GEOLOGY OF THE BRYANT POND QUADRANGLE MAINE

By

Charles V. Guidotti

QUADRANGLE MAPPING SERIES
NO. 3

DEPARTMENT OF ECONOMIC DEVELOPMENT
AUGUSTA, MAINE

FEBRUARY, 1965
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ABSTRACT

The Bryant Pond quadrangle, located in northwestern Maine, is bounded by north latitudes 44°30' to 44°15' and west longitudes 70°30' to 70°45'. It is a hilly to mountainous area and has been subjected to continental glaciation during Pleistocene time. Steep hillsides and cliffs are abundant.

The rocks have been highly metamorphosed and intruded by igneous rocks. The metamorphosed sedimentary rocks have been divided into two lithologically dissimilar sequences that are separated by a large normal fault.

The southern sequence consists of four conformable units. The Patch Mountain formation, the lowest unit, is made up of at least 2000 feet of calc-silicate granulite and lesser amounts of interbedded quartz-plagioclase-biotite schist. The Noyes Mountain formation, next above it and 800 feet thick, is made up of migmatitic gneiss in which the dark fraction consists of quartz-feldspar-two mica-sillimanite schist and quartz-feldspar-biotite granulite. The light fraction consists of pegmatitic quartz and plagioclase in bands and lenses ranging from 1 to 4 inches thick. It is overlain by the Berry Ledge formation, which, only 150 to 250 feet thick, consists of quartz-calcite-plagioclase-diopside granulite with interbedded impure marble and minor amounts of biotite schist. The preceding three units are probably Silurian in age. Conformably above them is the Moody Brook formation, which consists of migmatitic gneiss in which the dark fraction consists of quartz-feldspar-biotite granulite and quartz-feldspar-two mica-sillimanite schist. The light fraction is composed of pegmatitic quartz and plagioclase. It is 600 feet thick and is probably Silurian in age.

The northern sequence has been divided into six conformable units. The lowest, the Thompson Mountain formation, is very poorly exposed in the Bryant Pond area and consists of migmatitic rusty-weathering gneisses in which the dark fraction is made up of pyrrhotite-bearing, coarse quartz-feldspar-two mica-sillimanite schist with minor quartz-feldspar-biotite granulite. The light-colored fraction consists mainly of pegmatitic quartz and feldspar and a granitic fraction which is similar to the Speckled Mountain quartz monzonite. This unit is overlain by
the Shagg Pond formation, which is 3000 feet thick and is composed of gray-weathering migmatitic gneiss. The dark fraction consists largely of coarse quartz-feldspar-two mica-sillimanite schist with only minor quartz-feldspar-biotite granulite. The light-colored fraction consists of bands and lenses of pegmatitic quartz and some Speckled Mountain quartz monzonite type. Muscovite megacrysts produce a spangled appearance on fresh surfaces. The next unit is the Billings Hill formation, which is 2000 feet thick and is composed of rusty-weathering migmatitic gneiss. It is similar to the Shagg Pond formation except that it contains significant amounts of pyrrhotite and greater amounts of quartz-feldspar-biotite granulite which in some cases is present in massive beds.

The Thompson Mountain, Shagg Pond, and Billings Hill formations are considered to be Siluro-Devonian in age and to underlie conformably the Devonian Littleton formation. The dividing surface between the three Siluro-Devonian units and the Littleton formation has been set arbitrarily in order to avoid making the Littleton formation into an exceedingly thick unit with no evidence to support a Devonian age for the lower numbers. The Siluro-Devonian units are considered to be older than any part of the lower Littleton formation described by Billings et al. (1946).

The Concord Pond member is the lowest member of the Devonian Littleton formation in the Bryant Pond area. It is 2500 feet thick and consists of gray-weathering migmatitic gneiss and some horizons of quartz-feldspar-biotite granulite. It is very similar to the Billings Hill formation but does not contain significant amounts of pyrrhotite. It is overlain by the Wilbur-Mountain member, which is 300 to 500 feet thick and is composed of rusty-weathering migmatitic gneisses. These two units are members of the lower part of the Littleton formation as described by Billings et al. (1946). The highest unit exposed in the Bryant Pond area is the Howard Pond member. It is about 600 feet thick and consists of an upper and lower part made up of interbedded, pyrrhotite-rich micaceous schist and quartzose calc-silicate beds. A series of about 150 feet of interbedded quartzose calc-silicate rocks and quartz-feldspar biotite granulite is present between the upper and lower parts of the member. The Howard Pond member is correlated with the Boott member of the Lower Devonian Littleton formation described by Billings et al. (1946) in the Mount Washington area.

