

THE GEOLOGY OF THE
BURKE QUADRANGLE, VERMONT

By

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VERMONT GEOLOGICAL SURVEY

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Published by

VERMONT DEVELOPMENT DEPARTMENT

MONTPELIER, VERMONT

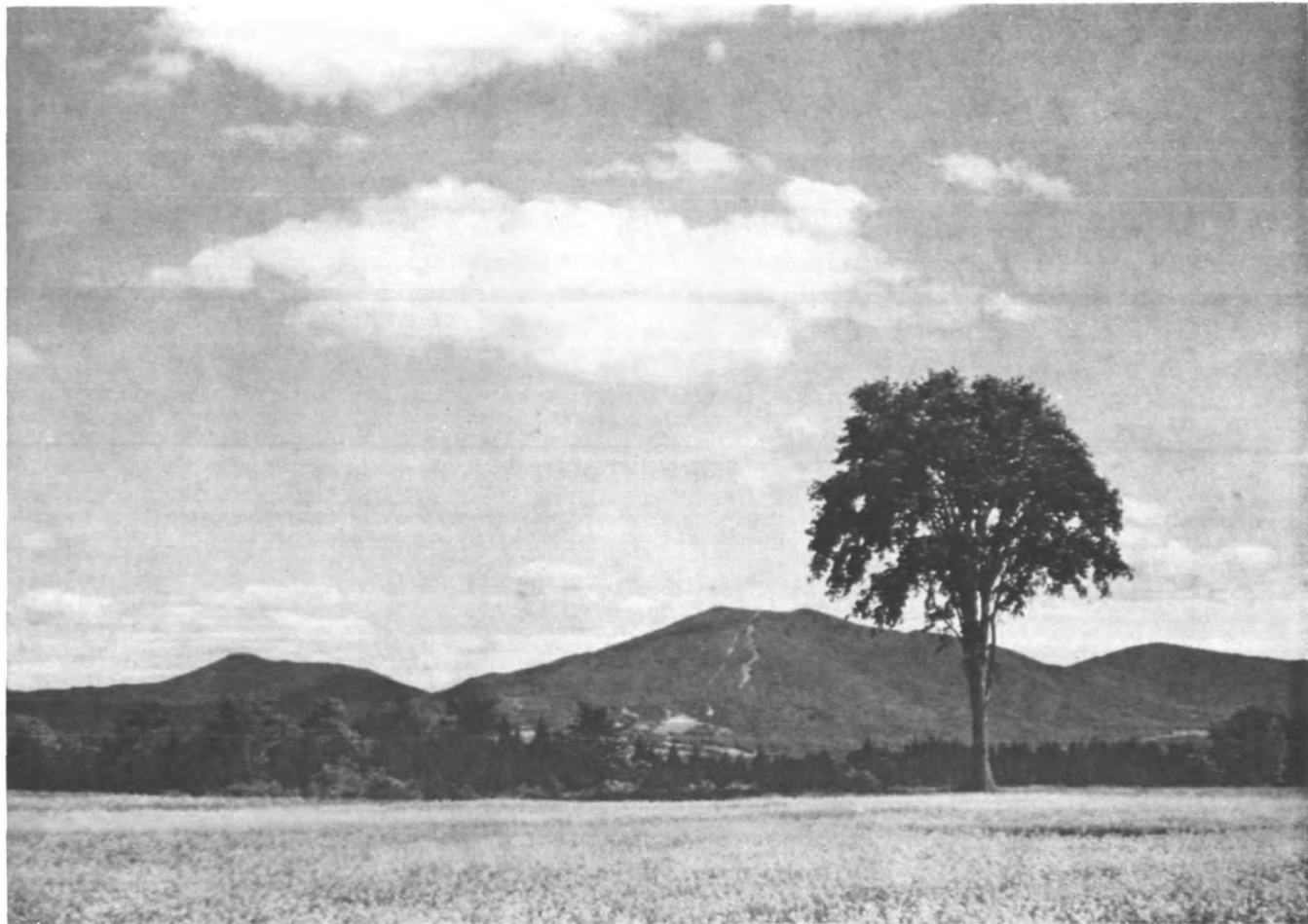


Figure 1. (Frontispiece). Burke Mountain, Umpire Mountain (left), and the northern spur of Kirby Mountain (right) are resistant masses of hornfelsed Gile Mountain formation and granite. View southeastwards from near Marshall School.

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By
BERTRAM G. WOODLAND*

ABSTRACT

The Burke quadrangle is located in northeastern Vermont, 18 miles south of the International Boundary and about 10 miles west of the New Hampshire boundary. It is largely a mountainous region lying east of the Green Mountain anticlinorium. The considerable amounts of granitic rocks link the Burke area with the New Hampshire plutonic belt.

Three metamorphosed sedimentary formations are found in the quadrangle: the Albee formation, of Ordovician age, which is part of the New Hampshire sequence of metamorphic rocks, and the Waits River and Gile Mountain formations, of presumed Silurian and Devonian age, which belong to the Vermont sequence of rocks and which have a wide distribution in the central and eastern parts of the state. The two sequences are separated by the Monroe fault, but an important unconformity may also occur between them. The Albee formation consists of schist, quartzite, and amphibolite. The Waits River formation is made up of impure limestone, phyllite, schist, quartzitic beds, amphibolite (representing volcanic tuff, lava, and intrusives), and acid volcanics and intrusives. The Gile Mountain formation is comprised of alternating bands of phyllite, schist, quartzose phyllite, and amphibolite. Granite has intruded the sedimentary rocks and in places the two form mixed zones of hornfels and granite.

The meta-sedimentary rocks are strongly deformed and show in their minor structures evidence of two phases of movement: (1) an earlier folding accompanied by a well-developed schistosity, which is generally parallel to the bedding, and (2) a later one which folded the schistosity and produced a slip cleavage. The two movements are probably successive phases of the Acadian orogeny. The strata have a steep dip mainly to the east-southeast, but they dip more gently in the

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western part of the quadrangle and dip steeply to the east-northeast in the northern part. The rocks are part of a major north-northeasterly-plunging structure which extends westwards for some 10 to 15 miles and the central zone of which is the Willoughby arch, an anticlinal structure in the Waits River formation which may mark the culmination of a large-scale recumbent fold; the axial plane of the fold parallels the arch.

Metamorphism has occurred in two distinct phases: (1) a lower grade phase, which accompanied the formation of the earlier folds and the schistosity, and (2) a later higher grade phase, which was post-deformational in the Burke area. It was caused by an increase in temperature of the rocks which were at moderate depth in the earth's crust, and resulted in the growth of porphyroblasts of actinolite-tremolite, biotite, garnet, and staurolite. In the final stage granitic magma was intruded and the adjacent rocks were thermally metamorphosed to a high grade, with the appearance of andalusite and sillimanite.

INTRODUCTION

Location

The Burke quadrangle is located in northeastern Vermont between 44° 30' and 44° 45' North Latitude and 71° 45' and 72° 00' West Longitude (Fig. 2). It has an area of approximately 216 square miles and includes parts of Caledonia, Essex, and Orleans counties.

Methods of Study

Thirty-eight weeks were spent mapping the Burke quadrangle during the summers of 1955, 1956, and 1957, and an additional two weeks were spent in the field in June of 1958. Field mapping was carried out by means of traverses spaced as frequently as time and terrain permitted. Streams in particular were followed as much of the area is mantled with glacial deposits. Often, however, the streams had not eroded down to bedrock so the hillslopes had to be searched for outcrops. Locations were determined by sighting or by pace and compass. In the forested country compass traverses were made; locations were found by the use of a pocket altimeter and by frequent stereoscopic examination of air photographs along the route. Outcrops were located on three inches equal one mile photographic enlargements of the one inch equals one mile U. S. Geological Survey topographic map of the Burke quadrangle, published in 1951.

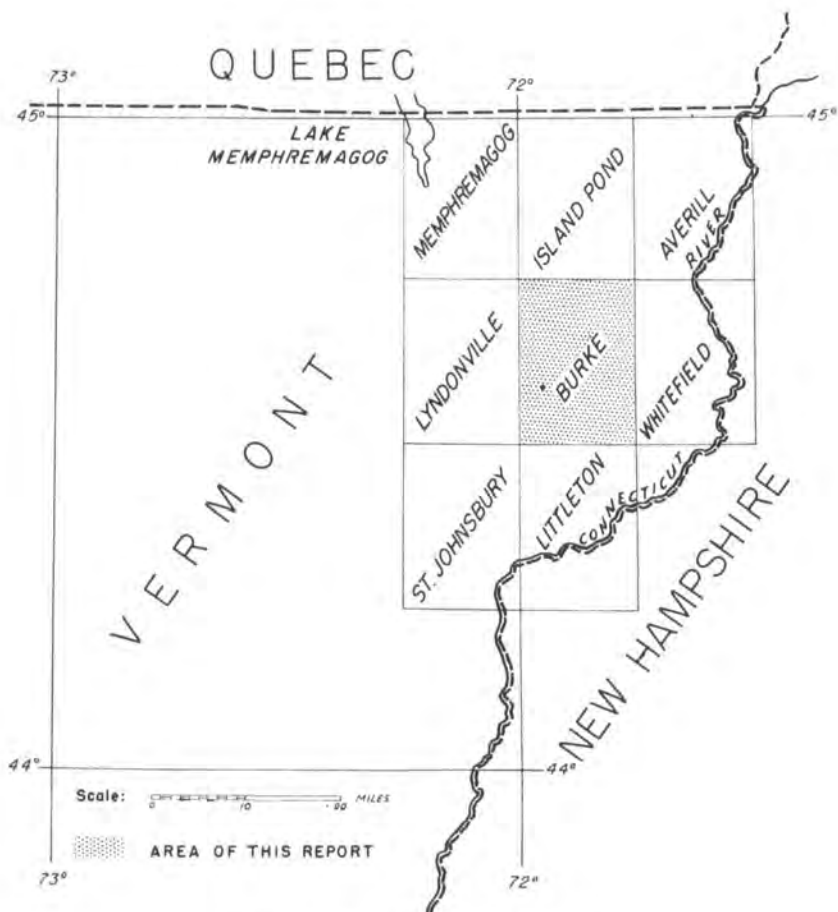


Figure 2. Location of Burke quadrangle.

A total of 3856 measurements of structural features were made in the field, including the strike and dip of bedding, schistosity, cleavage, and joints and the trend and plunge of fold axes and lineations. In addition, 31 trends of glacial striae were recorded. Hand specimens totalling 1819 were collected for detailed lithologic study; of these, 1029 were marked with their field orientation so that they could be re-oriented in the laboratory for further structural determinations.

In the laboratory all of the hand specimens were examined under a binocular microscope. Many planar and linear structures were measured

on the oriented specimens. Re-orientation was done by placing an oriented specimen in a clamp attached to a universal ball joint which permitted the setting of the specimen in its correct geographic position. A Brunton compass was used to position the specimen and to measure additional planes and fold axes; for some measurements an adaptation of a contact goniometer was used. When planes were inaccessible or only showed as lineations on two or more non-parallel surfaces, their orientation was determined by sighting along a plate, mounted on a ball joint, and placing the plate parallel to the plane to be measured. The attitude of the plate could then be determined readily by Brunton compass. The pitch of linear features was measured on oriented (or re-oriented) surfaces and their plunge determined by construction on a Wulff stereonet. The structural data were plotted on maps (see Plates 2, 3, and 4); they were also plotted on the lower hemisphere of a Schmidt stereonet for the purpose of preparing stereograms for the quadrangle as a whole and for twelve subareas, whose arbitrary boundaries are shown on Plate 4. Nearly all of the stereograms were contoured and used to obtain additional structural data, the details of which are given in the section on Structure.

A total of 201 thin sections were examined under the polarizing microscope.

All directions of strike of planar surfaces are stated in this report from 0° to 180° east of north.

Reference to localities is made by a letter and four-number grid system. The letter refers to the particular ninth of the quadrangle (Fig. 3), the first two numbers are inches and tenths of an inch from the western edge of the ninth, and the last two numbers are inches and tenths of an inch from the southern edge of the ninth.

Acknowledgments

The field work was carried out with the financial assistance of the Vermont Geological Survey. The advice and support of Dr. Charles G. Doll, State Geologist, who also accompanied the writer into the field on a number of occasions, are gratefully acknowledged. The writer had the benefit of discussions in the field several days in 1957 with Dr. James B. Thompson, Jr., of Harvard University. Discussions and joint field trips were held with Dr. Warren Johansson, who was mapping the adjoining Guildhall quadrangle, to consider problems of the mutual boundary. Similar consultations were held with Dr. Bruce Goodwin, who was mapping the adjoining Island Pond quadrangle, and with Dr.

A	B	C
D	E	F
G	H	I

Figure 3. Index letters to Burke quadrangle ninths.

John Dennis, who completed mapping of the Lyndonville quadrangle during the author's first field season and was later revising the Littleton quadrangle mapping.

The writer is pleased to acknowledge discussions and advice given freely by Dr. John C. Haff of Mount Holyoke College and by Dr. Walter H. Newhouse of the University of Chicago. He has also benefited from the assistance of his colleagues at the Chicago Natural History Museum, Dr. Eugene S. Richardson, Jr., Dr. Rainer Zangerl, Dr. Albert Forslev, who made a number of X-ray diffraction determinations, and Miss Maidi Wiebe, who helped in the preparation of the illustrations.

Mr. Malcolm Hibbard in the summer of 1956 and Mr. Christos Alex in the summer of 1957 provided valuable field assistance.

The residents of the Burke area were always extremely helpful; in particular the writer would like to mention Mr. and Mrs. Thomas Gould of East Burke, Mr. Edward Lund, Fire Warden of Granby, and Mr. and Mrs. Kenneth Hoffman of Kirby.

Last, but not least, the writer wishes to express his great debt to his wife, Dr. Mary Vogt Woodland, who was a constant source of encouragement and provided much help in all aspects of the work.

The writer gratefully acknowledges the provision of 54 thin sections by the Vermont Geological Survey and 128 thin sections by the Department of Geology, Mount Holyoke College.

Physiography

The Burke quadrangle lies in the northern portion of the New England Division of the Appalachian Highlands. Approximately the western third of the area consists of partially forested hills and valleys trending north-south; almost all of the farm land of the quadrangle is restricted to this section. Its relief is subdued, with the lowest point 660 feet, in the Passumpsic riverbed in the southwestern corner of the quadrangle, and with the land surface rising gradually northwards to over 1900 feet around Newark Street (Fig. 4). This area is underlain by the calcareous and associated rocks of the Waits River formation, and the trend of the hills and valleys follows in general the regional strike of the strata. Some of the lower hills are composed of glacial deposits, but most have a rock core and are not unlike rock drumlins, e. g., Graves Hill.

The eastern two-thirds of the quadrangle together with the northwestern part is mountainous and virtually under continuous forest cover. Burke, Umpire, East Haven, East, and Seneca mountains all rise to over 3000 feet, East Mountain being the highest with an elevation of 3420 feet. An extensive low-lying boggy area, largely below 1200 feet and ringed with mountains, occupies much of the Township of Victory in the southeastern part of the quadrangle. Burke and Umpire mountains (Fig. 1) stand above the surrounding country like monadnocks and both have a characteristic radial drainage pattern. The rugged nature of this terrain can be correlated with its underlying rock types. Thus, Kirby, Burke, Umpire, and East mountains are composed of a resistant hornfels-granite complex; East Haven and Seneca mountains are composed of rocks of the Gile Mountain formation, in part hornfelsed, and biotite granite. The aforementioned low-lying area in the southeast is believed to be mainly underlain by amphibole granite, an earlier and presumably more easily eroded granite than that in Burke, Kirby, and the other mountains.

The drainage of the area is mainly to the south via the East and West branches of the Passumpsic River which join just north of Lyndonville and via the Moose River. These are all tributaries of the Connecticut River. The northern flank of Bull and Seneca mountains in the north-northeastern part of the quadrangle drains north into the Nulhegan River in the Island Pond quadrangle and thence into the Connecticut River. In the east-northeastern part of the Burke quad-



Figure 4. Subdued topography developed on the Waits River formation in the northwestern portion of the Burke quadrangle. View from the road to the summit of Burke Mountain. Mount Pisgah (in the Lyndonville quadrangle) in left background; Bald Mountain (in the Island Pond quadrangle) in right-center background.

range the North Branch of Paul Stream, Madison Brook, and Stony Brook flow eastwards into Paul Stream in the Guildhall quadrangle and thence into the Connecticut River.

There are two dams and accompanying hydro-electric plants on the Passumpsic River, one just west of Red Village and the other in the extreme southwestern corner of the quadrangle.

There are several ponds in the area, Newark, Bald Hill, and Center ponds being the largest. The ponds are chiefly the result of poor drainage impeded by thick glacial deposits. Some of the small ponds are kettle lakes, such as Duck Pond, northwest of East Burke, and Lily Pond, southeast of Lyndonville.

The whole of the Burke quadrangle was glaciated during the Pleistocene. The evidences of glaciation are many; they include the thick glacial deposits which mantle most of the ground and the erratics, amongst which one of the most interesting is the large anvil stone of amphibolite now standing at the west side of the road north of Bundy School (D3847). Areas extensively covered with glacial deposits are shown on Plate 1. The presence of glacial till on the highest peaks such as Burke Mountain proves that the area was completely submerged beneath a thick cover of ice. Glacial striae and grooving can be seen in places, and their trends are recorded on Plate 1. These show that the ice moved generally in a south-southeasterly direction. Roche moutonnées are found in several areas; some of the best examples are on the summit of Spur Ridge (G2907) and southeast of Gallup Mills (I2749). Their long axes are oriented in a north-northwest-south-southeast direction, which is in keeping with the striae trends.

Late glacial deposits of sand and gravel abound in the valleys. Sinuous eskers parallel much of the length of the East and West branches of the Passumpsic, e. g., northeast of White School (E0628) and southeast of West Burke (D1336). Kames are also found along the sides of the valleys and give rise to characteristic hummocky topography, such as along Route 114 north of East Burke (D3409) and northwest of Ward School (D1214). They are composed of crudely stratified sand and gravel, often with many large boulders mainly of local origin. Large quantities of stratified sand with minor gravel occur in the lower reaches of the East and West branches of the Passumpsic and east of Lyndonville; about 50 feet are exposed in the bank east of the Canadian Pacific Railroad just north of Lyndonville (G0231). These deposits which choke the Passumpsic valley represent the remnants of outwash, the upper elevation of which lies at about 860 feet (probably rising to the

north up the valleys), and which were deposited when the ice paused in its recession at some point to the north, possibly just south of the south end of Lake Willoughby (Lyndonville quadrangle) at the divide between the St. Lawrence and Connecticut drainage systems. The Passumpsic River has now cut deeply into these deposits; a meander scarp in the latter rises above the present floodplain of the East Branch north-northeast of Lyndonville (G0837). In the Egypt area (e. g., G0553) the sand has been and still is subject to wind action and small active sand dunes are present. Varved clay deposits occur at widely scattered points in the southwestern and west-central parts of the Burke quadrangle (e. g., G1606, D1616) and at one locality in the southeast (I0412). The varved clay passes upwards into sandier deposits and finally into sand and gravel. It represents deposits laid down in temporary lakes and perhaps underlies the outwash sands, which were the final deposits laid down. Varved clay, however, is found at higher elevations up the brooks draining the west slopes of Kirby Mountain; the highest elevation at which the writer found it is 1620 feet, in Hawkins Brook (H0213). These higher occurrences were probably laid down in small lakes impounded between the ice that lay in the valley and the mountain side. The one example in the southeastern part of the quadrangle suggests that the low-lying area of Victory also was the site of a late glacial lake.

The "biscuit-board" pattern of the contours on Seneca, East, East Haven, Umpire, and Kirby mountains is remarkably similar to that formed by cirque development in areas of mountain glaciation. Cirque forms are well shown in the headwaters of North Branch, Paul Stream (C2028), the headwaters of the tributaries on the northern side of Madison valley (particularly C2318), the headwaters of the west branch of Greer Brook (E3930), the headwaters of Dish Mill Brook (H2654), and the headwaters of Barnes and Simpson brooks on the western slopes of Kirby mountain (H0628; H0333). They were probably eroded by mountain glaciers after the main continental glaciation. The glacial Madison valley is a hanging valley which plunges steeply eastwards via a series of waterfalls near the eastern boundary of the Burke quadrangle and extending into the Guildhall quadrangle to the over-deepened valley of Paul Stream.

Settlement

Settlement is essentially restricted to the western third of the Burke quadrangle coinciding roughly with the subdued topography developed

on the Waits River formation. It includes the eastern outskirts of Lyndonville and the villages of East Burke and West Burke which are the principal areas of population and the communities of Burke Hollow, Newark Street, East Haven, East Lyndon, and Red Village.

Dairy farming, lumbering, and tourism are the major industries of this settled area. Near East Burke Darling State Forest Park, embracing an area of more than 1745 acres of Burke Mountain and the northern part of Kirby Mountain, is a scenic attraction during the summer and a skiing center in the winter. Extensive panoramic views are obtained from the summit of Burke Mountain (Fig. 4).

The remainder of the Burke quadrangle is heavily forested and much more inaccessible. The small population is restricted to a few isolated communities such as Gallup Mills, Granby, and Victory. The cutting of lumber and pulp wood is the only significant industry; practically this whole area has been cut over at least once. Gallup Mills was a thriving community at the turn of the century when lumbering in the Moose valley was an important operation and a railroad extended northward up the Moose valley from North Concord.

Evidences of more extensive farming are found throughout the quadrangle in the pastures and orchards encroached upon by forest and by the presence of cellar holes, stone walls, and corduroy roads in the woods, e. g., north of Newark Hollow, north of Granby, around Lost Nation, and around Kirby Pond.

Previous Work

Early geological work in the region was carried out by Hitchcock (1861) and S. R. Hall (1861). Richardson (1906) made a number of references to the area; he described the pink granite of Newark and referred to fibrolite on Burke Mountain and to diabase dikes in Newark. Dale (1909, 1923) described the granite quarries in Newark and Kirby. The small part of the Township of Westmore that falls in the Burke quadrangle was included in the geological map of Jacobs (1923). Detailed geological accounts of the neighboring quadrangles (Memphremagog, Doll, 1951; Lyndonville, Dennis, 1956; Littleton, Eric and Dennis, 1958; and St. Johnsbury, L. M. Hall, 1959) have been published by the Vermont Geological Survey. Geological mapping of the Guildhall, Island Pond, and Averill quadrangles was in progress during the preparation of this report and reports have now been published (Johansson, 1963; Goodwin, 1963; Myers, 1964).

STRATIGRAPHY

General Statement

Three formations have been mapped in the Burke quadrangle—the Albee, Waits River, and Gile Mountain—all of whose contacts have in general a north-northeasterly strike. The latter two occur within the "Vermont sequence" while the Albee formation is part of the "New Hampshire sequence" (White and Jahns, 1950). The two sequences are separated by the Monroe fault, which was recognized to the south (Eric, White, and Hadley, 1941) and has been mapped, with some gaps, for over 80 miles from the Hanover quadrangle (Lyons, 1955) northwards into the Littleton quadrangle (Eric and Dennis, 1958), and is believed to occur at North Stratford, New Hampshire (Billings, 1956). In the Burke quadrangle the line of the fault is largely interrupted by granite intrusion and high-grade metamorphism; the contact is recognized only for a distance of a little over a mile on the eastern border in the vicinity of Granby. No actual contact was discovered as the exposures are not plentiful along its assumed line, which appears to follow a small valley and strikes northeastward into the Guildhall quadrangle. In the latter the Meetinghouse slate is found immediately west of the Albee formation (Johansson, 1963) as it is in the Littleton quadrangle (Eric and Dennis, 1958). It has been suggested that the Monroe contact between the Vermont and New Hampshire sequences may be an unconformity; the arguments for and against a fault or an unconformity and the structural and stratigraphic implications have been fully discussed by Eric and Dennis (1958), L. M. Hall (1959), Cady (1960), and Johansson (1963). No new information bearing on this problem has been discovered by the writer in the Burke quadrangle.

Thicknesses: The only one of the three formations exposed in the Burke quadrangle that has both an eastern and a western contact is the Gile Mountain. There is but a small portion of the Albee present along the eastern border of the quadrangle; the western contact of the Waits River formation lies in the Lyndonville quadrangle to the west (Dennis, 1956, Plate 1). White and Billings (1951) give a figure of 10,000 feet for the thickness of the Waits River in the Woodsville quadrangle.

The Gile Mountain formation has a maximum width of outcrop of about eight miles measured normal to the strike and along a line from a point between East Burke and Hartwellville in the west and the

Monroe fault near Granby in the east. The average dip of this section is between 60° and 70° southeastwards. Thus the apparent thickness of the formation at this latitude is 36,000 to 39,000 feet. Estimates of true thickness are exceedingly difficult to make in highly deformed metamorphic rocks. This is true in the Burke area and a study of the literature shows it to be so for eastern and central Vermont in general. The prevalence of minor folding suggests that repetition of beds by folding is probably important. Geologists working in other areas of eastern and central Vermont have applied a rough factor to correct for this thickening of beds. The writer is certain that the great apparent thickness of the Gile Mountain formation in the Burke quadrangle is the result of considerable repetition, but there are insufficient data to estimate its extent. Doll (1944) gives a figure of 6,500 feet for the Gile Mountain (excluding the Meetinghouse slate) in the Strafford quadrangle and White and Billings (1951) estimate a thickness of 6,000 to 7,000 feet in the Woodsville quadrangle. Thickening of the order indicated suggests repetition by major folding. Although there is no direct evidence for this, it is possible that it exists; this is discussed in the section on Major Structural Features.

Age of formations: No data bearing on the age of the three formations were found in the Burke quadrangle. The Albee formation was considered Ordovician in age by Billings (1934, 1937) based on its relations in New Hampshire. The determination of the age of the Waits River and Gile Mountain formations has been fraught with difficulties. Two possibilities have been discussed and supported by various geologists. An Ordovician age was suggested by Richardson (1902) and in more recent years by Cady (1950), White and Billings (1951), Lyons (1955), and Billings (1956). The rocks of southern Quebec along the line of strike of the Waits River and Gile Mountain formations are now called the St. Francis formation and assigned to the Ordovician (Cooke, 1950, 1957). A Silurian or Devonian age has been given to the Waits River and Gile Mountain formations by Doll (1943a, 1943b, 1944, 1951), Hadley (1950), Cady (1956), Dennis (1956), Eric and Dennis (1958), and L. M. Hall (1959). Billings (1956), although he placed the formations in the Ordovician, considers a Silurian or Devonian age a distinct possibility. In keeping with most of the recent views, the Waits River and Gile Mountain formations are assigned a Silurian and/or Devonian age in this report.

Full discussion of the age problem is given in Billings (1956), Dennis (1956), and Cady (1960).

Sequence: The correlation and order of the formations of the Vermont

sequence have been given a number of interpretations (see L. M. Hall, 1959, pp. 37-41 and Fig. 15, and Cady, 1960, for a complete summary of views). The most favored order is that the rocks are progressively younger eastwards from the Green Mountains so that the Waits River is followed upwards by the younger Gile Mountain and by Meetinghouse slate. The reverse order has also been suggested by White and Jahns (1950), who reject it, by Billings (1956), who considers it unlikely, and by Murthy (1958, 1959a, 1959b), who favors it. A further complication is the suggestion that the Northfield slate, a formation that underlies the Waits River to the west in the Hardwick quadrangle, is the equivalent of the Meetinghouse slate. This involves difficulty in explaining the eastern contact of the Gile Mountain formation and Meetinghouse slate or the western contact of the Northfield slate and the Waits River formation depending upon whether the Waits River is regarded as being older or younger respectively than the Gile Mountain. In the first case the Waits River would be apparently missing and in the latter case the Gile Mountain; these omissions may be explained tectonically or by facies change, which would mean equivalence in part of the two formations. No evidence has yet been forthcoming to support either of these possible explanations or the equivalence of the Northfield and Meetinghouse slates. The relationship between the Waits River and Gile Mountain formations in the Burke quadrangle will be referred to again in the section on Major Structures.

ALBEE FORMATION

The Albee formation was named for Albee Hill, New Hampshire, in the Littleton quadrangle (Billings, 1934). In the Burke quadrangle it has very restricted outcrops, all of high metamorphic grade. The exposures northeast of the village of Granby present the more typical lithologies, as described by Billings (1937) and by Eric and Dennis (1958). Here the rocks include quartzose and pelitic beds in the staurolite grade—not unlike parts of the Gile Mountain formation. They are fine-grained, gray to green rocks, composed of quartz, abundant biotite, in places sericite, garnet (up to 1.15 mm.), staurolite (up to 1 mm. long), sparse to abundant chlorite, and rare plagioclase (Fig. 5).

Outcrops on Lees Hill and Miles Mountain are of high-grade sillimanite hornfels associated with granite. It is possible that these outcrops are part of the Albee formation, but at their high grade of metamorphism there is no way to distinguish them from the Gile Mountain formation, at least not those examined in detail by the writer. Possibly one dis-



Figure 5. Photomicrograph ($\times 20$) of staurolite-garnet-mica-quartz schist from the Albee formation near Granby (I4251). Biotite and sericite form a wavy schistosity (S_1) and are in part oriented into a cross-cutting cleavage (S_2). Staurolite, garnet, and biotite occur as porphyroblasts. Note linear trend of garnets with opaque inclusions subparallel to S_2 .

tion, which may reflect the metamorphic history rather than the composition of the original strata, is that these rocks contain in general more prominent sillimanite and fewer or no sericitic aggregates after andalusite.

Amphibolite: Northeast of Granby amphibolitic bands are common in the limited exposures of the staurolite zone. Quite distinctive from the amphibolites of the Waits River and Gile Mountain formations, these are green, fine-grained, foliated rocks, often with small white or pink elongate spots or streaks up to 7 mm. long, which may represent deformed amygdules. The amphibolite is composed of abundant blue-green hornblende 0.1 to 0.9 mm. long, some altered to chlorite, and of very fine-grained interstitial areas (grain size 0.01 to 0.04 mm., seldom up to 0.2 mm.) containing rare, poorly twinned plagioclase (var. labradorite?), untwinned plagioclase, and quartz. Some of the interstitial quartz is

rather larger, up to 0.36 mm. Poikiloblastic garnet is common (up to 0.9 mm.), apatite is sparse; there is some rare sericite, and iron ores are common. Thin quartz veins cross-cut the foliation.

These amphibolites represent metamorphosed diabase sills, dikes, or flows.

WAITS RIVER FORMATION

Richardson (1906) named the thick sequence of limestone and schist in eastern Vermont the Waits River limestone, re-named the Waits River formation by Currier and Jahns (1941). Doll (1951) mapped calcareous rocks in the Memphremagog quadrangle as the Ayers Cliff and Barton River formations and Dennis (1956) mapped the southward continuation of the Barton River formation in the Lyndonville quadrangle as Waits River. Eric and Dennis (1958) show this formation to outcrop in the northwestern corner of the Littleton quadrangle. Lithologically similar rocks outcrop, as continuations of the Lyndonville and Littleton occurrences, throughout the western part of the Burke quadrangle, essentially west of Route 114, but also south of East Burke in the structurally complex area of East Lyndon and Simpson Brook (Plate 1).

Lithologic types include impure recrystallized limestone, calcareous phyllite, calcareous quartzose phyllite, calcareous quartzite, phyllite, quartzose phyllite, amphibolite, meta-quartz keratophyre, calc-silicate hornfels, and staurolite-garnet-biotite schist. These occur interbedded in varying proportions and thicknesses, in bands from a few inches to several feet thick, more rarely as much as nine to ten feet thick. Some outcrops contain considerable impure limestone with subordinate phyllite or schist and quartzose rocks, while others contain little or no limestone. Outcrops believed to be typical of the general occurrence in the quadrangle are well-exposed in Mountain Brook (G2446), in Dish Mill Brook (D3104), in Calendar Brook (D0116), and in Roundy Brook, north of Burke Hollow (D3033). North of East Sutton Ridge and Packer Cemetery, granitic sheets, sills, and tongues are prevalent, intruded mainly parallel with the limestone bedding, with the consequent development of calc-silicate minerals and occasionally of calc-silicate rocks and skarn. Amphibolite and meta-volcanics are confined to the more easterly outcrops.

Impure limestone: This is a light gray to blue-gray and dark gray rock which weathers to a buff-brown color. It is compact and fine-to medium-grained, with a poor foliation formed by sericite, phlogopite, or biotite. It frequently contains much quartz and mica, thus grading into cal-

calcareous quartzose phyllite. Occasionally it has a fissile structure and weathers to a loose brown sand. Thin, recrystallized segregation veins of quartz, with some calcite, less than one-eighth inch thick parallel the foliation. Quartz veins, a few millimeters to a few feet thick, commonly occur, but in general are not so prevalent as in the Gile Mountain formation.

Between West Burke and the northern outskirts of Lyndonville the limestone outcrops are frequently well-bedded and regular, e.g., in Quimby Brook, at the western edge of the Burke map (G0054). Farther to the east and southeast the limestone bands are usually strongly deformed internally and are discordant at their contacts with phyllite, sometimes in a very intricate, complex fashion. Indeed, in some occurrences 2 to 3 inches thick, tightly folded limestone bands or lenses have external contacts which are tectonic in origin rather than original bedding, the calcareous material having flowed into its present position. These occurrences are typical of calcareous rocks and marble flow-folded under the influence of strong deformation. Flow folding is well exhibited in Mountain Brook (G2446), in Simpson Brook (G2727), and west of Burke Hollow (D2031). (See Figs. 6, 7, 8, and 9).

Microscopically the limestone contains calcite and quartz in varying proportions, the latter often about equal in amount to the calcite, and ranging in grain size from about 0.06 mm. to 0.2 mm. and up to 0.35 mm. Sericite is common, particularly in the more phyllitic limestone; phlogopite or biotite is common (up to 0.3 mm.). Chlorite flakes, up to 0.87 mm. and randomly oriented, are present in some thin sections. In one slide cut from a flow-folded specimen the flakes are bent and deformed, with undulose extinction. Tremolite porphyroblasts (sometimes actinolitic) are commonly developed in radial, brush-like aggregates, in places up to 2 cm. in diameter (Fig. 10). Opaque minerals are ubiquitous; fine graphite dust and sulfide grains are common. In the higher metamorphic grades, e. g., on Burke Mountain and adjacent to the granite masses in the Newark area, the impure limestone has been converted into calcisilicate hornfels with diopside and grossularite. The rocks range in texture from those with grains less than 0.5 mm. to those with garnet dodecahedrons up to 5 mm., amphibole nearly 5 mm. long, and diopside up to 3 mm. long.

Pelitic rocks: Interbedded with the calcareous rocks, the pelitic rocks are usually well-foliated, biotite-rich, and dark-colored, with an abundance of fine opaques and much pyrite, which gives rise commonly to rusty-colored outcrops upon weathering. In general, the mineralogy of



Figure 6. Sharp minor folds in Waits River impure limestone in Mountain Brook (G2446), looking south.

the pelitic rocks is similar to that of the pelite described below in the Gile Mountain formation. Garnet and staurolite occur in the higher metamorphic grades. Andalusite and kyanite are very rare; kyanite occurs at B0347 and both are found together in an unusual occurrence on the east side of Route 114 (B1848) associated with abundant biotite and pegmatitic material (this is further described under Metamorphism). Tourmaline schist occurs at D4053 near pegmatitic veins.

The pelite bands in the southwestern part of the Burke quadrangle have been intensely folded and often appear as discontinuous outcrops surrounded by flow-folded limestone. They are also strongly cleaved, usually possessing one or more cleavages cutting the main schistosity.

Psammitic beds: Quartzose rocks are commonly interbedded with the limestone and pelitic strata and are similar to the psammitic beds in the



Figure 7. Flow folding in Waits River impure limestone in Simpson Brook (G2728), looking west.

Gile Mountain formation. The quartzose types are more plentiful in the zone transitional to the Gile Mountain than in the Waits River proper. They are fine-grained (the quartz grains are 0.03 to 0.3 mm.), usually with a well-developed sericite foliation, and have biotite both sub-parallel to the foliation and as porphyroblasts up to 0.45 mm.; opaque dust streaked in the foliation is sometimes present, and sulfides are ubiquitous, often elongated in the foliation. Chlorite occurs occasionally in the foliation in flakes 0.06 mm. across, but more commonly as porphyroblasts up to 0.65 mm. Small garnets are occasionally present. At one locality (G2552) in the more pelitic bands transitional to phyllite staurolite has been observed. Plagioclase (albite-oligoclase) is a minor constituent in



Figure 8. Flow-folded Waits River impure limestone cut by joints, which are healed by quartz. Note quartz rodding, caused by deformation of a quartz vein; rodding trends 152° and plunges 40° to the southeast (parallel to penknife, three and one-half inches long). In Simpson Brook (G2727), looking west.

some of the thin sections. Calcite may occur in small quantities in the bands transitional to the impure limestone. Poikiloblastic muscovite is present in some psammitic rocks, notably in the East Passumpsic valley between Hartwellville and White School. Accessory minerals are minute greenish tourmaline, zircon, sphene, and apatite. Quartz veins, from 0.5 mm. to a few centimeters thick, are frequent, usually paralleling the foliation and in places containing calcite; they are often folded and sometimes are cut by a cleavage.

Amphibolite: Amphibole-rich rocks occur in general in the belt of country comprising the transition between the Waits River and Gile



Figure 9. Flow-folded Waits River impure limestone with veins and masses of quartz. On hillslope west of Burke Hollow (D2031), looking east.

Mountain formations, e. g., along the East Passumpsic valley. The easternmost bands closest to the Gile Mountain formation probably can be correlated with the Standing Pond amphibolite described by Doll (1944) in the Strafford quadrangle and mapped by L. M. Hall (1959) in the St. Johnsbury quadrangle. The rocks are greenish in color and normally have a rather crude foliation parallel to that of the adjacent phyllite or limestone. The outcrop on Spur Ridge (G2908), however, is a massive sill-like exposure at least 12 feet thick, with only one boundary visible. Other similar outcrops nearby indicate a comparatively thick occurrence of amphibolite in Spur Ridge. The weathered surfaces show a marked reticulate structure, produced by the more resistant material of thin cross-cutting veins (Fig. 11). These veins probably represent metamorphic differentiation by secretory growth of the central quartzose-feldspathic material and enrichment of hornblende along the borders (Eskola, 1932; Ramberg, 1952, pp. 216-218 and Fig. 105).



Figure 10. Photomicrograph ($\times 20$) of Waits River impure limestone with calcite, quartz, and radiating "spongy" porphyroblasts of actinolitic tremolite. West slope of Graves Hill (G1326).

Usually the amphibolite is poorly exposed in scattered outcrops a few feet across which sometimes may be parts of a band 10 to 20 feet thick but which may also be bands 3 to 10 feet thick interbedded with phyllite and limestone. The weathered surfaces of amphibolite are frequently pitted as a result of solution of carbonates.

