.005
CoO	.01	--	--	--	.01	.01	0	.01	.002
MnO	.18	--	--	--	.18	.18	.03	.18	.03
CuO	.02	--	--	--	.02	.02	.005	.02	.003
PbO	0	--	--	--	0	0	0	0	0
ZnO	0	--	--	--	0	0	0	0	0
$GaO_{3/2}$	.001	--	--	--	.001	.001	.002	.001	.0002
VO_2	.02	--	--	--	.02	.02	.02	.02	.003
$ScO_{3/2}$	.010	--	--	--	.010	.010	.012	.010	.0016
$YO_{3/2}$	0	--	--	--	0	0	0	0	0
$YbO_{3/2}$	.0001	--	--	--	.0001	.0001	.0001	.0001	.00002
ZrO_2	.0005	--	--	--	.0005	.0005	.002	.0005	.00008
BeO	0	--	--	--	0	0	0	0	0
SrO	0	--	--	--	0	0	0	0	0
BaO	.0004	--	--	--	.0004	.0004	0	.0004	.00006
Total	107.92	.12	.03	.05	107.72	107.94	103.68	107.98	17.11
Less ($HO_{1/2}$+F)	7.92	--	--	--	7.92	7.94	3.68	7.98	1.27
Cations	100.00	.12	.03	.05	99.80	100.00	100.00	100.00	15.84

[1] Estimated mode of analyzed material:

Amphibole	98.80
Epidote	1.00
Albite	.05
Sphene	.12
Rutile	.03
	100.00

$$C_a = 1.0103C_{ae} - 0.0103C_{ea},$$

$$C_e = 1.0156C_{ea} - 0.0156C_{ae}.$$

Substituting appropriate values of C_{ae} and C_{ea}, constituent by constituent, in these equations, yields calculated analyses of pure amphibole and pure epidote. Thus, for SiO_2 in the pure amphibole:

$$\begin{aligned} SiO_2 &= 1.0103 \times 41.97 - 0.0103 \times 36.97 \\ &= 42.402 - 0.381 \\ &= 42.02. \end{aligned}$$

And so on (see tables 10 and 11).

AMPHIBOLE FORMULA

1. Let the amphibole be represented by the following formula:

$$(\mathrm{Na,K})_x\mathrm{Ca}_s\left[\mathrm{R}^{+2}{}_t\mathrm{R}^{+3}{}_{y+z+u}(\mathrm{VI})\frac{u+w}{2}\right]^{\mathrm{VI}} \left[\mathrm{Si}_{8-w-x-y}\mathrm{P}_w\mathrm{Al}_{x+y}\right]^{\mathrm{IV}}\mathrm{O}_{22}\left[\mathrm{O}_{2-v}\mathrm{F}_v\mathrm{H}_{2-v-z}\right],$$

where:

IV and VI at the upper right of brackets designate coordination numbers, (VI) in the formula represents vacancies in six-coordination position.

s = total Ca,

t = total divalent ions, except Ca,

$u = R^{+3}$ in direct substitution for R^{+2} in six-coordination position, the excess charge being balanced by $u/2$ vacancies in the six-coordination positions. Thus, u is equal to the total amount of R^{+3} minus the R^{+3} in coupled substitution in six- and four-coordination positions, minus the Al^{IV} balanced by Na and K, and minus the $(R^{+3})^{VI}$ balanced by deficiency of H.

That is,

$$u = R^{+3} - 2Al^{IV} - (z + Na + K),$$

TABLE 11.—*Calculation of epidote formula*

[Sample No. A-BM-53-139a, Lab No. 14330b. Analyses from table 1. 0, looked for and not found]