Most of the plutonic rocks in the Bryant Pond area belong to the New Hampshire magma series. The Songo granodiorite, the most important body, forms a large dome-shaped pluton in the western part of the area and generally is concordant but locally discordant. Binary granite, exposed in a small area along the southern boundary of the quadrangle, is part of the large Sebago batholith that underlies much of south-central Maine. Pegmatites and Speckled Mountain quartz monzonite which occur as numerous small discordant and discordant bodies are probably late magmatic differentiates of the Songo granodiorite. In some areas these small bodies are so abundant that the metamorphosed sediments are present only as remnants and inclusions.

Post-metamorphic dikes, mainly diabase, are present throughout the area. They are considered to be co-magmatic with the White Mountain magma series of Triassic (?) age. It is believed that they formed in association with the post-metamorphic normal faults and joints.

The structural features within the metamorphosed sediments can be divided into two groups that are separated by a large normal fault (Moll Ockett fault). North of the fault the folds are generally large, relatively open, and only slightly overturned toward the west. In the northeast part of the area the folds trend in a northeasterly direction, which coincides with the normal structural trends in northern New England. But near the Songo granodiorite the fold axes swing around to the northwest and become parallel with the border of this large pluton.

South of the Moll Ockett fault the folds trend about N. 30° W. and are strongly overturned toward the southwest. The north-plunging Norway anticlinorium is the major structural feature in this area.

The Songo granodiorite, structurally the most important pluton, forms a large dome that is more or less discordant along its northern boundary but is cut off on the south by the Moll Ockett fault. It is believed to be syntectonic.

Folding, believed to have taken place during the Acadian Orogeny of Middle to Late Devonian time, was caused by north-
west-southeast-oriented compressive forces. However, the syn-
tectonic intrusion of the Songo granodiorite and Sebago granite
probably set up a local stress field and caused the anomalous
northwest-trending folds in much of the Bryant Pond area.

Faulting is evidenced by one large normal fault and several
smaller ones. The faults are paralleled by the major joint set in
the area and the majority of the post-metamorphic dikes. These
structural features formed during Late Triassic time and prob-
rably resulted from a decrease in the horizontal compressive
stresses, possibly associated with regional doming.

The strata in the Bryant Pond quadrangle have been meta-
morphosed to the sillimanite and sillimanite plus potassium feld-
spar grades of metamorphism. One isograd, based upon the
mineralogy of the pelitic schists, has been drawn, separating
these two grades. In the rocks of sillimanite grade the pair
sillimanite-microcline is prevented by the coexistence of the pair
muscovite-albite. Attainment of the sillimanite plus potassium
feldspar isograd is indicated by a change in the compatibilities
which allows the coexistence of microcline and sillimanite. The
migmatitic gneisses in which the dark fraction is commonly
pelitic schist are probably the result of partial melting of the
quartz-feldspar fraction in original shales.

Originally dolomitic sediments have been metamorphosed to
diopside granulite, amphibolite, and various other types of calcsilicate rocks. Some impure marbles are present also. No iso-
grads can be drawn on the basis of the mineralogy in these calcsilicate rocks.

The conditions of metamorphism have been considered by
comparing the mineral assemblages present in the rocks of the
Bryant Pond area with experimental data on pertinent minerals
and assemblages. It is concluded that the temperatures were
on the order of 650 to 750 degrees centigrade and pressures on
the order of 6 to 14 kilobars.

INTRODUCTION

Location

The Bryant Pond quadrangle, located in western Maine south-
west of the city of Rumford, has an area of slightly more than
two hundred square miles and is bounded by north latitudes 44°
30’ and 44°15’ and west longitudes 70°30’ and 70°45’ (Fig. 1).
It includes most of the towns of Hanover, Woodstock, Milton,
Greenwood, large portions of Bethel, Rumford, Norway, Peru,
Paris, and small segments of Sumner, Waterford and Albany.
The area is readily accessible via Routes 5 and 26 from the south
and north, Route 2 from the east and west, and by the Grand
Trunk Railway.

Physiography

The northern and western sections of the quadrangle are hilly
to mountainous with Spruce Mountain (El. 2420 feet) the highest
point. These parts of the quadrangle are characterized by a
rather knobby steep topography, particularly on southern slopes.
The northern slopes of the hills are more gentle and deeply
covered with glacial debris.

The southern part of the quadrangle has more gentle topog-
raphy with a strong northwesterly grain compared to the in-
definite topographic lineation in the northern and western sec-
tion.

The Androscoggin River and Little Androscoggin River are
the main streams. The southernmost parts of the quadrangle
drain into the Little Androscoggin via the northern part of the
Norway quadrangle. By far the major portion of the quadrangle
drains into the Little Androscoggin River.