Amphibole is always a major constituent, medium to coarse grained and up to 19 mm. in length; it is usually a green, poikiloblastic, xenoblastic hornblende, although in some cases it may be actinolitic (Z varying from blue-green to dark green). Sometimes the amphibole has altered in part to chlorite. The groundmass of the amphibolite is fine-grained (0.02 to 0.1 mm.), generally of untwinned plagioclase and quartz (Fig. 12). Definite determination of the plagioclase is usually not possible because of the small grain size and diffuse grain boundaries. Twinned plagioclase (up to 0.5 mm.) is sometimes common, but more usually is rare, and the twinning is often indistinct. The twinned plagioclase seems



Figure 11. Reticulate weathered surface of amphibolite on the northern slope of Spur Ridge (G3008).

to vary in composition from a fairly sodic type to labradorite. Biotite (0.02 to 0.36 mm.) is usually present in small amounts, but in some occurrences it is common, giving a slightly purplish hue to the fresh rock surface. Chlorite (0.1 to 0.36 mm.), often penninite, is rare, but in some thin-sections it is common. A small amount of clinozoisite and epidote occurs very seldom. Opaque minerals include sulfides and ilmenite, altered in part to leucoxene. Accessory minerals include apatite (in grains up to 0.2 mm.) and sphene (up to 0.78 mm.), partly altered to leucoxene. Calcite is often present in small quantities, particularly when the amphibolite is interbedded with limestone. Tourmaline (pleochroic colorless to smoky gray, blue, or brown) occurs in a few thin sections; in one coarse-grained calcareous "inclusion" in amphibolite tourmaline is abundant in a stellate pattern and is associated with axinite and quartz.

Two unusual occurrences at G1010 and at G1015 merit special mention. These are green rocks with a well-developed schistosity, which has a



Figure 12. Photomicrograph ($\times 18$) of Waits River amphibolite showing poikiloblastic hornblende in a groundmass of plagioclase, quartz, calcite, apatite, chlorite, and iron ores. Northwest of Graves Hill (G0930).

silky, greasy feel. They are composed of abundant blue-green or colorless amphibole, which is granulated and much reduced; abundant minute flaky talc and chlorite (some chlorite in aggregates and lenticles); abundant dusty opaques and grains of iron ores; a little fine-grained felsic material; and occasional garnet and apatite. A thin section (Fig. 13) of this rock shows it to be made up of exceedingly fine-grained, ground-up minerals, streaked out in places into curved shear planes, sometimes paralleled by lenticular areas of chlorite and amphibole. These two occurrences are of amphibolite highly sheared to the point of mylonitization. In the exposure at G1015 normal amphibolite outcrops within 100 yards of the sheared rock.

Light-colored hornblende quartzose bands: These are characteristically developed as bands a few inches thick intercalated among phyllite and limestone beds in the transition zone between the Waits River and Gile Mountain formations. As they are more common in the latter, a description is deferred to the section dealing with the Gile Mountain.

Meta-acid igneous rocks: Metamorphosed acid lavas and perhaps tuffs

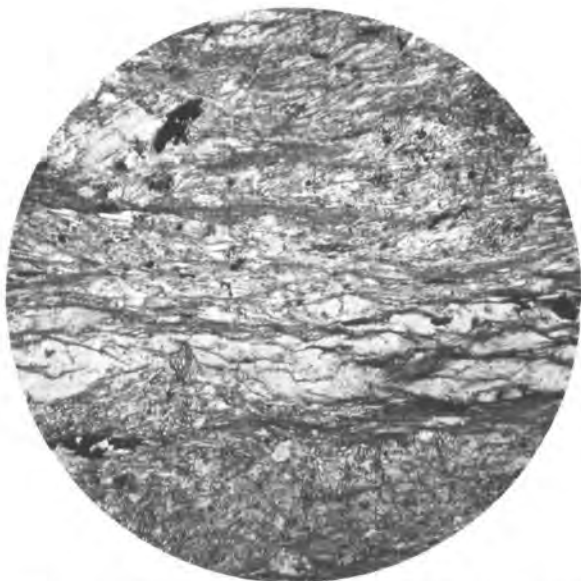


Figure 13. Photomicrograph ($\times 18$) of sheared amphibolite showing granulated amphibole, chlorite, garnet, opaques, ? talc. From the southern slope of Shonya Hill (G1015).

are exposed at a number of localities (Plate 1), and appear to represent a single volcanic episode. They occur in the transition zone between the Waits River and Gile Mountain formations. Limestone lies near most of the meta-acid outcrops; just south of one occurrence, at G3050, there is amphibolite. These meta-acid types are placed, therefore, within the Waits River formation.

Eric and Dennis (1958) refer to meta-rhyolite tuff associated with amphibolite just within the Gile Mountain formation of the Littleton quadrangle. It is probable that this occurrence should be correlated with the acid lavas and tuffs of the Burke quadrangle.

These meta-acid igneous rocks are gray, fine-grained, crudely foliated, and composed mainly of sodic plagioclase, quartz, and biotite. The plagioclase is albite or oligoclase, which in some thin sections is in micro-lites, often bent, 0.2 to 0.3 mm. long, in a felted or almost bostonitic texture (Fig. 14). In other sections the plagioclase is xenomorphic granular, often less than 0.05 mm., both twinned and untwinned. Some euhedral blastophenocrysts occur up to 0.5 mm., with inclusions of biotite. Quartz may be sparse to common; it is usually fine grained although some



Figure 14. Photomicrograph ($\times 18$) of a crudely-foliated meta-quartz keratophyre containing oligoclase in a felted texture, quartz, and biotite. From a small stream northeast of Mount Hunger (G2950).

coarser grained pods also occur. Biotite (or lepidomelane) is either scattered throughout, in flakes 0.02 to 0.25 mm., or is aggregated into trains producing a crude foliation, or appears as abundant porphyroblasts up to 4 mm. Muscovite is sparse, but in one thin section it is concentrated in linear aggregates. Calcite occurs sparsely in the groundmass and also in amygdules, sometimes with quartz. A little sphene and apatite are present in some specimens. Opaque dust may be plentiful; iron ores, particularly sulfides, are common in some instances. Some microcline and cryptoperthite occur in one thin section. This same slide also has an area of a coarser grained aggregate of microcline, cryptoperthite, and quartz which may represent a cognate xenolith. A small sheared lenticle in Mountain Brook (G2645) may also be representative of these meta-acid flows. It is highly altered, with many opaques and abundant small tourmaline; the feldspar is nearly all altered to muscovite.

These meta-acid occurrences represent lava flows with an affinity towards quartz-keratophyre. The sodic nature of the feldspars may be secondary, resulting from the metamorphism, in which case they would have originally been more calcic.

A series of outcrops in the East Passumpsic River at D3918 presents varied metamorphosed igneous rock types. The exposures are poor and the relations to the country rock and of the types one to another are not known. They are gray, medium- to fine-grained rocks with a crude biotite foliation, and are composed predominantly of feldspar, biotite, and quartz, with garnet and accessory minerals. The kind of feldspar (up to 6 mm.) varies from specimen to specimen. In some it is mainly plagioclase, in laths and sometimes in larger blastophenocrysts of albite or albite-oligoclase; the laths are frequently bent. The twin lamellae are often very fine and indistinct, and the grains are usually full of indeterminate granular alteration products, including calcite and possibly zoisite. A little myrmekite is sometimes found as well as microantiperthite. There are also patches of fine granular material which is probably untwinned plagioclase (albite?) replacing feldspar. Cryptoperthite is present, usually in small quantities, but in one thin-section it is common. Quartz occurs sparsely to abundantly. Biotite is a common constituent, often in reticulate aggregates. Garnets (up to 3 mm.) are sparse to abundant, poikiloblastic and xenoblastic to idioblastic, altered slightly to chlorite and epidote. Chlorite is present in some slides; calcite is often sparse, but is common in a few thin sections. Accessories are apatite and sulfides. One specimen from a varied outcrop contains abundant ragged poikiloblasts of blue-green hornblende.

It is difficult to interpret these occurrences, particularly as the outcrops are small and their relationships obscure. They may represent metamorphosed intrusives of a quartz diorite type, or they may represent varied acid flows (now sodic) to be correlated with those described above, or there may be both flows and intrusives present.

Two other obscure exposures may also represent correlatives of the flows. At G1112 there is a light-colored, banded, felsitic rock composed of fine-grained feldspar (0.01 to 0.06 mm.), including rare twinned plagioclase, and of quartz, with biotite (partly altered to chlorite) in aggregates, chlorite streaks, and sparse apatite and opaques. There is a suggestion of a folded, sheared foliation. This is possibly a sheared metamorphosed rhyolite or tuff. The other exposure is part of an outcrop at G2717 composed of abundant biotite, with mesh-like inclusions of rutile, anhedral feldspar, abundant quartz, calcite, apatite, pyrite, and a little muscovite and sphene.

GILE MOUNTAIN FORMATION

Gile Mountain schists was the name given by Doll (1944, p. 18) to rocks well-exposed on a mountain of that name in the Strafford

quadrangle. In the Burke quadrangle the Gile Mountain formation occupies a wide belt of country approximately east of Route 114 and extending from the southern border to the northern border of the quadrangle except where interrupted by areas of granite (Plate 1). The eastern boundary is mappable for only a very short distance in the vicinity of Granby village, where the lithologic trend is northeast-southwest in the small stream valley about three-quarters of a mile northeast of the Granby postoffice. The trend turns north-south before disappearing into the probable granite area just southwest of the point where the Granby road crosses the Granby-Victory town line. The western contact of the Gile Mountain with the Waits River formation is arbitrarily mapped as extending in a general northeast-southwest direction, apart from the complication of the boundary in the Shouya Hill—Graves Hill area produced by severe folding and shearing. The contact is interrupted by the granite intrusion north of East Haven. Where it reappears to the north of the granite mass it has a northwest trend (Plate 1).

The rock types comprising the Gile Mountain formation include fine-grained phyllitic quartzite, quartzose phyllite, dark gray phyllite, quartz-sericite-biotite schist, garnet-sericite-biotite schist, staurolite-garnet-mica schist, andalusite- and sillimanite-bearing hornfels, amphibole quartzite, and amphibolite. The characteristic lithology of much of the formation is a banded alternation of light gray quartzose phyllite and dark gray pelitic or slaty phyllite (Fig. 15). This lithology is well-exposed in accessible outcrops east of the road in the Kirby Pond area (e. g., H 0311) and in the Moose River and adjacent road cuts between Moccasin Mill and the confluence of the East and West branches.

The individual bands range generally from a few millimeters to several centimeters in thickness but occasionally as much as 3 to 4 feet; very thinly banded or pin-striped outcrops occur at I4157. Banding of this type is not always developed. Quartzose phyllite may predominate in the outcrops, with the pelitic layers subordinate; this appears to be more noticeable in the outcrops close to the boundary with the Waits River formation.

Highly metamorphosed, bare exposures are to be seen on top of Burke Mountain around the foot of the fire tower and particularly in the vicinity of West Peak.

Quartzose phyllite: The quartzose rocks are, in general, light-colored, foliated, and never composed purely of quartz. They occur from the lowest metamorphic grade of the area (the biotite zone) to the highest grade (the sillimanite zone), maintaining their characteristics except



Figure 15. Typically banded quartzose and pelitic phyllite of the Gile Mountain formation northeast of Kirby Pond (H0211).

when metasomatized and granitized. Close inspection reveals a foliation parallel or closely parallel to bedding contacts, where the latter are observed by lithologic contrast. This schistosity may be due to sericite or to microlaminae of biotite which may represent original clay partings in some cases but in others are due to crystallization segregation along closely spaced shear planes. These alternative origins produce very similar results and care is needed to discriminate between them; sometimes it is impossible to decide in a restricted outcrop. Often the rock possesses microlaminations about a millimeter thick composed of alternating lighter and darker quartzose layers, obviously an original sedimentary feature.

Microscopically these rocks are composed predominantly of a quartz mosaic (individual grains vary in size from 0.01 to 0.35 mm., averaging 0.06 to 0.15 mm.), usually, but not always, with sericite in flakes (0.02 to 0.2 mm.) which are oriented to form a poor to fairly well developed schistosity and which are accompanied occasionally by chlorite flakes (0.02 to 0.06 mm.). The sericite schistosity is microfolded and cross-cut by a cleavage related to the limbs (usually the short limbs) of the



Figure 16. Photomicrograph ($\times 18$) of a quartz schist from the Gile Mountain formation showing a microfolds sericite schistosity and cross-cutting slip cleavage with sericite; biotite and chlorite porphyroblasts. South of Spur Ridge (G2703).

microfolds and with sericite aligned parallel to it. This structure is not so well developed as in the more pelitic bands but in very narrow bands it may be dominant (Fig. 16). Often the false cleavage cutting poorly developed schistosity in the sericite-poor beds is a fracture cleavage (close joints) (Fig. 38).

Red-brown, pleochroic biotite, in flakes 0.03 to 1.1 mm., is ubiquitous. Sometimes it lies parallel with the main schistosity, forming the latter, particularly in the absence of sericite. Sometimes it is aggregated into streaks within which the flakes are not well oriented. Occasionally it is also in part parallel with the cleavage. Frequently it is porphyroblastic without obvious orientation and cutting both schistosity and cleavage. Inclusions of opaque minerals in the biotite parallel the main schistosity. Chlorite porphyroblasts up to 0.63 mm. are common. Some are sub-parallel to the foliation and occasionally to the slip cleavage, but frequently they show no visible orientation. Opaque minerals are common and include fine dust and grains of sulfides up to 0.8 mm. across. The latter are sometimes elongated in the schistosity, while the dust is frequently in streaks parallel to both the main schistosity



Figure 17. Photomicrograph ($\times 18$) of a folded quartz vein. The schistosity (S_1) parallels the vein; both were folded together. Slip cleavage (S_2) is well developed and has ruptured the quartz vein in places. Biotite porphyroblasts. West of Shonya Hill (G0821).

and the slip cleavage. Accessory minerals are tourmaline (0.03 mm.), zircon (up to 0.04 mm.), rare plagioclase (albite-oligoclase, 0.08 to 0.24 mm.), rare perthite (0.18 mm.). Calcite (0.08 to 1.5 mm.) occurs sparsely near the Waits River contact.

In thin section quartz veins are common, with their thickness ranging from 0.1 mm. to a few millimeters and with grains up to 1.4 mm. In the field veins up to 5 feet thick occur. The veins often parallel the foliation, but also crosscut it; they are intricately folded and sheared by the movements that caused the slip cleavage. In Figure 17 the sericite foliation can be seen to cut the quartz vein. Some calcite also occurs with the quartz.

Higher grades of metamorphism are shown by the rare development of garnet (up to 0.9 mm.) and of staurolite poikiloblasts. In the highest grade, the sillimanite zone, the quartzose rocks still remain remarkably recognizable; the grain size has coarsened a little and the texture is hornfelsic and "sugary." The foliation is often but by no means always

destroyed. Microscopically varying amounts of metasomatism may be seen; this is described in the section on Metamorphism and in Woodland, 1963.

Before metamorphism the quartzose phyllites were coarse siltstones and fine sandstones mainly of a quartz wacke type.

Pelitic rocks: Intercalated with more quartzose rocks the pelitic rocks occur in bands from less than 1 mm. to several inches thick, more rarely several feet thick. Typically, they are gray to dark gray and fine-grained, with a well-developed schistosity, frequently cut by a later cleavage and sometimes strongly lineated. The development of porphyroblasts in the higher grades of metamorphism coarsens the texture, but the groundmass remains surprisingly fine-grained even into the highest grades (sillimanite). The pelitic rocks thus include slaty phyllite, phyllite, and schist. Much of their occurrence is of a grade higher than that normally associated with phyllite, yet their field appearance is not that of typical schist.

Sericite-muscovite is an abundant constituent together with increasing amounts of fine-grained quartz (0.03 to 0.08 mm.) as the rock grades into more quartzose types. Chlorite laths 0.06 to 0.18 mm. long occur in some pelite of the lowest grade. Red-brown biotite is ubiquitous, usually porphyroblastic, with cleavage flakes up to 1 mm. apparently not oriented, but cutting across the schistosity and cleavage where the latter are developed (Fig. 31). Sometimes, however, it is oriented in and is responsible for the schistosity, and sometimes it is partially oriented in the cleavage planes, showing the structural control of post-tectonic crystallization. Rarely some thin pelitic bands (up to 2 to 3 cm. thick) are exceedingly rich in biotite, being composed almost exclusively of this mineral.

Minute carbonaceous specks frequently occur; in some thin-sections they are abundant and streaked out in the schistosity and cleavage. Within porphyroblasts of biotite, chlorite, garnet, and staurolite the trace of the schistosity is often clearly marked by the opaque streaks. Sulfide grains are common (up to 0.2 mm.), as well as some ilmenite altering to leucoxene. The pyrite is sometimes elongated and oriented within the schistosity. Minute tourmaline occurs sparsely.

Increasing grade of metamorphism is indicated by the appearance of idioblastic garnet (usually up to 1 mm., but more rarely 4 to 8 mm.) and staurolite (up to 12 mm. long) which is often idioblastic (Fig. 18), with inclusions of quartz and opaque minerals, but is sometimes xenoblastic

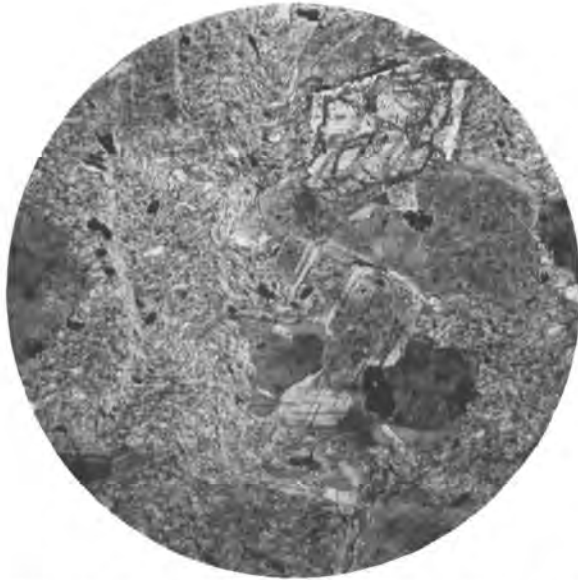


Figure 18. Photomicrograph ($\times 44$) of staurolite-biotite schist from the Gile Mountain formation showing euhedral staurolite and biotite with many opaque dusty inclusions in a groundmass of sericite, quartz, and iron ores; the sericite forms a sharply microfolded schistosity; shadowy chlorite. From the Moose valley (F1523).

and "spongy" with inclusions. The staurolite may have a "corona" of sericite or may even be almost or completely replaced by sericite-muscovite. Occasionally it is partly replaced by chlorite. Some outcrops on the lower western slopes of Kirby Mountain have bands of spotted phyllite. The spots are ovoid (up to 9 mm. long) and consist of aggregates of biotite, chlorite, muscovite, and quartz. These may represent altered cordierite, although this mineral has not been identified in any thin-section. Higher grade pelite contains andalusite (Fig. 19), often replaced completely by sericite, and sillimanite. These occur close to the granite contacts and in rocks that are clearly distinguishable in the field as having undergone a "contact" type of metamorphism. They are more fully described later in the section on Metamorphism and in Woodland, 1963.

Chlorite porphyroblasts (up to 6 mm.) occur plentifully in some areas in rocks of the garnet-staurolite grade (Fig. 65). In the more quartzose pelites accessory minerals are minute zircon and olive-green tourmaline. Quartz veins paralleling and cutting the schistosity are common.



Figure 19. Photomicrograph ($\times 18$) of Gile Mountain andalusite hornfels showing idioblastic andalusite (var. chiastolite) with inclusions of opaques, quartz, biotite, and muscovite in a fine groundmass of quartz, biotite, muscovite, and opaques. From East Mountain (F2947).

Tourmalinization of phyllite adjacent to a granitic dike has been noted at G3627. The phyllite has been converted to a tourmaline-quartz rock; the tourmaline is oriented lengthwise in bands and accompanied by considerable magnetite.

Amphibolite: Bands of amphibolite parallel with the bedding foliation outcrop at several places in the Burke quadrangle, as shown on Plate 1. Total thicknesses are not exposed, but some bands must be over 10 feet thick. They are crudely foliated and composed of abundant blue-green or green hornblende; plagioclase (andesine to labradorite), sometimes poorly twinned but more commonly untwinned and fine-grained; quartz; some biotite (usually rare); occasionally chlorite; and, in two thin sections, sparse garnet. Accessory minerals are iron ores, sphene, and apatite.

One exposure at I0941 contains a band of dark greenish-gray rock which has abundant chlorite aggregates, common biotite aggregates, much iron ore, common quartz and epidote, and a fine-grained ground-

mass (0.02 to 0.04 mm.), which probably is mainly untwinned plagioclase and iron ores. It is possibly a hornfelsed amphibolite which has suffered diaphoresis.

The amphibolite in general is interpreted as representing metamorphosed basaltic tuff and lava.

In the northern part of the area there are closely similar occurrences of a gray-green, massive, well-jointed but not foliated amphibolite. The exposures are all relatively small and the contacts are poorly exposed, so that their actual dimensions and relationship to the country rock are not known, but they appear to represent local cross-cutting intrusions. They are composed of abundant green or blue-green hornblende up to 4 mm. long, altered in part to chlorite; much minute acicular amphibole; badly altered plagioclase, poorly twinned and often full of needle inclusions of amphibole; and fine, granular, diffuse areas presumably of untwinned plagioclase and quartz. The plagioclase is exceedingly difficult to determine, but some of the albite twins in one thin section are labradorite and in another thin section appear to be about the composition of oligoclase-andesine. A small amount of myrmekite is present in one thin section. A little biotite and accessory sulfides, ilmenite, sphene (altering to leucoxene), and apatite also occur. In one specimen a little calcite is present in the more felsic portions and there are calcite amygdules. This nonfoliated amphibolite is either metamorphosed diabase or metamorphosed gabbro.

Amphibole-bearing quartzite: A characteristic rock type which occurs in thin beds about 1 to 3 inches thick, rarely a few feet thick, is a fine-grained, light-colored quartzose rock with prominent porphyroblasts of amphibole. It is interbedded with phyllite and quartzose rocks, particularly in the lower western slopes of Kirby Mountain and to the west in the zone transitional to the Waits River formation. It is also interbedded with impure limestone and amphibolite of the Waits River formation as mentioned above (p. 31).

This amphibole-bearing quartzite is composed of a quartz mosaic (grain diameters 0.01 to 0.15 mm.) with large porphyroblasts, up to several centimeters long, of green or dark green amphibole which may lie in the crude foliation, but frequently do not, and which are often plumose or feathery in habit. The amphibole is usually strongly pleochroic, with Z = dark green, i. e., hornblende; sometimes Z = bright blue-green or occasionally yellow-green, perhaps representing an actinolitic hornblende. Sometimes the amphibole is partly altered to a chloritic material.

Twinned plagioclase is a rare constituent. Untwinned plagioclase is also believed to be present but the fineness of grain and commonly diffuse nature of the grain boundaries make it extremely difficult to identify, so that its quantity and composition have not been determined. Chlorite sometimes occurs as minute flakes (0.03 to 0.21 mm.) in the foliation and frequently as porphyroblasts (var. penninite) up to 0.63 mm., in places aggregated in streaks. Biotite is present in variable quantity, sometimes sub-parallel with the foliation and poorly oriented within streaks; it is occasionally associated with hornblende and shows alteration to chlorite. Xenoblastic poikiloblasts of garnet often occur, at times notably "flattened" in the foliation. Calcite is a constituent of some bands and is accompanied by "dusty" opaque minerals marking the foliation. Sulfides are always present, and sphene and apatite may be accessory.

These rocks may be interpreted as altered tuffaceous sediments or dolomitic quartzose sediments. Their frequent association with amphibolite points to a contamination with volcanic ash in some occurrences. In one outcrop similar-appearing material occurs as an incomplete shell around a pod-shaped, calc-silicate core, indicating a reaction between aluminous material and the dolomitic content of the original impure carbonate core, which may have been a concretion.

Dark hornblende-rich bands: Dark hornblende-rich bands are sparsely distributed throughout the Gile Mountain formation. Characteristically they are about 1 inch thick, but may vary from $\frac{3}{4}$ to 2 inches thick. They contain abundant well-developed poikilitic hornblende (up to 6 mm. long) sometimes length-parallel to the boundary of the bands, but frequently plumose and bent (Fig. 20); X = yellow to yellow-green, Y = yellow-green to green, Z = green to blue-green. It is often partly altered to chlorite. Pink idioblastic garnets up to 2.5 mm. are a common constituent. These are poikilitic, with inclusions of many opaque minerals and of some quartz, but rarely of amphibole and of chlorite (altered amphibole), and are sometimes outlined with opaques. Chlorite, var. pennite (in flakes 0.8 mm.) may be abundant. In one thin-section a little clinozoisite occurs as well as a little biotite, associated with the hornblende. The groundmass is made up of very fine, diffuse grains 0.01 to 0.04 mm., and is very "dirty" with opaque minerals and inclusions. In part some twinned plagioclase (0.12 to 0.36 mm.) can be identified. The remainder may be untwinned plagioclase and quartz. Opaque minerals are common and include pyrite and much fine dust, sometimes in aggregated streaks. Sphene, in part altered to leucoxene, is common in one



Figure 20. Photomicrograph ($\times 18$) of a dark hornblende-rich band from the Gile Mountain formation showing plumose, porphyroblastic hornblende and idioblastic garnet in a very fine groundmass of plagioclase, quartz, and iron ores. From northeast of Kirby Pond (H0211; see Fig. 15).

case, and sparse apatite (in grains 0.12 to 0.15 mm.) in another.

These thin hornblende-rich bands are interpreted as originally basic tuff or tuffaceous sediment.

MEETINGHOUSE SLATE

Meetinghouse slate was the name given by Doll (1944) to a band of slate in the Strafford quadrangle as a member of the Gile Mountain formation. The name was raised to formation rank by White and Jahns (1950).

No exposures of this formation were found during the field work in the Burke quadrangle. Eric and Dennis (1958, Plate 1) on their geologic map of the Littleton quadrangle show Meetinghouse slate extending north-eastwards to the southern border of the Burke quadrangle at I0200. The terrain to the north of their map is obscured by glacial deposits. The nearest exposures, on Mitchells Knoll and on the northwestern slopes of Miles Mountain, are part of a granite-hornfels complex of high metamorphic grade (sillimanite zone). Whether the slate formation is present

could not be determined. Moreover, the low ground of Cold Brook below about 1200 feet and the area to the north are considered to be underlain by granite (p. 52).

In the vicinity of Granby exposures of the Albee formation occur within 100 feet, in some cases, of typical outcrops of the Gile Mountain formation which have a marked pin-striped lithology (I4154). There is no evidence that the Meetinghouse slate is present between the two.

INTRUSIVE ROCKS

Granitic Rocks and Hornfels-Granite Complexes

Granitic rocks are found extensively throughout the northern, north-eastern, and south-central parts of the Burke quadrangle. They all cut and intrude the metamorphic rocks, i. e., the Albee, Waits River, and Gile Mountain formations, and are post-tectonic. Although referred to in general as "granite", the rock types include adamellite, granodiorite, and quartz diorite as well as true granite. As their modes of occurrence differ, each general area will be described separately.

Kirby-Burke-Umpire-East Haven Mountains and Hobart Ridge: Kirby, Burke and Umpire mountains owe their elevation and ruggedness directly to the resistance of their rocks, a complex of granite and hornfels. Because of the nature of this complex no boundaries between granite and country rock can be drawn. The continuous forest cover and presence of glacial drift on all of the lower slopes also mitigate against any attempt to map in detail even the larger areas of mainly hornfels or mainly granite within the complex. The complex is bounded on the north, west, and south by the Gile Mountain formation; the eastern boundary is more complicated and will be considered below. The approach to the complex is marked by the increasing metamorphic grade of the Gile Mountain formation—the appearance of staurolite, then andalusite, almost invariably retrogressed to muscovite "shimmer aggregates", and sillimanite, also muscovitized, followed by the appearance of granitic stringers and dikes.

Traverses within the complex reveal a very mixed suite of rocks. Sometimes the granite exposures are relatively large, e. g., several hundred yards across, as in the vicinity of I0952, or the hornfels may extend over broad areas, the hornfels always being of high metamorphic grade and usually with some granitic stringers, although occasionally with none. More often, however, the granite and hornfels occur together in the same exposures frequently in alternating zones or as agmatite



Figure 21. Mixed granite and hornfels (agmatite) on eastern side of the parking area, summit of Burke Mountain (H1149).

(Fig. 21). The contact between the two although intricate is normally sharp (Fig. 22), often following bedding foliation, and there is no difficulty in discriminating between the granite and the metamorphic rock. On Burke Mountain there are excellent exposures on the road to the summit, on the summit, in the area around West Peak, and along the fire road, which extends along the western flank at an elevation of 1900 to 2300 feet. The structures are generally well preserved in these sillimanite-grade rocks; the bedding, schistosity, and cleavage are still evident and antedate the high-grade metamorphism and intrusion. In places there are, however, zones of dark, fine-grained, equigranular, biotite-rich rock



Figure 22. Contact between Gile Mountain hornfels and granite. Note the original sedimentary banding with the schistosity parallel to it still preserved, although the hornfels is in a high grade of metamorphism (sillimanite zone). Western slope of Kirby Mountain (G4028).

with no internal structure and a somewhat granitic appearance, but still distinct from the normal granite. The contact between it and the granite is usually very intricate and not sharp as is the case with the normal hornfels. Apparently it is a granitized sediment and represents a mixed rock; its total volume appears to be very subordinate.

Another type of mixed rock which is more prevalent is migmatite (Fig. 23), an intimate mixture of granitic stringers and patches with intervening hornfels and the previously described mixed rock. Sometimes the hornfels has been thoroughly shattered and broken up and cut by granitic stringers. A more unusual occurrence is for the hornfels, including some granitized rocks, to be brecciated and contorted wildly and to contain few granite stringers—all in all, giving the appearance of having been rendered to a semi-plastic condition and stirred up. The



Figure 23. Photograph of a hand specimen of migmatite from outcrop on road to the summit of Burke Mountain (H1156).

granitic hornfels breccia is presumably what S. R. Hall (1861) and Richardson (1906, p. 84) describe as occurring in Granby, Victory, and Concord.

Xenoliths are abundant in the granite proper and are commonly ellipsoidal, biotite-rich pods from a few millimeters up to several inches in length. (The larger masses of hornfels alternating with the granite and together comprising the complex are not considered xenoliths in the strict sense of the term). Other xenoliths are angular masses of calc-silicate hornfels up to several feet long and 1 to 6 inches wide, representing sections torn off original bands of limestone. The biotite-rich pods may be sharply delineated or fade into the granite, but the calc-



Figure 24. Contact between hornfels and granite showing the regular banding of calc-silicate hornfels. Note biotite-rich pods in granite. Eastern side of the parking area, summit of Burke Mountain (H1149).

silicate xenoliths invariably have sharp contacts and show no macroscopic evidence of assimilation (Fig. 24).

East of Umpire Mountain granite crops out over an extensive area of Hobart Ridge and the southern slopes of East Haven Mountain. On the latter the contact between the granite and the Gile Mountain formation is irregular and can be mapped only approximately. There is little of the granite-hornfels complex as developed on Burke and Umpire mountains. South of Umpire Mountain there is apparently a large continuous granite mass, but farther south the hornfels and granite recur in the Stanley Brook area and represent here the southern boundary of the granite.

The granite in the quarries on the south slope of Kirby Mountain, just south of the Burke quadrangle boundary, have been described by

Richardson (1906) and by Dale (1909, 1923). It is possible that the Grout quarry, described by Dale (1909, p. 29), is the one found during the field work approximately at H2302. Dale describes the rock in this quarry as a gray biotite granite, whereas in the quarries just south of the quadrangle and to the west it is quartz monzonite. Eric and Dennis (1958) also describe the rock from the northern boundary of the adjacent Littleton quadrangle as Kirby quartz monzonite and state that “. . . in a few places where microcline is abundant (more than 67% of the feldspar) it is a true granite” (p. 28).

A detailed investigation of the petrography of the granite of the area extending from Kirby to East Haven Mountains has not been carried out, but the rock is mainly a fine- to rarely medium-grained, sometimes porphyritic, hypidiomorphic granular, biotite or biotite-muscovite quartz monzonite. It contains oligoclase, kaolinized and sericitized, occasionally zoned, sometimes overgrown with orthoclase and with some myrmekite borders against microcline; turbid orthoclase, kaolinized and sericitized; slightly clouded microcline, occasionally cryptoperthitic (?) and enclosing a little plagioclase; quartz; reddish-brown biotite, sometimes bleached and containing rutile needles and ore specks; sometimes muscovite; and accessory zircon and apatite.

In places a medium-grained melanocratic phase is present, e.g., at H1050 at the top of the ski lift on Burke Mountain. This rock is composed of oligoclase or oligoclase-andesine, often zoned, cloudy with kaolin, and with inclusions of apatite, microcline, and amphibole; microcline; very subordinate orthoclase; reddish-brown biotite; a little green biotite; hornblende with patchy pleochroism ($X =$ yellow green, Y and $Z =$ olive green), and in some cases altering slightly to biotite; quartz; and accessory tourmaline, sphene, apatite, and zircon. A little carbonate is present as a secondary product. The relationship of these granodioritic phases to the normal leucocratic granite and quartz monzonite is not known; they may represent contaminated zones, but are considered more likely to represent separate intrusive material.

The East Haven Mountain medium-grained biotite granite contains microcline, orthoclase, antiperthite, oligoclase, biotite (chloritized in part), quartz, myrmekite, and accessory apatite.

Southeast Victory: The low-lying Victory area is covered with mounded sand and gravel and numerous boulders of granite, as well as extensive swamps. The higher ground northwest and southwest of Damon Crossing and on Mitchells Knoll has exposures of high-grade banded hornfels and granite, with granite conspicuous in some places on the lower slopes.

These occurrences are similar to the granite-hornfels complex of Burke Mountain. The outcrops northwest and southwest of Damon Crossing may be the complex zone forming part of the roof of the pluton. The hornfels is identical to that on Burke Mountain and is believed to be of the Gile Mountain formation.

The northern slopes of Miles Mountain are heavily covered with glacial deposits and definite outcrops are located only on the higher slopes. These exposures are of sillimanite-bearing hornfels and of granitic rocks similar in occurrence to the Burke Mountain complex. Eric and Dennis (1958) describe about half of the outcrops on Miles Mountain in the Littleton quadrangle as granitoid dikes and sills in sillimanite-grade rocks of the Albee formation. The granitic rocks range from light gray granite to dark gray tonalite. A thin section from one specimen of quartz diorite collected by the writer is composed of acid andesine, zoned and antiperthitic, a little cloudy and sericitized; biotite, partly chloritized; quartz; rare orthoclase; rare muscovite; sphene; zircon; and apatite.

The low hill at I3205 is covered with large blocks of banded hornfels-granite, none of which are *in situ*, but are considered locally derived. Not a single outcrop within the Burke quadrangle was found on Temple Mountain; the entire slopes are littered with granite boulders. An outcrop on the northernmost peak of the mountain about one-quarter mile east of the Burke quadrangle boundary (i. e., in the Guildhall quadrangle) is a fine-grained quartzose phyllite assigned to the Albee formation (W. I. Johansson, personal communication, 1957). It is not possible to determine metamorphic grade from such a small occurrence, but its appearance is of a considerably lower grade rock than the sillimanite hornfels of Miles Mountain.

Lees Hill has outcrops of sillimanite hornfels and biotite granitic rock which are very similar to those on Miles Mountain, and may represent a small area of injected rock of the Albee formation.

In the vicinity of Little Roundtop and southwest of Gallup Mills there is an area of hornfels-granite complex where no contact between the Gile Mountain formation and the granite could be mapped, with the exception of one small knob of granite northwest of Little Roundtop.

Many of the boulders on the lower slopes of Miles Mountain and along the crude oil pipeline are of a melanocratic rock type. Some along the pipeline east of the Concord road and in the Moose River probably came from bedrock cut into during the laying of the pipeline. One thin section from a representative specimen is composed of microcline (? some cryptoperthitic); oligoclase, some zoned, some cloudy with kaolin and

antiperthitic; scarce orthoclase; myrmekite; dark brownish-green biotite with inclusions of sphene, apatite, and zircon; hornblende (X = yellow-green, Y = olive green, Z = dark green); and accessory epidote, sphene, and apatite. The relationship of this monzonitic rock to the leucocratic granite and quartz monzonite is not known.

A similar rock type is present in the low-lying ground north and northeast of Seneca Mountain, e. g., at C4059. Where the Grand Trunk Railroad crosses the Nulhegan River in the Averill quadrangle this same rock type is seen to be cut by biotite granite. Doll (1951, p. 44) states that dark, biotite-hornblende granite is cut by later, light-colored, coarser granite in the Memphremagog quadrangle. It is possible that much of the Victory basin-like area may be underlain by this melanocratic phase of the intrusion and may account for the erosion of the area to a low basin in contrast to the surrounding mountains of hornfels, leucocratic biotite granite, and quartz monzonite. It may be a situation analogous to that of the basin-like area north of Seneca Mountain and extending into the Island Pond and Averill quadrangles.

Newark: Considerable outcrop and, indeed, cliffs of granite are present on Packer Mountain, Hawk Rock, Walker Mountain, Deer Hill, and in La Pawac and Jack brooks. The western and northern granite boundary is in contact with the Waits River formation, and the eastern and southern one with the Gile Mountain formation. In Mill Brook (B1703) there is in contact with the Waits River a microgranite, which may be a fine-grained offshoot from the border of the main mass. Unfortunately the boundaries of the main mass are not well exposed, but they are probably relatively well defined. A mixed granite-hornfels zone is absent except along the northern contact, where there is a confused alternation of granite, hornfels, schist, phyllitic limestone, and quartzose rocks. In the East Passumpsic River north of B2045 and in the two nearby streams flowing into the Passumpsic, one from the east and the other from the west, thermal metamorphism of the impure limestone is scarcely noticeable, even in outcrops close to the granite. Calc-silicate hornfels is present, however, at a number of localities along the northern contact zone. Schist and pelite show characteristics of the higher grades of metamorphism and include the sillimanite zone, particularly along the northern contact and in a narrow zone against the granite's eastern boundary, e. g., on the mountain east of Lost Nation.