	Molecular (cation) present								
	Analyzed material	Impurities[1] (except amphibole)			C'_{ea} (epidote plus amphibole)	$C_{ea}=\frac{100\,C'_{ea}}{\Sigma C'_{ea}}$	C_{ae} (from table 10)	C_e (pure epidote) $=1.0156\,C_{ea}-0.0156\,C_{ae}$	Formula Nos. (Si=6)
		Sphene	Rutile	Albite					
SiO_2	36.90	0.36	--	0.03	36.51	36.97	41.97	36.89	6.00
$AlO_{3/2}$	29.22	--	--	.01	29.21	29.57	13.42	29.82	4.85
$FeO_{3/2}$	6.81	--	--	--	6.81	6.89	2.53	6.96	1.13
FeO	.73	--	--	--	.73	.74	9.10	.61	.10
MgO	.70	--	--	--	.70	.71	16.27	.46	.07
CaO	24.11	.36	--	--	23.75	24.05	9.62	24.27	3.95
$NaO_{1/2}$	.60	--	--	.01	.59	.60	5.26	[2].53	0
$KO_{1/2}$	.10	--	--	--	.10	.10	.88	[2].09	0
$HO_{1/2}$	3.61	--	--	--	3.61	3.65	7.88	3.58	.58
TiO_2	.46	.36	0.10	--	.00	.00	.51	.00	.00
$PO_{5/2}$	.14	--	--	--	.14	.14	.14	.14	.02
F	.03	--	--	--	.03	.03	.06	.03	.005
$CrO_{3/2}$	.08	--	--	--	.08	.08	.03	.08	.01
NiO	.002	--	--	--	.002	.002	.03	.002	.0003
CoO	0	--	--	--	0	0	.01	0	0
MnO	.03	--	--	--	.03	.03	.18	.03	.005
CuO	.005	--	--	--	.005	.005	.02	.005	.0008
PbO	0	--	--	--	0	0	0	0	0
ZnO	0	--	--	--	0	0	0	0	0
$GaO_{3/2}$	.002	--	--	--	.002	.002	.001	.002	.0003
VO_2	.02	--	--	--	.02	.02	.02	.02	.003
$ScO_{3/2}$	.012	--	--	--	.012	.012	.010	.012	.002
$YO_{3/2}$	.004	--	--	--	.004	.004	0	.004	.0006
$YbO_{3/2}$	.0001	--	--	--	.0001	.0001	.0001	.0001	.00002
ZrO_2	.002	--	--	--	.002	.002	.0005	.002	.0003
BeO	0	--	--	--	0	0	0	0	0
SrO	.07	--	--	--	.07	.07	0	.07	.01
BaO	0	--	--	--	0	0	0	0	0
Total	103.64	1.08	.10	.05	102.41	103.68	107.94	103.61	16.74
Less ($HO_{1/2}+F$)	3.64	--	--	--	3.64	3.68	7.94	3.61	.58
Cations	100.00	1.08	.10	.05	98.77	100.00	100.00	100.00	16.16

[1] Estimated mode of analyzed material:

Epidote	97.27
Amphibole	1.50
Sphene	1.08
Rutile	.10
Albite	.05
	100.00

[2] Dropped as contaminants.

v = total F, substituting for OH,

w = total P,

x = total (Na,K), and Al^{IV} balanced by (Na,K),

$y = Al^{IV}$ in coupled substitution for $(R^{+3})^{VI}$,

z = deficiency of $H = 2 - F - H$.

(All values are calculated on the basis of $O + OH + F = 24$.)

2. From the analysis of pure amphibole, calculate the number of cations associated with each oxide on the basis of $(O + OH + F) = 24$.

3. Ions are assigned to different positions in the structural formula as follows:

Na, K, Ca, R^{+2}, Si, P, H, and F occur in only one position each in the structural formula, so the assignment of values is unambiguous. Each is assigned the total amount of that ion.

O outside of the OH position is equal to 22.

O in the OH position is equal to $2 - F$.

$Al^{VI} = 8 - Si - P$.

$R^{+3}_{y+z+u} = R^{+3} - Al^{IV}$.

Assignment of other values is evident from the structural formula and from the explanation of the symbols used in it (see 1, above).

EPIDOTE FORMULA

1. The structural formula of epidote, in terms of the contents of the unit cell determined by Ito, Morimoto, and Sadanaga (1954), is

$$[Ca_4]^{VIII}[Al_4(Al,Fe^{+3})_2]^{VI}[Si_6]^{IV}O_{24}(OH)_2,$$

where the Roman numeral superscripts indicate the coordination position of the cations. The term $(Al, Fe^{+3})_2$ in six-coordination position indicates that Fe^{+3} can substitute for Al only at particular sites in the lattice, and only to a maximum ratio of Fe^{+3}/Al of 1:2. Small amounts of R^{+3} may probably substitute for Al, and small amounts of R^{+2}, for Ca.

2. The formula composition of epidote can be calculated from a chemical analysis in several ways—for example, by setting $O+OH$ equal to 26, or by setting Si equal to 6. As the determination of H_2O^+ appears commonly to be of doubtful accuracy, and as Si is present very nearly or precisely in stoichiometric proportions, it appears generally to be better and more convenient to calculate the formula on the basis of $Si=6$.

3. In table 11, the calculated analysis of pure epidote, column headed "C_e (pure epidote)," is derived by methods described above and is recalculated to the basis of $Si=6$ by multiplying each term by 6/36.89. The subscript value for O in the formula (oxygen not combined with H) is determined by summing the oxygen associated with the cations in their oxide formulas and subtracting the equivalent of ½ H (to provide the additional O for OH determined as H_2O). The values for the different ions are then assigned to their proper place in the structural formula.

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