Purpose of Study

The current study was undertaken by the writer to extend into
the Bryant Pond quadrangle the stratigraphy as determined by
Billings et al. (1946) and Billings (1956) in the Mount Washing-
ton area, by Fisher (1952) in the Bethel quadrangle, and by
Milton (1961) in the Old Speck Mountain quadrangle. It was
hoped that this work would narrow the gap separating the stratig-
raphy of the Mount Washington area from the fossiliferous
strata in the Waterville area in south-central Maine (Perkins and Smith, 1925). The units of the Waterville area had been traced by Fisher (1941) to the Lewiston area and by Hanley (1939) to the Poland area, which lies just southeast of the Bryant Pond quadrangle. In general this attempt has been successful, but the exact stratigraphic relations have remained elusive due to faulting and complex stratigraphic problems in the area between Waterville and Bryant Pond.

A second major purpose was to determine the structural relations within the metamorphosed strata and how these are related to the granitic rocks of the quadrangle.

Finally the writer wished to work out the igneous and metamorphic petrology of the area. The whole area is of high-grade metamorphism with only one isograd present (Guidotti 1963); hence the mineralogy of the pelitic schists and biotite schists is monotonous. The calc-silicate rocks are considerably more interesting but lack extensive mineralogical variability.

Methods of Study

The writer worked in the field a total of forty-five weeks during 1959, 1960, and 1961. Field mapping was conducted by the author for a period of forty-five weeks using a two-times enlargement of a United States Geological Survey topographic map (original on a scale of 1 inch = 1 mile). Traverses were made along streams, ridges, and hilltops, as well as on set compass bearings over nondescript slopes. Outcrops were found to be abundant on ridges, hilltops, south-facing slopes, and streams with steep gradients. Valley bottoms and north-facing slopes have very poor exposures of bedrock.

The only available United States Geological Survey topographic map of the Bryant Pond quadrangle, published in 1911 and very unreliable, made station locations and field orientation difficult. Aerial photographs helped to alleviate this difficulty in some instances.

A total of 550 specimens were collected and approximately 300 thin sections were studied and the modes estimated. The composition of plagioclase was determined in thin sections by the twin relations and relief. The presence of potassium feldspar was ascertained by thin sections and optics of crushed fragments. The optics of crushed fragments of feldspar and other minerals were determined by means of a Leitz Micro-refractometer with sodium light for illumination.

Previous Work

The earliest published geological works that mention the Bryant Pond area include those by Jackson (1838, 1839), Hitchcock (1861, 1862, 1878), and Keith (1933), who compiled a geological map of Maine showing the Bryant Pond area to be metamorphosed Precambrian strata and intrusive bodies.

Specialized studies of some of the pegmatites and mineralogy in the area have been made by Cameron et al. (1954), Bastin (1911), Cummings (1955), Glass and Fahey (1933), Landes (1925), Trefethen (1945). Dale (1907, 1923) has described the
"granite" quarry within the granodiorite body in the quadrangle, while Leavitt and Perkins (1934) and Crosby (1922) have considered some aspects of the surficial geology. A gravity survey of New Hampshire and adjacent areas (Joyner, 1958, 1963) includes two stations within the Bryant Pond quadrangle.


Other publications relating to the present work are the contributions by Barker (1961) on the Hallowell granite near Augusta, Patterson (1942) on the geology of the Lake Cobbossecontee area, Perkins (1924) on the fossil locality near Waterville, Perkins and Smith (1925) on the geology to the east of the Waterville area, Katz (1918) in southwestern Maine, Pavlides et al. (1961) in Aroostook County, and Billings (1937) in the Littleton-Moosilauke area in New Hampshire.

At present, several geological studies are being made in adjacent areas. Dr. Kost Paniwskyj is mapping the Dixfield quadrangle. Mr. Jeffrey Warner is mapping the Buckfield quadrangle and Dr. Philip Osberg of the University of Maine is continuing and extending his mapping in the Waterville-Augusta area.

Acknowledgements

During the summer of 1959 funds were provided by the Reginald and Louise Daly Fund of Harvard University. In the summers of 1960 and 1961 the writer was employed by the Maine Geological Survey under the direction of the State Geologist, Mr. Robert G. Doyle. Funds for thin sections were generously provided by the Department of Geological Sciences at Harvard University. Jeffrey Warner served as field assistant in the summer of 1960 and Edward S. Gaffney during the summer of 1961.

Professor Marland P. Billings visited the writer for a total of five days during the summer of 1961 and Professor James B. Thompson, Jr. for two days in the summer of 1961. The writer wishes in particular to express his gratitude to Professor Billings who has given freely of his time and aid in many discussions with the writer throughout the course of this work. Moreover, he has kindly allowed the writer to include his unpublished information on the Gorham quadrangle.