The Newark granite has been described by Dale (1909, 1923) from one quarry east of Center Pond at approximately B0329. This was not definitely located during the field work, but there are numerous large ledges

in the vicinity, and one of them may have been the site of a small quarry. Dale (1923, pp. 112-113) describes the rock as a light pinkish biotite-granite with very sparse porphyritic feldspar up to 1.5 by 0.5 inch in a groundmass of coarse-grained (up to 0.8 inch) orthoclase and microcline (with some perthitic intergrowths), medium-grained quartz, kaolinised albite to oligoclase-albite, biotite, and accessory magnetite, pyrite, sphene, and allanite.

There is a pink granite at a number of places west of the East Passump-sic River; the best locality is an old quarry behind Shepard's farm (B2040), where many large quarried blocks remain. Not all of the rock is pink in color, for it grades into a normal light-colored biotite granite. The distribution of the pink-colored rock is very difficult to determine in the field because the weathered granite outcrops tend to have a pinkish color even when the fresh rock is white or creamy-colored. The pink color is possibly caused by finely divided hematite in the microcline. Some outcrops on the eastern shore of Center Pond show surfaces weathered to a deep pink or red. The pink rock at Shepard's quarry is composed predominantly of slightly cloudy microcline micropertthite phenocrysts up to 13 mm. long, with plagioclase inclusions which are kaolinized and sericitized. The microcline is also often cryptoperthitic with irregular vein-like intergrowths and with disturbed indistinct twinning. Plagioclase (up to 3.0 mm.) is common; it is much kaolinized and sericitized, with some zoning. There is little orthoclase. Myrmekite occurs in grains up to 0.65 mm. and as borders to the oligoclase, which in turn borders the microcline. The sparse biotite is green; some is partly chloritized. Quartz (up to 4.75 mm.) is common and has slightly undulose extinction. Accessory minerals are apatite, sphene, iron ores (associated particularly with biotite), epidote, and very rare allanite and zircon.

Thin sections from other localities show a very similar composition: abundant microcline micropertthite and cryptoperthite, and cloudy oligoclase, occasionally antiperthitic. The biotite is either green or a dirty brownish-green and partly chloritized. A small amount of tiny garnet is present in a specimen from Deer Hill and rare fluorite in specimens from near the top of Packer Mountain (B0524) and from a granite ledge on the east side of Route 114 near Benchmark 1162.

The Newark granite is strikingly different from the granitic rocks of Kirby, Burke, and East Haven mountains in the presence of abundant perthitic microcline and green, partially chloritized biotite.

East Mountain: Granitic rocks are exposed extensively on East Moun-

tain and on the spur which extends southward. Their contacts with the Gile Mountain formation are difficult to determine because of the extensive glacial deposits which mantle all of the lower slopes. However, on the south and west the contact is relatively sharp, with little or no "mixed zone." For example, in an excavation just off the new summit road at F2947 granite overlies andalusite schist (Fig. 19). On the lower eastern slopes of the southern spur hornfelsed Gile Mountain strata occur. The northeastern spur has many large outcrops of hornfels as well as of granite and of mixed rock, and comprises a hornfels-granite complex similar to that of Burke Mountain. However, the easternmost knob of the spur is entirely granite, as is the adjacent area of the Guildhall quadrangle. Hornfels also occurs on the top of East Mountain and on the lower slopes of the northwestern spur, with granite on the upper slopes. The northern boundary of the granite is not known, for glacial drift obscures the lower slopes into Madison Brook; the inaccessibility of the northern slope of East Mountain precluded a more detailed search there for outcrops in the time available. The granite may be continuous across Madison valley with the granite of the rugged southern spur of Seneca Mountain (at C3014), but no outcrops were found in the valley in this vicinity. It is possible that much of the northern flank of East Mountain is a Gile Mountain hornfels-granite complex. The relationship between the granite and hornfels of the northeastern spur to the granite, gneiss, and hybrid-like rocks in Madison Brook around C4112 is also not known.

The main igneous mass of East Mountain appears to be a medium- to fine-grained, often porphyritic, biotite or biotite-muscovite granitic rock similar to the normal Burke and Kirby quartz monzonite or granite. Aplite dikes cut it sporadically. A coarse leucocratic phase is present and is well exposed on the new road to the summit at F3047. It is composed of abundant, slightly cloudy microcline (3.2 mm.), some of which is cryptoperthitic and microperthitic with cloudy plagioclase inclusions (up to 0.4 mm.); abundant quartz; sericitized and kaolinized oligoclase, some antiperthitic and with myrmekite borders; a little orthoclase; sparse green biotite, in part bleached and altered to chlorite; rare muscovite with symplectic borders against microcline; and accessory sphene and garnet. Some of the granite has a slightly pinkish color.

Melanocratic diorite is exposed on the road at approximately F3356. Its constituents are oligoclase or calcic oligoclase with a little zoning, highly altered to kaolin, sericite and carbonate; hornblende (X = yellow-green, Y = dark green, Z = green), much altered to biotite and chlorite

and with sphene inclusions common; reddish-brown biotite (apparently mainly after amphibole) often bent and strained and with inclusions of sphene and, in some, of reticulated rutile; some cloudy orthoclase (it is difficult to discriminate its quantity as the feldspars are so altered, but it appears to be very subordinate in amount); a little myrmekite; sparse quartz; and accessory sphene, altering to leucoxene, iron ores, and apatite. Hornblende may exceed feldspar; the color index of the rock then rises to about 50. The diorite is associated with dark hybrid rock, hornfels and granitized hornfels, contaminated granite, and some pegmatitic masses composed of apatite, biotite, tourmaline, amphibole, feldspar, and quartz.

Seneca Mountain: Granitic rock comprises the whole of the southeastern spur of Seneca Mountain, and it also occurs in a hornfels-granite complex on the extreme northeastern spur. The granite of the southeastern spur is in contact with hornfelsed Gile Mountain formation on the northern flanks and also in the North Branch of Paul Stream just east of the Burke quadrangle boundary. To the south, as mentioned above, it is not known whether the granite is continuous under the broad Madison valley to East Mountain. The rock is fine to medium in grain and similar in hand-specimen to the granite and granodiorite of the Burke-Kirby area. A thin section from C4129 is composed of abundant microcline (up to 4 mm.) with some inclusions of cloudy plagioclase and perthitic intergrowths; kaolinized and sericitized oligoclase, some with myrmekite borders against microcline; a little orthoclase; abundant quartz with undulose extinction; reddish-brown biotite; and a little secondary muscovite. In contrast, a specimen from C4217 is composed of abundant plagioclase (oligoclase to acid andesine), finely twinned and often zoned, slightly altered to kaolin and sericite, and occasionally antiperthitic; abundant reddish-brown biotite, slightly altered to chlorite along cleavages and some bleached; quartz with fine needle inclusions; sparse symplectic muscovite; sericite aggregates in streaks up to 7.3 mm. long; and accessory apatite, iron ores, and zircon. This rock is a quartz diorite of the trondhjemite variety.

A series of exposures occurs in Madison Brook near the eastern boundary of the Burke quadrangle. The rocks are quite varied and, as they are restricted in outcrop to the brook, their relationships to the granite of Seneca Mountain on the north and to that of East Mountain on the south are not known. They include biotite "granite," gneiss, meladiorite, basic intrusives, and very dark basic material with considerable pyrite. Downstream in the Guildhall quadrangle biotite gra-

nitic rocks crop out together with masses of Gile Mountain hornfels, although the ridges both to the north and south are apparently entirely of "granite."

The melanocratic dioritic rocks are composed of euhedral to subhedral hornblende (X = colorless, Y = yellow, Z = light yellow green or X = yellow green, Y = green, Z = green), partly altered to biotite and chlorite, and with inclusions of sphene and apatite; oligoclase to acid andesine, slightly kaolinized and sericitized, rarely zoned, with occasional overgrowths of orthoclase and rare myrmekite rims; a little orthoclase and quartz; and accessory sphene, apatite, epidote, and tourmaline (as inclusions in the plagioclase). Reddish-brown biotite, with inclusions of apatite and sphene, may be abundant. Lighter colored quartz diorite contains hornblende (X = yellow green, Y = green, Z = bluish-green), altering to chlorite and containing rutile needles; andesine; quartz; and accessory sphene, apatite, and epidote. One hand-specimen is cut by veins (up to 2.5 mm. thick) of a gray-colored glass which appears under the microscope as light brown isotropic material containing rounded quartz crystals, many opaque granules, and, in some zones, incipient spots of crystallization and minute crystallites. These veins presumably represent chilled offshoots from an adjacent basic dike. The host rock shows quartz with undulose extinction, cracks, and dusty lines of inclusions; much cloudy and sericitized plagioclase (oligoclase-andesine) with some carbonate and with the twins ruptured, bent, and poorly defined; some cloudy orthoclase; brownish-green bent biotite; colorless amphibole; a little chlorite; common sphene altering to leucocene; apatite; and pyrite. Near the veins the alteration increases and aggregates of biotite and sphene occur at a high angle to the contact. One dark green basic zone contains abundant masses of pyrite (up to 9.5 mm.); abundant reddish-brown biotite, bleached and chloritized; much chlorite and light green amphibole; rare plagioclase; apatite; quartz; and common colorless mica, particularly adjacent to and penetrating the pyrite masses.

The gneiss is composed of a regular alternation of dark schistose bands 8 to 17 mm. thick and white granulose bands 3 to 7 mm. thick dipping to the southwest. The dark bands are composed of euhedral and subhedral green hornblende, biotite (mainly green but a little dark brown), plagioclase, quartz, and accessory sphene, apatite, epidote, and sulfides. The light-colored bands have minor biotite and hornblende, abundant acid andesine (2.6 mm.), which has perthitic intergrowths of microcline (0.15 mm.) and cryptoperthitic patchy intergrowths, occasional oscil-

latory zoning and patchy kaolinization, rare microcline (up to 0.36 mm.), and rare myrmekite. This occurrence of quartz diorite-gneiss is unique in the Burke quadrangle; its origin is obscure, but it may represent metasomatized banded Gile Mountain strata.

McSherry Mountain-Newark Street: The northwestern section of the Burke quadrangle, around Bald Hill Pond and Newark Pond and extending southward to East Sutton Ridge and eastward to Walker Pond and Newark Street, has many exposures of granitic rock, particularly in the high ground such as McSherry Mountain. Unfortunately large areas are also covered with glacial drift, and the contacts of granite with the Waits River formation are, in general, impossible to map. However, a characteristic feature is that many of the granite exposures overlie Waits River strata and the latter dip at low angles beneath the granite. Examples of this relationship occur on the lower eastern slope of McSherry Mountain (A0751), north of Newark Pond, (A0842), and northeast of Bald Hill Pond (A1558). In addition, granitic exposures alternate with metamorphosed calcareous rocks, for example, along the road north of Bean School and on the slopes northwest of Newark Street. Elsewhere sills and dikes frequently cut the Waits River strata, as in the East Sutton Ridge area. The granitic rocks are considered to have essentially a sheeted form following the structure of the Waits River formation in the northwestern area; some of the sheets are thin offshoots while others must be several hundred feet thick, such as that on McSherry Mountain. The intrusions extend from the granitic areas of Bald Mountain to the north (Island Pond quadrangle), Haystack Mountain (Lyndonville quadrangle), and the area south of Lake Willoughby (Lyndonville quadrangle) mapped by Dennis (1956, Plate 1) as containing an abundance of granitic dikes. There is, in general, little evidence of advanced thermal metamorphism in the exposed rocks of the Waits River formation except locally where there are bands of calc-silicate hornfels.

The granitic rocks are very variable in texture; the larger masses are normal, fine- to medium-grained biotite granite or granodiorite. They pass frequently into a very coarse pegmatitic phase, which may often comprise the whole of many exposures. Fine-grained, very leucocratic white phases also occur as well as dikes and sills up to one or two feet thick.

A leucocratic specimen from A2316 is composed of slightly clouded microcline (up to 2.1 mm.), micropertthitic and cryptopertthitic, with muscovite inclusions; kaolinized and sericitized oligoclase, some anti-

perthitic; quartz (up to 1.2 mm.), a little orthoclase; sparse muscovite; rare myrmekite; and iron ores. A sheeted rock overlying the Waits River formation at A1622 has abundant quartz; oligoclase (up to 1.8 mm. occasionally), normally zoned, with some antiperthitic; orthoclase; a little microcline; sparse brown biotite; and sphene. Quartz also occurs in rounded grains dotted sparsely throughout the feldspars; it appears to be of secondary metasomatic origin.

At A2455 an exposure of "granite" has a platy flow structure trending 154° and dipping 15° northeast.

Contact Relations and Mode of Emplacement

All of the granitic rocks appear to be related and perhaps co-magmatic. The melanocratic amphibole-bearing granodiorite to monzonite varieties were intruded earlier than the leucocratic quartz monzonite and granite. They are all post-tectonic intrusions and probably are all members of the New Hampshire magma series (Billings, 1934), one of the seven series recognized in New Hampshire (Billings, 1956). The New Hampshire magma series is believed to be Middle Devonian in age (Lyons, et al., 1957), although according to the new time scale (Kulp, 1961) the radioactive age falls within the Mississippian or Pennsylvanian time span.

The contact relations of the granitic rocks are not known in detail because of the poor exposures. Over large areas there is not a sharp contact, but rather a broad zone of mixed granite and hornfels. In general the granite cross-cuts the regional structure, but in local exposures the contacts are seen to both parallel and cut across the bedding schistosity. In the northern part of the Burke quadrangle the metamorphosed sediments dip beneath the granite; this suggests that at least in this area the intrusion is in the form of sheets and tongues more or less parallel to the bedding.

There is little, if any, evidence of large-scale deflections by the granite of the strike of the bedding of the metamorphic rocks. The east-west strike in Ferdinand, in the northeastern part of the quadrangle, may be related to the intrusion of the granite to the north; the bedding of the Gile Mountain strata dips steeply northwards into the granite. East-west strikes are also prevalent in the Gile Mountain near the granite contacts on East Haven and Hobart Ridge, and there is a slight suggestion of a bending of the strike in the area northwest of Burke Mountain. The hornfels-granite complex zones provide evidence of much disruption of the intruded rocks and rotation of the resultant blocks. On the

northern flank of Burke Mountain east-west strikes are common in the hornfels of the complex. These relations imply much fracturing and much penetration of the metamorphosed sediments by the granite and an intrusion mechanism by stoping. The granite is not, however, surrounded everywhere by complex zones.

One point of interest is that the hornfels on Burke Mountain includes a noticeable number of thin calc-silicate bands. Limestone bands are rarely observed in the Gile Mountain formation; perhaps the metamorphism of the calcareous bands to calc-silicate rocks makes them more apparent than in their lower metamorphic grade outside of the complex. However, it could be that these Burke Mountain hornfelsed rocks belong to the transition zone between the Gile Mountain and the Waits River formations and have been brought to their present elevation by the granite or by a previous fold structure which has been disrupted by the granite.

The rocks of the complex also show some evidence of assimilation and granitization; these are believed to be on a small scale. The mechanism of intrusion appears to have been by a combination of forcible intrusion, utilizing essentially the structural weaknesses of the intruded rocks, mainly the bedding schistosity, by intimate penetration of the rocks and stoping, and to a smaller extent by assimilation and granitization.

Granitic Dikes

Granitic dikes cut the country rock usually near the main "granite" masses or near the granite-hornfels complexes. They are prevalent in the area north and south of East Sutton Ridge. The majority of the dikes are thin, usually less than one foot thick, but very occasionally up to 25 feet thick. Some are simple pegmatites, while others are fine-grained, sometimes porphyritic, representatives of the larger granitic intrusions ranging in type from microgranite to microdiorite.

One occurrence at G3627 contains rare brown tourmaline, and the phyllite in contact with it has very abundant brown tourmaline. Sparse green hornblende together with brown biotite is present in a dike of microgranodiorite at A0313, and abundant hornblende altering to biotite and chlorite occurs in a porphyrite dike six feet thick at I2147. On East Mountain (F3356) there is a more unusual dike. This is a microsyenite: greenish, mottled with pink; one foot thick, with slickensided walls and joints. It is composed of abundant blue-green hornblende; common green biotite, slightly altered to chlorite; microcline (crypto-

perthitic); a little orthoclase; some kaolinized and sericitized feldspar (probably plagioclase); epidote; scarce interstitial quartz; and accessory sphene, apatite, and iron ores.

Mafic Dikes

Thirty-seven post-metamorphic mafic dikes have been encountered, mainly in the northern half of the quadrangle; they are described in detail in Woodland, 1962. The exposures are generally poor and it was often impossible to determine either the thickness or the attitude of the intrusions. Richardson (1906, p. 111) mentions briefly the occurrence of diabase dikes in Newark.

The dikes have a gray to greenish-gray or dark gray color; they range in thickness from less than one inch to several feet and appear to dip very steeply. They are aphanitic to fine- to medium-grained rocks, occasionally with a narrow light-colored chilled border, often with amygdules of calcite and chlorite (rarely analcite), and generally with sparse to abundant phenocrysts. Microscopically they comprise a varied suite which possesses one common characteristic—they have all suffered considerable deuteric alteration. Carbonates and chlorite are ubiquitous. The rock texture is intersertal with the latter minerals, together with other groundmass minerals, occupying the interstices between the feldspars.

The fine grain and altered nature of the rocks usually make it very difficult or impossible to identify all of the constituents, particularly in the mesostasis. Plagioclase often is one of the major constituents, but in a rare instance it is present in only small quantities. It may be in small laths 0.04 to 0.06 mm. long or in larger laths up to 1.25 mm. long, felted, rarely pilotaxitic, but commonly in ocelli which are crudely variolitic. The plagioclase is much altered to sericite, kaolin, and calcite; its determination normally is not possible but in some cases the feldspar appears to be andesine and perhaps occasionally labradorite. Phenocrysts of plagioclase sometimes occur up to 4.5 mm. long, also much altered, and containing inclusions of apatite. Alkali feldspar is present in some dikes in discrete crystals as well as mantling the plagioclase, and may be present in the groundmass in other dikes. Phenocrysts of olivine (up to 1.5 mm.) occur in about half of the thin sections examined; sometimes they are very abundant and almost invariably completely altered to serpentine-carbonate or talc-carbonate-serpentine aggregates. Pyroxene (augite) is of frequent occurrence both as phenocrysts (up to 1 mm. or even 3.5 mm.) and in the groundmass; it is also often altered to carbonate-

serpentine aggregates. Brown amphibole occurs in about half of the sections, sometimes as phenocrysts (up to 3.7 mm. long) but more often in the groundmass; rarely it may be after pyroxene. It is a brown hornblende or barkevikite. Biotite is present in over half the sections but usually sparsely. In some it is common (olivine is then absent and pyroxene may or may not be present) and is pleochroic to a dark red-brown. It is chloritized in two occurrences. Analcite occurs in a few slides usually interstitially but rarely as larger patches (up to 1.5 mm.) with many minute inclusions; it is also in amygdules with calcite. Quartz, in small quantities, is very rare. The interstitial material is abundant, sometimes preponderantly so, and often difficult to resolve, but includes chlorite (often in spherulites), carbonate, antigorite, and chlorophaeite. Apatite is sometimes present and is occasionally very abundant as needles throughout the thin section. Sphene occurs sparsely. Opaques are abundant, and in some cases are so prevalent as to make the thin section very dense; included are magnetite, ilmenite (altering to leucoxene), leucoxene, and sulfides. The latter are sometimes very noticeable even in hand specimens. In two thin sections the opaques occur in a mesh-work of fine rods; in one the mesh is very irregular throughout the abundant groundmass; in the other the mesh forms a very regular network, with angles of approximately 87° and 93° in small patches, superimposed on the groundmass throughout the slide. Granular opaque minerals also occur abundantly.

The metamorphic effects on the intruded country rock are not noticeable in the field. In thin section a limestone and a schist contact have been examined. In one case (A3742) the immediate limestone contact (about 0.35 mm. thick) is composed of fine-grained quartz (0.04 to 0.36 mm.), phlogopite, calcite, and many opaque grains. The individual grains are larger away from the contact and contain masses and streams of opaques; microcline, apatite, chlorite, and tremolite also occur. In the other case (B1948) the schist against the dike is composed of quartz grains with fine needles and felted masses of sillimanite; abundant biotite, much of which is altered considerably (in part to sillimanite); abundant sericitic aggregates; and opaques. The fibrolitic sillimanite, however, was probably present before the intrusion of the dike.

The mafic dikes are much altered but appear to range in type from amygdaloidal diabase to olivine diabase, analcite diabase, and lamprophyric types. The latter predominate and include intrusions of camptonitic affinities, although none possess the typical panidiomorphic granular texture. They may be related to the White Mountain plutonic-volcanic

series of New Hampshire or to the intrusive necks of the Monteregian Hills of Quebec (see Woodland, 1962, p. 1107-1109).

Metamorphosed Mafic Dikes: A few metamorphosed dikes cut the Gile Mountain formation in a number of localities, e. g., at F2802 and H0513. These dikes contain plagioclase (often untwinned), hornblende, biotite, a little quartz, and iron ores; chlorite is common in some and rare epidote occurs in one specimen. They were intruded before the metamorphism of the area.

STRUCTURE

Minor Structural Features

The structural features described herein include all planar and linear structures and mesoscopic (Weiss and McIntyre, 1957, p. 577) and microscopic folds. Mesoscopic or minor fold structures are those that occur within the limits of an average hand specimen or small outcrop in the field, and range in size from about one-quarter inch up to several yards across; generally they are between one-half inch and three inches across. In contrast, major fold structures cannot be viewed directly in the field in the Burke area, and their presence and form can only be determined by indirect means; they may range in size up to regional structures miles across.

The minor structural features of the Burke quadrangle have been made the subject of intensive study so that they could be grouped by type and relative size, their geometry described, regional preferred orientations, if any, defined, and any areal change of preferred orientations discovered. It was hoped this would lead to and support the description of major structures and the interpretation of their origins.

To this end, many planar, linear, and fold features were measured in the field and a total of 1029 geographically oriented specimens were collected. The latter have all been studied in the laboratory with a binocular microscope to determine the types of planar and linear features present, and whenever necessary the specimens have been re-oriented in a simple apparatus which enabled the field orientation to be reproduced and the orientation of additional structural surfaces and linear elements to be measured, as described under Methods of Study. In this way many more data have been obtained to supplement the field readings. One important result of the laboratory examination has been a more detailed appreciation of the geometry, origin, and mutual relationships of the structural features. The accumulation of such orientation

data in many cases would have been difficult or impossible to obtain in the field, particularly in the relatively poor light in the forests and in the stream courses completely overshadowed by trees.

Most of the measurements have been plotted in the conventional way on Plate 2 (planar features map), Plate 3 (linear features map), and Plate 4 (way-up of bedding and shear sense map), but in some areas there were too many readings to plot on a map of the scale of one inch equals one mile. All of the measurements have been plotted on the lower hemisphere of the Schmidt stereonet (Lambert equal-area projection) or on the lower hemisphere of the Wulff stereonet. (See Plate 5 and Fig. 56).

PLANAR FEATURES

These include all parallel planes, regardless of origin, which can be discerned in the rock because of a compositional or mechanical inhomogeneity, for example, repeated planes along which a rock tends to part. Planar features of the rock fabric have been designated "s"-surfaces by Sander (1930). In the Burke area "s"-surfaces include bedding, schistosity, false cleavage, and joints.

PRIMARY PLANAR FEATURE

Bedding (S_0): Compositional banding which represents original depositional layers is easily recognizable throughout the Burke quadrangle. It is particularly well developed in the Gile Mountain formation, where the alternating layers vary from a fraction of an inch to several feet in thickness. Characteristically it provides a striped or banded appearance of alternating light gray quartzose phyllite and dark gray slate, phyllite, or schist. A well-developed schistosity is nearly always present, and in most exposures this is sensibly parallel to the bedding layers and is termed a bedding schistosity (S_1). In the quartzose rocks the schistosity is marked not only by the alignment of sericite but often by a fine striping a millimeter or two apart caused by small biotite flakes also aligned parallel to the main compositional banding and appearing to represent original silt or clay partings. Although the biotite laminae may well be derived from such partings, they are essentially of mechanical origin. Comparatively rarely the bedding is seen to be tightly folded, with the schistosity and fine striping paralleling the axial plane and the limbs, but cutting across the bedding layers at the nose of the folds, proving the mechanical nature of the schistosity and of the fine biotite laminae. These folds are designated as earlier or B_1 folds; by far the more abundant folds in the area deform the schis-

tosity, and these are called later or B_2 folds (see the section on Minor Folds). It can not be assumed, therefore, that schistosity is always parallel to original bedding in outcrops that show no distinctive bedding layers, although more often than not it is likely to be parallel or sub-parallel. The great majority of bedding orientations recorded in the field were actually measured on foliation planes sensibly parallel to the primary compositional banding and are referred to in this paper as bedding schistosity or S_1 . These, together with the relatively small number of bedding (S_0) readings where the latter are not parallel to S_1 , are plotted on the planar features map, Plate 2, which thus shows the present attitude of the bedding in the Burke quadrangle.

The bedding contacts maintain remarkable constancy of attitude over large areas and give the impression of a simple structure with a steep east-southeast dip, becoming gentle and even horizontal in the western part of the quadrangle and dipping steeply to the east-northeast in the northern part. The monotonous nature of the strata within the Waits River and Gile Mountain formations obscures any repetition of beds caused by isoclinal folding or strike faulting. However, in the southwestern portion of the quadrangle there is evidence of considerable repetition. This is indicated by the successive occurrence across the strike of outcrops of Waits River lithology alternating with outcrops of the Gile Mountain type. Unfortunately the evidence is provided, not by repetition of a sharp, clearly defined boundary between the two formations, but by repetition in the transitional zone between the two. Structural complexity is very evident in this area; minor folds abound, and often outcrops show the rocks to have been strongly deformed and contorted, sometimes to a point that defies geometric reconstruction. Limestone bands are infolded sharply with phyllite and die out repeatedly along the strike as do bands of phyllite interbedded in more quartzose rocks; the intricate nature of the limestone folds, particularly, and their disharmonic relation to their envelopes imply a large measure of rock flowage. Shears are also evident at some localities, particularly at G1010 and G1015. It is likely that other tectonic breaks also occur, perhaps representing sheared-out limbs of folds. This complexity of structure in the southwest, the gradational lithology, and the scattered nature of the outcrops (due to glacial and post-glacial cover) render the mapping highly conjectural in places; the pattern represented on the geologic map (Plate 1) is thus a generalized interpretation of the field data.

Elsewhere in the Burke quadrangle the evidence for repetition is not available. Minor folds occur throughout the area, but in general outcrops

do not show the structural complexity of those in the southwest, with the exception of the outcrops in the granite-hornfels complexes. North of East Burke the evidence for any complexity of the boundary between the Waits River and Gile Mountain formations is obscured by thick, widespread drift deposits.

Evidence bearing on the original order of superposition of the strata is exceedingly scanty and is rendered difficult to interpret because of the strong deformation and recrystallization of the rocks. Repeated isoclinal folding is believed to affect the beds, with both reversed and normal limbs present. However, it is reasonable to suggest that, if the formations as a whole are in their normal order, that is, if they dip and face to the east, indications of a normal order should be more prevalent than those of a reversed order in a random occurrence of appropriate outcrops. A complicating factor is the presence of later folds which may result in local reversals of the original attitudes of limbs of the earlier folds. However, the later folds appear to be mainly of a minor character and are not accompanied by large-scale structures so that inversion effects are restricted to a very local scale, which can be recognized; the way-up data in such cases are rejected.

The features that were examined in the field include cross-lamination, graded bedding, attitude of drag folds, and the relationship of bedding (S_0) to schistosity (S_1), and are shown diagrammatically on Plate 4. Structures similar to cross-lamination are rare throughout the area. Secondary cleavages often simulate this primary structure and usually it is not possible to be sure whether it is cross-lamination or cleavage which is present. Of the ten cases of accepted cross-lamination in the area, seven indicate that the bedding is right-way-up and faces eastwards, and three indicate that the bedding is upside-down and faces westwards. Graded bedding is present in the banded pelitic and psammitic beds of the Gile Mountain formation. Rarely a psammitic bed grades upwards through a semi-pelitic zone into a succeeding pelitic band within the space of two or three inches. Normally the alternating psammitic and pelitic bands have sharply differentiated boundaries on both sides so that it is impossible to determine top and bottom, although the banding probably represents a type of rhythmically graded bedding. Only in five instances was the way-up determined; in three the beds dip and face eastwards and in the other two the beds dip eastwards and face westwards.

The attitude of minor drag folds has long been used in determining the correct order of deposition. White and Jahns (1950, p. 197) suggest

the use of the terms sinistral and dextral to describe the form of folds with one limb shorter than the other as exposed on a horizontal (or vertical) surface. L. M. Hall (1959, pp. 61-63) shows that, if for a horizontal exposure the plunges of the folds are also known, the shear sense that produced the folds can be deduced and thus the position of the minor folds in the major structure and the way-up of the beds can be determined. However, the minor folds so used must be congruous folds related to the major structure. In the Burke quadrangle only the earlier folds (B_1) can be used safely, as the later folds (B_2) are superimposed on a folded structure (see section on Minor Folds below). A further complication is that early folds may be rotated bodily around the B_2 axis or the plunge may be inverted by rotation around an axis at a high angle to the B_1 axis. However, a total of thirteen early minor folds not deformed or rotated by later folds (so far as could be ascertained) have been analyzed; nine indicate that the strata face and dip eastwards and four that they dip eastwards and are upside-down (see Plate 4).

The relationship of the attitude of bedding to that of axial plane cleavage, slaty cleavage, and fracture cleavage in folds is well-known (e. g., Billings, 1942; Wilson, 1947). In the Burke area the mutual relations of bedding (S_0) and schistosity (S_1) have been noted wherever an angular relation could be discerned and where later folds do not cause complications—a total of thirty-one observations. In seventeen cases the beds are right-way-up (they dip eastwards in twelve instances and westwards in five), and in fourteen they are upside-down (they dip eastwards in thirteen instances and westwards in one). (See Plate 4).

Thus, out of a total of fifty-nine determinations the strata are right-way-up in thirty-six localities (thirty-one with an easterly dip) and upside-down in twenty-three (twenty-two with an easterly dip).

SECONDARY PLANAR FEATURES

Schistosity (S_1): The dominant planar feature of the metamorphic rocks of the area is a well-developed schistosity or foliation, that is, a structure caused by the parallel orientation of platy or prismatic minerals throughout the rock and producing an infinite number of parallel planes along which the rock tends to part. This agrees with the definition of what Leith (1905, p. 105) describes as flow cleavage, which is developed parallel to the axial plane of folds and which is parallel to the plane containing the greatest and intermediate axes (AB) in the strain ellipsoid. Fairbairn (1935, p. 592) refers to this

structure as "axial plane" foliation and Mead (1940, p. 1010) calls it flow cleavage developed perpendicular to the least axis of strain. As noted above, in the great majority of outcrops the schistosity is sensibly parallel to the primary bedding, and is referred to as a bedding foliation or schistosity. The schistosity is seen to be parallel to the axial plane of the rare earlier minor folds. It is accordingly not to be interpreted in the sense of bedding foliation as defined by Mead (1940, p. 1009).

Schistosity is plotted on Plate 2; poles to schistosity are plotted on the lower hemisphere of a Schmidt stereonet in Figure 56A and on Plate 5 (Figs. 1 to 11).

The schistosity is defined mainly by the subparallel orientation of mica (sericite, muscovite, and biotite) and is best developed in the pelitic and semi-pelitic rocks. Quartz and calcite often appear "flattened" and show a dimensional orientation with their longer direction parallel to the schistosity. The quartzose rocks usually have sericite in subparallel orientation throughout and frequently it is also somewhat concentrated in narrow zones; the quartzose phyllite thus usually has a well-developed foliation. It is not possible to determine the origin of the schistosity (S_1) surface, but the fact that it plainly cuts the bedding layers at the noses of minor isoclinal folds and is axial plane to these folds indicates that it was formed during the deformation producing the tight isoclinal folds (thus explaining why it is nearly always subparallel to bedding) and is accordingly an axial plane schistosity. The rocks have been elongated parallel to the schistosity and normal to the B_1 or earlier fold axis direction, as is indicated by the usually considerable "thickening" of the isoclinal minor fold hinges and by rare boudinage of original bedding parallel to B_1 caused by movement on the schistosity. It is probable that movement parallel to the schistosity occurred throughout the area and resulted in the obliteration of original bedding and destruction of fold hinges; it may explain the scantiness of the observations of the latter. The development of the schistosity essentially parallel to the bedding may have been partly caused by translatory movements being initiated and continued on the original bedding surfaces. The latter must have been prominent, particularly in the banded shales and siltstones of the Gile Mountain formation.

False Cleavage (S_2): Later cleavages transect the schistosity very commonly in the Burke quadrangle. Collectively these are called false cleavage and include slip cleavage and fracture cleavage. They are plotted on Plate 2; poles to cleavage are plotted on the lower hemisphere of a Schmidt stereonet in Figure 56B and on Plate 5 (Figs. 12 to 19).



Figure 25. Drawing of a hand specimen of Gile Mountain banded phyllite, showing bedding (S_0), striking 40° , dipping 63° to the southeast; schistosity (S_1), striking 39° , dipping 72° to the southeast; and cleavage (S_2), striking 41° , dipping 50° to the southeast. From the southwestern slope of Sugar Hill (G3306).

Cleavage occurs throughout the area, but is not everywhere equally developed as structural surfaces, and even in an area such as the southwestern section (subareas I and II—see Plate 4), where it is, in general, prominent, it is not present in some exposures while in others it is the dominant structure. Cleavage is best developed in the pelitic and semi-pelitic rocks where it sometimes nearly obliterates the schistosity (Fig. 25). An outcrop or even a hand specimen frequently shows two or more later cleavages cutting the schistosity. Often these may be caused by differences in lithology so that the attitude of the cleavage surfaces in quartzose bands varies from that in pelitic layers. This refraction of the cleavage is clearly seen in Figure 26. In other cases the relationship

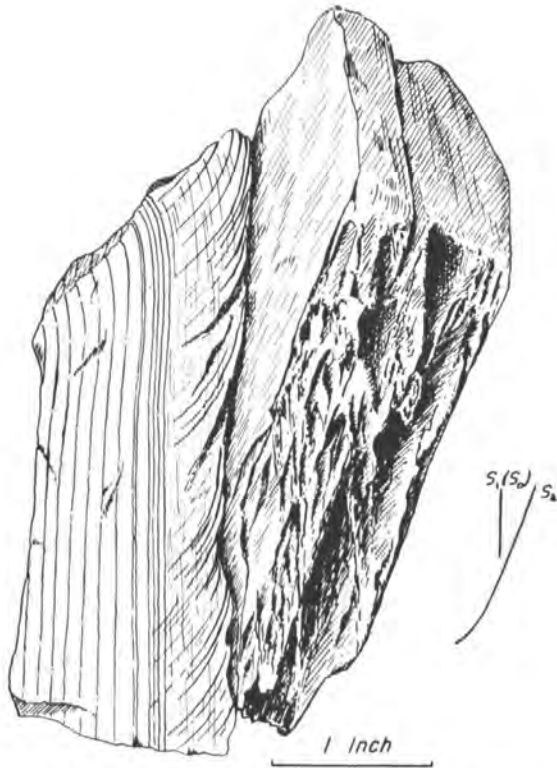


Figure 26. Drawing of quartzose and pelitic phyllite showing "refraction" of cleavage (S_2) in the quartzose band. From western foot of Sugar Hill (G3107).

between two or more cleavages is not clear, although one is usually dominant and much better developed than the others. A few cleavages (S_3) apparently later than S_2 , are plotted on Plate 2. Where later B_2 folds deform the schistosity (S_1) surfaces the dominant S_2 surfaces are parallel to the axial planes (Figs. 27 and 37). Displacement on and parallel to the cleavages is often evident; it is more apparent in the quartzose bands and also in the finely banded quartzose and pelitic or semi-pelitic beds (Fig. 30, a and b). The amount of movement rarely exceeds two or three millimeters. Quartz veins, often parallel to the schistosity, are disrupted by the cleavage and, when displaced (Fig. 28), give rise to cleavage mullions (Wilson, 1953).

The characteristics of the cleavage and its development vary with the lithology. The pelitic and quartzose beds offer contrasting situations and



Figure 27. Typically banded Gile Mountain strata deformed by later (B_2) folding. Schistosity (S_1) is parallel to the bedding (S_0); slip cleavage (S_2), parallel to penknife, is axial plane to the B_2 fold. Note drag folds in the quartzose band and folded quartz veins (bottom center and on right). On east side of road up the Moose valley (F1232; see Fig. 42).

are described separately below; intermediate types are formed in semi-pelitic rocks.

Cleavage in pelitic beds: The stages in the development of cleavage can be established by study of hand specimens from different outcrops. Examination of rough fractured surfaces approximately normal to the schistosity and lineation, as well as examination of polished surfaces cut with this orientation, frequently shows the deformation of the schistosity and the formation of the later cleavage. Examination of thin sections, also prepared normal to the schistosity and the lineation, provides detailed evidence of mineral orientations and of the nature of the structural surfaces.

The earliest stages of deformation show the schistosity (S_1) with regular zones of "crinkling," which produce a rippling of the surfaces. The "crinkle" zones, S_2 , are spaced about one to two millimeters apart



Figure 28. Quartz vein disrupted into cleavage mullions by movement on slip cleavage (S_2), which parallels penknife. In Gile Mountain formation, Sheldon Brook (G2302).