The writer would also like to acknowledge the cooperation of Dr. Daniel J. Milton, Dr. Kost A. Pankiwskyj, and Mr. Jeffrey Warner, all of whom, having mapped in areas adjacent to the Bryant Pond quadrangle, have provided much aid and stimulation to the writer. Professor Philip H. Osberg of the University of Maine has given much aid to the writer by freely discussing his work which is in progress on the stratigraphy and geology of the Waterville-Augusta area. Moreover, he has allowed the writer to discuss briefly the sequence of units which he has worked out.

Dr. Daniel S. Barker has allowed the writer to quote from his work on the area near Augusta, Maine. W. T. Forsythe has granted similar permission with regard to the Rumford quadrangle, John B. Hanley—the Poland quadrangle, I. S. Fisher—the Bethel quadrangle, D. J. Milton—the Old Speck Mountain quadrangle, K. A. Pankiwskyj—the Dixfield quadrangle, J. Warner—the Buckfield quadrangle. The writer appreciates the kindness of these people.

Professors Marland P. Billings of Harvard University, Cordell Durrell of the University of California at Davis, Arthur Hussey of Bowdoin College, Philip H. Osberg of the University of Maine, and Mr. Robert G. Doyle of the Maine Geological Survey read and criticized early drafts of this report. The time and effort of these people are sincerely appreciated.

Finally, the writer is very grateful to his wife who has typed parts of this report and assisted in taking several photographs.
STRATIGRAPHY AND LITHOLOGY

General Statement

Two-thirds of the Bryant Pond quadrangle is underlain by intensely deformed, metamorphosed, and intruded sedimentary strata that range in age from Ordovician (?) to Devonian (?) (Table 1). In a strictly descriptive sense, most of the metamorphosed sediments are migmatites. The writer follows the definition of migmatite used by Turner and Verhoogen (1960, p. 370) who state,

"It is generally agreed that the most typical rocks of this class are those in which a granitic component (granite, aplite, pegmatite, granodiorite, or the like) and a metamorphic host rock are intimately admixed on a scale sufficiently coarse for the mixed condition of the rock to be megascopically recognizable."

Small to large intrusive bodies of Middle to Upper Devonian age make up approximately one-third of the area. Pegmatites are exceedingly common and have been used to a considerable extent as sources of feldspar, mica, beryl, and semi-precious gems. A detailed discussion of the ages and correlations of the formations is given in a later section.

The map units have been divided into two distinct groups which may, for convenience, be referred to as a northern sequence and a southern sequence. The southern sequence, which is probably 3650 feet thick, is the older one and consists of four formations of presumed Ordovician to Silurian age as shown in Table 1. The northern sequence includes six units (see Table 1) which are Silurian to Devonian in age, and is approximately 9,000 feet thick.

The intrusive rocks are assigned to two series. The older series, belonging to the Devonian New Hampshire magma series, includes the Songo granodiorite, binary granite, Speckled Mountain quartz monzonite, and probably most of the numerous pegmatite bodies. The second series, which includes basaltic, post-metamorphic dikes, is of very minor importance and is probably of Triassic age.
Within the southern sequence bedding is easily recognized, and is parallel to the foliation. In the northern sequence bedding can usually be found and only rarely is it not parallel with the foliation. To a considerable extent, within the northern sequence, the writer has encountered the problem of defining true stratigraphic units. Fisher (1952, 1962), who mapped similar rocks in the adjacent Bethel quadrangle, encountered this same problem. The writer, like Fisher, has drawn the boundaries of units on the basis of the pyrrhotite content of the unit, which is indicated in the field by rusty or gray-weathered outcrops. The validity of this approach seems to be verified by structural data and the outcrop patterns on the geologic map. However, some uncertainty remains, especially with regard to the Wilbur Mountain member. Another bothersome feature in defining units is the presence of very large amounts of pegmatite throughout the quadrangle.

SOUTHERN SEQUENCE

Patch Mountain Formation

General Statement.—The Patch Mountain formation is composed of thin to massive-bedded green quartz-plagioclase-diopside granulite interbedded with medium-grained, purplish-gray quartz-feldspar-biotite schist. A few very thin impure marble beds are present. The calc-silicate rocks account for one-half to two-thirds of the unit. The Patch Mountain formation underlies a broad belt along the southern border of the Bryant Pond quadrangle and extends northward about five miles into the quadrangle (Plate I). Exposures are good on the higher hills but many are associated with large amounts of pegmatite. In many areas the volume of the pegmatite exceeds the volume of the metamorphosed sedimentary rocks. The name Patch Mountain formation is proposed here for the first time. The type locality is found on the eastern slopes of Patch Mountain, where both of the major lithologic types may be seen. Very good exposures are found on much of the west side of Noyes Mountain, the east side of Patch Mountain, and in the vicinity of Hedgehog Hill.

Lithology.—The two major rock types grade into each other, producing numerous intermediate lithologies. They are medium-grained, green quartz-plagioclase-diopside granulite and quartz-feldspar-biotite schist with the former the dominant type. Beds two to six inches thick are probably most common and bedding surfaces are well defined by alternations of the calc-silicate layers with the quartz-feldspar-biotite schist layers, which are usually one to three inches thick, but much thicker in some cases. In the thinner-bedded calc-silicate rocks the bedding is emphasized by the presence of interbedded thin calcite-rich layers and in this case the weathering surfaces have a characteristic ribbed, punky appearance. Migmatitic quartz-feldspar lenses are abundant throughout this unit.

The quartz-plagioclase-diopside granulite is a medium-grained homogenous, granular rock. Biotite is generally absent, but in a few places, owing to a more argillaceous character, biotite is common and the rock is dull grayish-green to black in color. Such rocks represent a transition to the quartz-feldspar-biotite schist.

The diopside granulite is primarily a diopside-quartz-feldspar rock with equigranular texture. Average estimated modes are given in Table 2 along with the assemblages recorded for the unit. The feldspar is well-twinnecalc plagioclase (ca. An40) but in some specimens microcline is also abundant. Diopside is present as anhedral, commonly poikilitic equant grains. Hornblende is a common minor constituent. In many specimens calcite forms up to 10 per cent of the rock. Impure marbles account for less than 1 per cent of the whole unit. Sphene is nearly ubiquitous and the opaques are mainly pyrrhotite and graphite. Zoisite is locally present in small amounts and appears to be an alteration of the calcic plagioclase. Scapolite is common in some sections and its fresh, clean appearance suggests that it is primary. Chlorite is present in small amounts in some specimens but is a retrograde mineral after hornblende or diopside.

The quartz-feldspar-biotite schist interbedded with the diopside granulite is medium grained and only moderately well foliated. It is dark but commonly with a purplish tinge. The rock consists chiefly of quartz, biotite, and plagioclase (ca. An40) (Table 2, mode 2). Garnets and hornblende are minor constituents while opaques, apatite, zircon (in the biotite), and
tourmaline are accessories. Chlorite (in some cases with sprays of rutile) is a retrograde mineral.

Minor rocks within the Patch Mountain formation are conformable beds of amphibolite and very minor amounts of pelitic schists. Such amphibolite beds attain a maximum thickness of 20 feet and are most common in the upper part of the formation.

The upper boundary of the Patch Mountain formation is gradational into the Noyes Mountain formation; hence the upper limit is somewhat difficult to define. It is drawn where the calc-silicate rocks make up less than 5 per cent of the rock.

Thickness.—The thickness of the Patch Mountain formation is difficult to obtain. It is involved in complex structures, many minor folds are present, and it is impossible to make a reason-

### TABLE 2A

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An content of plagioclase: 67 53 25 64 26

(1) Patch Mountain formation: Calc-silicate granulite
(2) Patch Mountain formation: Quartz-feldspar-biotite schist
(3) Noyes Mountain formation:
(4) Berry Ledge formation:
(5) Moody Brook formation:

### Table 2B

Assemblages in the units of the Southern Sequence
(Numerals indicate number of specimens)

(1) Patch Mountain formation
   Calc-silicate granulite type
   qtz-plag-diop-hnb-micr-calc-garn (1)
   qtz-plag-diop-hnb-micr-calc (1)
   qtz-plag-diop-hnb-sil-calc (4)
   qtz-plag-diop-hnb-calc (2)
   qtz-plag-diop-hnb-micr (3)
   qtz-plag-hnb (1)
   qtz-plag-diop-hnb (2)
   qtz-plag-diop-hnb-micr-calc (6)
   qtz-plag-diop-hnb-micr-calc-Fetrem (1)
   qtz-plag-diop-hnb-micr-sil-calc (1)
   diop-sil-calc (1)
   qtz-plag-diop-hnb-sil-calc (1)
   qtz-plag-diop-hnb-micr-sil-calc (1)

(2) Patch Mountain formation
   Quartz-feldspar-biotite schist
   qtz-plag-bio-garn (2)
   qtz-plag-bio (2)
   qtz-plag-bio-micr (1)
   qtz-plag-bio-hnb (2)
   qtz-plag-bio-hnb-diop (1)

(3) Noyes Mountain formation
   qtz-plag-bio-musc-sil-garn (3)
   qtz-plag-bio-sil-garn (3)
   qtz-plag-bio-garn (5)
   qtz-plag-bio-micr (1)
   qtz-plag-bio-musc-sil-garn (2)
   qtz-plag-bio-musc-sil-garn (1)
   qtz-plag-bio-micr (1)
   qtz-plag-bio-sil-garn (1)
   qtz-plag-bio-musc-sil-garn (1)