(Fig. 29a), and the mica is oriented oblique to the trend of the zones. At this stage S_2 is not a well-developed structural surface and does not give rise to a cleavage. Further deformation results in the development of microfolds in S_1 (Fig. 29, b and c), the short limbs of which line up and form the S_2 surfaces. As the micas of the short limbs lie within and parallel or sub-parallel to the trends of the limbs, the S_2 is a well-developed cleavage surface. The sense of movement on both S_1 and S_2 surfaces is evident at this stage and is in the same direction on both (Fig. 29b), although no slip is discernible on S_2 . Apparently further deformation and slip on S_1 result in rotation of S_2 as the short limbs of the microfolds become more overfolded. A stage is then reached when the rotation of S_2 surfaces brings them into a favorable position for actual slip to take place, and displacement on S_2 surfaces occurs. Shears, however, may arise before the overturned limbs have produced "through-going" surfaces; the shears (S_2) transect the limbs at a small angle and movement along them drags the mica into parallelism with S_2 (Fig. 30, a, b, and c, and Fig. 31). Meanwhile, the S_2 surfaces have become closer together

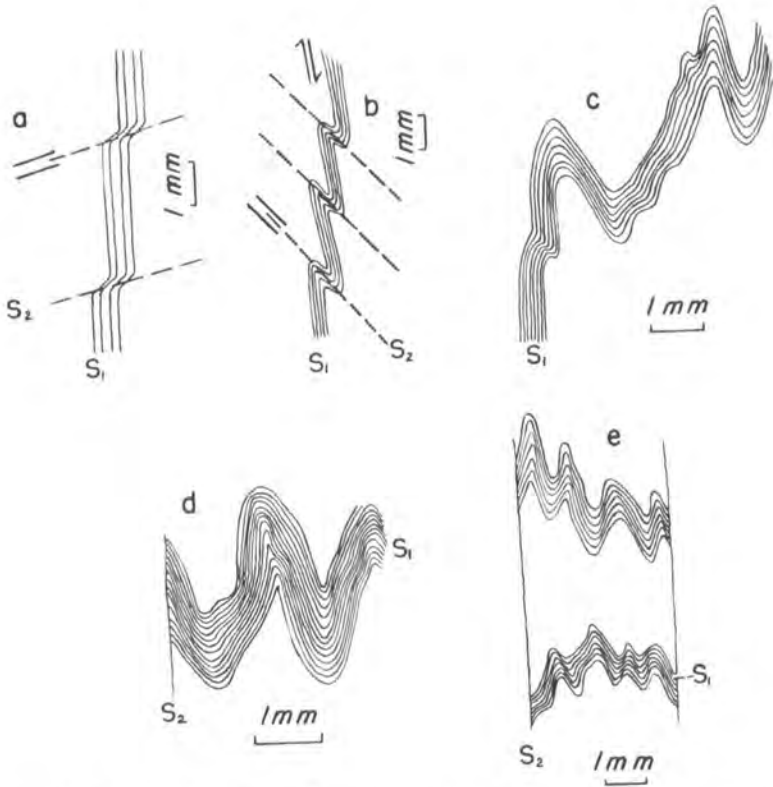


Figure 29. Series of drawings from one specimen showing the deformation of schistosity (S_1) and the development of cleavage (S_2) in the Gile Mountain formation, From the western slope of Shonya Hill (G1020).

- a. Crinkling of S_1 and incipient development of S_2 .
- b. Microfolding of S_1 and development of S_2 on aligned short limbs.
- c. Isoclinally folded S_1 .
- d. Tight isoclinal folds, sheared in places by S_2 .
- e. Minutely flexured S_1 between parallel S_2 planes.

due to the microfolding of the S_1 surfaces (Fig. 29, d and e). Initiation of displacement on the S_2 surfaces limits further slip on S_1 . As a result of the stresses between adjacent S_2 surfaces, the S_1 surfaces are thrown into further microfolds, the limbs of which are sub-parallel to S_2 (Fig. 29e and Fig. 32). Continued movement results in more and more mica becoming oriented parallel to the S_2 surfaces so that S_1 is almost obliterated. Its presence at this stage is shown by relict noses of microfolds



Figure 30a. Photograph of polished surface of finely banded Gile Mountain phyllite; dark spots are biotite porphyroblasts. From the summit of Shonya Hill (G1120).
 a. Microfolds in S_1 , which is parallel to S_0 , and development of S_2 cleavage. Note thinning of long limb (on the right).

lying between the S_2 surfaces, which are now extremely well developed throughout and form a true schistosity (Fig. 33). The sense of movement on S_2 is now not possible to determine. This apparent development of S_2 surfaces is very similar to that described by Sorby (1880), White (1949), Brace (1953), De Sitter (1956), Weiss and McIntyre (1957), and L. M. Hall (1959).

Heim (1878; quoted in Harker, 1886) describes cleavage related to minute faults resulting from microfolding as "Ausweichungscleavage." Sorby (1858, 1880) calls it "close-joints cleavage," which, he says, could



Figure 30b. Enlargement of a portion of (a) to show development of S_2 along the short limbs of the microfolds.

develop into a true slaty cleavage. Bonney (1886) calls a similar structure "strain-slip cleavage," and Harker (1886) uses the term "false cleavage." Dale (1892, 1894) uses the term "slip cleavage," which he equates with "ausweichungslivage" and strain-slip cleavage. Leith (1905, p. 120), under the general context of "fracture cleavage," describes "false cleavage" as ". . . usually the result of closely spaced parallel overthrust folds grading into minute faults. . ." and equates it with "ausweichungslivage," slip cleavage, and strain-slip cleavage. Mead (1940, p. 1010) describes shear cleavage as consisting of ". . . roughly spaced surfaces of shear displacements on which platy minerals have developed and into which they may have been dragged." He contrasts it with fracture cleavage

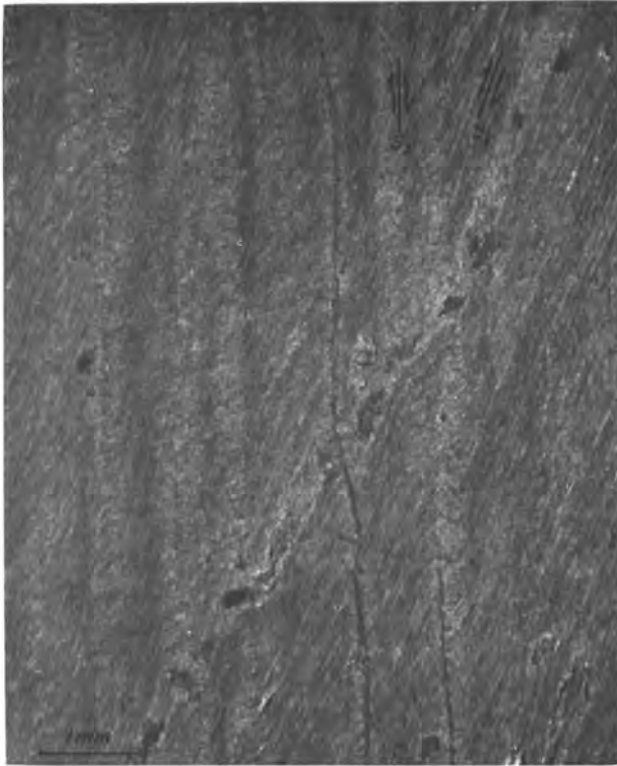


Figure 30c. Enlargement of a portion to the left of (a) to show development of S_2 in more homogeneous pelitic rock.

by suggesting that the former is particularly “. . . a phenomenon of rock flowage rather than of fracture” (p. 1011). He further states, however, that “. . . there is little importance in distinguishing between these two varieties of cleavage (that is, shear cleavage and fracture cleavage) because they have the same relationship to causal stresses and accomplish the same deformational results” (p. 1017). White (1949, p. 593) describes the deformation of rock during the formation of slip cleavage as comprising two components, a flattening of thin plates bounded by slip planes and a distortion by simple shear along the slip planes. De Sitter’s ideas (1956, p. 97) are essentially similar.

White (1949, pp. 592–593) prefers not to use Leith’s term “fracture cleavage” because:

“. . . it is questionable that slip cleavage is entirely independent of

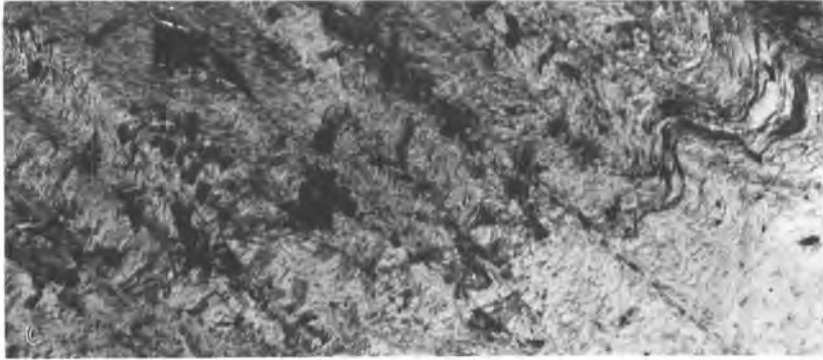


Figure 31. Photomicrograph ($\times 13$) of Gile Mountain banded phyllite showing microfolded schistosity (S_1) well developed in the pelitic layer and less well developed in the quartzose layer (bottom right-hand corner). S_2 cleavage is developed and transects the limbs of the microfolds in S_1 . Biotite porphyroblasts (some altered to chlorite) are common. In Barnes Brook (G2817).

mineral orientation. . . . Fracturing on the theoretical shear planes of the stressed rock as a whole does not seem to be the primary cause of the localization and orientation of the slip cleavage . . . (it) places undue emphasis on one aspect of this planar element, whereas actually there is clear evidence of appreciable flowage in the process leading to its formation."

He thus calls the structure "slip cleavage" after Dale, and this is the term widely used for a similar structure in all of the recent reports on Vermont geology.

A cross-cutting structure which may give rise to S_2 surfaces has also been observed to occur in some cases when the schistosity is deformed into minute chevron folds, the alignment of whose sharp angular hinges produces a parallel structure (S_2) crossing the general trend of the schistosity. There are no platy minerals oriented parallel to these structures, which are called "Knickungsebene" by Brögger and referred to by Harker (1886).

S_2 surfaces are also developed in the Burke area in a kinematically different way to that outlined above. In Figure 34 the S_2 surfaces are prominent but relatively widely spaced. The S_1 schistosity is folded (note the sharp fold on the left) and is displaced along the S_2 shear surfaces. The sense of slip is opposite to that on S_1 —which is the reverse of the position previously described (see Fig. 30b). Figure 35 also shows the same relationship, although this occurrence is complicated by the pres-



Figure 32. Photomicrograph ($\times 18$) of banded phyllite, with minor calcite, from the transition zone between the Waits River and Gile Mountain formations, showing folded sericite schistosity (S_1) cut by S_2 cleavage, biotite porphyroblasts (some lying parallel to S_2), and opaques in pelitic band (lower area). Note that S_2 surfaces are also developed in the quartzose band (upper area). From south of East Haven village (E1651).

ence of a prominent S_2 surface, the sense of slip on which is opposite to that on the other S_2 surfaces and appears to be related to a minor fold visible in the quartzose band. This fold structure and the related slip surface are thus analogous to those described above and shown in Figure 30b. The opposite slip sense on S_1 and S_2 is well shown, too, in Figure 36; the development of a new flexure with a new incipient S_2 surface, the sense of slip on which is opposite to that on the other S_2 surfaces, can also be seen.

Hoeppener (1956) outlines the development of S_2 structures whose orientation and movement sense are similar to those of Figure 34. He states that S_2 surfaces first form at angles of 115° to 145° to S_1 surfaces, and the slip sense on S_2 is opposite to that on S_1 , in each case towards the obtuse angle between them. Rotation of S_2 is accompanied by a flattening and flexure of the S_1 surfaces into microfolds between the S_2 planes, and new S_2 surfaces form, the movement direction on which is opposite to



Figure 33. Photograph of polished surface, etched with hydrofluoric acid, showing well-developed S_2 cleavage and relict schistosity (S_1), the latter appearing as isolated fold noses. From the southern slope of Bemis Hill (G1440).

that on the old S_2 surfaces; then it becomes no longer possible to determine the movement sense on the S_2 surfaces. The S_2 surfaces are finally rotated into planes perpendicular to the least strain direction. The development of S_2 surfaces in the pelitic portion of Figure 35 and also in Figure 36 appears to be in agreement with Hoepfener's description.

It is not clear why these two kinematically different S_2 surfaces should arise together, but it is probable that the lithology of the rocks and the initial attitude of the S_1 surfaces in the stress field causing the new structures are both responsible. The lithology of the specimen in Figure



Figure 34. Drawing of a hand specimen of Wait's River banded phyllite showing folded bedding schistosity (S_1) cut by S_2 cleavage. Note small isoclinal fold in S_1 on the left. From Mountain Brook (G2446).

30a is a finely banded quartzose and pelitic phyllite, and it is possible that the micro-overfolds and attendant shears on the overturned limbs are controlled by flexural slip folding of the thin competent bands.

Brace (1953) describes different patterns in the development of slip cleavage:

"... one if west limbs (of microfolds) are thinned and brought into sub-parallel arrangement, another if the east limbs are so affected. ... Both limbs may become aligned, forming "Totfalten" (Ampferer, 1938)."

His first two cases are identical with the two varieties described above,

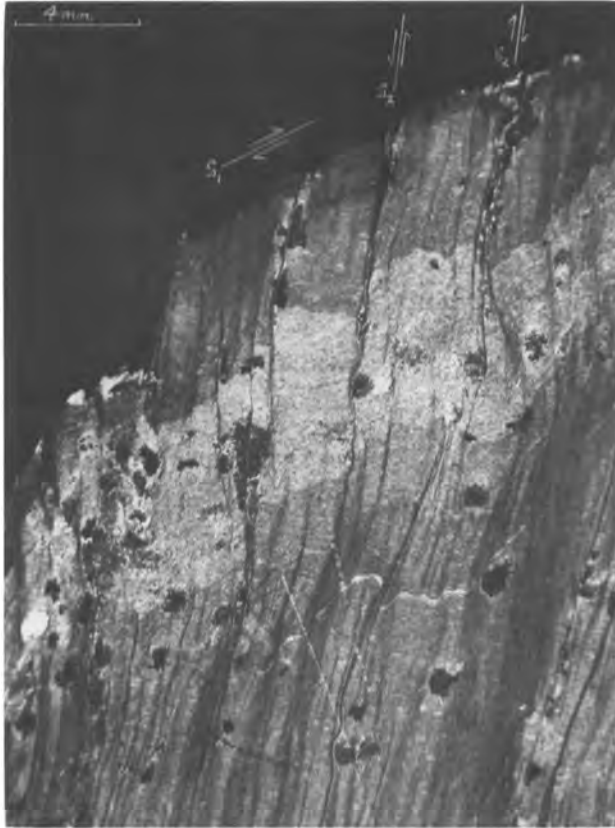


Figure 35. Photograph of polished surface, etched with hydrofluoric acid, of Gile Mountain phyllite showing microfolded bedding schistosity (S_1) cut by a well-developed S_2 cleavage. Note that the cleavage (S_2) associated with the fold in the quartzose band in the right-center has a reverse sense of slip to that on the other S_2 surfaces. From the hill north of Sheldon Brook (G2204).

that is, S_2 surfaces related to micro-overfolds, with movement sense on S_1 and S_2 the same (Fig. 30) and S_2 surfaces accompanied by micro-folding, with the movement sense on S_1 and S_2 opposed and directed into the obtuse angle between them (Fig. 34). Alignment of both limbs of microfolds will arise in both cases when flattening between the S_2 surfaces causes flexuring of the S_1 surfaces and continued slip on S_2 shears out the limbs, leaving the detached noses of S_1 as relicts between the S_2 planes (Fig. 33) analogous to the "Totfalten" of Ampferer. As the S_2 surfaces

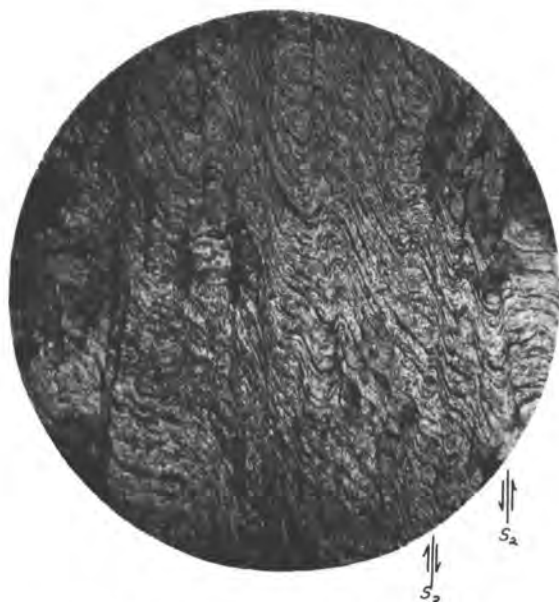


Figure 36. Photomicrograph ($\times 17$) of microfolded schistosity (S_1) in phyllite, with the development of slip cleavage (S_2). Note the sharp new flexures, the limbs of which parallel incipient S_2 cleavage and the slip movement on one of which is in an opposite direction to that on the other S_2 surfaces. From the northern slope of Spur Ridge (G3008).

are frequently subparallel to the axial planes of minor folds in the schistosity, their ultimate attitude would appear to be in the AB plane of the strain ellipsoid and normal to the direction of greatest stress. When this stage is attained, no further slip on S_2 is possible. Rotation of S_2 planes into an axial plane attitude is clearly demonstrated in Figure 37. Recrystallization of mica parallel to the S_2 surfaces has often accentuated them. However, although post-deformational growth of biotite porphyroblasts is to be seen many times as preferential along S_2 planes, their (001) cleavages are frequently oriented at a high angle to S_2 , thereby producing "edgewise" biotite as seen on the cleavage surfaces.

Cleavage in quartzose rocks: S_2 cleavage surfaces are not often present in the quartzose rocks, but some outcrops do have a cleavage cross-cutting the schistosity (S_1). These cross-cutting surfaces are generally relatively widely spaced (one-eighth to one-half inch apart) and usually show as shear surfaces with visible displacement of the bedding schistosity (Fig.



Figure 37. Photomicrograph ($\times 18$) of a minor fold in Gile Mountain banded phyllite showing microfolded bedding schistosity (S_1) and the development of S_2 cleavage, which has been rotated into an axial plane position in the core of the fold. From northeast of Granby (I4157).

38). Sometimes the planes contain biotite porphyroblast concentrations, which have formed after the deformation and which make the cross-cutting S_2 structures very prominent (Fig. 39) and similar to Dale's (1896) "cleavage banding".

The S_2 cleavage of quartzose rocks has the characteristics of Leith's (1905) "fracture cleavage". It is, however, also axial plane to later folds (Figs. 38, 40, and 41). In Figure 41 it is clearly seen as a shear cleavage with displacement of the bedding schistosity, the fold being essentially a shear fold.

Cleavage in calcareous rocks: S_2 cleavage structures are rarely observed in the impure limestone. Intergranular and intragranular adjustments were apparently continuous during the movements so that the material deformed plastically (Fig. 48) without the development of shear surfaces. The paucity of mica and the relatively poor development of S_1 surfaces did not provide the conditions for inter-layer slip and microfolding as described under the pelitic rocks. Calcareous phyllites deformed much like the latter, however.

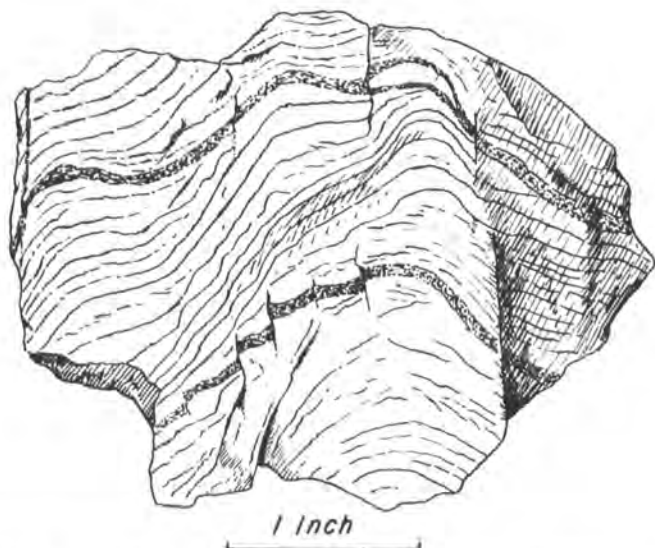


Figure 38. Drawing of a hand specimen of Gile Mountain quartzose phyllite, showing an open fold in the bedding schistosity and quartz veins (dotted) cut by fracture cleavage, movement on which has caused displacement of the quartz veins. From west of Sugar Hill (G2708).

Analysis of attitudes of S_2 cleavage: Sometimes the divergent attitudes of the S_2 cleavage cannot be explained by differences of lithology, as described in the preceding section, but may be due to: (1) the development of microfolds in S_1 , the limbs of which are not parallel but are more or less equally developed and transect the general trend of S_1 , or (2) rotation of the surfaces during the course of deformation or external rotation of fold limbs which would bodily rotate the slip surfaces relative to the stress field. It is difficult to explain why there are two kinematic variations of the S_2 surfaces, particularly as the actual distribution of each is not known. The writer is inclined to believe that the micro-overfold type is the more common in the Burke quadrangle because it occurs more often in the hand specimens.

It is probable that this micro-overfold type of S_2 cleavage is the result of initial deformation of S_1 by microfolding with concomitant slip on S_1 and flexural slip folding of the more competent rock layers. Evidently the initial attitude of the well-developed S_1 surfaces in relation to the applied stress favored slip on them (simple shear in a homogeneous field). As the microfolds develop, the micas on the over-



Figure 39. Drawing of a hand specimen of Gile Mountain quartzose phyllite showing slightly flexed bedding schistosity (S_1) cut by fracture cleavage (S_2). From the western slope of Shonya Hill (G0822).

folded limbs line up and produce the S_2 surfaces, which, in the earliest stages, are oriented so that slip (also oblique or simple shear relative to the greatest stress direction) takes place on them in the same direction as on the S_1 surfaces. In Figure 30a it can be seen that the long limbs of the very asymmetric folds have been thinned and the short overturned limbs have been preserved; slip on the S_1 surfaces of the long limbs and on the S_2 surfaces was in the same direction, but it was apparently opposite in direction on the short limbs. Early slip on S_2 surfaces, however, would lead to shearing-out of the overturned limbs. As the S_2 surfaces are rotated towards the AB plane, slip on S_2 diminishes and further deformation takes place by new microfolding of S_1 and the rotation of more mica into parallelism with the S_2 planes. The sense of slip is opposite on opposite limbs of the microfolds.

The other main type of S_2 surfaces is developed as shears, the attitude and slip movement of which are complementary to those of the S_1 surfaces; the initial mutual relation of the S_1 and S_2 surfaces conforms to that predicted by the strain theory of rock deformation (Swanson, 1927). The S_2 surfaces are gradually rotated into a sub-parallelism with the axial planes of minor folds and thus into the AB plane, and further deformation takes place by flexuring of the S_1 surfaces.

The microfolding of the schistosity in the pelitic bands and the development of the new slip surfaces (S_2) result in an extension of the



Figure 40. Photograph of polished surface of Gile Mountain quartzose phyllite showing schistosity (S_1) cut by shear cleavage (S_2). Biotite is aligned in S_2 , while the biotite in the S_1 laminae is oriented parallel to the S_2 direction. From west of Brookside School (G2923).

rock mass initially parallel to a direction bisecting the acute angle between the S_1 and S_2 surfaces, but, as the S_2 is rotated into parallelism with the AB plane, the extension direction lies parallel to it.

Slip sense on S_1 : The relationship between the direction of slip sense on S_1 and on S_2 surfaces can sometimes be determined by examination of hand specimens and particularly by the form of small, even microscopic, folds in the S_1 surfaces. The geometry of sixty-nine of these folds in the Burke quadrangle is shown on Plate 4; an analysis of them indicates that in thirty-seven the slip movement was upwards on the



Figure 41. Photograph of polished surface of Wait's River quartzose phyllite showing shear fold in bedding schistosity (S_1) caused by slippage on S_2 cleavage. Biotite is aligned in both S_1 and S_2 . From west of Sugar Hill (G3009).

west side of the predominantly easterly-dipping S_1 surfaces and in the remaining thirty-two it was upwards on the east side.

The slip sense on S_1 in any specimen may also be determined from the attitude of S_2 , if it is assumed that the angular relation between S_1 and S_2 surfaces conforms in general to the strain theory of deformation (Swanson, 1927), i. e., when the slip sense on S_1 is towards the obtuse angle formed by the intersection of the S_1 and S_2 surfaces. This is the relationship exhibited by all of the Burke hand specimens where slip sense on S_1 could be determined; it is also the relationship widely accepted by geologists and used for a long time to determine the po-

sition of an outcrop in a fold and to determine the way-up of bedding (see, for example, Leith, 1913; Billings, 1942; Wilson, 1947; Shrock, 1948). The direction of apparent slip sense may also be ascertained if it is taken as being approximately normal to the line of intersection of S_2 with S_1 . Using these criteria, a number of oriented specimens have been analyzed and the slip sense determined by construction on the lower hemisphere of a Wulff stereonet; the plots of the slip sense are shown in the eight stereograms of Plate 4, together with the geographic area covered by each plot. The stereograms indicate that in the majority of cases the slip sense is downwards on the east side of the S_1 surfaces, but upward slip on the east is common in areas A, C, and E. This is in agreement with the data derived from the folds in S_1 , although the diagrams show slip sense downwards on the east more preponderantly than do the folds. Downward slip on the east has mainly a south-eastward plunge in areas B, C, and D, a southwestward plunge in area F, a southward plunge in areas E and G, and a southeastward to southwestward plunge in area A. Upward slip directions (on the east side of S_1) from the east and southeast have gentler plunges in area A; likewise, those from the southeast in area C and those from the southeast, east, and northeast in area E have gentler plunges. Slip movement is thus mainly downwards on the east side of the schistosity (S_1) in a general southerly direction; upward slip on the east side may represent movement on the gentler, less common, opposite limbs of isoclinal folds in the schistosity (S_1).

The spread of the directions shown on the plots may indicate differences in the main stress vectors; for example, the diagrams of areas D and F suggest this. However, some spread of points would be expected even in a homogeneous stress field, because the initial attitude of the S_1 surfaces would have varied somewhat around the preferred orientation. Slip directions plotted in the northeastern quadrant of the stereogram occur in areas B and E and to a lesser extent in A; these are to be correlated with southeastward- and southward-plunging lineations (B_2) and with S_2 surfaces differently oriented to those associated with northeastward-plunging lineations (B_2) and southward-plunging slip vectors. The former S_2 surfaces either strike east of and dip more gently than the S_1 surfaces or they strike west of and dip more steeply than S_1 , while in the latter case the attitude of the S_2 surfaces relative to S_1 is the reverse.

Relationship of S_1 and S_2 : The relationship between the schistosity and cleavage surfaces is shown in Figure 57A for the Burke quadrangle as

a whole and in Figures 57B to K for the subareas. Over the entire quadrangle the S_2 maximum lies west of and dips more gently to the southeast than the S_1 maximum. This is true also for subareas I, II, III, V, and VI. Insufficient data are available for preparing a stereogram of S_2 in subarea VII; however, the measurements available show that S_2 trends in part west of and dips more steeply than S_1 , and in part trends east of and has about the same dip as the S_1 surface. In subarea VIII the S_2 cleavage tends to parallel S_1 or else to strike east and dip more gently. S_2 surfaces strike west of and dip more steeply to the east than S_1 in subareas IX and XI, while in subarea X they have a more easterly strike and a gentler dip. The relationships in subareas VII, VIII (in part), IX, X, and XI are correlated with B_2 lineation plunging in an easterly or southerly direction, while in the other subareas the lineation maximum plunges to the northeast.

Joints: A detailed study of all of the joints in the Burke quadrangle has not been carried out, but the attitudes of the joints in three areas have been analyzed. Contoured stereograms of poles to all of the joints measured in the metamorphic rocks of subarea II, in the hornfels of subarea IV, in the granite of subarea IV, and in the Newark granite are shown on Plate 5, Figures 31 to 34, respectively. The metamorphic rocks of subarea II have a joint plane maximum trending 127° and nearly vertical and a sub-maximum trending 58° and dipping 70° to the northwest. The hornfels of subarea IV has two maxima of nearly vertical joint planes trending 78° and 143° ; this relationship strongly suggests a conjugate shear joint system. The granite of the same area has a different and rather scattered joint pattern. The majority of joint planes either strike 53° and are vertical or strike 32° and dip 16° to the northwest; the latter represents sheeting. The origin of the joints in the granite thus appears to be quite different from that of the joints in the hornfels. The steep cliffs developed on the west-southwestern flank of Burke Mountain and associated with an extensive rock fall appear to be controlled by jointing in the hornfels trending 143° . The vertical cliffs on the northeastern slope of the northern spur of Umpire Mountain also may be due to northwest-southeast-trending joints in the hornfels and phyllite. The data are insufficient to explain as due to jointing the precipitous eastern flank of Umpire Mountain which trends north-south.

The Newark granite has a maximum joint plane attitude trending 48° and dipping 75° to the northwest, a second maximum trending 166° with a near-vertical dip, a third maximum trending 156° and dipping 78° northeast, a sub-maximum trending 106° and dipping 80° to the

southwest, and a second sub-maximum of horizontal joints; the latter represents sheeting. The steep cliffs on Packer Mountain and Hawk Rock are probably due, in part, to the joints striking 166° , while the south-facing cliffs on Walker Mountain may have been developed as a result of the joints striking 106° .

Acid dike trends: The trends of 87 acid dikes, which lie mainly in the northern half of the Burke quadrangle, have been measured. Of these 56% lie between $N.18^\circ W.$ and $N.10^\circ E.$

Poor exposures almost invariably prevented measurement of their dips.

Faults: Only a few faults have been observed in the Burke quadrangle (e. g., at A0503), and all have small displacements of the bedding (not exceeding three inches). It is possible that the phyllonite that occurs in the southwest (G1010 and G1015), south of Shonya Hill, represents a moderate dislocation, but the extent of displacement, if any, cannot be estimated. It is also possible that large-scale strike-faulting may occur, yet be undetected, because of the absence of marker horizons in the Waits River and Gile Mountain formations.

MINOR FOLDS

Folds are common throughout most of the quadrangle, but are apparently particularly numerous in the southwestern part where the contact between the Waits River and Gile Mountain formations is complex, as a result of folding on a relatively large scale. The Albee formation has so few exposures in the quadrangle that little opportunity is offered for observing fold structures; no fold axes have been measured in this formation. All of the folds which have been observed in the field are small-scale, normally less than three to four inches across, but some a few feet across do occur (Figs. 42 and 43). Two generations of folds can be distinguished, and they are described below separately as earlier and later folds. The time interval between the deformations causing the two series of folds is not known; the series probably belongs to the same general orogeny and represents successive phases.

Earlier folds (B_1): Very few folds of the earlier deformation have been identified. The considerable minor folding, particularly in the southwest, nearly all belongs to the later deformation. No early folds have been discovered in the Albee formation.

Early folds are recognized by the bending around of the beds of differing composition that are cut by the extremely well-developed schistosity (S_1), which is axial plane to the folds. Slip cleavage (S_2),



Figure 42. A large, relatively open, later (B_2) fold in banded Gile Mountain strata. The schistosity (S_1) parallels the bedding (S_0); the slip cleavage (S_2) is axial plane to the fold and is parallel to the penknife. The fold axis plunges 45° northeastwards and parallel to the hammer handle. On the east side of the road up the Moose valley (F1232; see Fig. 27).

if present, cuts the schistosity often at a small angle, and is not axial plane to the earlier fold. The folds are isoclinal and often very tight, with the result that the axial plane schistosity is practically parallel to the limbs. The schistosity is so well developed that it is the dominant structural surface of the rock; the bedding is not present as structural surfaces. The bedding around the nose of the fold may be flayed out along the schistosity because of slippage along the latter (Fig. 44). Some of the folds are so tight as to be represented by very narrow interdigitations of bands, as seen in plan view (Fig. 45). The bedding bands may be rendered inconspicuous by the pervading schistosity and show up merely as shadowy lines (Fig. 46). In some cases translation along the schistosity has left noses of folds as relicts suspended between the new mechanically-produced surfaces (Fig. 47). It is probable that translation on the schistosity has obliterated much of the evidence of early folds, particularly in the layers which are more homogeneous in composition, where they would be obscure at best.



Figure 43. A steeply-plunging, later (B_2) fold in banded Gile Mountain strata. The schistosity (S_1) parallels the bedding (S_0). The fold axis plunges 70° northeastwards and parallel to the hammer handle. On the east side of the road up the Moose valley (F1234).

Undoubtedly the later deformation has also served to obscure the earlier folds as the later folds have refolded the earlier ones. Examples of refolded folds have been rarely identified in the Burke quadrangle, presumably because the relationships in the outcrops would be complex. Highly deformed rocks crop out fairly often in the southwestern area, and these usually defy geometric analysis. Figure 48 is an example of a refolded fold in banded limestone of the Waits River formation. The earlier fold is a small, tight, isoclinal drag fold which is now folded over the crest of the later fold. There is no visible cleavage parallel to the axial plane of the later fold. Another example of a refolded fold is shown in Figure 49. Here the earlier fold is tightly isoclinal with a quartzose phyllite core and a well-developed axial plane schistosity (S_1), which is parallel with the limbs but cuts the nose and flays out the bedding contact. The later fold is an open fold which has a cleavage, sub-parallel



Figure 44. Plan view of plunging earlier (B_1) folds in the Gile Mountain formation. The schistosity (S_1) is parallel to the axial planes of the folds (parallel with hammer handle). Note how the pelite of the cores has flayed out along the schistosity. Northeast of Kirby Pond (H0512).

to its axial plane, developed in the quartzose core and also developed in the pelitic envelope, where it has a different attitude.

Of the thirteen earlier folds which could be analyzed, six have a sinistral pattern in plan view (see White and Jahns, 1950; L. M. Hall, 1959), five plunging northwards and one plunging southwards; seven have a dextral pattern, three plunging northwards and four plunging southwards. Most of these earlier folds occur in the southwestern ninth (G) of the quadrangle.

Later folds (B_2): Abundant in the southwestern part of the Burke quadrangle (Figs. 6, 7, and 8), later folds are also prevalent to the north, around Burke Hollow (Fig. 9), Hartwellville, and in the Moose valley (Figs. 42 and 43), and along the eastern border at I4258. At the latter locality the schistosity of the Albee formation is deformed into small folds; this deformation is equated with the later deformational episode which affects the Waits River and Gile Mountain formations.

Later folds are identified by the criterion that the schistosity (S_1)



Figure 45. Interdigitations of pelite in quartzose phyllite which represent very tight, isoclinal, earlier (B_1) folds. The schistosity (S_1), parallel to the penknife, is axial plane to these folds. The earlier (B_1) folds are flexed into an open later (B_2) fold, the axial plane of which, parallel to the pencil, marks the slip cleavage (S_2). West of Sugar Hill (G2907).

is folded. It is actually the S_1 surfaces which define the form of the fold. As the S_1 surfaces are parallel to the sedimentary banding, the latter is usually folded congruently with the former. (However, Figure 48 and Figure 49 are contrary examples).

The later folds vary from relatively open ones (Fig. 38 and Fig. 42), usually in the more quartzose bands, to relatively tight, isoclinal ones (Fig. 48 and Fig. 50). Figure 50 shows clearly the mechanism of folding in a series of bands of differing competence. The quartzose bands are folded by concentric shear along the well-developed S_1 surfaces, while the pelite band is microfolded and has developed a slip cleavage which is parallel to the axial plane of the minor fold. This has resulted in the thickening of the hinge by slippage on the S_2 surfaces. The pelite folds



Figure 46. Photograph of a shadowy relict of an earlier (S_1) fold cut by an extremely well-developed schistosity (S_1), which is axial plane to the folds. Note the rippling of the schistosity at the bottom. From Barnes Brook, northeast of East Lyndon (G2816).

are, therefore, similar folds, which have been formed partly by shear folding. The same relationship is shown in Figure 37, where the S_2 cleavage is more pronouncedly developed in the core of the fold and is parallel to the axial plane, while out on the limbs it is poorly developed and has not been rotated into parallelism with the axial plane.

Figure 51 shows a characteristic drag-like fold form in a more quartzose band and intercalated quartz vein, while the pelitic rock above and below has been microfolding and has developed dominant S_2 surfaces, although the relicts of the S_1 surfaces are clearly evident.

Some of the later folds are entirely shear folds. These are observed in



Figure 47. Photograph of a relict earlier (B_1) fold now seen only as an isolated fold core. A perfect schistosity (S_1) has developed axial plane to the fold. From north of Flower Brook (E1627).

quartzose rocks which have a well-developed S_2 cleavage. In Figure 41 the flexuring of the S_1 surfaces between the S_2 surfaces is slight, but it is more pronounced in Figure 40 where a considerable amount of biotite has recrystallized along the S_2 surfaces. Quartz veins are deformed congruently with the folding upon shearing of the S_1 surfaces (Figs. 17, 51, 52, and 53), proving that the veins belong to an earlier phase of regional metamorphism. Earlier folds have been refolded by the later folds; only rarely are examples actually seen, although they must be common in the southwestern part of the quadrangle. Figure 45 shows a broad flexuring of the very tight earlier isoclinal folds marked by the



Figure 48. Photograph of folds in Waits River impure limestone and phyllite, with quartz veins. Note the discontinuous beds over the arch of the fold; these represent tight isoclinal earlier (B_1) folds which have been refolded by the more obvious later (B_2) fold. Note, also, the absence of axial plane cleavage in the limestone. From Simpson Brook (G2728).

interdigitating pelitic bands. Figure 48 illustrates a refolded fold in limestone where no axial plane S_2 cleavage has developed. Figure 49 shows an isoclinal earlier fold deformed by an open later fold; the fold axes of the two folds are coincident, but the S_2 cleavage, sub-parallel to the axial plane of the later fold, cuts across the S_1 axial plane schistosity of the earlier fold.

LINEAR FEATURES

Fold Axes: The trend and plunge of fold axes have been measured wherever possible in the field; additional axes have been measured on oriented specimens in the laboratory. Still more fold axes have been constructed on the Wulff stereonet by the method described by Wegmann

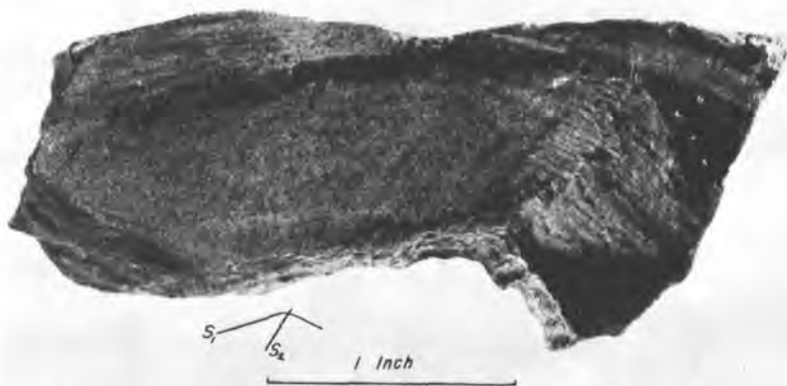


Figure 49. Photograph of an isoclinal earlier (B_1) fold with a quartzose core and a pelitic envelope. The schistosity (S_1) is axial plane to the fold. This earlier fold has been deformed by a later open fold with the development of axial plane cleavage (S_2). S_2 has a different orientation in the quartzose and in the pelitic layers. Note how the biotite porphyroblasts outline the lithologic boundary. From west of Sugar Hill (G2908).

(1929): A series of readings of strike and dip are made in the field or on a hand specimen at intervals across a folded surface. Poles to these planes are plotted on a Wulff stereonet. If the fold is cylindroidal, these points will lie along a zone corresponding to a great circle. The pole to this great circle defines the fold axis.