(4) Berry Ledge formation
   qtz-plag-diop-hnb-micr-calc (3)
   qtz-plag-diop-hnb-calc (5)
   qtz-plag-diop-micr-calc (1)
   qtz-diop-micr-calc (1)
   qtz-plag-diop-hnb-micr-calc-sep (1)
   qtz-plag-hnb-micr-calc-sep (1)
   plag-diop-hnb-micr (1)
   qtz-plag-diop-hnb (2)
   qtz-plag-diop-hnb-micr-calc-sep (1)
   qtz-plag-diop-micr (1)

(5) Moody Brook formation
   qtz-plag-bio-musc-garn (2)
   qtz-plag-bio-garn (5)
   qtz-plag-bio-musc-sil-garn (3)
   qtz-plag-bio-sil-garn (4)
able estimate of the volume of the unit now occupied by pegmatite. Finally, the bottom of the unit is nowhere exposed in the Bryant Pond quadrangle. Only a minimum value for the thickness can be determined and the best estimate was obtained on the west side of Patch Mountain where the western limb of an overturned anticline is well exposed and dips uniformly eastward. A minimum value of 2000 feet is obtained here.

Origin.—The original pre-metamorphic rocks of this unit are sedimentary. The quartz-plagioclase-diopside granulites are the metamorphosed equivalents of dolomitic, argillaceous sandstones. The calcite marbles, with minor amounts of calc-silicate minerals, represent dolomites with small amounts of quartz and argillaceous material. The protolith of the quartz-feldspar-biotite schist was most likely an argillaceous feldspathic sandstone possibly with a small amount of dolomite, as reflected by the rather calcic nature of the plagioclase (An40). An alternate explanation of the rather calcic nature of the plagioclase is that it reflects metasomatic movement of calcium from the enclosing calc-silicate bands into the biotite schist bands.

The thin amphibolite beds could possibly be volcanic in origin but the writer believes they are sedimentary for several reasons: 1) They do not form widespread stratigraphic marker beds; 2) in some cases they contain considerable amounts of diopside and even grade into diopside granulite; 3) the plagioclase is calcic (An20); and 4) the close association with undoubted calc-silicate rocks indicates that it, too, is likely to be a metamorphosed dolomitic sediment.

Correlation.—The writer suggests a possible correlation with the Mayflower Hill formation of the Waterville area and with the Androscoggin formation in the Lewiston area.

Noyes Mountain formation

General Statement.—The Noyes Mountain formation is composed mainly of migmatitic quartz-feldspar-biotite gneiss and migmatitic quartz-feldspar-two mica-sillimanite gneiss. In a few specimens the migmatitic banding is much reduced and the Noyes Mountain formation can be described as quartz-feldspar-biotite schist, or quartz feldspar-two mica sillimanite schist. Minor thin beds of calc-silicate rocks are also present in this unit.

The Noyes Mountain formation underlies large areas in the southern third of the Bryant Pond quadrangle (Plate I). Exposures are generally good on the steeper and higher hills but are commonly associated with large amounts of pegmatite. The name Noyes Mountain formation is used here for the first time. The type locality is on Noyes Mountain where the migmatitic quartz-feldspar-two mica-sillimanite gneiss is well exposed on the southern and eastern slopes. The migmatitic quartz-feldspar-biotite gneiss can be seen at the road outcrops just east of the north end of Hicks Pond. Very good exposures are also present on the Nubble, Shaw Ledge, the lower parts of the first brook south of Stearns Hill, and on the two hills immediately northwest of Moose Pond.

The Noyes Mountain formation grades into the Patch Mountain formation below and the upper limit is the Berry Ledge formation, which is a thin but distinctive quartz-calcite-plagioclase-diopside granulite.

Lithology.—The Noyes Mountain formation is composed mainly of various types of migmatitic, quartz-feldspar-mica gneisses in which bedding is easily seen due to textural and compositional differences (largely micaceous versus less micaceous beds). The light-colored bands in these migmatitic gneisses consist of pegmatitic quartz-oligoclase (and sometimes perthitic microcline) bands and to a lesser extent quartz monzonite bands. Generally they are 1 to 3 inches thick and occur at irregular intervals of 1 to 6 inches within the rock (Photo 1). They probably account for one-fifth to one-third of most outcrops but occasionally make up a much larger per cent of a given outcrop. Some stringers and anhedral megacrysts of microcline also occur.