B₁ folds: The trend and plunge of the 13 earlier (B_1) fold axes measured are shown on Plates 3 and 4 and have been plotted on the lower hemisphere of a Schmidt stereonet together with B_1 lineations, which represent the intersections of bedding (S_0) and schistosity (S_1), and are mainly in the southwestern area. The resultant dot plot has been contoured, as shown in Figure 56C. The maximum trends 24° and plunges 42° to the northeast. This indicates a northeasterly plunge for the earlier fold structures, and probably represents the attitude for the quadrangle as a whole.

B₂ folds: The trend and plunge of the 168 B_2 folds measured are shown on Plate 3. The pattern, axial plunge, and attitude of the axial planes of 63 of these later folds are shown on Plate 4. A total of 38 folds have a dextral pattern in plan view, 28 plunging northwards and 10 plunging southwards; 25 folds have a sinistral pattern in plan view, 21 plunging northwards and four plunging southwards.



Figure 50. Photograph of a sharp later (B_2) fold in Gile Mountain banded phyllite. Note the slip cleavage (S_2) developed in the pelite band, which is "thickened" in the core of the fold parallel to S_2 . The quartzose phyllite retains the schistosity (S_1), which is folded by concentric shear. From west of Sugar Hill (G3107).

The trend and plunge of the B_2 folds have also been plotted on the lower hemisphere of a Schmidt stereonet for the quadrangle as a whole and for subareas I and II. These dot plots have been contoured, and the resultant stereograms are shown in Figure 56D and on Plate 5, Figures 20 and 21, respectively. The composite diagram shows that the maximum of the B_2 fold axes trends 21° and plunges 23° to the northeast. The axes lie in a girdle striking 14° and dipping 70° to the southeast; this orientation is similar to the attitude of the maximum of the S_1 surfaces (see Fig. 57A). The B_2 fold axes in subarea I have a very similar attitude to that shown on the composite diagram, but in subarea II the maximum plunges 39° due south, although there is a sub-maximum trending 27°

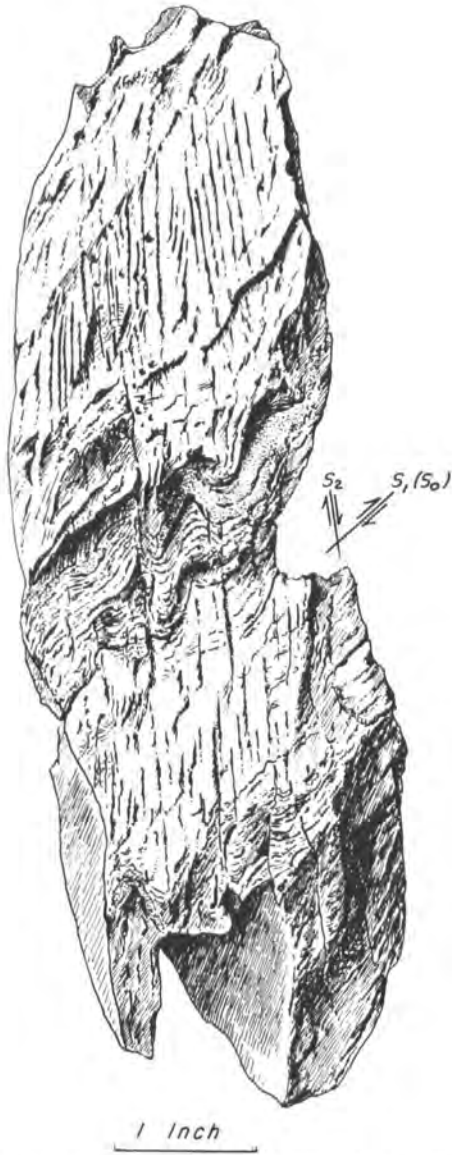


Figure 51. Drawing of a hand specimen of phyllite showing drag folding of intercalated quartzose band and quartz vein (later folds) and development of slip cleavage (S_2) in the pelitic bands. The cleavage is sub-parallel to the axial planes of the folds and almost obliterates the schistosity (S_1), which is parallel to the bedding (S_0). From the southern slope of Bemis Hill (G1440).



Figure 52. Photograph of later (B_2) folds in Gile Mountain banded phyllite. Axial plane cleavage (S_2) is weakly developed and masked by biotite porphyroblasts. Note the folded quartz vein, which is showing incipient boudinage. From Sheldon Brook (G0703).

and plunging 20° to the northeast. The axes lie in a girdle striking 12° and dipping 76° to the southeast, which is very similar to that of sub-area I.

Lineation: Lineation is frequently observed on schistosity (S_1) and cleavage (S_2) surfaces. It is usually prominent on these surfaces in pelitic rocks, but fainter in quartzose rocks, and absent in the impure limestone. The great majority of lineations noted in the Burke area are caused by the intersection of planar features and by the corrugation of surfaces due to microfolding. Mineral alignment, other than that produced by intersecting surfaces, is quite subordinate; amphibole is



Figure 53. Photomicrograph ($\times 18$) of folded quartz veins in phyllite. The schistosity (S_1) of the phyllite has been microfolded and is cut by slip cleavage. From south of Red Village (G1501).

sometimes aligned in the foliation of amphibolite and biotite sometimes occurs in streaks.

The pitch of some lineations has been measured in the field, but the great majority of determinations have been made on oriented specimens in the laboratory, where the nature of the lineation can be ascertained more readily and where fine lineations, which might be undetected in the field, are more visible. The pitch of the lineations has been converted to plunge by construction on a Wulff stereonet, and the plunges are plotted on Plate 3. The plunges have also been plotted on a Schmidt stereonet and the dot plots contoured; the resultant stereograms are shown in Figures 56E and F and on Plate 5, Figures 22 to 30.

B₁ lineation: Frequently individual hand specimens possess more than one lineation. Rarely one of the lineations is caused by the intersection of bedding (S_0) and schistosity (S_1); these instances are designated B_1 lineations, for they are the "b" lineations of an earlier deformation. As they are few in number, they are included with the earlier (B_1) folds on the aforementioned stereogram, Figure 56C.

B₂ lineation: The most prominent lineations are caused by the inter-

section of S_2 cleavage surfaces and schistosity; these are designated B_2 lineations, the "b" lineations of a later deformation. They are shown for the quadrangle as a whole in the stereogram, Figure 56E, and for subarea I in Figure 22 on Plate 5. The "b" lineations are often very coarse and mark the axes of microfolds in S_1 . Frequently, however, it has not been possible to determine the type of lineation. The undetermined ones have been called undifferentiated lineations on the maps and stereograms; in many cases they are probably B_2 lineations. The stereogram, Figure 56F, includes all of the 330 B_2 lineations and all of the 618 undifferentiated lineations measured. The two diagrams, Figures 56E and F, are very similar; there is no evidence that the undifferentiated lineations are statistically different from the B_2 lineations. This is also true in subarea I, where only the B_2 lineations are included in its stereogram (Plate 5, Fig. 22), whereas in all of the other subarea lineation stereograms (Plate 5, Figs. 23 to 30) undifferentiated lineations are included with the B_2 lineations because of the scarcity of data. It is possible that some "a" lineations, i. e., fine striations parallel to the direction of transport of the deformed rock, are included among the undifferentiated lineations. An analysis of the stereograms shows that in subarea V there are two concentrations, one of which is mainly of B_2 lineations and the other mainly of undifferentiated lineations; they lie 60° apart on the great circle defining a girdle of lineation. In subarea XI a similar relationship is found, but in the remaining subareas there is no mutually exclusive preferment of B_2 or undifferentiated lineations. The stereograms on Plate 5, Figures 23 to 30, thus probably define the B_2 lineation directions.

When specimens show two or more lineations, it is sometimes possible to determine their order of development. Thus, in Figure 54 a coarse lineation marks the intersection of the prominent S_2 surface with the schistosity (S_1); a later lineation crosses both the S_1 and S_2 surfaces, but no measurable surface coincides with it. It appears to be caused by an incipient cleavage formed later than the S_2 cleavage and presumably in response to a changed stress field. Figure 55 exhibits three lineations. Firstly, there is a coarse one caused by S_2 intersecting S_1 . This is cut by a prominent ripple-type lineation which appears to be due to the intersection of an incipient cleavage, but which is not seen as a structural surface. The third is a finer but more generally developed ripple-type lineation caused by another incipient cleavage. The two ripple-type lineations do not appear to cut one another and they may be contemporaneous.

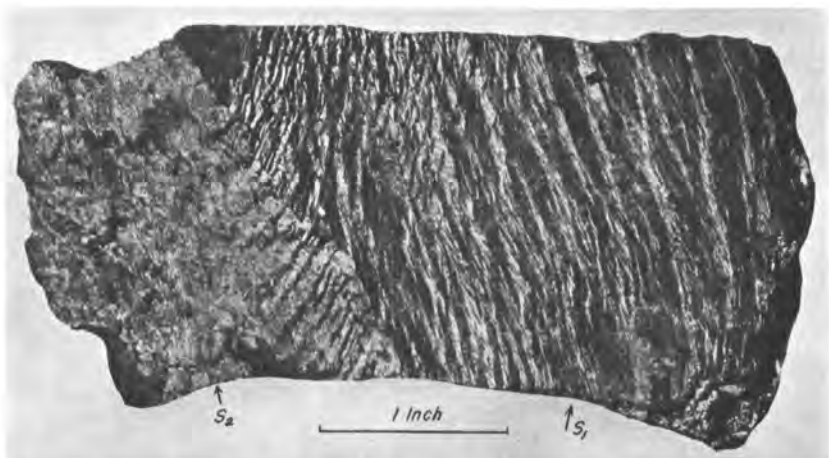


Figure 54. Photograph of phyllite showing two lineations. One lineation (B_2) is caused by the intersection of schistosity (S_1) and S_2 cleavage and produces a strong rippling of S_1 . The other lineation cuts both S_1 and S_2 , and is caused by a later incipient cleavage (S_3) intersecting them. From Red Village (G1005).

All multiple lineations have been measured and plotted on the lineation stereograms.

The B_2 lineations (Fig. 56E) have a maximum concentration trending 40° and plunging 45° to the northeast, which is somewhat different from the maximum concentration of the B_2 folds (Fig. 56D). The lineations fall on a girdle striking 13° and dipping 65° to the southeast, which, as with the girdle of B_2 folds, is similar to the attitude of the maximum of the S_1 planes. The subarea diagrams (Plate 5, Figs. 22 to 30) show, however, a diversity in the attitude of the lineation maxima. In subareas I and II the maxima are similar to that for the composite plot; in III the maximum is to the north-northeast and much gentler; in V there are two maxima, one plunging gently to the northeast and the other plunging gently to the east-southeast. In subareas VI to X the lineation maxima lie much farther to the east, even to the east-southeast in some cases.

Lineation girdles: Although the lineation stereograms have a distinct maximum in all cases, a well-developed girdle is also often present. Weiss and McIntyre (1957) give an excellent account of the geometry of superimposed fold systems and, in particular, point out that an early rectilinear lineation, if folded in a later deformation, will produce a girdle of points around the later fold axis direction, whereas lineations

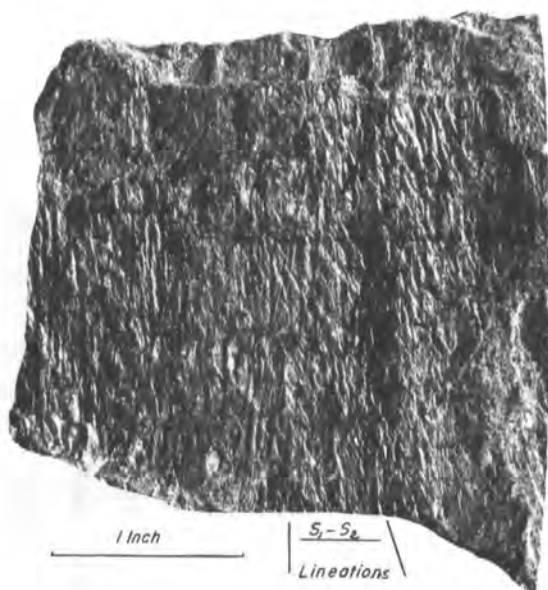


Figure 55. Photograph of schistosity (S_1) of Gile Mountain phyllite showing multiple lineations. The intersection of S_1 and S_2 surfaces produces a coarse horizontal lineation. A rippling of the S_1 surface in two other directions may be caused by incipient cleavages. From north of Granby (F3701).

produced by the intersections of a later cleavage across earlier folded surfaces will fall on a great circle defined by the later cleavage surface. The lineation girdles in the Burke quadrangle do not conform to the former of these geometric relations, as the poles to the lineation girdles do not coincide with the B_1 , the B_2 , or the β_2 directions (Fig. 57A). The girdles do, however, coincide approximately with the attitude of the S_2 plane maximum in subareas III, V, VI, VII, VIII (in part), IX, and XI, although the girdles lie nearer to the plane of the S_1 maximum for the quadrangle as a whole and for subareas II and VIII (in part) and between the maxima of S_1 and S_2 planes for subarea I. The girdles may thus represent the intersections of an S_2 cleavage across previously folded S_1 surfaces. However, the axes of folds in S_1 also define a girdle with a similar attitude to the corresponding lineation girdle (Figs. 57A, B, and C), and S_2 is often seen to be axial plane to folds in S_1 . The B_2

lineations thus are mainly contemporaneous with the B_2 folds in the S_1 surface, not later.

It is suggested that the explanation for the lineation girdle is that the generation of S_2 cleavages at varying attitudes around β'_2 would produce different intersections of S_2 on S_1 . Likewise, varying attitudes of S_1 would produce a similar spreading of the lineation plunges. Why some subarea girdles parallel the S_1 surface maximum and others parallel the S_2 surface maximum is not completely understood. It may be that in the former instance the attitudes of the S_1 surfaces vary less than the S_2 and in the latter instance the converse is true.

Axial directions: Regional axial directions may be constructed (Wegmann, 1929) from a stereonet plot of the poles to a form surface by determining the normal to the great circle along which these poles lie (i. e. a π -diagram). The result is similar to a β -diagram (Sander, 1942). Axial directions have been constructed in this way from the strikes and dips of the S_1 surfaces (β_2 directions) and of the S_2 surfaces (β'_2 directions) for the whole of the Burke quadrangle (Fig. 57A) and for the subareas (Figs. 57B to K). β_2 directions plunge a moderate or small amount to the northeast in the quadrangle as a whole and in subareas I, II, III, V, VI, and VIIIA; they plunge steeply eastwards in subareas VIII (and VIIIB) and gently southwards in subareas VII, IX, X, and XI.

The S_2 planes do not form a folded structure; the β'_2 directions simply indicate the plunge of the line of intersection of all S_2 surfaces in the quadrangle. On the assumption that all of the S_2 surfaces are genetically related, their intersections define the intermediate strain direction for the field concerned. β'_2 directions plunge a moderate to small amount to the northeast in the quadrangle as a whole and in subareas I, II, III, V, and X; moderately to the east in subareas VI and VIII, and gently to the southeast in subarea IX.

Regional "b" directions can also be constructed statistically from the plunge of the intersections of the maxima of the S_1 and S_2 surfaces, which are plotted in Figures 57A and K. Northeasterly plunges are recorded for the quadrangle as a whole and in subareas I, II, V, and VI; southeasterly plunges in subareas VIII and IX; southerly plunges in subarea III, and southwesterly plunges in subareas X and XI.

Rodding: Occasionally quartz rodding is present in some outcrops (e. g., Fig. 8). It is caused by the close folding and shearing of quartz veins during the later phase of deformation. The rods represent the cores of nearly detached folds.

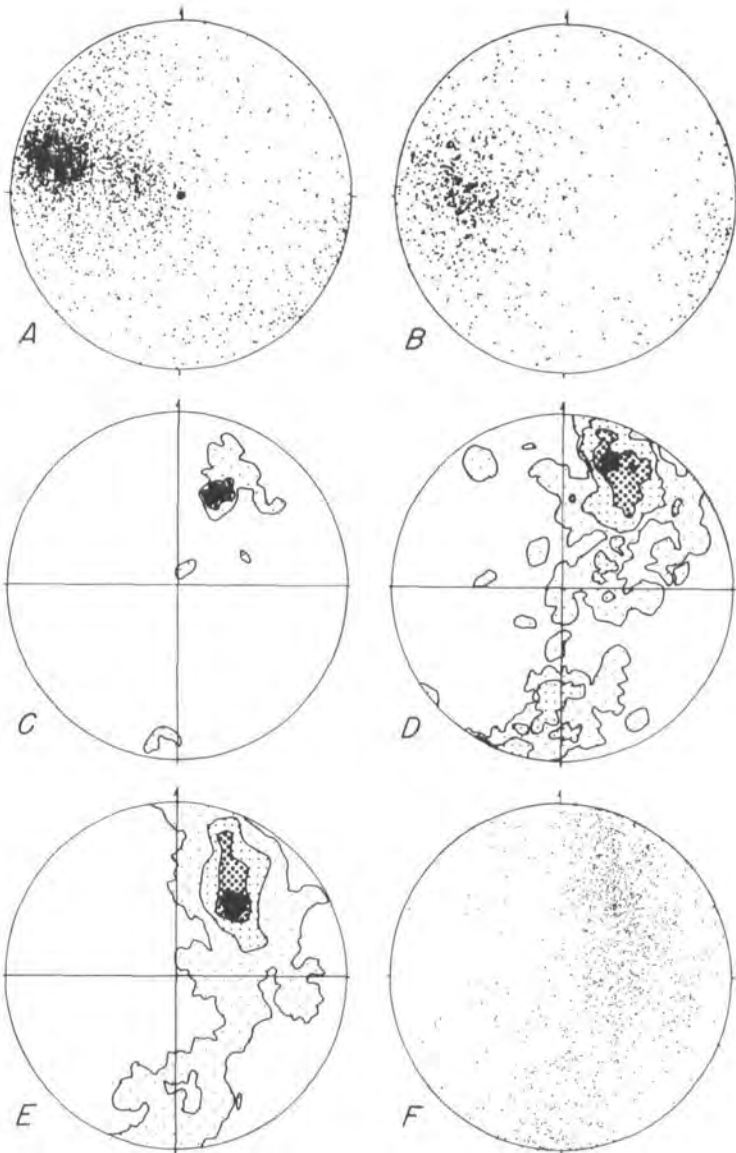


Figure 56. Composite stereograms of Burke quadrangle structural data.
 A. Plot on the lower hemisphere of a Schmidt stereonet of poles to 1811 schistosity (S_1) planes in the Waits River and Gile Mountain formations.
 B. Similar plot of poles to 766 cleavage (S_2) planes.

Boudinage: Boudins have been observed at a small number of localities in the Burke quadrangle. The most common variety is produced from quartz veins by cross-cutting S_2 cleavage surfaces, along which some movement has occurred (Fig. 28). These boudins are similar to what Wilson (1953) describes as cleavage mullions. Comparable structures, but oval in cross-section, are well seen in a disrupted band of calc-silicate hornfels, intercalated in a fold and cut by cleavage, on Burke Mountain (H1254).

More interesting boudins are found in the large outcrops on the east side of Route 114 near the northern boundary of the quadrangle (B1848 and B1751). These have the characteristic elliptical cross-sections, the long axes of which vary in length from five inches to five feet. These boudins are composed of calcareous amphibole-quartz rock. A thin section of one specimen contains, in addition, clinozoisite, sphene, and plagioclase and is cut by narrow, folded quartz veins. Garnet occurs in some of the boudins. The outer surface of the boudins has a sheen of mica. One series of boudins, two and one-half inches thick, is intercalated in impure limestone. The plunge of the line between the boudins at these localities seems to be approximately down the dip of the strata; the boudins thus trend between 69° and 79° and plunge 56° to 59° to the northeast. This orientation is parallel to one of the two B_2 lineation maxima of subarea VIII and is also parallel to β_2 ; the boudins thus lie parallel to the "b" direction of the subarea.

RELATIONSHIPS OF BURKE STRUCTURAL DATA

The relationships of the various structural data are shown on the composite stereograms in Figures 56A to F and are shown synoptically on Schmidt stereonet plots in Figures 57A to K and in plan on the outline map in Figure 58. The simplicity of the data for the quadrangle as a whole (Fig. 57A) is misleading, as is evident from a study of the subarea data (Figs. 57B to K). The synoptic plot shows β_2 closely coincident with the maximum of B_2 fold axes and with the maxima due

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- C. Similar plot of 33 earlier (B_1) fold axes and lineations; contoured at 6-12-15% per 1% area.
 - D. Similar plot of 168 later (B_2) fold axes; contoured at 1-3-6-9% per 1% area.
 - E. Similar plot of 330 B_2 lineations, due to the intersection of schistosity (S_1) and cleavage (S_2); contoured at 1-4-7-10% per 1% area.
 - F. Similar plot of 948 lineations, including the 330 B_2 lineations in (E) and 618 undifferentiated lineations.

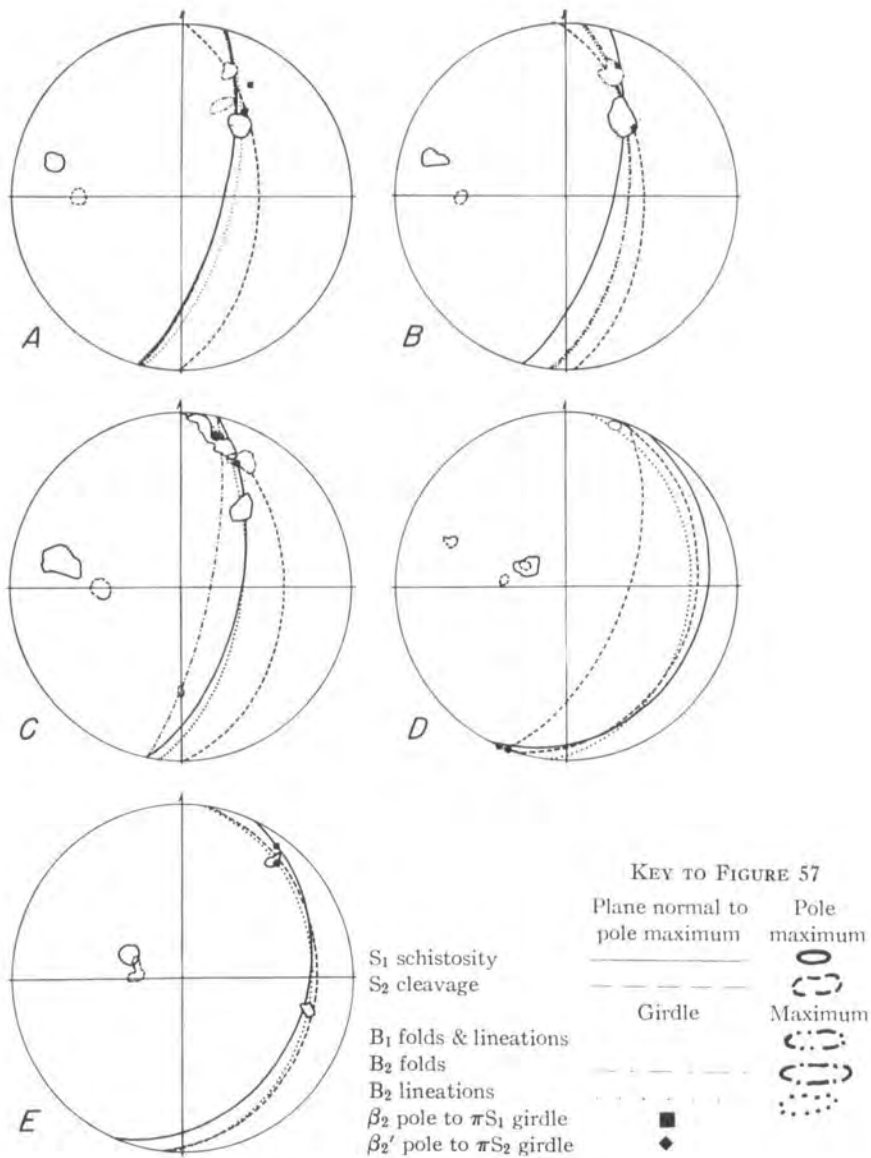


Figure 57. Continued on page 109.

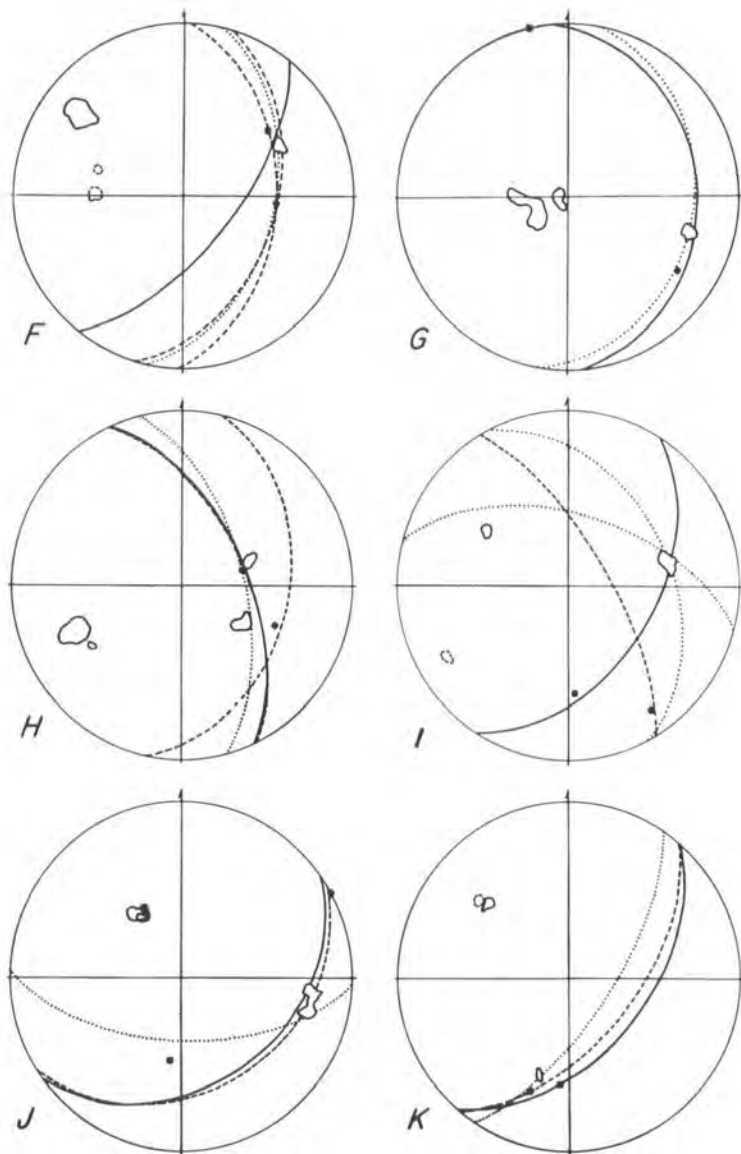


Figure 57. Synoptic stereograms of Burke quadrangle structural data.
 A, entire area; B, subarea I; C, subarea II; D, subarea III; E, subarea V;
 F, subarea VI; G, subarea VII; H, subarea VIII; I, subarea IX; J, subarea X;
 K, subarea XI.

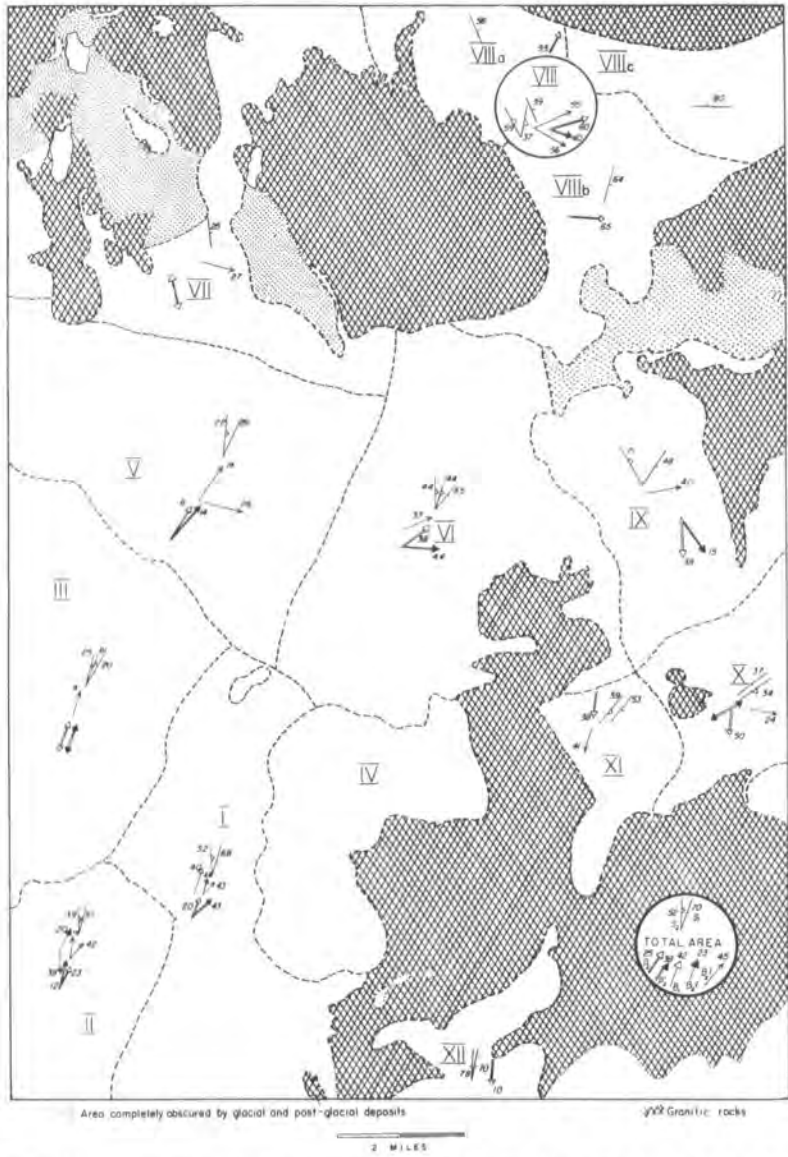


Figure 58. Map of generalized, statistically-derived structural data of the Burke quadrangle and its subareas. Roman numerals refer to subareas. All symbols as used for total area diagram: S_1 , schistosity; S_2 , cleavage; β_2 , pole to πS_1 girdle; β'_2 , pole to πS_2 girdle; B_1 earlier fold axes and lineations; B_2f , later fold axes; B_2l , later lineations.

to the intersection of S_1 and S_2 planes. β'_2 lies close to the maximum of the B_2 lineation, and the girdle of B_2 folds and B_2 lineations is coincident with the S_1 plane maximum.

The schistosity (S_1) has a strong preferred orientation throughout; its strike, however, shows a distinct swing to the northwest in the northern part of the quadrangle and a more easterly swing in the eastern part. The S_2 cleavage also shows a strong preferred orientation. Over much of the area it has a constant relation to S_1 , namely, it trends west of and has a gentler dip than the latter; in three subareas (III, IX, and XI) it has a steeper dip, while in one subarea (X) it trends slightly east of S_1 . The β_2 and β'_2 directions do not coincide. The north-northeast plunge of β_2 steepens from 20° to 23° in the southwestern part of the quadrangle to 55° in the northeastern part, and the β_2 trend varies from northeasterly in the western part of the quadrangle to easterly in the center and northern parts and to southerly in the eastern part. β'_2 plunges northeasterly in the southwestern subareas (I and II), but plunges easterly in subareas VI and VIII and southeasterly in subarea IX. These variations indicate an inhomogeneity in the structure of the quadrangle as a whole.

The small amount of detail available on B_1 folds and lineations indicates that they plunge 42° to the northeast. This is the regional plunge of the structures formed during the earlier phase of deformation. Although these structures have been affected by the later deformation, it is considered that the later movements superimposed mainly smaller scale structures on the earlier structures, so that the present regional B_1 plunge is closely similar to its original attitude and represents, therefore, the axis of intermediate strain of the earlier deformation. The axis of greatest strain may lie in the schistosity (S_1) maximum and normal to the B_1 direction; this axis would trend about 172° and plunge 48° to the southeast. It may lie at a small angle to the S_1 plane if the deformation that produced the schistosity was substantially one of simple shear.

The regional B_2 fold maximum trends 21° and plunges 23° northeast; plunge to the south is important only in subarea II. The regional B_2 lineation trends 40° and plunges 45° northeast, but the trend actually varies throughout the area, as noted above. In particular, the northern and eastern subareas show an easterly or southerly plunge, compared to the northeasterly plunge in the western subareas. The regional β_2 axis is close to the B_2 fold maximum, while the β'_2 axis is close to the B_2 lineation maximum. The subareas, however, again show greater varia-

tions. The northern and eastern subareas have β_2 axes plunging in an easterly or southerly direction.

The girdle of B_2 lineations may be explained by the varying attitude of S_2 surfaces superimposed on strongly preferred, but also varying, S_1 surfaces. Later S_2 surfaces may perhaps also have developed in part after the folding of S_1 surfaces. The lack of coincidence between the β_2 and β'_2 axial directions may be explained by the inclined and somewhat varying attitude of the S_1 surfaces to the axis of greatest stress; β'_2 would tend to develop parallel to the axis of intermediate strain, but the β_2 axis (and the B_2 fold axis) would develop at varying trends and plunges depending upon the attitude of the S_1 surfaces. For the area as a whole the attitude of the maximum of S_2 planes may approximate the AB plane of the strain ellipsoid, and may contain the greatest and intermediate axes of strain. The intermediate axis would be defined by the β'_2 axis, which trends 37° and plunges 38° northeast. The least strain axis would then coincide with the S_2 pole maximum, which trends 271° and plunges 38° northwest.

Major Structures

The constancy of strike and dip of the bedding schistosity (S_1) suggests that the overall structure in the Burke quadrangle is a simple one. The bedding schistosity strikes north-northeast and dips steeply to the southeast over much of the quadrangle. The dip is more gentle to the west and in the northwestern area it is horizontal; in the northern part the strike swings to the northwest and the dip is steep to the northeast. (The deflection of strike to a more easterly direction in subareas VIIIIC, X, and XI may be explained as the effect of nearby granitic intrusions). The rocks may thus be interpreted as homoclinal and, as the dip flattens to the west, as being on the eastern limb of a very large anticlinal structure, the axis of which lies mainly to the west of the Burke quadrangle but trends north-northeastwards through the northwestern part of the quadrangle, where horizontal or gentle dips to the northeast or north are characteristic. This large structure would plunge north-northeast parallel to the B_1 folds and B_1 lineations. The abundant B_2 folds and the S_2 cleavage were formed during a later deformation; the plunge of these later folds is mainly to the northeast, but southward plunges also occur. The slip movement appears to have been mainly upwards to the northwest on the west side of the bedding schistosity.

The plunge of the minor folds in the southwestern part of the quadrangle, particularly the infold of Gile Mountain strata surrounded by

Waits River strata in the Shonya and Graves hills area, appears to be abnormal. If the Gile Mountain formation is younger than the Waits River, the Gile Mountain must lie in a stratigraphic syncline, but structurally it seems to behave as an anticline, plunging northwards at its northern end and southwards at its southern end. Possibly, however, the plunge of the later folds may obscure the true plunge of the major structure, which is an earlier structure, as shown by the trace of its axial surface paralleling the schistosity (S_1). Alternatively, it may be that the folds and their plunges have been inverted, as L. M. Hall (1959) suggests for the isoclinal folds involving the Waits River-Gile Mountain contact in the eastern part of the St. Johnsbury quadrangle. Reversal of plunges by oversteepening, that is, by rotation around an axis normal to the axial plane is, however, unlikely, as no evidence to support such rotation has been found in the Shonya-Graves area. If the Gile Mountain formation is older than the Waits River, the folds have a normal form for the inverted limb of a major structure.

Turning to the surrounding regions, the northwestern trend of the northern part of the Burke quadrangle continues into the Island Pond quadrangle (Goodwin, 1963); the Gile Mountain formation strikes into the Westmore formation mapped by Doll (1951) in the Memphremagog quadrangle, indicating that the two formations are to be correlated with one another. Westwards, in the Lyndonville quadrangle, the dip of the Waits River formation is gentle over the crest of the major structure called the Willoughby arch (Dennis, 1956), and then steepens to the west and is overlain by a band of Gile Mountain strata which is continuous with the Westmore formation of Doll. The Gile Mountain strata are in a syncline overturned to the east; this has been called the Brownington syncline by Doll (1951) and is recognized by Dennis (1956). To the south, the northeasterly trends of the bedding schistosity of the Burke quadrangle continue into the Littleton quadrangle (Eric and Dennis, 1958) and into the St. Johnsbury quadrangle (L. M. Hall, 1959) where the Willoughby arch and the Brownington syncline are also recognized.

The Willoughby arch has been interpreted by Dennis (1956) and by L. M. Hall (1959) as a later structure and to be a cleavage arch. Hall considers the schistosity over much of the arch to be a later schistosity, of the same age as the slip cleavage to the east of the arch. This is in agreement with the structural interpretations made in the Woodsville quadrangle by White (1949) and by White and Billings (1951). Farther south, in the Strafford area, White and Jahns (1950) describe a broad

cleavage arch (which includes the Strafford cleavage dome) in the Waits River formation. They ascribe its origin to a later upward flowage of the calcareous rocks superimposed on a homoclinal sequence which produced the later reverse drag folds on the flanks and also large folds which were subsequently rolled over the arch into a recumbent position. White and Jahns tentatively correlate the band of phyllite to the west of the cleavage arch with the Gile Mountain formation and they consider the band to be a homocline, the sequence being repeated by a large strike fault in the calcareous rocks just east of the band of phyllite. They also consider the evidence for whether the western band of Gile Mountain is a syncline or the western limb of a larger syncline; in the latter case the Waits River formation of the cleavage arch would lie stratigraphically above the Gile Mountain band and the cleavage arch would be superimposed on the syncline. White and Jahns reject both of these alternatives. Murthy (1957), however, in his description of the geology of the East Barre quadrangle, favors a major synclinal structure. He also suggests that facies changes are important and account for the southern termination of the western band of Gile Mountain in the Randolph quadrangle and for the northern termination of the Waits River of the cleavage arch in the Island Pond quadrangle. (See the discussion of Murthy's explanation by White, 1959, and by Dennis, 1959).