In about one-half of these migmatitic gneisses the dark fraction consists of dark-gray to black, quartz-feldspar-biotite schist.
All gradations exist between this biotite schist and a more micaceous schist in which muscovite and sillimanite are abundant. The plagioclase generally ranges in composition from An25 to An30. Small 1 to 2 mm purplish-red garnets are common and in some cases account for as much as 5 per cent of the rock. The rocks are fine to medium grained with a granular to schistose texture. Some of the quartz-feldspar-biotite schist displays a distinct one-half to one inch thick banding in shades of gray that reflects variations in the amount of quartz, which is probably a feature of the original sediments. Such banding also has been observed by the writer in some of the schists mapped by Fisher (1941) as overlying the Androscoggin formation in the Lewiston area.

The quartz-feldspar-biotite schist fraction of the gneiss is more schistose as the amount of muscovite or sillimanite or both increases. In nearly half of the Noyes Mountain formation the dark fraction of the migmatitic gneiss is more micaceous than the quartz-feldspar-biotite schist. The more micaceous rocks include quartz-feldspar-biotite-sillimanite schist, quartz-feldspar-two mica schist, and quartz-feldspar-two mica-sillimanite schist. Small purplish-red garnets are common in some rocks. The most distinctive aspect of these micaceous schists is the presence of thin wisps of fibrolitic sillimanite parallel to the foliation. In hand specimen the cross sections of these fibrolitic aggregates appear as thin white wisps or lenses with rather frayed ends. They are somewhat less than 1 cm long and 1 mm or less thick and about equi-dimensional in the plane of the foliation. Where sillimanite is abundant, the lenses and wisps coalesce and isolate the quartz-feldspar groundmass into lenses of interlocking granular texture. In thin section the fibrolite appears as ragged lenses and wisps consisting of many needles of sillimanite. Commonly the sillimanite is intergrown with biotite and in some specimens with muscovite.

The quartz-feldspar-two mica-sillimanite schist fraction of the migmatitic gneiss is usually medium to coarse-grained with the micas generally larger than the other minerals in the rock. Biotite is usually coarser and more abundant than muscovite and usually shows some degree of alteration to chlorite. The groundmass consists of an interlocking granular aggregate of quartz and feldspar. In the more micaceous rocks, the granular aggregate of quartz and feldspar locally becomes divided into isolated lenses of granular quartz and feldspar. The feldspar is plagioclase (An25 to An30) but microlite is occasionally present in the groundmass also. Table 2 includes the average estimated modes and assemblages of the dark fraction of the migmatitic gneisses. The chlorite shown in the modes is always a retrograde product of biotite.

Thin beds of calc-silicate granulite account for less than 1 per cent of the Noyes Mountain formation and are essentially quartz-plagioclase-diopside granulite with numerous reddish to flesh-colored garnets. In many cases notable amounts of hornblende are present also. The calc-silicate beds are 1 to 4 inches thick and many lense out within the limits of a single outcrop. In a few outcrops calc-silicate granulite beds are as much as 10 to 20 feet thick.

A minor rock in the Noyes Mountain formation is a graphite-rich and sulfide-rich quartz-feldspar-biotite schist that weathers to a deeply stained, reddish-orange to black surface. Graphite probably accounts for 15 to 20 per cent of some thin beds. Good exposures of this type are present on the southern slopes of Shaw Ledge and the brook just southeast of Eastman Hill.
Another very minor rock within the Noyes Mountain formation is quartz-garnet-sulfide-biotite schist that occurs in thin beds only 1 to 2 inches thick. Some beds are as much as 50 modal percent garnet.

Thickness.—Reasonable estimates for the thickness of the Noyes Mountain formation can be obtained from the west side of Patch Mountain or the east side of Noyes Mountain. Both the top and bottom of the unit is exposed at these two localities. Uncertainty lies in the thickening due to minor folding, poor topographic control, and the locations of the exact contacts. The maximum thickness is about 1150 feet. An average value of the thickness is on the order of 800 feet.

The very broad outcrop of the Noyes Mountain formation just south of West Paris is mainly due to the low dip, but faulting may also be a factor.

Origin.—The protolith for the Noyes Mountain formation is shale interbedded with argillaceous feldspathic sandstone. The thin calc-silicate beds represent original dolomitic feldspathic sandstones. Commonly the thin calc-silicate bands have been boudined into concretion-like nodules but can be distinguished from true concretions by a lack of any concentric structure. The thin pyrrhotite-rich and graphite-rich schists indicate periods during which reducing conditions prevailed in local basins. Two facts support a marine origin for the protolith of the Noyes Mountain formation: 1) The unit is very distinctly bounded above and below by originally dolomitic units which are almost certainly marine in origin. 2) It is more likely that the sulfide-rich and carbonaceous layers that gave rise to the thin but extensive lenses of pyrrhotite-rich and graphite-rich beds originated in an euxinic marine environment.