The interpretation of the major structure of an area like the Burke quadrangle is fraught with difficulty because Burke comprises only a relatively small part of the major structure. An interpretation must be consistent with the detailed minor structural data described under Minor Structural Features, and it should also be reasonably in accord with the information available from the aforementioned neighboring quadrangles. The northwestward swing in strike of the bedding schistosity (S_1) proves that the Waits River formation of the Willoughby arch indeed terminates in the Island Pond quadrangle by structural closure. The arch, according to White and Jahns (1950) and to Dennis (1956), is formed by the later S_2 cleavage produced by upward and outward rock flowage; however, the bedding schistosity (S_1) is also deformed into the north-northeastward-plunging arch structure. This deformation of the earlier-formed schistosity was accompanied by the formation of the B_2 folds and S_2 cleavage surfaces. If these structures are to be related to upward flowage caused by the rise of subjacent granitic rocks (Lyons, 1955; Dennis, 1956; L. M. Hall, 1959), then the regional structure of northeastern Vermont would be similar to one of the two structures shown diagrammatically by Eric and Dennis (1958, Fig. 13, p. 58). Because of the great thickness

of Gile Mountain strata exposed in the Burke quadrangle, the presence of a syncline (Ibid, Fig. 13B) immediately west of the Monroe contact is attractive. No direct evidence for such a syncline has been detected, but, as pointed out above, the minor structures of the eastern subareas, as exemplified in the B_2 , β_2 , and β'_2 directions, are different from those of the western subareas. The absence of the Waits River formation on the eastern limb of the syncline could be explained by facies change, for rapid facies changes may be expected normal to the length of an orthogeosyncline such as the one in which the sediments of the Waits River and Gile Mountain formations were deposited. If the contact between the Gile Mountain formation and the Albee formation in this area is a fault, the presence of a major unconformity between the Vermont and New Hampshire sequences is still not excluded (see L. M. Hall, 1959, p. 86).

The writer is, however, attracted to another interpretation of the major structure. This considers large-scale recumbent folding to be present, with the Brownington syncline representing the downward-facing nose of such a fold, the axial plane of which lies across the Willoughby arch. Eric and Dennis (1958, p. 61) briefly mention this as a possibility; see also Goodwin (1963) and Ern (1963). The following data from the Burke quadrangle support the presence of this type of structure:

(1) The schistosity (S_1) is nearly everywhere sensibly parallel to the bedding (S_0), but is seen to be axial plane to the few earlier (B_1) folds recognized in the area. This relationship suggests that translatory movements parallel to the bedding were important in the development of the schistosity and that the folds that were formed in the bedding were to a considerable extent obliterated by continued movement on the mechanically generated schistosity, thereby explaining why few such folds are recognized in a sequence that must be greatly thickened.

(2) The direction of the maximum of the B_1 folds and lineations is similar, but not identical, to that of the B_2 folds and B_2 lineations (Fig. 57A). This suggests that the deforming forces that produced both the B_1 and B_2 folds were operating in general from the same direction.

(3) If it is assumed that the S_2 surfaces represent the AB plane of the strain ellipsoid, then the S_2 pole maximum corresponds to the axis of least strain and corresponds approximately with the direction of greatest stress.

(4) Likewise, the pole maximum of the S_1 surfaces may approximate the least strain axis of the rotational strain that accompanied the development of the S_1 surfaces and that resulted in rotation of the latter

almost into the AB plane. Such a relation between the S_1 maximum and the least strain axis would be expected if considerable translation had occurred.

(5) The statistical axes, i. e., the pole maxima of the S_1 and S_2 surfaces, for the Burke area as a whole (Fig. 57A) lie relatively closely together. The attitudes of the pole maxima vary from subarea to subarea (Figs. 57B to K), but they all lie in a general westerly direction with variable plunges. This may be indicative of the rotational nature of recumbent folding.

(6) The attitudes of the B_2 , β_2 , and β'_2 axial directions also vary from subarea to subarea. In general, however, there is a relatively consistent variation of each of the three axial directions between the western subareas and the northern and eastern subareas. This might be due to difference in tectonic level, for example, the eastern subareas might be on the opposite limb of a large-scale recumbent fold.

(7) If a large-scale recumbent fold is present, then the contact between the Waits River and Gile Mountain formations is an inverted one, for the latter would be in the core of an anticline. The apparently inverted northward-plunging isoclinal folds that affect the contact in the southwestern area of the quadrangle (see section CD on Plate 1) would then be normal for an inverted limb. L. M. Hall (1959) reports abnormal fold structures in the Waits River-Gile Mountain contact in the St. Johnsbury quadrangle and interprets them as folds, the plunge of which has been inverted.

(8) The shear sense, which is mainly up on the west side of the S_1 surfaces, would also be normal, as would dextral minor folds plunging northwards and sinistral minor folds plunging southwards.

(9) In subarea VIII the relationships (Fig. 57H) are interesting in that the β_2 direction trends 75° and plunges 60° northeastwards. This suggests that a major fold plunging in this direction may be present in the area rather than a north-northeastward-plunging one. The β'_2 direction is even more to the east, trending 112° and plunging 40° southeastwards.

Large-scale recumbent folds across the Strafford dome have been mapped by White and Jahns (1950), who interpret them as later stage folds which were rolled over the arch by continued upward flowage of the calcareous Waits River formation. Lyons (1955) has also mapped recumbent folds in the Hanover quadrangle. Murthy (1957, pp. 62-63) describes recumbent minor folds on the top of the arch in the East Barre quadrangle; he regards these as rotated earlier folds (but see White, 1959, and Murthy, 1959a and b). L. M. Hall (1959, Plate 2) also maps recum-

bent minor folds, on the crest of the Willoughby arch in the St. Johnsbury quadrangle.

Unfortunately the data on way-up of bedding in the Burke quadrangle (as described under Bedding, S₀) do not provide evidence of large-scale repetition in the Gile Mountain formation. On balance, the data favor the interpretation that a small majority of the outcrops concerned are the right-way up. However, the author considers these data too sparse to be conclusive one way or the other. On Plate 1, the Gile Mountain is shown as being younger than the Waits River formation in keeping with the generally accepted interpretation of the sequence; however, as is described above, it is possible that the Waits River is the younger of the two.

In summary, large-scale recumbent folding is a distinct possibility. The earlier and later deformation phases recognized in northeastern Vermont would then be successive phases in a continuously developed tectonic framework and the major structure would be related to both. The Strafford-Willoughby arch would represent a culmination, and its steeply-plunging northern end in the Island Pond quadrangle would be a major crossing structure producing a plunge depression transverse to the major recumbent fold.

The relationships at the Monroe contact remain obscure, but perhaps major faulting is present. Further difficulty in building up the overall structural and stratigraphic picture is because of probable facies changes, both normal to and along the strike of the structures. If the Brownington syncline is the downward-facing nose of a recumbent anticline, the structural relations west of the syncline would be more complex than present interpretations indicate. The writer is not able to comment on this problem as he is not well enough familiar with the ground in the neighboring areas.

If the hypothesis of major recumbent folding has to be ruled out, then the interpretation applied to the structures by Eric and Dennis (1958), and indicated in their diagrammatic section in Figure 13B, is a distinct possibility.

METAMORPHISM

The sedimentary rocks and associated extrusive and intrusive rocks of the Burke quadrangle have experienced a complex metamorphic history. There is evidence for two distinct phases of progressive metamorphism and perhaps for a final retrogressive stage. Differences in metamorphic environments are expressed in the resultant mineral assemblages,

and these have been described in the geologic literature and grouped into a series of metamorphic facies broadly related to the kinds and intensity of the metamorphism (Eskola, 1915, 1921; Fyfe, Turner, and Verhoogen, 1958). Metamorphic grade is also indicated by the appearance of index minerals in sediments which are essentially isochemical (Barrow, 1893; Tilley, 1924). The appearance of new minerals corresponds to changes in the conditions of metamorphism, particularly temperature, so that zones of metamorphism can be recognized and traced in the field. Thus, a line drawn on a map marking the first appearance of a particular index mineral, representing the passage from a lower to a higher grade of metamorphism, defines the outer limit of the higher grade zone. Such lines of equal grade are called isograds (Tilley, 1924). Careful recording of the mineralogic variation of a particular widespread rock type enables the zone of progressive metamorphism to be mapped. Pelitic rocks commonly have been chosen for this purpose as they are sensitive to changes in metamorphic conditions.

The isograds of pelitic sediments throughout the Burke quadrangle are shown on Plate 1. The index minerals of progressive metamorphism utilized are biotite, almandine garnet, staurolite, andalusite, and sillimanite. These isograds are very generalized, particularly in some areas, because of the glacial drift cover, the lack of outcrops of the appropriate chemical composition, and the impossibility of covering every bit of the ground, especially in the more inaccessible areas, in sufficient detail in the time available. Furthermore, although the index minerals are usually developed as porphyroblasts (except sillimanite, which is often only present microscopically), it is not always possible to recognize them in the field. Even though many specimens were collected for later laboratory examination it is probable that occurrences of index minerals were missed. Much of the area of the Waits River formation west of the East Branch of the Passumpsic River presents difficulties because of the paucity of outcrops, particularly of rocks of the appropriate composition. It is possible that the garnet zone is more wide-spread than shown on Plate 1. However, although the lines may be amended by further detailed work, they show the generalized distribution of zones of progressive metamorphism.

Mineralogy of the Metamorphic Rocks

Biotite: Deep brown biotite is ubiquitous in the pelitic and quartzose rocks of the area. It occurs as small flakes aligned in the schistosity and, less commonly, in the false cleavage. Occasionally it occurs in narrow,

biotite-rich layers parallel to the false cleavage where frequently the highly perfect basal cleavage of the biotite is oriented at high angles to the rock cleavage. Occasionally layers one to two centimeters thick and paralleling the main schistosity are composed almost entirely of biotite. Porphyroblasts of biotite occur plentifully throughout the area; they cut both the schistosity and false cleavage and are undeformed. Characteristically they are pleochroic straw-yellow to red-brown, they possess prominent pleochroic halos and frequently lines of dust inclusions oriented parallel to the surrounding schistosity, and they sometimes show alteration to chlorite. In the hornfels zone the biotite often has the characteristic hornfels pleochroism to a very dark brown in contrast to the red-brown biotite of the lower grades. It is sometimes bleached and partly altered to chlorite; in some thin sections it is seen to be replaced by felted fibrous sillimanite.

Feldspar: Feldspar occurs in most of the thin sections. In the lower grades of the pelitic rocks it is inconspicuous; in the psammitic rocks plagioclase (usually oligoclase, sometimes albite), untwinned or very rarely with albite twinning, is very sparse. Plagioclase is present in the amphibole quartzite bands and is an important constituent, usually fine grained and untwinned, of the amphibolite (see below, under amphibolite assemblages). Twinned and untwinned plagioclase, fine grained and difficult to determine, but identified as oligoclase in some slides, is present, usually in small amounts in the hornfels; rare myrmekite is present in a few slides. Microcline is present in only two sections out of thirty. A number of thin sections of what in hand specimen appears to be granitized sediment contain notable amounts of feldspars which have very irregular, poorly defined boundaries against the groundmass, and contain abundant bleb-like inclusions, particularly of quartz. In some sections they attain a size up to 1.58 mm., and are untwinned or simply twinned. Optical determination is very difficult, but in several specimens the maximum refractive index indicates oligoclase. It is possible, however, that some orthoclase is also present. These poikiloblastic feldspars appear to have grown metasomatically in the groundmass incorporating much of the latter within the structure, and point to the action of soda- and possibly potash-bearing solutions. Microcline and plagioclase occur also in some thin sections of calcisilicate hornfels.

White mica: White mica occurs as very small flakes (sericite) forming the schistosity in all zones of metamorphism. It is most abundant in the pelitic layers but occurs also in the quartzose rocks where it also forms

the schistosity. Individual grains may be bent when the schistosity is thrown into microfolds. The microfolding results in the sericite forming the secondary false or slip cleavage; whether any new recrystallization of sericite takes place in the latter is difficult to determine but in the extreme examples when the cleavage pervades the rock mass and is indistinguishable from a schistosity some recrystallization is likely to have taken place.

White mica is a very common constituent in hornfels occurring in the great majority of specimens examined in amounts varying from rare to abundant. It occurs in two characteristic ways—as aggregates replacing porphyroblasts of other minerals and in some rocks as very common large porphyroblastic plates up to 12 mm. across. The latter have a peculiar metallic luster and dominate a type of hornfels in which there is little or no relict foliate structure and biotite is much reduced in amount. This hornfels rarely has porphyroblasts of aluminum silicates, although sillimanite is present as microscopic needles and fibers in the quartz and mica and in some cases as partially muscovitized fibrolite. This rock type thus seems to have been extremely muscovitized. The pseudomorphous replacement aggregates of muscovite are discussed below under “shimmer aggregates”. White mica appears also to replace biotite in some specimens in what were apparently biotite-rich laminae and is accompanied by granular iron ores, fibrolite, and tourmaline.

Garnet: Idioblastic, wine-colored to deep red almandine garnets (up to 6 mm.) occur frequently, and their appearance marks the garnet zone. They occur also in the staurolite zone (Fig. 65), but in the high grade zone are usually small, xenoblastic, and rarely are conspicuous. Some are altering to chlorite and in one instance to green biotite. The refractive index of some idioblasts is 1.795 (± 0.004) and the specific gravity is 4.05; these data indicate 70% almandine and 30% pyrope (Winchell and Winchell, 1951, pp. 485-488).

The garnets occur in the pelites and semi-pelites, also in amphibole quartzite bands and rarely in amphibolite. They usually have inclusions of quartz, which are sometimes arranged helicically. They are sometimes fractured and occasionally granulated. Grossularite occurs in calc-silicate hornfels bands, particularly on Burke Mountain, in some cases up to 33 mm. in diameter.

Staurolite: Staurolite has a wide occurrence in the pelitic rocks of the area. It is frequently idioblastic (Figs. 18 and 59); cruciform twins and twins crossing at 60° both occur, but one type seems to characterize a particular area. Large cruciform twins (up to 22 mm. long) are very



Figure 59. Photomicrograph ($\times 18$) of idioblastic staurolite in Gile Mountain hornfels. Note the arrangement of the inclusions of quartz and opaques and the small, lath-like, inclusion-free areas, which represent previous biotite flakes. The groundmass contains quartz, abundant biotite, muscovite, and opaques. From the western slope of East Mountain (F2355).

abundant on Bull Mountain and its northern slopes. The porphyroblasts are usually full of inclusions, sometimes arranged helicically and parallel to the surrounding schistosity and cleavage. They often show complete or partial alteration to sericite and chlorite.

Idioblastic staurolite occurs rarely within the sillimanite zone; xenoblastic grains are often present, however, and over 25% of the thin sections contain them. These occur sparsely throughout the groundmass and are also included within "shimmer aggregates." The former apparently are relicts, but the latter occurrence is more difficult to explain. Some aggregates are undoubtedly pseudomorphous after staurolite and the included grains are relict; others are not after staurolite, and this occurrence is discussed below under "shimmer aggregates."

Andalusite: Porphyroblasts of andalusite have been identified in four thin sections of hornfels and in a number of other specimens it was found by microscopic examination of crushed "shimmer aggregates." At locality F2948 perfect crystals up to 17 mm. long are abundant in a

large exposure of Gile Mountain schist underlying granite. In thin section (Fig. 19) the porphyroblasts are pleochroic colorless to rose and contain opaque inclusions. In the other sections the andalusite is less well formed or is xenoblastic and associated with much muscovite (Fig. 60) and in some cases with sillimanite needles and fibers.

Sillimanite: Sillimanite is of frequent occurrence in the thin sections of the high-grade hornfels. It is presumed to have a general distribution throughout these more highly altered rocks and it seems justified placing them in the sillimanite zone. It occurs both as porphyroblasts up to 22 mm. long, as needles, and as matted fibrous aggregates. The porphyroblasts are well developed on Burke Mountain and are often associated with muscovite; they may be seen lying within muscovite aggregates. Some also occur as parallel aggregates of fibers (fibrolite). Large porphyroblasts are also common on Lees Hill (Fig. 61) and Miles Mountain in rocks believed to be in the Albee formation. Minute needles are frequently observed in the quartz grains; these are assumed to be sillimanite as they occur only in rocks bearing sillimanite or within the sillimanite zone. Matted fibrous aggregates of sillimanite (fibrolite) also occur within muscovite (Fig. 62) and also as sinuous streaks and patches associated with the penetrating dark brown biotite and obviously replacing the latter (Fig. 63). Needle and fibrous sillimanite (fibrolite) inclusions have also been noted in andalusite. In some staurolite xenoblasts, particularly those from "shimmer aggregates," there are curious matted anastomosing fibrous inclusions which may also be sillimanite.

Kyanite: Large brilliant blue kyanite porphyroblasts up to 87 mm. long have been found at one locality, in the roadside outcrop at B1848 of schists of the Waits River formation in the zone transitional to the Gile Mountain formation. The kyanite (Fig. 64) occurs embedded in coarse quartz or pegmatitic vein material which is associated with contorted masses of biotite-rich schist. Coarse pink andalusite crystals also occur with the kyanite, and sillimanite needles (fibrolite) are present in the associated quartz and plagioclase and as fibrolite masses after biotite.

The only other occurrence of kyanite found is at B0347 where it is in small plates visible only with a hand lens.

Cordierite: No cordierite has been definitely identified in any thin section, but some "spotted" rocks are present in a few localities, the spots of which may represent pseudomorphs after this mineral. At locality B1848, near the aforementioned kyanite occurrence, are irregular spots composed of a highly birefringent aggregate which may

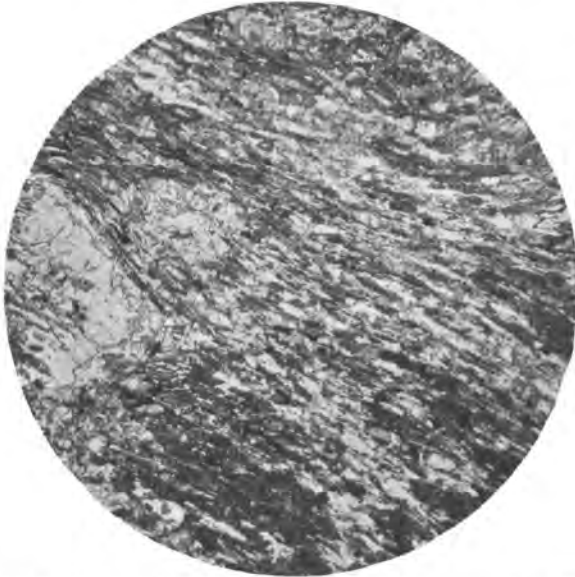


Figure 60. Photomicrograph ($\times 18$) of andalusite-sillimanite schist. The andalusite (on the left) has inclusions of quartz, biotite, and opaques and is surrounded by a corona of muscovite. Sillimanite (fibrolite) occurs in quartz and after biotite (near top and right-hand border). Staurolite xenoblasts are also present. Groundmass contains quartz, biotite, and muscovite. From the eastern side of Route 114 (B1751).

be after cordierite. On the western slope of Kirby Mountain phyllite contains prominent ovoid spots up to 9 mm. long which are composed of aggregates of biotite, chlorite, muscovite, and quartz; these may also be after cordierite but more likely after garnet.

Chlorite: Chlorite is of frequent occurrence throughout the area. Much of it occurs as an alteration product of biotite, amphibole, staurolite, and garnet in all stages of retrogression in all of the zones of metamorphism and in the granitic rocks. In addition, it occurs as porphyroblasts (Fig. 65), sometimes ghost-like, but usually well-defined and large (up to 6 mm. across). This occurrence is also widespread, but it is particularly noticeable in the garnet and staurolite zones of the Gile Mountain formation in the Moose valley east of East Haven Mountain. It does not appear to be replacing any particular mineral. Optically it is slightly pleochroic colorless to light green and full of opaque inclusions. Most of the specimens examined have a refractive index of $N\beta = 1.638 (\pm 0.002)$ and very low anomalous birefringence, and are



Figure 61. Photomicrograph ($\times 18$) of sillimanite-garnet hornfels, showing idiomorphic sillimanite (one surrounded by garnet) containing inclusions of quartz, biotite, and muscovite; some fibrolite. The other minerals are biotite, muscovite, a little microcline, oligoclase, and iron ores. From Lees Hill (I2725).

nearly uniaxial positive. According to Winchell and Winchell (1951, p. 383) this would be ripidolite, an alumina-iron-rich chlorite; one specimen is probably diabantite. Penninite, with ultra-blue polarization colors also often occurs, apparently as an alteration of biotite or amphibole.

It is difficult to determine the relationship of the chlorite porphyroblasts to the biotite, garnet, and staurolite porphyroblasts. They are of late growth and are later than the development of the slip cleavage. It is possible that they represent a retrograde development during falling temperature contemporaneous with the chlorite formed by the alteration of biotite, hornblende, staurolite, etc., or else they were formed contemporaneously with the other porphyroblasts during progressive metamorphism. Dennis (1956, p. 74) describes euhedral porphyroblasts of chlorite occurring "... randomly distributed over the whole (Lyndonville) quadrangle, but comparatively rare." Doll (1951, p. 71) also records large chlorite porphyroblasts in the Memphremagog

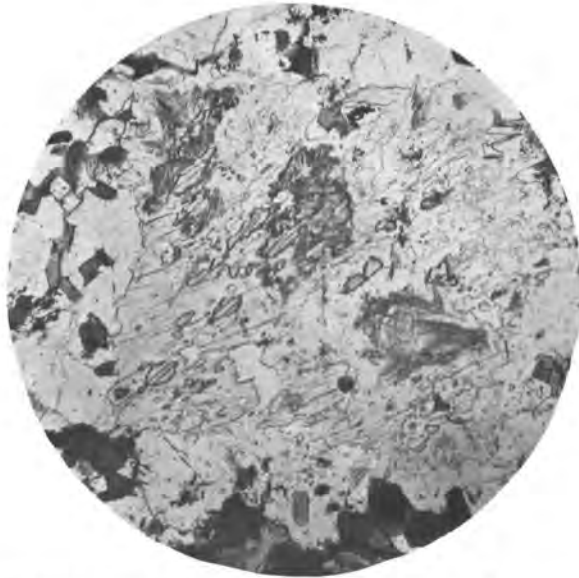


Figure 62. Photomicrograph ($\times 44$) of hornfels showing the large square cross-section of a "shimmer aggregate," composed of muscovite, with abundant felted fibrolite inclusions, and prisms of sillimanite; some biotite and quartz occur within the pseudomorph. The groundmass contains quartz, biotite, plagioclase, and iron ores. From Burke Mountain (H0653).

quadrangle. Porphyroblastic chlorite thus has a wide regional development in northeastern Vermont.

Tourmaline: Tourmaline is an occasional accessory mineral, some of which may be recrystallized detrital material. However, in some cases larger prisms occur up to 5 mm. long. Tourmaline has also been noted in the "shimmer aggregates," where it may indicate metasomatic introduction of boron. It is pleochroic brownish (e) to olive-green or bluish-green (o). Tourmalinized phyllite occurs adjacent to a granitic dike on the lower western slope of Kirby Mountain (G3627) and near pegmatitic material in the roadside outcrop north of Bundy School at D4053. Tourmaline is conspicuous with axinite in amphibolite at D3016; the tourmaline is pleochroic colorless (e) to blue and smoky gray (o).

"*Shimmer aggregates*": Perhaps the most characteristic macroscopic feature of many of the outcrops of the sillimanite zone, as well as part of the staurolite zone, is the presence of abundant large aggregates of

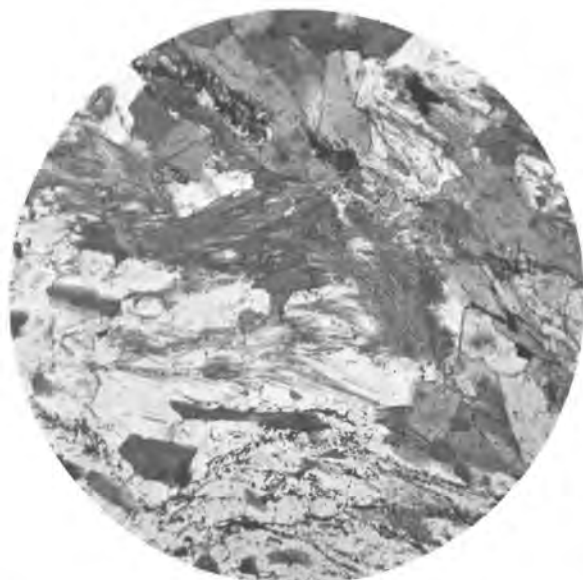


Figure 63. Photomicrograph ($\times 44$) of hornfels showing abundant biotite, in part altered to flamboyant masses of fibrolite; abundant muscovite, with inclusions of fibrolite, biotite, and iron ores; xenoblastic staurolite (right-hand border); quartz; and sparse oligoclase. Garnet and coarser sillimanite occur in an unphotographed portion of the slide. From the col between Burke and Kirby mountains (H0843).

muscovite, called "shimmer aggregates" by Barrow (1893). In the Burke area the mica making up these is often quite large, certainly larger in size than many typical "shimmer aggregates" described by others. These aggregates weather out prominently on relict bedding or schistosity surfaces and have diverse orientations; although many lie with their long direction parallel with the surfaces, many others do not. The aggregates retain the shape of the replaced mineral, and range in size from 8 mm. to 40 mm. in length and 2 mm. to 7 mm. square in cross-section. Occasionally some stubbier masses occur, with pairs arranged in orientation similar to that of staurolite twins; such aggregates are replacing staurolite and are present more usually in the staurolite zone (e. g., at B1353), and rarely in the high-grade hornfels zone.

The form of the characteristic aggregates indicates that they are pseudomorphous after andalusite or sillimanite. In cross-section they often possess a dark core, caused by inclusions of opaques or biotite,



Figure 64. Photomicrograph ($\times 18$) showing an andalusite porphyroblast with inclusions of kyanite (top left), a kyanite porphyroblast (top right), and biotite, in part altered to fibrolite (right). The large area in the center is partially kaolinized oligoclase with inclusions of fibrolite; quartz with fibrolite inclusions is also present. Xenoblastic staurolite occurs elsewhere in the slide away from the quartz-oligoclase vein. Muscovite is abundant in the vein and elsewhere. From east of Route 114 (B1848).

which is strikingly similar in appearance to chialstolite. In thin section the aggregates are seen to be composed of randomly oriented muscovite flakes up to 0.7 mm. across. In one thin section (Fig. 62) from Burke Mountain (H0653) the aggregates also contain a little quartz, biotite, and sillimanite. The latter occurs as needles, felted fibers, and larger prisms up to 1.26 mm. long and 0.18 mm. square. In cross-sections of the pseudomorphs the sillimanite grains lie normal to the "c"-axis and the several separate grains of a pseudomorph extinguish together under crossed nicols; they are never seen to occupy the cores, but in some cases are in a discontinuous ring between the center and the periphery. In longitudinal sections of the pseudomorphs the sillimanite prisms are parallel to the "c"-axis, their ends splaying out into fibers penetrating the muscovite. The needles and fibrous masses occurring within the muscovite plates have a random orientation, and give the appearance of replacing the muscovite. Crushed "shimmer aggregates" from other



Figure 65. Photomicrograph ($\times 18$) of staurolite-garnet schist, showing idioblastic staurolite and garnet with traces of relict wavy schistosity. Biotite porphyroblasts are abundant; chlorite porphyroblasts also occur (center). The groundmass is composed of quartz, sericite, and iron ores. From the Moose valley (F1519).

specimens reveal the presence of grains of tourmaline, staurolite, and rare andalusite, as well as sillimanite. The staurolite fragments contain matted anastomosing fibers believed to be sillimanite, which appear to be replacing the staurolite.

The pseudomorphs present some difficulties in determining their original identity. The section described above indicates that they were sillimanite initially; it is perhaps strange, however, for the felted fibrous variety to occur as well as the coarser prisms as relicts of a single porphyroblast. Furthermore, no unaltered sillimanite porphyroblasts as large as the usual pseudomorphs are found on Burke or Kirby Mountain, although some as long as 22 mm. are prominent in places. The presence of andalusite grains in some of the pseudomorphs suggests that they were originally that mineral. Andalusite, altering to sericite, occurs in a number of thin sections of hornfels, e. g., in the Moose River at J2147 and near the northern boundary of the Burke quadrangle at B1349 and B1850 (Fig. 60); the andalusite sometimes contains sillimanite needles. No fresh andalusite porphyroblasts are found on Burke or Kirby Moun-

tain; fresh andalusite is prominent on the southern spur of East Mountain. The evidence suggests that the "shimmer aggregates" were originally in many cases andalusite crystals which were in the process of inversion to sillimanite, and the former were altered to muscovite by potash metasomatism which the sillimanite mainly resisted. Some of the pseudomorphs, however, are probably after sillimanite, as partially muscovitized sillimanite is present.

Muscovitization of andalusite and sillimanite is a common phenomenon. Billings (1938) describes such an alteration of sillimanite in western New Hampshire and suggests that it was caused by metasomatic introduction of potash derived from intrusions of the New Hampshire magma series. Heald (1950), too, describes replacement of sillimanite by muscovite in western New Hampshire. Billings (1937) also records the retrograde alteration of sillimanite by reaction with biotite (and water) to produce staurolite and muscovite. It is possible that the staurolite grains in the "shimmer aggregates" are the result of such a reaction, although the presence of included fibrous masses suggests that the staurolite is more likely relict in the process of altering to sillimanite.

Hornblende: Hornblende occurs abundantly mainly in amphibolite but also in thin amphibole quartzite bands. It is often poikiloblastic and xenoblastic, but sometimes develops large prismatic forms, and has a dark green to black color in hand specimen. Another characteristic occurrence is as feathery sprays on foliation planes. The pleochroism varies from X = yellow, light green, or nearly colorless to Z = blue-green, dark green, or light green. No detailed study has been made of the hornblende to determine whether its properties vary systematically with metamorphic grade. Intensely blue-green hornblende is found throughout the area in thin sections of amphibolite from the Waits River, Gile Mountain, and Albee formations, ranging from the upper biotite zone to the staurolite zone (see Eric and Dennis, 1958, p. 50). Hornblende often shows patchy alteration to chlorite and occasionally to biotite.

Tremolite: Porphyroblasts of tremolite are found sporadically in the impure calcareous rocks of the Waits River formation. In some areas it is abundant and characteristically weathers out on the rock surfaces in large, radial, brush-like aggregates. Light green actinolite takes its place in some exposures.

Diopside: Light green, granular diopside, colorless in thin section, is abundant in calc-silicate hornfels bands on Burke Mountain and adjacent to granite in the northwestern area of the Burke quadrangle. It is accompanied by tremolite or actinolite, grossularite, an epidote

mineral (xenoblastic, with ultra-blue interference colors, apparently mainly clinozoisite), and rarely vesuvianite.

Sphene: Sphene occurs sparsely in calc-silicate rocks and often abundantly in amphibolite. It is xenoblastic, "drop-like" in form, and frequently has rims of leucoxene. In a rare instance in an amphibolite sphene surrounds ilmenite.

Characteristic Mineral Assemblages

In pelitic and semi-pelitic rocks (see, also, Woodland, 1963):

<i>Biotite zone</i>	quartz-sericite-biotite-plagioclase \pm groundmass chlorite (\pm chlorite porphyroblasts)
<i>Garnet zone</i>	quartz-sericite-biotite-garnet-plagioclase \pm chlorite
<i>Staurolite zone</i>	quartz-sericite-biotite-garnet-staurolite \pm chlorite quartz-sericite-biotite-staurolite \pm chlorite quartz-chlorite-sericite-garnet-staurolite
<i>Sillimanite zone</i>	quartz-biotite-plagioclase-muscovite-garnet \pm sillimanite quartz-biotite-muscovite-plagioclase-staurolite- sillimanite-garnet quartz-biotite-plagioclase-staurolite-sillimanite- garnet quartz-biotite-muscovite-plagioclase-staurolite- andalusite-sillimanite quartz-biotite-muscovite-plagioclase-andalusite- sillimanite \pm garnet quartz-biotite-plagioclase-staurolite-andalusite- sillimanite quartz-biotite-muscovite-andalusite

Kyanite is present in a few localities; in one specimen it occurs with andalusite and fibrolite. Accessory minerals include tourmaline, apatite, sphene, and opaques.

In quartzose rocks: (i. e., those mainly composed of quartz in all zones and which grade into semi-pelitic rocks upon an increase of mica):

<i>Biotite zone</i>	quartz-biotite-sericite-plagioclase (albite) \pm chlorite \pm calcite quartz-biotite-sericite
<i>Garnet zone</i>	quartz-biotite-sericite-garnet-plagioclase (albite- oligoclase or oligoclase) quartz-biotite-sericite-oligoclase

	quartz-biotite-chlorite-garnet
	quartz-biotite-sericite-chlorite-garnet-calcite
<i>Staurolite zone</i>	quartz-biotite-sericite-plagioclase \pm garnet \pm staurolite
<i>Sillimanite zone</i>	quartz-biotite-muscovite-sillimanite
	quartz-biotite-plagioclase-microcline
	quartz-biotite-plagioclase-garnet
	quartz-biotite-muscovite-plagioclase-sillimanite
<i>In amphibole quartzite (light-colored variety):</i>	
<i>Garnet zone</i>	quartz-hornblende-plagioclase (oligoclase-andesine)-garnet-biotite-chlorite \pm calcite
	quartz-hornblende-plagioclase (oligoclase-andesine)-biotite-chlorite \pm calcite
	quartz-hornblende-plagioclase-garnet-calcite

The amphibole varies from very pale green (Z) to green and to blue-green in different specimens, and is often partly altered to chlorite and calcite. It is believed to be aluminous hornblende, but may be actinolitic in some cases. The biotite is sometimes partly altered to chlorite. Chlorite (often var. penninite) is usually seen to be pseudomorphous after amphibole or biotite. The plagioclase is invariably very fine grained (less than 0.05 mm.) and almost entirely untwinned; its refractive index is higher than that of quartz. Accessory minerals include apatite, sphene, and iron ores.

In amphibole quartzite (dark-colored variety):

<i>Garnet zone</i>	quartz-hornblende-garnet-plagioclase-chlorite-clinozoisite
	quartz-hornblende-garnet-plagioclase-biotite-chlorite

The hornblende has Z = blue-green or green, but is often bleached with patchy polarization and prominent pleochroic halos around inclusions; the alteration is possibly metamictization by the inclusions. Accessory minerals include apatite, sphene, and abundant opaques, including sulfides and ilmenite.

In amphibolite:

<i>Biotite zone</i>	hornblende-plagioclase-biotite-quartz \pm calcite \pm chlorite
	hornblende-plagioclase-quartz-clinozoisite-axinite-tourmaline-calcite
<i>Garnet zone</i>	hornblende-plagioclase-biotite-quartz-clinozoisite-calcite \pm chlorite
	hornblende-plagioclase-biotite-quartz \pm chlorite

	hornblende-plagioclase-garnet-quartz \pm chlorite
	hornblende-plagioclase-biotite-garnet-quartz \pm chlorite
<i>Staurolite zone</i>	hornblende-plagioclase-quartz \pm chlorite
	hornblende-plagioclase-biotite \pm chlorite
	hornblende-plagioclase-garnet-quartz
<i>Sillimanite zone</i>	hornblende-biotite-plagioclase-quartz

Hornblende forms the major part of the rock and ranges in color from green to bright blue-green (Z). The plagioclase varies in composition from oligoclase-andesine to labradorite; it is generally fine grained and untwinned. Biotite is rare to common in amount and is often seen to be forming from hornblende. The chlorite is apparently always retrograde, forming after hornblende particularly. Axinite and tourmaline occur in only one locality, at D3016, in the Waits River formation; they are abundant and represent metasomatic introduction of boron. Accessory minerals are apatite, sphene (altering to leucoxene), epidote, ilmenite, and sulfides.

In impure limestone:

<i>Biotite and garnet zones</i>	calcite (dolomite)-quartz-phlogopite-sericite-chlorite
	calcite-quartz-tremolite and/or actinolite-phlogopite-sericite-chlorite
	calcite-tremolite-quartz-chlorite-sphene
<i>Staurolite and sillimanite zones</i>	(interpreted as representing the hornblende hornfels facies of contact metamorphism in thin bands of the Gile Mountain formation on Burke Mountain and in the transition zone of the Waits River-Gile Mountain formation)
	diopside-grossularite-tremolite-labradorite-calcite-quartz
	tremolite-actinolite-diopside-microcline-clinozoisite-calcite-quartz
	actinolite-grossularite-vesuvianite-sphene-quartz
	diopside-grossularite-actinolite-hornblende-calcite-quartz-sphene
	tremolite-calcite-clinozoisite-microcline-plagioclase-diopside-quartz-sphene

In addition, the following assemblages occur adjacent to granite or in areas of granitic dikes in the Waits River formation in the northwestern part of the quadrangle:

diopside-tremolite-calcite-clinozoisite-plagioclase-quartz-sphene
diopside-actinolite-oligoclase-quartz-calcite-clinozoisite-sphene
tremolite-microcline-andesine-quartz-phlogopite-calcite-clinozoisite
tremolite-quartz-microcline-plagioclase-clinozoisite-phlogopite-
muscovite-sphene

Scapolite is present in a calc-silicate rock adjacent to granite in the north (B1850), and vesuvianite is found in the skarn rocks of this area.

Discussion

TEXTURES AND METAMORPHIC GRADE

The texture and mineralogic composition of much of the pelitic, of the semi-pelitic, and, to a lesser extent, of the quartzose rocks of the area strongly suggest that the metamorphism took place in two main stages separated by a deformational episode. These events were probably not unrelated, but represented a continuous sequence and may have even overlapped in time in different areas. The groundmass of the pelites in the biotite, garnet, and staurolite zones is fine grained and shows a well-developed schistose structure, the mineral assemblage remaining essentially similar, e. g., quartz-sericite-biotite. The schistosity is disrupted by a slip cleavage, and in places nearly obliterated by its development. Porphyroblasts of biotite, garnet, staurolite, and chlorite have grown in the rock cutting both the schistosity and cleavage, which are frequently seen as relicts traced by inclusions in the porphyroblasts (Fig. 65). In some cases biotite, in addition to unoriented porphyroblasts, has formed parallel to the later slip or fracture cleavage presumably by post-cleavage crystallization controlled by structural surfaces. However, rarely, a biotite porphyroblast may be deformed by the slip cleavage. Some garnets show extensive fracturing, which occurs rarely in parallel planes at a high angle to the schistosity and which may also have been produced during the later movements. Doll (1951, p. 73) describes fractured and broken staurolite from the Memphremagog quadrangle. Normally the porphyroblasts show no evidence of deformation and are not granulated, which indicates that they were formed during post-cleavage crystallization.

The grade of the initial metamorphism is indicated by the groundmass of the porphyroblast-bearing pelitic and semi-pelitic rocks in all but the high-grade hornfels. (The later recrystallization of the latter has been so extensive as to obliterate or greatly change the early metamorphic assemblages.) Biotite appears to have been a constituent of the pelites

throughout the area, and it is possible garnet may have been present in places; generally, however, the rocks seem to have been in the biotite zone of progressive metamorphism, which places them in the greenschist facies of regional metamorphism (quartz-albite-epidote-biotite subfacies), and perhaps locally in the garnet zone (quartz-albite-epidote-almandine subfacies).

The impure calcareous rocks of the Waits River formation carry tremolite (actinolite) porphyroblasts sporadically throughout, as a result of the reaction between dolomite and quartz. The calcite-quartz-tremolite assemblage is stable over a wide range of conditions—the whole greenschist facies, including chlorite, biotite, and low to medium garnet zones (Turner, in Fyfe, Turner, and Verhoogen, 1958). The accompanying pelites appear to be mainly in the biotite zone, although the garnet zone is also present, and higher grades with garnet, staurolite, kyanite, and sillimanite occur locally near the large Newark granite mass when the impure calcareous rocks carry diopside. It is not possible to determine whether all or part of the tremolite appeared in the early metamorphism or at the time of the development of the porphyroblasts in the pelites of the Gile Mountain formation. Certainly the knob-like tremolite-quartz aggregates which characterize some of the limestone outcrops show no evidence of distortion, even though the rock has been strongly deformed into flowage folds with concurrent recrystallization. This indicates that at least these porphyroblast occurrences are post-deformational.

The growth of porphyroblasts during the later metamorphism resulted in the assemblages of the biotite, garnet, and staurolite zones, which are characteristic of the developments described from many normal regionally metamorphosed areas (Turner, in Fyfe, Turner, and Verhoogen, 1958). One notable feature is that the groundmass of the rocks in these three zones shows little evidence of recrystallization and the grain size remains small; the minerals are, however, the ones to be expected in equilibrium assemblages in these zones.

One exception is a chlorite schist which is exposed adjacent to amphibolite at G0931. This is the sole outcrop of this type found within the Burke quadrangle, but L. M. Hall (1959, p. 24) reports garnet-chlorite schist at the top of the Standing Pond amphibolite in the St. Johnsbury quadrangle. The schist (Fig. 66) is composed of fine laths of chlorite forming a perfect schistosity, with interstitial quartz and in places much sericite; iron ores occur throughout aligned in the schistosity. Rare green biotite also occurs. The chlorite flakes are bent into microfolds



Figure 66. Photomicrograph ($\times 18$) of staurolite-garnet-chlorite schist, showing poikiloblastic garnet (right) associated with poikiloblastic staurolite (below garnet), which is surrounded by muscovite aggregates and chlorite porphyroblasts. The groundmass is composed of chlorite, quartz, and sericite aligned into a wavy schistosity; rare green biotite and tourmaline are also present. Note "halo" around garnet and staurolite poikiloblasts where groundmass chlorite is almost absent. From northeast of Lyndonville (G0930).

which produce a false cleavage parallel to the short limbs or to the limbs of sharp sigmoidal folds. This cleavage is developed imperfectly at irregularly spaced intervals and often fails to cross the width of the thin section. Sericite and opaques are, in part, aligned in the false cleavage. Interrupting both schistosity and false cleavage are aggregates of sericite and quartz with a rim of large twinned chlorite; the aggregates have no apparent orientation and lines of opaques parallel to the false cleavage cross them. Around the aggregates the groundmass is almost devoid of chlorite and is composed of quartz; this produces a halo-like effect. Large porphyroblasts of garnet (up to 6 mm. across) also interrupt the schistosity and cleavage; they have a very spongy peripheral zone with abundant inclusions of quartz and some chlorite. Staurolite is often associated with garnet. Sometimes staurolite twins penetrate to the center of garnet from its periphery; or an individual crystal may be

enclosed in garnet. The staurolite always has a rim of sericite alteration; large chlorite flakes occur around the periphery of both staurolite and garnet. The garnet and staurolite also have a halo-like zone (0.2 to 1.4 mm. wide) poor in groundmass chlorite surrounding them. Some green tourmaline is present. The sericite aggregates are probably pseudomorphous after staurolite.

The paragenesis appears to have been: (1) chlorite schist formed during early deformation, with possibly the development of some garnets (now the inclusion-free cores of the porphyroblasts), (2) false cleavage produced, (3) further garnet growth, with staurolite porphyroblasts then growing in the garnet, and (4) partial retrograde alteration of staurolite to sericite and chlorite and slight alteration of garnet to chlorite. It is difficult to explain why the groundmass remained chloritic during stage (3) unless the rock was potash-deficient or the chlorite was originally alumina-rich. High alumina content stabilizes chlorite into higher metamorphic grades (Harker, 1939, p. 216; Ramberg, 1952, pp. 59, 141). The staurolite and garnet do not appear to be relict in a retrograde chlorite schist. The chlorite schist may be the result of differential shearing stress which affected the margins of the more resistant amphibolite mass more strongly than elsewhere during the early regional metamorphism and formation of schistosity. The porphyroblasts developed during the post-cleavage metamorphism, which was caused particularly by a rise in temperature.

The siliceous phyllitic marble of the Waits River formation, excluding the calc-silicate rocks, has a very similar texture throughout the Burke area except for the erratic appearance, both in quantity and size, of tremolite (actinolite), which is probably related more to varied original composition than to varied metamorphic conditions. Because of its mineral composition the impure marble probably suffered much greater recrystallization during the later movements, which caused considerable internal deformation, as is evidenced by the prevalence of flowage folds, the majority of which are believed to belong to this stage.

Amphibolite also has a similar texture throughout the quadrangle with the exceptions of: (1) the sheared and mylonitized schist in the southwestern part, (2) the granoblastic outcrops in the northern part, which probably represent small gabbroic intrusions that have retained their texture because of their size and shape, and (3) the coarser textured outcrop at B1855 which is adjacent to calc-silicate rock and close to granite. No petrographic evidence has been obtained to determine whether the amphibolite was altered from a lower grade during the later

phase of metamorphism but it probably was. The amphibolite appears to have a broadly similar mineralogy throughout the metamorphic grades; even in the biotite zone there is an absence of chlorite (other than that pseudomorphous after hornblende), epidote minerals, and albite. (It is not possible to determine whether any of the amphiboles are aluminum-poor without chemical analysis). In this respect they may be compared with similar amphibolites of Banffshire, Scotland (Read, 1923, and Wiseman, 1934). Sutton and Watson (1951) suggest that the abnormalities of the Banffshire rocks, where the pelitic rocks also do not conform to the normal zones of regional metamorphism (see below), are the result of temperature control rather than of deficient stress. This may also apply to the Burke area, where temperatures relative to depth of burial were probably higher than normal because of the rise of large subjacent igneous masses. The mylonitized amphibolite with a chlorite-actinolite-? talc assemblage is characteristic of the greenschist facies, produced in this case by shearing stress. If this was developed during either the main regional metamorphism or the later slip cleavage phase, there is no evidence of recrystallization that might have been caused during the post-cleavage (porphyroblastic) metamorphism.

METAMORPHISM IN THE HORNFELS

The highest grade zone of andalusite and sillimanite has some distinct characteristics. The distribution of this zone is closely related to the occurrence of granitic rocks, and much of it is incorporated within a complex of granite and hornfels. The hornfels has a distinctive appearance, although varied, according to the original rock composition and to the extent of metamorphism and metasomatism. Banding caused by alternating beds of contrasting lithology is frequently well preserved, as is the schistosity parallel to it and a cross-cutting slip cleavage. The bedding foliation of hornfels masses may be regular but usually it is highly variable, folded, and contorted; where granitic patches and tongues occur throughout a mass the relict foliation of the hornfels in one part may bear no apparent relation to that in other parts. Occasionally an outcrop shows the rock to be highly contorted and broken up, with the fractured segments wildly disarranged (e. g., Fig. 21). The pre-existing structural surfaces within the complex are thus drastically altered, but in no definable geometric fashion. (The stereogram, Plate 5, Figure 4, of poles to bedding foliation for the Burke Mountain-Kirby Mountain-Umpire Mountain subarea is triclinic). The rocks are noticeably indurated and toughened compared to similar lithologies in the

lower metamorphic grades and, although the general grain size is still small (approximately 0.04 to 0.2 mm.), it is noticeably coarsened and in some cases it is very coarse grained, e. g., with very abundant white mica plates up to 12 mm. across. Adjacent to granitic occurrences the hornfels has often lost its schistose structure and has a "sugary" granulo-appearence, particularly in the quartz-rich types. In part, these may fade gradationally into the normal hypidiomorphic granitic rock, although normally a sharp, but highly irregular, line can be drawn between the latter and the metamorphosed sediments. Hornfelsed quartzose phyllite has characteristically developed a purplish hue apparently due to the crystallization of numerous minute disseminated biotite flakes. The texture and appearance of the rock are typical of thermally metamorphosed rock within an aureole of an igneous intrusion.

As mentioned above, andalusite is present as fresh porphyroblasts in the vicinity of East Mountain and as relicts in "shimmer aggregates"; it occurs along with staurolite and fibrous sillimanite. Both fibrous and coarser prismatic sillimanite is found with staurolite and garnet. Andalusite is normally associated with thermally metamorphosed aureoles in the hornblende hornfels facies (Turner, p. 207, in Fyfe, Turner, and Verhoogen, 1958), where it is often accompanied by sillimanite adjacent to granitic intrusions. Bosworth (1910) describes contact altered rocks of a granite impregnation zone around the Ross of Mull granite where sillimanite occurs as fibers and prisms, often in parallel intergrowths with andalusite. The sillimanite prisms within the andalusite behave almost as a single crystal. (See also Harker, 1939, Fig. 170B and 172B). This is identical with the interpretation given to some of the "shimmer aggregates" occurring in the granite-hornfels complex in the Burke area. Read (1927, p. 324) also describes sillimanite and andalusite in parallel intergrowths in pelitic schists of an injection complex in Cromar, Aberdeenshire, Scotland.

The presence of staurolite and garnet accompanying sillimanite is not uncommon; for example, Billings (1937, pp. 493, 551-552) refers to staurolite as occurring in the sillimanite zone as a retrograde mineral after sillimanite. In most cases in the Burke area the xenoblastic staurolite occurring in the same thin section (Fig. 67) as sillimanite appears to be in relict crystals, indicating incomplete reaction. It is possible that the staurolite grains that are present in the "shimmer aggregates" represent a retrograde product after sillimanite similar to the occurrence described



Figure 67. Photomicrograph ($\times 44$) of staurolite-garnet-sillimanite hornfels, showing garnet (bottom left), staurolite (left and bottom right), and poorly-oriented biotite porphyroblasts in a groundmass of quartz, sericite, opaque dust, and scarce apatite and tourmaline. Fibrolite (right) occurs as tufted aggregates after biotite. From Victory (H3410).

by Billings; the evidence is not equivocal, however, and these, too, may be relict grains.

Not all of the staurolite of the aureole rocks is xenoblastic. In one thin section (Fig. 59) large porphyroblasts occur within a fine-grained hornfels-type groundmass, which is composed mainly of quartz, decussate biotite, and opaque minerals. The staurolite contains many inclusions, apparently of quartz and some opaques, arranged in a regular manner similar to that of chiasolite. The areas of the inclusions are interrupted by lath-shaped, inclusion-free areas which have a size and arrangement similar to the biotite outside of the porphyroblasts. It appears in this case that the staurolite developed late and grew in a rock already possessing a hornfelsic texture. In contrast with this, another thin section shows large staurolite porphyroblasts with inclusions of much finer grain size than the groundmass grains and arranged helicitically but with no relationship to the wavy schistosity of the groundmass. The

staurolite thus grew before recrystallization of the groundmass and has been rotated so that its relict schistosity is no longer parallel with the existing schistosity. This rock contains poikiloblastic garnet, which may be of later origin than the staurolite, and fibrolite after biotite. The garnet of the hornfels is often quite granulated and sometimes shows evidence of rotation, as lines of inclusions are not parallel and do not lead into the surrounding schistosity.

The mutual occurrence of garnet, staurolite, andalusite, and sillimanite (fibrolite) is less common. It may represent polymetamorphism or a type of regional metamorphism described by Turner (in Fyfe, Turner, and Verhoogen, 1958, p. 211) as transitional between the hornblende hornfels facies and the almandine amphibolite facies or the Buchan type described by Read (1952, p. 278). It also fits into the low-pressure intermediate facies of Myashiro (1961). Read (1923, p. 59) describes the regional occurrence of andalusite schists in Aberdeenshire and Banffshire, Scotland, in which staurolite, garnet, and sillimanite also occur, together with cordierite. Regional thermal metamorphism is the explanation given by Read, 1952, p. 278) while Turner (in Fyfe, Turner, and Verhoogen, 1958, p. 206) suggests high temperatures, developed at higher than normal levels, as the cause. Harker (1939, p. 233), however, writes:

"If we suppose, after the production of staurolite, a decided falling off of shearing stress, while the temperature still remained high, the conditions would seem suitable for the formation of andalusite. . . This, it should be observed, is quite different from a second metamorphism, as, e. g., when a staurolite-schist is invaded by a later granite intrusion."

Doll (1951, p. 73) also records the occurrence together of staurolite and andalusite in the Memphremagog area and suggests that the latter formed at a later time than the staurolite, i. e., when the temperature was falling.

In the Burke area the assemblages of the sillimanite-andalusite zone have a close spatial relation to the intrusion of granite and are most widely developed in the mixed granite-hornfels complex. The texture and mineral assemblage of these rocks suggest that high temperature was the important factor in their metamorphism, together with metasomatism by fluids emanating from the granite. The country rocks were already in an advanced stage of metamorphism, caused by a rise in temperature, which was produced by a subjacent mass of granitic magma; it is believed that the development of porphyroblasts was initiated during this stage. Granitic magma was then injected into the

rocks, and the effect of heat and solutions on the immediately adjacent strata was marked by their alteration to a hornfels, bearing andalusite and sillimanite. Rocks with andalusite and no sillimanite may perhaps represent zones little affected by metasomatism, while those with extensive porphyroblastic development of sillimanite may represent the most altered and soaked rocks. This explanation thus does not involve a separate metamorphic episode nor yet is it Read's Buchan type of metamorphism nor Harker's interpretation, although stress effects would decline as the rocks became more plastic as the result of injection. With reference to the facies concept, the sequence may be described as a culminating metamorphic phase, producing the hornblende hornfels facies on rocks already at high temperatures and already in the almandine amphibolite facies of regional metamorphism (see Woodland, 1963, for further discussion of the metamorphism of pelitic rocks).

One further minor comment may be made about sillimanite. It has been noted that both fibrolite and coarser prismatic varieties occur within the same thin section. It might be that these represent two stages of development. Harker (1939, p. 328) holds that a similar occurrence of fibrolite and prismatic sillimanite in hornfelsed schist enclosures in the Ross of Mull granite, Scotland, represents two stages, older fibrolite and newer sillimanite, although Bosworth (1910) and Bailey and Thomas (1925) regard both as belonging to the thermal metamorphism caused by the granite. No positive evidence of age difference has been obtained in the Burke area; the two varieties are probably more or less contemporaneous. Much of the fibrolite occurs as sheafs within large muscovite plates and some replaces biotite. The large increase in muscovite in some of the rocks is probably of late metasomatic origin, produced by circulating solutions emanating from the granite, and it is possible that some or all of the fibrolite may be of similar origin, as is suggested by Watson (1948) for fibrolite in migmatites of Kildonan, Scotland. G. J. Williams (1934, p. 338) in describing roof pendants of biotite-muscovite schist on Stewart Island, New Zealand, explains muscovite containing knotted bunches of fine sillimanite (Ibid, Fig. 2A), apparently identical to occurrences on Burke Mountain, as representing muscovitization of sillimanite, the latter having been formed by the loss of silica by endogenous secretion. Late muscovitization may also explain the fibrolite-muscovite association in the Burke area, but the evidence from thin sections is not definite on this point; however, it appears more likely that the sillimanite is contemporaneous with or perhaps replacing the muscovite.

The occurrence of kyanite in the northern part of the Burke quadrangle is quite unusual. It is found as rare flakes in a few specimens and as large, beautiful crystals up to 8.7 cm. long at B1848. The latter occurrence is in a coarse pegmatitic lens associated with large pink, slightly pleochroic andalusite and with muscovite (Fig. 64). Some of the oligoclase crystals have been deformed and show strained lamellae, and the quartz has undulose extinction. Some pyrite is present. The kyanite is, in part, altered to muscovite; one crystalloblast in thin section is strained and fractured. The associated contorted host rock is biotite-rich. Close to the margins of the veins the biotite is altering to flamboyant masses of fibrolite which also penetrate the associated quartz and plagioclase. The biotite is often bleached and appears to be altering to muscovite with the release of much iron ore; rarely it is also partly altered to chlorite. Ilmenite and sphene are present, both much altered to leucoxene. Fine opaques trace a helicitic structure in the biotite marking a relict wavy schistosity. Xenoblastic staurolite with inclusions of quartz and opaques is a common constituent. Tourmaline and zircon occur as accessories. Some small rare kyanite has also been found at this locality and at B0347.

Dennis (1956) records kyanite associated with granite plutons (all occurrences within 1,500 feet of a granite contact) and suggests that it formed as a contact aureole mineral because of low water pressure and concentration in the pelites. However, kyanite is generally regarded as the high pressure polymorph of Al_2SiO_5 (Fyfe, Turner, and Verhoogen, 1958, p. 165; Clark, et al, 1957), although the stability relations of the three polymorphs (kyanite, andalusite, and sillimanite) are not known for certain. The three have been recorded in the same rock but have not been regarded as contemporaneous (Barth, 1936, p. 790), and Hietanen (1956) suggests that their mutual occurrence in cordierite-mica schist is due to fluctuation of temperatures and stress. Bosworth (1910) describes kyanite on the Island of Mull, Scotland, occurring in pegmatite and passing into tourmaline-bearing kyanite gneiss and also, in another occurrence, kyanite associated with large garnet, staurolite, and fibrolite enveloped in muscovite and fringing kyanite. In the latter area Clough (1911) reports that Flett also detected andalusite besides sillimanite and staurolite, while MacKenzie (1949) states that this kyanite-bearing band extends into the aureole of the Ross of Mull granite where kyanite is transformed into andalusite, particularly in quartz-rich segregations. Read (1934) records kyanite with chloritoid produced from andalusite and staurolite by a later metamorphism, and

Tilley (1935) suggests that the kyanite which occurs in part as replacement of andalusite and cordierite in the hornfels zone of the Carn Chuinneag granite, Ross-shire, Scotland, has been formed during the initial stages of the dynamic metamorphism of the hornfels rocks. Barth (1936, p. 789) describes the occurrence of needle kyanite in shear zones in schists which carry porphyroblasts of staurolite and kyanite, and states that it ". . . crystallized directly from a solution, and apparently in this case did not depend on the action of shearing stress for its formation."

The occurrence of kyanite in the Burke quadrangle is very restricted and must represent specialized local conditions of formation. It appears, however, to be of late development, and the main large porphyroblastic example is associated with and presumably contemporaneous with pegmatitic material. The andalusite is slightly altering to muscovite and is in part intergrown with kyanite, but it is not possible to determine whether one mineral is replacing the other. The fibrolite is after biotite and seems to be directly related to the occurrence of the pegmatitic vein, although biotite altering to fibrolite is surrounded by andalusite in one instance. Clifford (1958) suggests that in the relatively low pressure field andalusite replaces kyanite with increasing temperature. It is possible that the andalusite and kyanite in this occurrence crystallized from aluminous-rich fluids under pressure and temperature conditions at or near the triple point for the three polymorphs, and the fibrolite may have formed contemporaneously from biotite while other biotite close by was muscovitized (see Woodland, 1963).

MUSCOVITIZATION

In many places in and adjacent to the granite there is evidence of late potash metasomatism of the hornfels. This is demonstrated by certain occurrences of abundant large muscovite plates (often containing fibrolite) and also by the extensive muscovitization of the andalusite, sillimanite, and staurolite porphyroblasts. The presence of very abundant biotite in an amphibolite of the hornfels zone may be due, too, to potash metasomatism; a rock from Hobart Ridge (I0941) composed of abundant masses of biotite, masses of chlorite presumably after biotite, plagioclase, and quartz may be another metasomatized amphibolite. Muscovitization of aluminum silicates is a common phenomenon; Billings (1938) describes such an alteration of sillimanite in western New Hampshire which he shows to have been caused by the action of potash-bearing solutions which he suggests were derived from nearby granite plutons.

Heald (1950) describes a similar effect in the Lovewell Mountain quadrangle, New Hampshire. In the Burke quadrangle the source of the potash-bearing solutions could have been the residual fluids of the adjacent granitic rocks.

SUMMARY OF THE GEOLOGIC HISTORY OF THE BURKE QUADRANGLE

Sediments of Ordovician age, represented by the Albee formation in the eastern part of the Burke quadrangle, were laid down in what Kay (1951) has termed the Magog eugeosyncline. These were presumably affected by the Taconic orogeny. Further subsidence of the geosynclinal belt during Silurian and early Devonian time was accompanied by deposition of a considerable thickness of sediments, which, when indurated, included arenaceous and argillaceous limestone, shale, silty shale, quartz siltstone, and rocks grading from quartz wacke to quartz arenite. The calcareous rocks with interbeds of shale and arenite now comprise the Waits River formation, which crops out over the western portion of the quadrangle and extends westwards into the Lyndonville quadrangle. The Gile Mountain formation, which occurs in a wide belt east of the Waits River formation, was laid down as sediments which became indurated into a series of alternating thin bands of shale, siltstone, quartz wacke, and quartz arenite—a lithology characteristic of a rapidly sinking eugeosynclinal belt. The alternation of shale and arenite suggests a thick sequence of mainly graded beds. The calcareous rocks of the Waits River and the non-calcareous sequences in the Gile Mountain are probably equivalent, at least in part, with the former being deposited to the west at the edge of the eugeosyncline (perhaps representing a miogeosynclinal environment), and the latter to the east in the eugeosyncline proper. The exact relations between the two are open to doubt; the Waits River is believed to be older than the Gile Mountain by most workers in northeastern Vermont, while this author considers, on the basis of an alternative structural interpretation, that it may be, in part, younger.

Volcanism was sporadic, and is represented by lava and tuff, both acid and basic, which occur intermittently throughout the Gile Mountain strata and particularly in the transition zone between the Waits River and Gile Mountain formations. Some intrusive activity, resulting in mainly basic dikes and plugs, accompanied the volcanism.

In the Middle Devonian subsidence ceased and the area was subjected to considerable deformation during the Acadian orogeny. The rocks were

strongly folded and underwent low-grade metamorphism (green-schist facies) during the early phase of the orogeny. The movements produced a well-developed schistosity in the rocks, and it is probable that recumbent folding took place as the rocks deformed under the influence of stresses which apparently operated in an east-southeast-west-northwest direction. Continuation of the movement then caused deformation of the earlier-formed schistosity, which was folded and cut by a newly-developed cleavage (slip cleavage) that almost or completely obliterated the schistosity in the more strongly deformed areas. The Willoughby arch structure resulted at this stage from a linear culmination zone.

Meanwhile, the temperature of the rocks of the area was increasing and towards the end of the movements the higher temperatures, accompanied probably by the rise of large subjacent masses of granitic magma, resulted in the thermal metamorphism of the rocks at what was probably a moderate depth in the earth's crust. Porphyroblasts of biotite, garnet, staurolite, tremolite, and actinolite grew at this stage, and cut both the schistosity and the cleavage. Granitic magma then broke into the structures, partly by following their structural surfaces and forcing the country rock aside and partly by wedging and intimate penetration of the rock. Adjacent to the granite the rocks were converted to hornfels, with the production of diopside, tremolite, and grossularite in the calc-silicate rocks developed from impure limestone and with the production of andalusite and sillimanite in the banded hornfels developed from the pelitic and semi-pelitic rocks. Sillimanite was abundantly produced in the intimately injected zones, now comprising the hornfels-granite complexes, where there is also evidence of the activity of fluids in the formation of migmatite and granitized rocks. Muscovitization was a late stage in events and led to the development of much muscovite in the hornfels and to the replacement of staurolite, andalusite, and, to a lesser extent, sillimanite by "shimmer aggregates" of mica.

As the rocks cooled, some retrograde metamorphic effects were produced; chlorite replaced hornblende, biotite, garnet, and staurolite and also appeared conspicuously as porphyroblasts in the pelitic and semi-pelitic rocks. At some later date, in the Mesozoic Era, basic dikes, mainly lamprophyres, were intruded into both the granitic and metamorphic rocks. Since that time, as far as it is known, the area has been subjected to continuous erosion, with periodic uplift enabling the removal of many thousands of feet of rock. The present configuration of the land surface broadly reflects the relative resistance of the various rocks to weathering and erosion. During the Pleistocene the area was completely

covered by ice, and since the withdrawal of the ice weathering and erosion have continued.

ECONOMIC GEOLOGY

The economic deposits of the Burke quadrangle are granite and sand and gravel. Granite has been quarried in the past on a small scale in Newark, e. g., at the northern end of Packer Mountain and northeast of Walker Mountain, and on Kirby Mountain, near the southern border of the quadrangle. There are many opportunities for granite quarrying; the most promising areas are in Newark, particularly east and west of Route 114. An attractive granite, some of it pink in color and within easy access of the highway, occurs west of Route 114 and north of East Newark School, in the slopes leading up to Walker Mountain and Hawk Rock.

Sand and gravel are obtained from eskers, kames, and late-glacial deposits in the valleys of the East and West branches of the Passumpsic River. Other sand and gravel deposits are found near Newark Hollow, Bean School, south of Umpire Mountain, and in the mounds throughout the low-lying area of Victory.

BIBLIOGRAPHY

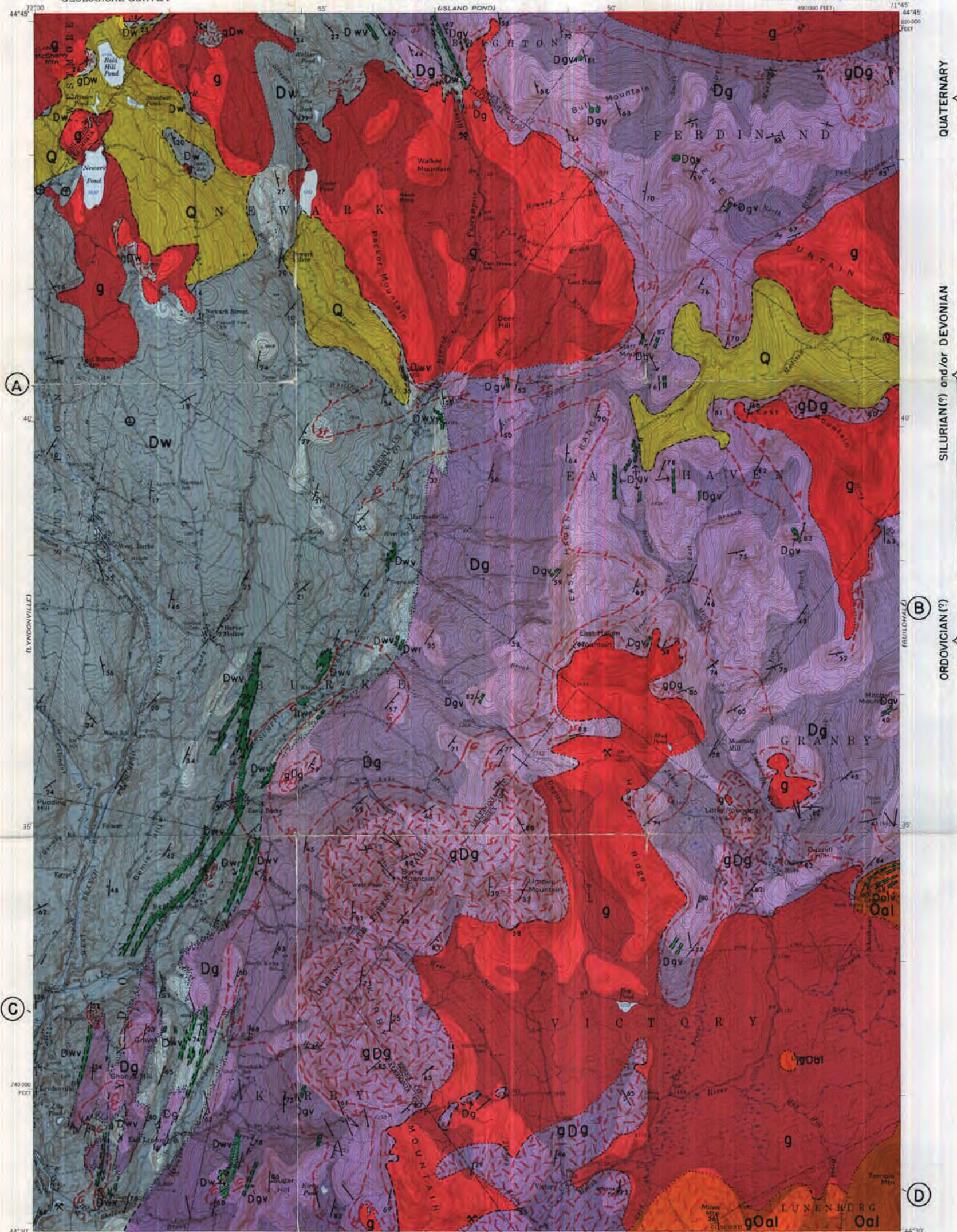
- ADAMS, C. B., 1845, First annual report of the geology of Vermont: Burlington, Vt., 92 pp.
- AMPFERER, O., 1938, Über die tektonischen Begriffe von Totfaltung: Wien Akad. Wiss., Sber., Abt. 1, Bd. 147, pp. 35-42.
- BAILEY, E. B., and THOMAS, H. H., 1925, See LEE, G. W., BAILEY, E. B., et al, 1925, The pre-Tertiary geology of Mull, Loch Aline, and Oban: Geol. Surv. Scotland, Mem., 140 pp.
- BALK, R., 1936, Structural and petrologic studies in Dutchess County, New York, Part I, Geologic structure of sedimentary rocks: Geol. Soc. Am., Bull., v. 47, pp. 685-774.
- BARROW, G., 1893, On an intrusion of muscovite-biotite gneiss in the Southeastern Highlands of Scotland and its accompanying metamorphism: Geol. Soc. London, Quart. Jour., v. 49, pp. 330-356.
- BARTH, T. F. W., 1936, Structural and petrologic studies in Dutchess County, New York, Part II, Petrology and metamorphism of the Paleozoic rocks: Geol. Soc. Am., Bull., v. 47, pp. 775-850.
- BILLINGS, M. P., 1934, Paleozoic age of the rocks of central New Hampshire: Science, n. s., v. 79, pp. 55-56.
- , 1937, Regional metamorphism of the Littleton-Moosilauke area, New Hampshire: Geol. Soc. Am., Bull., v. 48, pp. 463-566.
- , 1938, Introduction of potash during regional metamorphism in western New Hampshire: Geol. Soc. Am., Bull., v. 49, pp. 289-302.
- , 1942, Structural geology: New York, Prentice-Hall, Inc., 1st ed., 473 pp.; 1954, 2nd ed., 514 pp.
- , 1956, The geology of New Hampshire, Part II, Bedrock geology: Concord, N. H., N. H. State Planning and Development Comm., 203 pp.
- BONNEY, T. G., 1886, The anniversary address of the president: Geol. Soc. London, Quart. Jour., Proc., v. 42, pp. 38-115.
- BOSWORTH, T. O., 1910, Metamorphism around the Ross of Mull granite: Geol. Soc. London, Quart. Jour., v. 66, pp. 376-396.
- BRACE, W. F., 1953, The geology of the Rutland area, Vermont: Vt. Geol. Surv., Bull., no. 6, 120 pp.
- CADY, W. M., 1950, Fossil cup corals from the metamorphic rocks of central Vermont: Am. Jour. Sci., v. 248, pp. 488-497.
- , 1956, Bedrock geology of the Montpelier quadrangle, Vermont: U. S. Geol. Surv., Geol. Quad. Map GQ79.
- , 1960, Stratigraphic and geotectonic relationships in northern Vermont and southern Quebec: Geol. Soc. Am., Bull., v. 71, pp. 531-576.
- CLARK, S. P., ROBERTSON, E. C., and BIRCH, F., 1957, Experimental determination of kyanite-sillimanite equilibrium relations at high temperatures and pressures: Am. Jour. Sci., v. 255, pp. 628-640.
- CLIFFORD, T. N., 1958, A note on kyanite in the Moine series of southern Ross-shire, and a review of related rocks in the Northern Highlands of Scotland: Geol. Mag., v. 95, pp. 333-346.

- CLOOS, E., 1946, Lincation: a critical review and annotated bibliography: *Geol. Soc. Am., Mem.*, no. 18, 122 pp.
- CLOUGH, C. T., 1911, See CUNNINGHAM CRAIG, E. H., WRIGHT, W. B., et al, 1911, The geology of Colonsay and Oronsay, with part of the Ross of Mull: *Geol. Surv. Scotland, Mem.*, 109 pp.
- COOKE, H. C., 1950, Geology of a southwestern part of the Eastern Townships of Quebec: *Geol. Surv. Canada, Mem.*, no. 257, 142 pp.
- , 1957, Coaticook-Malvina area, electoral districts of Stanstead and Compton: *Quebec Dept. Mines, Geol. Surv. Branch, Geol. Rept.*, no. 69, 37 pp.
- CURRIER, L. W., and JAHNS, R. H., 1941, Ordovician stratigraphy of central Vermont: *Geol. Soc. Am., Bull.*, v. 52, pp. 1487-1512.
- DAHLSTROM, C. D., 1954, Statistical analysis of cylindrical folds: *Canadian Inst. Min. Met., Trans.*, v. 57, pp. 140-145.
- DALE, T. N., 1892, On plicated foliation: *Am. Jour. Sci.*, 3rd s., v. 43, pp. 317-319.
- , 1894, Mount Greylock, its areal and structural geology: *U. S. Geol. Surv., Mon.*, no. 23, pt. 3, pp. 119-203.
- , 1896, Structural details in the Green Mountain region and in eastern New York: *U. S. Geol. Surv., 16th Annual Rept.*, pt. 1, pp. 543-570.
- , 1899, The slate belt of eastern New York and western Vermont: *U. S. Geol. Surv., 19th Annual Rept.*, pt. 3, pp. 153-307.
- , 1909, The granites of Vermont: *U. S. Geol. Surv., Bull.*, no. 404, 138 pp.
- , 1923, The commercial granites of New England: *U. S. Geol. Surv., Bull.*, no. 738, 488 pp.
- DENNIS, J. G., 1956, The geology of the Lyndonville area, Vermont: *Vt. Geol. Surv., Bull.*, no. 8, 98 pp.
- , 1959, A revision of the Lower Paleozoic stratigraphy in eastern Vermont: a discussion: *Jour. Geol.*, v. 67, pp. 583-584.
- DESITTER, L. U., 1956, Structural geology: New York, McGraw-Hill Book Co., Inc., 552 pp.
- DOLL, C. G., 1943a, A Paleozoic revision in Vermont: *Am. Jour. Sci.*, v. 241, pp. 57-64.
- , 1943b, A brachiopod from mica schist, South Strafford, Vermont: *Am. Jour. Sci.*, v. 241, pp. 676-679.
- , 1944?, A preliminary report on the geology of the Strafford quadrangle: *Vt. Geol. Surv., 24th Biennial Rept. for 1943-44*, pp. 14-28.
- , 1951, Geology of the Memphremagog quadrangle and the southeastern portion of the Irasburg quadrangle, Vermont: *Vt. Geol. Surv., Bull.*, no. 3, 113 pp.
- , CADY, W. M., THOMPSON, J. B., JR., and BILLINGS, M. P., 1961, Centennial geologic map of Vermont, 1:250,000: *Vt. Geol. Surv.*
- ERIC, J. H., 1942, Geology of the Vermont portion of the Littleton quadrangle: Unpublished Ph. D. thesis, Harvard Univ., 101 pp.
- , and DENNIS, J. G., 1958, Geology of the Concord-Waterford area, Vermont: *Vt. Geol. Surv., Bull.*, no. 11, 66 pp.
- , WHITE, W. S., and HADLEY, J. B., 1941, Monroe fault of New Hampshire and Vermont (abstract): *Geol. Soc. Am., Bull.*, v. 52, p. 1900.
- ERN, E. H., JR., 1963, Bedrock geology of the Randolph quadrangle, Vermont: *Vt. Geol. Surv., Bull.*, no. 21, 96 pp.
- ESKOLA, P., 1915, On the relation between chemical and mineralogical composition in the metamorphic rocks of the Orijärvi region: *Comm. géol. Finlande, Bull.*, no. 44, pp. 109-145.

- , 1921, The mineral facies of rocks: *Norsk. Geol. Tidsskr.*, v. 6, pp. 143-194.
- , 1932, On the principles of metamorphic differentiation: *Comm. géol. Finlande, Bull.*, no. 97, pp. 68-77.
- FAIRBAIRN, H. W., 1935, Notes on the mechanics of rock foliation: *Jour. Geol.*, v. 43, pp. 591-608.
- , 1949, *Structural petrology of deformed rocks*: Cambridge, Mass., Addison-Wesley Press, 344 pp.
- FYFE, W. S., TURNER, F. J., and VERHOOGEN, J., 1958, Metamorphic reactions and metamorphic facies: *Geol. Soc. Am., Mem.*, no. 73, 259 pp.
- GOODWIN, B. K., 1963, Geology of the Island Pond area, Vermont: *Vt. Geol. Surv., Bull.*, no. 20, 111 pp.
- HALL, L. M., 1959, The geology of the St. Johnsbury quadrangle, Vermont and New Hampshire: *Vt. Geol. Surv., Bull.*, no. 13, 105 pp.
- HALL, S. R., 1861, See HITCHCOCK, E. et al, v. 2, pp. 718-732.
- HARKER, A., 1886, On slaty cleavage and allied rock structures, with special reference to the mechanical theories of their origin: *Brit. Assoc. Adv. Sci., Rept. for 1885*, pp. 813-852.
- , 1939, *Metamorphism*: London, Methuen & Co., 362 pp.
- HEALD, M. T., 1950, Structure and petrology of the Lovewell Mountain quadrangle, New Hampshire: *Geol. Soc. Am., Bull.*, v. 61, pp. 43-89.
- HIETANEN, A., 1956, Kyanite, andalusite, and sillimanite in the schist in Boehls Butte quadrangle, Idaho: *Am. Miner.*, v. 41, pp. 1-27.
- HITCHCOCK, E. et al, 1861, Report on the geology of Vermont: descriptive, theoretical, economical, and stratigraphical: Claremont, N. H., 2 v., 988 pp.
- HOEPPENER, R., 1956, Zum Problem der Bruchbildung, Schieferung und Faltung: *Geol. Rund.*, Bd. 45, H. 2, pp. 247-283.
- HOLMES, A., 1959, A revised geological time scale: *Edinburgh Geol. Soc., Trans.*, v. 17, pp. 183-216.
- JACOBS, E. C., 1923, The geology of Westmore, Brownington, and Charleston: *Rept. of Vt. State Geol. for 1921-22*, no. 13, pp. 93-108.
- JOHANSSON, W. I., 1963, Geology of the Lunenburg-Brunswick-Guildhall area, Vermont: *Vt. Geol. Surv., Bull.*, No. 22, 86 pp.
- JOPLIN, G. A., 1952, The granitization process and its limitations as exemplified in certain parts of New South Wales: *Geol. Mag.*, v. 89, pp. 25-38.
- KAY, M., 1942, Development of the northern Allegheny synclinorium and adjoining regions: *Geol. Soc. Am., Bull.*, v. 53, pp. 1601-1658.
- , 1951, North American geosynclines: *Geol. Soc. Am., Mem.*, no. 48, 143 pp.
- KULP, J. L., 1961, Geologic Time scale: *Science*, v. 133, pp. 1105-1114.
- LEITH, C. K., 1905, Rock cleavage: *U. S. Geol. Surv., Bull.*, no. 239, 216 pp.
- , 1913, *Structural geology*: New York, Henry Holt & Co., 1st ed., 169 pp.; 1923, revised ed., 390 pp.
- LYONS, J. B., 1955, Geology of the Hanover quadrangle, New Hampshire-Vermont: *Geol. Soc. Am., Bull.*, v. 66, pp. 105-146.
- , JAFFE, H. W., GOTTFRIED, D., and WARING, C. L., 1957, Lead-alpha ages of some New Hampshire granites: *Am. Jour. Sci.*, v. 255, pp. 527-546.
- MACKENZIE, D. H., 1958, The structure of the Grantown granite complex, Morayshire: *Geol. Mag.*, v. 95, pp. 57-70.
- MACKENZIE, W. S., 1949, Kyanite-gneiss within a thermal aureole: *Geol. Mag.*, v. 86, pp. 251-254.

- McKINSTRY, H. E., 1953, Shears of the second order: *Am. Jour. Sci.*, v. 251, pp. 401-414.
- MEAD, W. J., 1940, Folding, rock flowage, and foliate structures: *Jour. Geol.*, v. 48, pp. 1007-1021.
- MIYASHIRO, A., 1949, The stability relation of kyanite, sillimanite, and andalusite, and the physical conditions of metamorphic processes: *Geol. Soc. Japan, Jour.*, v. 55, pp. 218-223.
- , 1961, Evolution of metamorphic belts: *Jour. Petrology*, v. 2, pp. 277-311.
- MURTHY, V. R., 1957, Bed rock geology of the East Barre area, Vermont: *Vt. Geol. Surv., Bull.*, no. 10, 121 pp.
- , 1959a, A revision of the Lower Paleozoic stratigraphy in eastern Vermont: a reply to the discussion by Walter S. White: *Jour. Geol.*, v. 67, pp. 581-582.
- , 1959b, A revision of the Lower Paleozoic stratigraphy in eastern Vermont: a reply to the discussion by John G. Dennis: *Jour. Geol.*, v. 67, p. 584.
- MYERS, P. B., JR., 1964, Geology of the Vermont portion of the Averill quadrangle, Vermont: *Vt. Geol. Surv., Bull.*, no. 27, 69 pp.
- PHILLIPS, F. C., 1954, *The use of stereographic projection in structural geology*; London, Edward Arnold (Publishers) Ltd., 86 pp.
- RAMBERG, H., 1952, *The origin of metamorphic and metasomatic rocks, a treatise on recrystallization and replacement in the earth's crust*: Chicago, Univ. of Chicago Press, 317 pp.
- READ, H. H., 1923, *The geology of the country around Banff, Huntly and Turriff*: *Geol. Surv. Scotland, Mem.*, 240 pp.
- , 1927, *The igneous and metamorphic history of Cromar, Deeside, Aberdeenshire*: *Roy. Soc. Edinburgh, Trans.*, v. 55, pt. 2, pp. 317-353.
- , 1933, On quartz-kyanite rocks in Unst, Shetland Island, and their bearing on metamorphic differentiation: *Miner. Mag.*, v. 23, pp. 317-328.
- , 1934, *The metamorphic geology of Unst in the Shetland Islands*: *Geol. Soc. London, Quart. Jour.*, v. 90, pp. 637-684.
- , 1936, *The stratigraphic order of the Dalradian rocks of the Banffshire coast*: *Geol. Mag.*, v. 73, pp. 468-476.
- , 1952, *Metamorphism and migmatization in the Ythan Valley, Aberdeenshire*: *Edinburgh Geol. Soc., Trans.*, v. 15, pp. 265-279.
- RICHARDSON, C. H., 1902, *The terranes of Orange County, Vermont*: *Vt. State Geol. 3rd Rept.*, pp. 61-101.
- , 1906, *The areal and economic geology of northeastern Vermont*: *Vt. State Geol. 5th Rept.*, pp. 63-115.
- SANDER, B., 1930, *Gefügekunde der Gesteine*: Wien, J. Springer, 358 pp.
- , 1942, Über Flächen und Achsengefüge Westende der Hohen Tauern, III. Bericht: Reichsamts Bodenforschung, Zweigstelle Wien, Mitt., no. 4, pp. 1-94.
- SHROCK, R. R., 1948, *Sequence in layered rocks*: New York, McGraw-Hill Book Co., Inc., 507 pp.
- SORBY, H. C., 1858, On some facts connected with slaty cleavage: *Brit. Assoc. Adv. Sci., Rept. for 1857, Trans. of Sec.*, pp. 92-93.
- , 1880, On the structure and origin of noncalcareous stratified rocks: *Geol. Soc. London, Quart. Jour., Proc.*, v. 36, pp. 46-92.
- SUTTON, J., and WATSON, J., 1951, Varying trends in the metamorphism of dolerites: *Geol. Mag.*, v. 88, pp. 25-35.

- SWANSON, C. O., 1927, Notes on stress, strain, and joints: *Jour. Geol.*, v. 35, pp. 193-223.
- , 1941, Flow cleavage in folded beds: *Geol. Soc. Am., Bull.*, v. 52, pp. 1245-1264.
- THOMPSON, J. B., Jr., 1955, The thermodynamic basis for the mineral facies concept: *Am. Jour. Sci.*, v. 253, pp. 65-103.
- TILLEY, C. E., 1924, The facies classification of metamorphic rocks: *Geol. Mag.*, v. 61, pp. 167-171.
- , 1935, The role of kyanite in the "hornfels zone" of the Carn Chuinneag granite, Ross-shire: *Miner. Mag.*, v. 24, pp. 92-97.
- TURNER, F. J., 1948, Mineralogical and structural evolution of the metamorphic rocks: *Geol. Soc. Am., Mem.*, no. 30, 342 pp.
- , and VERHOOGEN, J., 1951, *Igneous and metamorphic petrology*: New York, McGraw-Hill Book Co., Inc., 602 pp.; 1960, 2nd. ed., 694 pp.
- WATSON, J., 1948, Late sillimanite in the migmatites of Kildonan, Sutherland: *Geol. Mag.*, v. 85, pp. 149-162.
- WEGMANN, C. E., 1929, Beispiele tektonischer Analysen des Grundgebirges in Finnland: *Comm. géol. Finlande, Bull.*, v. 87, pp. 98-127.
- WEISS, L. E., and MCINTYRE, D. B., 1957, Structural geometry of Dalradian rocks at Loch Leven, Scottish Highlands: *Jour. Geol.*, v. 65, pp. 575-602.
- WHITE, W. S., 1949, Cleavage in east-central Vermont: *Am. Geophys. Union, Trans.*, v. 30, pp. 587-594.
- , 1959, A revision of the Lower Paleozoic stratigraphy in eastern Vermont: a discussion: *Jour. Geol.*, v. 67, pp. 577-581.
- , and BILLINGS, M. P., 1951, Geology of the Woodsville quadrangle, Vermont-New Hampshire: *Geol. Soc. Am., Bull.*, v. 62, pp. 647-696.
- , and JAHNS, R. H., 1950, Structure of central and east-central Vermont: *Jour. Geol.*, v. 58, pp. 179-220.
- WILLIAMS, G. J., 1934, A granite-schist contact in Stewart Island, New Zealand: *Geol. Soc. London, Quart. Jour.*, v. 90, pp. 322-353.
- WILLIAMS, H., TURNER, F. J., and GILBERT, C. M., 1955, *Petrography*: San Francisco, W. H. Freeman & Co., 406 pp.
- WILSON, G., 1947, The relationship of slaty cleavage and kindred structures to tectonics: *Geologists' Assoc. London, Proc.*, v. 57, pp. 263-302.
- , 1953, Mullion and rodding structures in the Moine series of Scotland: *Geologists' Assoc. London, Proc.*, v. 64, pp. 118-145.
- WINCHELL, A. N., and WINCHELL, H., 1951, *Elements of optical mineralogy, Part II, Description of minerals*: New York, John Wiley & Sons, Inc., 4th ed., 551 pp.
- WISEMAN, J. H. D., 1934, The Central and South-west Highland epidiorites: *Geol. Soc. London, Quart. Jour.*, v. 90, pp. 354-416.
- WOODLAND, B. G., 1962, Lamprophyric dikes of the Burke area, Vermont: *Am. Mineral.*, v. 47, pp. 1094-1110.
- , 1963, A petrographic study of thermally metamorphosed pelitic rocks in the Burke area, northeastern Vermont: *Am. Jour. Sci.*, v. 261, pp. 354-375.
- YODER, H. S., and EUGSTER, H. P., 1955, Synthetic and natural muscovites: *Geochim. et Cosmochim. Acta*, v. 8, pp. 225-280.

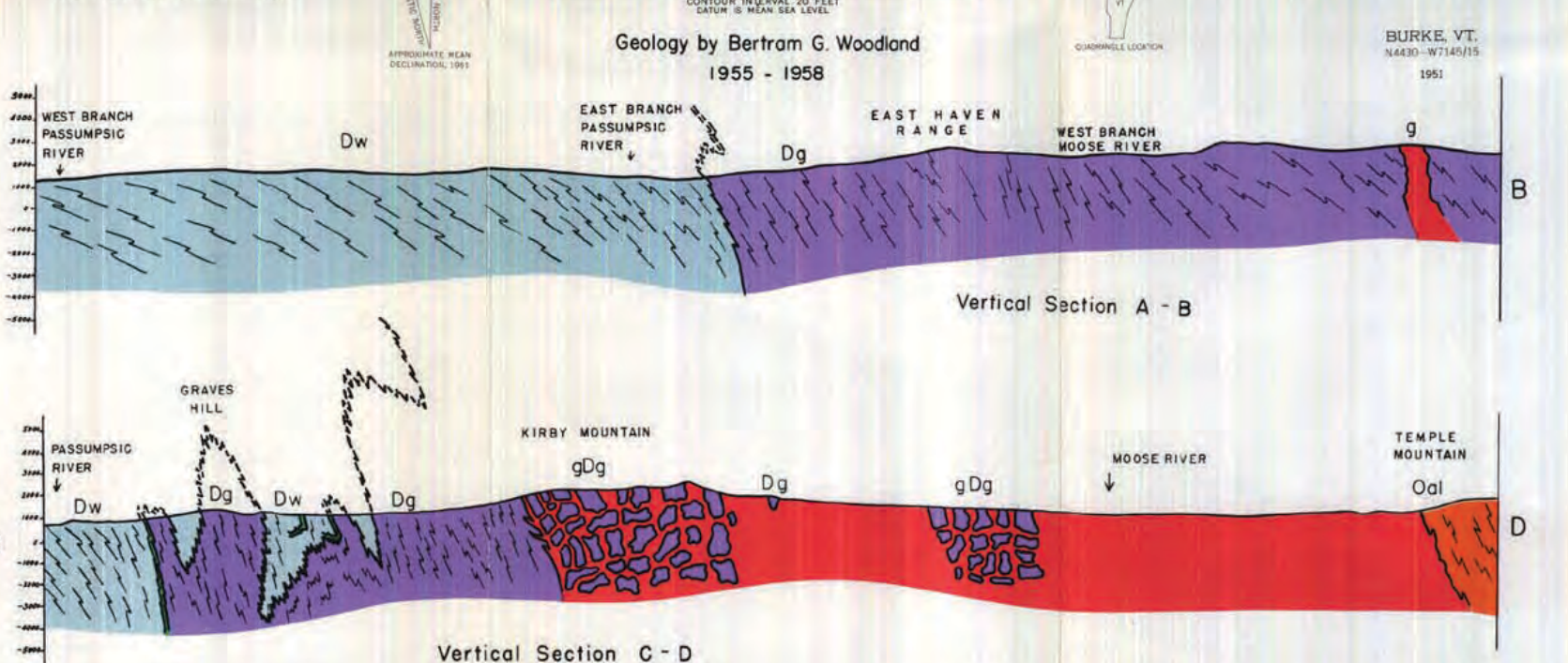


- The shaded areas indicate extensive glacial and post-glacial deposits (cover) where outcrops are often sparse.
- QUATERNARY**
- Q** Area completely obscured by glacial and post-glacial deposits, underlying bedrock can not be inferred.
- SILURIAN(?) and/or DEVONIAN**
- g** Granite, quartz monzonite, granodiorite, diorite, monzonite
 - Dw** Waits River formation with many granitic sills and dikes
- HORNFELS-GRANITE COMPLEXES: These have large outcrops composed entirely of granitic rock, of hornfels, or of both intimately mixed.**
- gDg** Gile Mountain hornfels and granite.
 - gOal** Albee hornfels and granite.
- GILE MOUNTAIN FORMATION**
- Dg** phyllite, quartzose phyllite, slate, quartzite, micaschist, garnet-mica schist, staurolite-garnet-mica schist, andalusite and sillimanite hornfels, amphibole quartzite
 - Dgv** amphibolite
- WAITS RIVER FORMATION**
- Dw** recrystallized quartzose limestone (often tremolitic), phyll. limestone, calcareous phyllite, phyllite, mica schist, quartzose phyllite, quartzite, amphibole quartzite, minor calc-silicate hornfels, and staurolite-garnet-mica schist
 - Dwv** amphibolite
 - Dwr** meta-soda rhyolite, keratophyre
- ORDOVICIAN(?)**
- Oal** ALBEE FORMATION
Oal: staurolite-garnet-mica schist, quartzose schist
Oalv: amphibolite
- Approximate contact
Inferred contact
Inferred fault
Shear zone
- Metamorphic isograd lettered on higher-grade side
G: garnet zone (biotite zone on lower-grade side)
St: staurolite zone
Si: sillimanite zone
A: andalusite occurrence
K: kyanite occurrence
- Strike and dip of bedding and bedding schistosity
Strike and dip of vertical bedding and bedding schistosity
Horizontal bedding and bedding schistosity
Strike and dip of main schistosity
Direction of glacial striae
Disused quarry

Mapped, edited, and published by the Geological Survey
Control by USGS and USC&GS
Topography from aerial photographs by multiple methods
Aerial photographs taken 1942 and 1947. Field check 1949 and 1951.
Photonic projection. 1927 North American datum
10,000-foot grid based on Vermont coordinate system

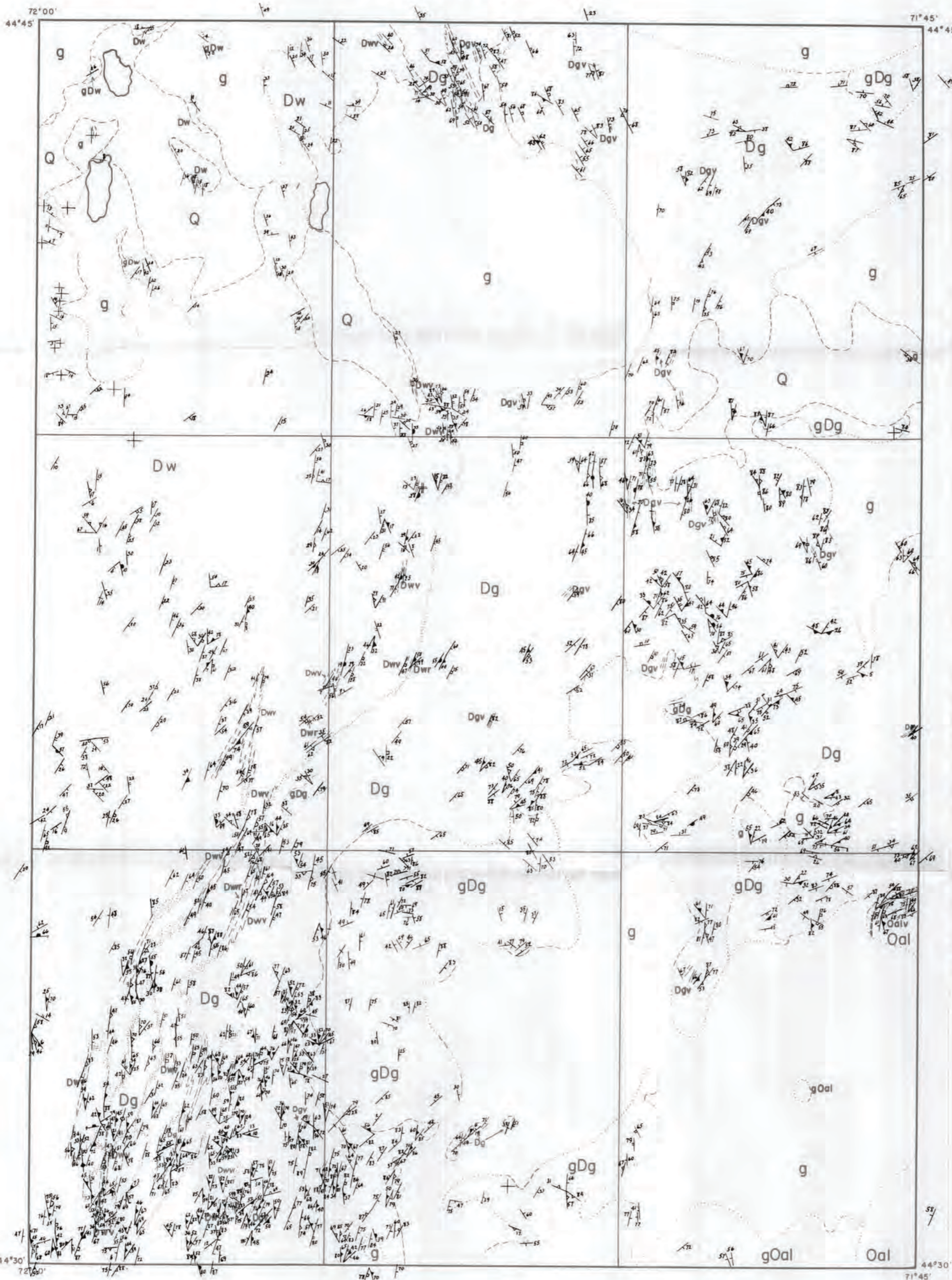
Geology by Bertram G. Woodland
1955 - 1958

BURKE, VT.
N4430-W7145/15
1951







PLANAR STRUCTURES
MAP OF BURKE QUADRANGLE


PLATE 2







LEGEND


- 




Horizontal Vertical
Attitude of bedding (S₀) or bedding schistosity (S₁)
- 




Horizontal Vertical
Attitude of main schistosity (S₁), not parallel to bedding
- 

Strike and dip of axial plane schistosity (S₂) of minor folds
- 


Attitude of false cleavage (S₂) including slip and fracture cleavage
- 


Attitude of axial plane false cleavage (S₂) of minor folds
- 

Strike and dip of false cleavage (S₂)
- 


Attitude of undifferentiated main schistosity
- 


Attitude of undifferentiated false cleavage

- Q: Area completely obscured by glacial and post-glacial deposits
- g: Granitic rocks
- gDw: Waits River formation with many granitic sills and dikes
- gDg: Gile Mountain hornfels and granite
- gOal: Albee hornfels and granite
- Dg: Gile Mountain formation
- Dgv: Amphibolite
- Dw: Waits River formation
- Dwv: Amphibolite
- Dwv: Meta-soda rhyolite, keratophyre
- Oal: Albee formation
- Oalv: Amphibolite

- 

Approximate contact
- 

Inferred contact
- 

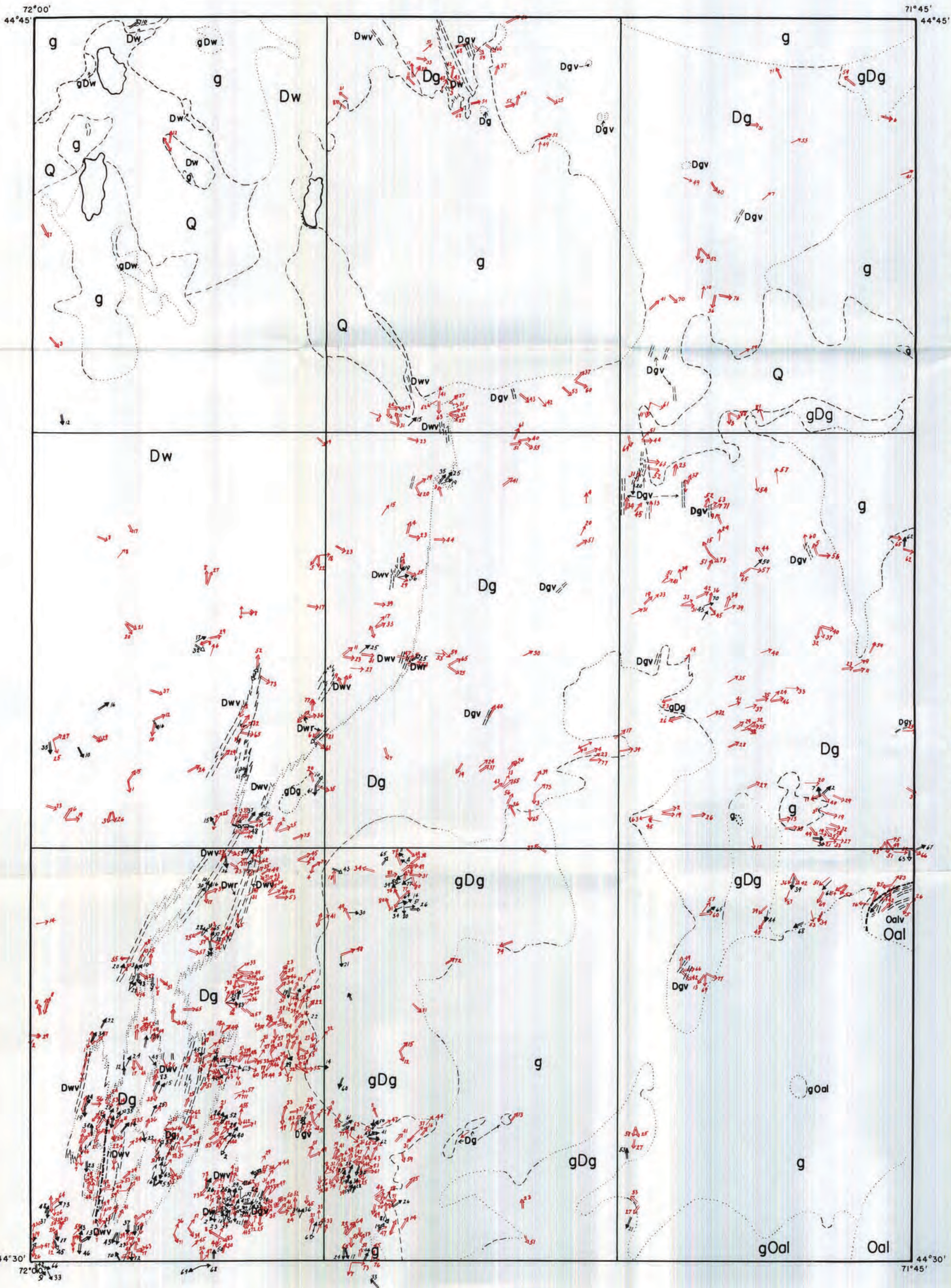
Inferred fault
- 

Shear zone



LINEAR FEATURES
MAP OF BURKE QUADRANGLE

PLATE 3



LEGEND

- Trend and plunge of (early) B₁ fold axis
- Trend and plunge of (late) B₂ fold axis
- Vertical B₂ fold axis
- Trend and plunge of undifferentiated fold axis
- Trend of horizontal undifferentiated fold axis
- Trend and plunge of constructed B₂ axial direction
- Trend of horizontal constructed B₂ axial direction
- Trend and plunge of b₁ lineation (intersection of bedding, S₀, and schistosity, S₁)
- Trend of horizontal b₁ lineation
- Trend and plunge of b₂ lineation (intersection of schistosity, S₁, and false cleavage, S₂), including crinkling on schistosity, S₁, and quartz rodding
- Trend of horizontal b₂ lineation
- Trend and plunge of undifferentiated lineation
- Trend of horizontal undifferentiated lineation
- Trend and plunge of mineral alignment on schistosity, S₁

- Q: Area completely obscured by glacial and post-glacial deposits
- g: Granitic rocks
- gDw: Waits River formation with many granitic sills and dikes
- gDg: Gile Mountain hornfels and granite
- gOal: Albee hornfels and granite
- Dg: Gile Mountain formation
- Dgv: Amphibolite
- Dw: Waits River formation
- Dwv: Amphibolite
- Dwr: Meta-soda rhyolite, keratophyre
- Oal: Albee formation
- Oalv: Amphibolite

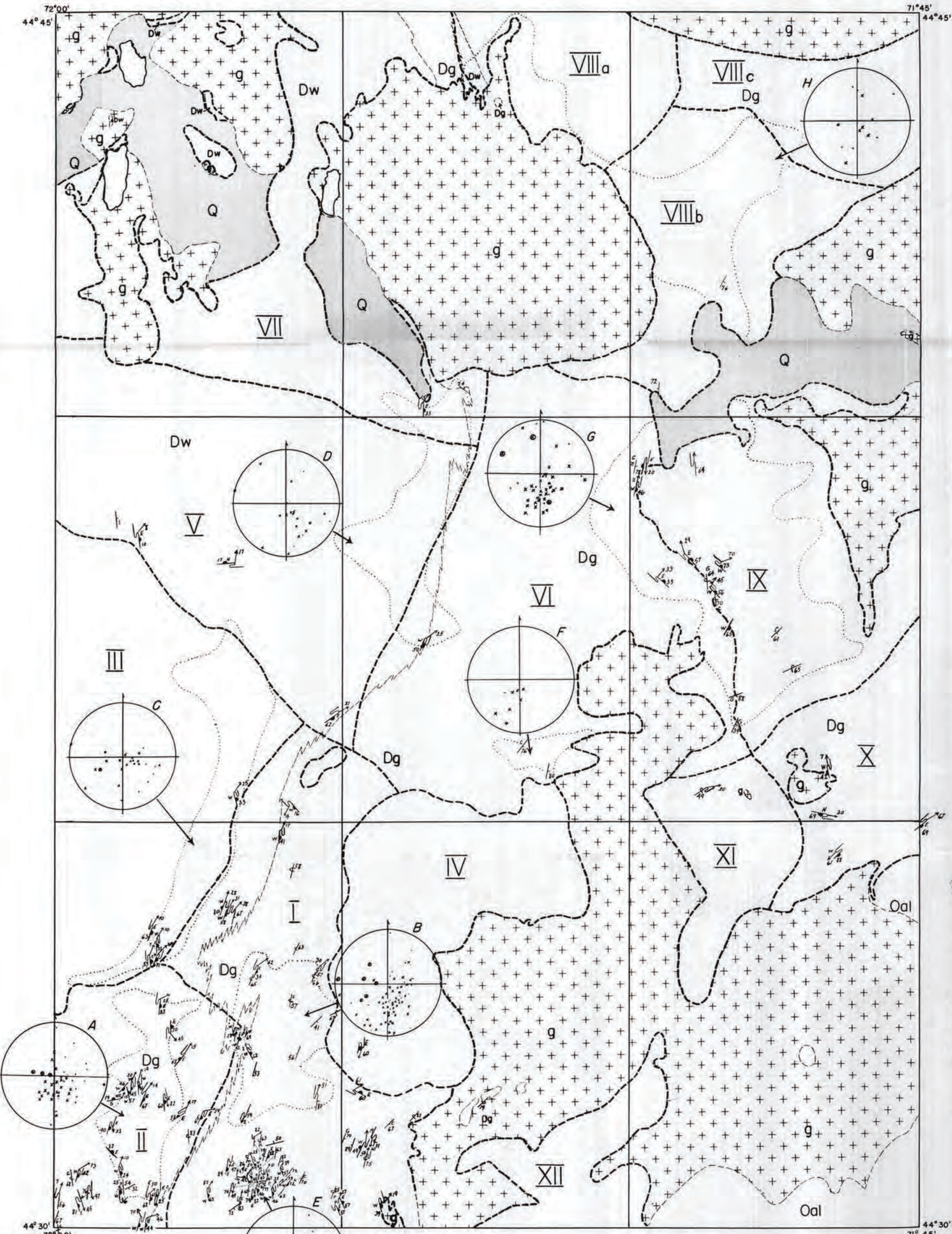
- Approximate contact
- Inferred contact
- Inferred fault
- Shear zone



VERMONT GEOLOGICAL SURVEY
Charles G. Doll, State Geologist
(Bulletin No. 28)

WAY-UP OF BEDDING (S_0) AND
SHEAR SENSE ON SCHISTOSITY (S_1) OF
BURKE QUADRANGLE

PLATE 4



LEGEND

$\frac{1}{50}$ Based on bedding (S_0)-schistosity (S_1) relationship
 $\frac{1}{50}$ Based on current bedding
 $\frac{1}{50}$ Based on graded bedding
 Strike and dip of bedding (S_0) right way up

$\frac{1}{50}$ Based on bedding (S_0)-schistosity (S_1) relationship
 $\frac{1}{50}$ Based on current bedding
 $\frac{1}{50}$ Based on graded bedding
 Strike and dip of bedding (S_0) upside down

$\frac{1}{50}$ Strike of vertical bedding (S_0) facing east

$\frac{20}{50}$ Bedding right way up
 $\frac{20}{50}$ Bedding upside down

Plan of minor B_1 fold showing strike and dip of axial plane (S_1) and trend and plunge of fold axis: general bedding dips in similar direction as S_1 .

$\frac{20}{50}$ Shear sense up on west side of S_1
 (E-N-S Shear sense up on east, north and south side of S_1 respectively)
 Plan of minor B_2 fold showing strike and dip of axial plane (S_2) and trend and plunge of fold axis: general schistosity (S_1) dips in similar direction as S_2 .

$\frac{12}{50}$ Shear sense up on east side of S_1
 $\frac{12}{50}$ Shear sense up on west side of S_1
 Strike and dip of axial plane cleavage (S_2), general schistosity (S_1) dips in similar direction as S_2 . Based on vertical section of fold.

Stereogram of shear sense on schistosity (S_1) (based on $S_1 - S_2$ relationship) plotted on lower hemisphere of Wulff stereographic net.
 \otimes Shear down shear up
 \circ Shear down shear up
 \times Shear up shear down
 \circ Shear up shear down
 Shear sense on east side of S_1
 Boundary of area for which stereogram of shear sense on S_1 is shown

Boundary of sub-area with number referred to in text

Q Area completely obscured by glacial and post-glacial deposits

+g+ Granitic rocks

Dg: Gile Mountain formation

Dw: Waits River formation

Oal: Albee formation

Approximate and inferred contact

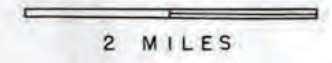


PLATE 5. CONTOURED STEREOGRAMS OF THE BURKE QUADRANGLE

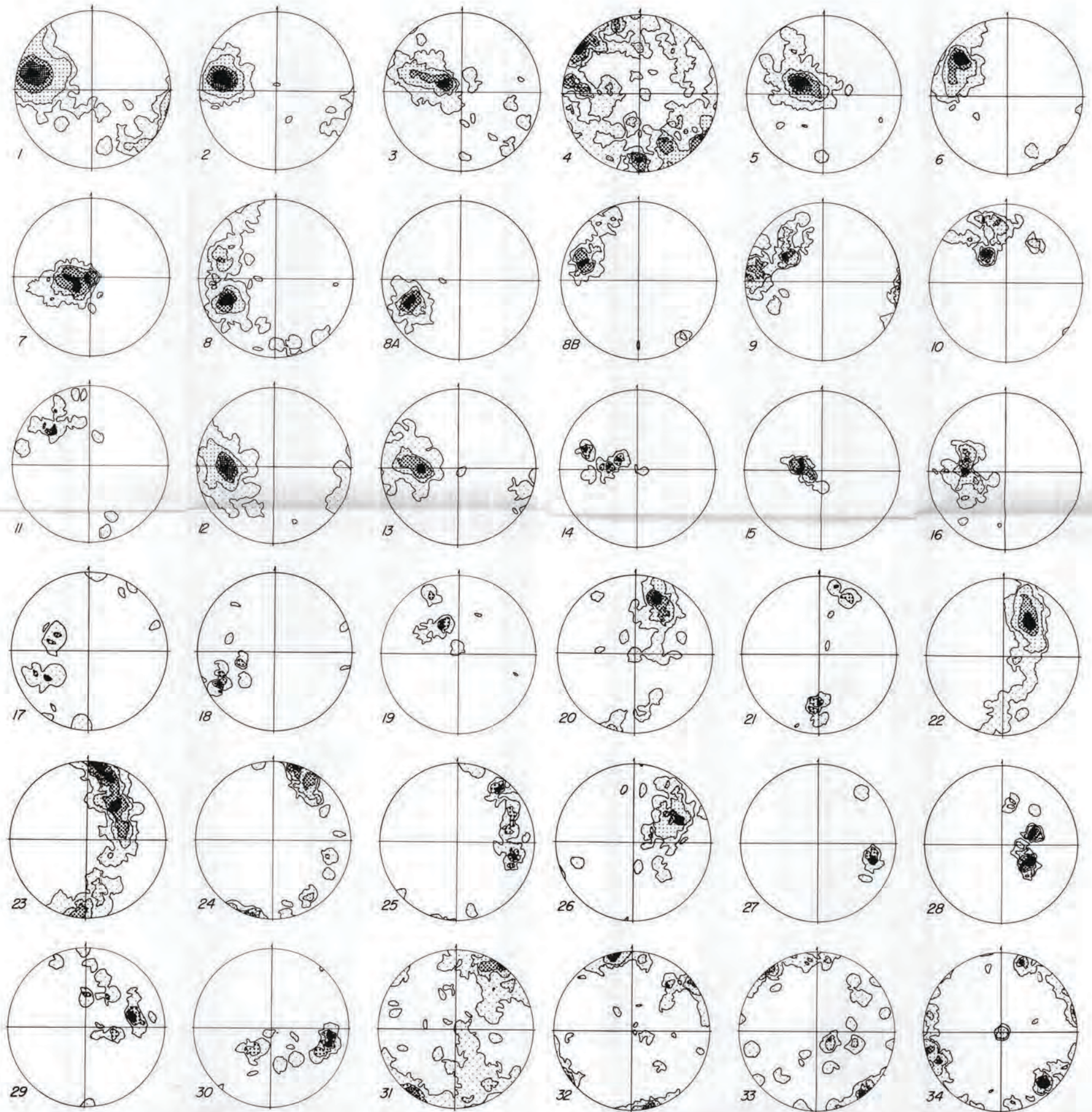


Plate 5. Contoured stereograms of the Burke quadrangle measurements of schistosity, cleavage, later fold axes and lineations, and joints.

Figures 1 to 11. Plot on the lower hemisphere of a Schmidt stereonet of poles to schistosity (S_1) by subarea (see Plate 4 for boundaries); contour percentages expressed per 1% area.

1. 686 measurements, subarea I; contoured at $\frac{1}{2}$ -1-5-10-18%
2. 256 measurements, subarea II; contoured at 1-4-8-12-16%
3. 143 measurements, subarea III; contoured at $\frac{1}{2}$ -5-10-15%
4. 244 measurements, subarea IV; contoured at 1-2-3-4%
5. 165 measurements, subarea V; contoured at 1-5-10-15-20%
6. 142 measurements, subarea VI; contoured at 2-5-9-12%
7. 141 measurements, subarea VII; contoured at 2-6-10-15%
8. 113 measurements, subarea VIII; contoured at 2-4-7-11%
- 8A. 48 measurements, subarea VIIIA; contoured at 4-10-16-24%
- 8B. 50 measurements, subarea VIIIB; contoured at 4-8-12-16%
9. 70 measurements, subarea IX; contoured at 3-6-9-12%
10. 58 measurements, subarea X; contoured at 3-7-10-14%
11. 37 measurements, subarea XI; contoured at 6-11-14%

Figures 12 to 19. Similar plot of poles to cleavage (S_2) by subarea; contour percentages expressed per 1% area, unless otherwise noted.

12. 253 measurements, subarea I; contoured at 1-4-8-12-16%
13. 117 measurements, subarea II, contoured at 2-6-10-15%
14. 32 measurements, subarea III; contoured at 6-13-19%
15. 25 measurements, subarea V; contoured at 8-16-24-32%
16. 56 measurements, subarea VI; contoured at 4-7-10-13%
17. 22 measurements, subarea VIII; contoured at 5-10-15% per 2% area
18. 31 measurements, subarea IX; contoured at 6-9-13%
19. 33 measurements, subarea X; contoured at 6-13-19%

Figures 20 to 21. Similar plot of B_2 fold axes by subarea; contour percentages expressed per 1% area.

20. 94 measurements, subarea I; contoured at 2-5-8-11%
21. 37 measurements, subarea II; contoured at 6-11-17%

Figures 22 to 30. Similar plot of B_2 lineations by subarea; contour percentages expressed per 1% area.

22. 165 measurements, subarea I; contoured at 1-3-9-15% (undifferentiated lineations not included)
23. 162 measurements, subarea II; contoured at 1-3-5-8% (undifferentiated lineations included)
24. 57 measurements, subarea III; contoured at 3-7-10-14% (B_2 folds and undifferentiated lineations included)
25. 51 measurements, subarea V; contoured at 4-8-12% (B_2 folds and undifferentiated lineations included)
26. 72 measurements, subarea VI; contoured at 3-6-10-13% (B_2 folds and undifferentiated lineations included)
27. 31 measurements, subarea VII; contoured at 6-13-19% (B_2 folds and undifferentiated lineations included)
28. 32 measurements, subarea VIII; contoured at 6-9-12-15% (B_2 folds and undifferentiated lineations included)
29. 49 measurements, subarea IX; contoured at 4-8-12% (B_2 folds and undifferentiated lineations included)
30. 47 measurements, subarea X; contoured at 4-8-13% (B_2 folds and undifferentiated lineations included)

Figures 31 to 34. Similar plot of poles to joint planes; contour percentages expressed per 1% area.

31. 201 measurements in the metamorphic rocks of subarea II; contoured at 1-4-7-10%
32. 144 measurements in the hornfels of subarea IV; contoured at 2-4-6-8%
33. 83 measurements in the granite of subarea IV; contoured at 2-5-7-10%
34. 140 measurements in the Newark granite; contoured at 2-4-6-7%