Correlation.—The writer suggests a possible correlation with the lower part of the Waterville formation in the Waterville area (Osberg, in preparation). This is considered more fully in a later section.

**Berry Ledge formation**

**General Statement.**—The Berry Ledge formation is a distinctive, thin, calcite-rich calc-silicate unit. The predominant lithology is a dark-green quartz-calcite-plagioclase-diopside granulite with interbedded impure marble. The name Berry Ledge formation is proposed here for the first time. The type locality is Berry Ledge, which lies one mile northeast of the village of West Paris. Here the unit is well exposed in cliffs in a shallow syncline plunging gently north. This unit is the most distinctive marker horizon in the southern sequence. It is well exposed along the west side of Stearns Hill, in the West Paris quarry, and on the hills on either side of Rock Dundee. It lies conformably above the Noyes Mountain formation and conformably below the Moody Brook formation and is distinguished from both by its distinctive calc-silicate nature as opposed to their micaceous lithology.

Lithology.—Outcrops of the Berry Ledge formation have a distinctive punky-weathering aspect that emphasizes the bedding. The thin impure marble tends to weather out to a much greater extent than the quartz-calcite-plagioclase-diopside granulite. The remnant of weathering is a loose aggregate of diopside and plagioclase grains in a furrow that is enclosed between beds of relatively unweathered green diopside granulite (Photo 3). The distinctive weathered character of the Berry Ledge formation and its small stratigraphic thickness serve to distinguish it from the Patch Mountain formation. Fresh specimens of the thin marble beds appear to be composed of 60 to 75 percent calcite, the rest being diopside and plagioclase. The marble beds are about 1 inch thick and the diopside granulite beds 1 to 2 inches thick. In places, however, the diopside granulite beds are up to 18 inches thick with few or no marble beds. The bedding in these marble-free zones is made evident by 1 to 2 inch beds of purplish-gray biotite schist. Diopside granulite accounts for at least 70 percent of the unit and impure marble about 20 percent.

In hand specimen the diopside granulite is a medium-grained dense dark-green rock. Diopside is the most obvious mineral. All variations exist between quartz plagioclase-diopside granulite and impure marble. Some fresh hand specimens contain pyrrhotite.

Table 2 presents the average estimated modes of the quartz-calcite-plagioclase-diopside granulite from the Berry Ledge formation. In general these granulites are richer in calcite than are those of the Patch Mountain formation. The diopside occurs as equant, anhedral, commonly poikilitic grains, and accounts for 30 to 40 percent of the rock. Hornblende, present in small
amounts in most samples, occurs as elongate, ragged grains, many of which seem to be altering to diopside. Quartz accounts for 10 to 20 per cent of the rock and plagioclase, (Ca. An₁₀), about 25 per cent. A few specimens have scapolite and/or microcline in notable amounts. The relations of zoisite suggest that it may be a retrograde product of calcic plagioclase. Sphene is an ubiquitous accessory. The opaque minerals are predominantly pyrrhotite and graphite.

Thickness.—The thickness of the Berry Ledge formation can be determined with reasonable accuracy in several places. At the type locality on Berry Ledge and along the west side of Stearns Hill, the bedding dips 5° or less. Altimeter traverses up these hills give thicknesses on the order of 150 to 250 feet thick.

In some areas this unit has a relatively large breadth of outcrop due to the low easterly dips (5 to 15°).

Origin.—The Berry Ledge formation is derived from dolomitic or calcareous sediments. By comparison with the stratigraphically equivalent Benton Falls limestone in the unmetamorphosed Waterville area it would seem that much of the original carbonate was calcite.

Correlation. The Berry Ledge formation may have a possible correlation with the Benton Falls limestone as described by Osberg (in preparation).

Moody Brook formation

General Statement.—The Moody Brook formation consists mainly of migmatitic gneiss with the dark fraction made up of dense quartz-feldspar-biotite schist and quartz-feldspar-two mica-sillimanite schist. In general it is very similar to the Noyes Mountain formation and cannot be distinguished from it at an isolated outcrop in the field or in thin section. However, a comparison of the modes of the Moody Brook formation and the Noyes Mountain formation (Table 2) shows that the former is less micaceous.

The name Moody Brook formation is proposed here for the first time. It is well exposed on the hills immediately north of West Paris, the hill just east of Stearns Hill and at the type locality along the upper part of Moody Brook (Plate 1).
Stearns Hill and north of Moody Brook. However, this is only a minimum estimate because the top of the unit is not exposed and structural complexities exist there. The strata dip at angles of 5 to 15° eastward and lie within the core of a syncline strongly overturned to the southwest so that much minor folding and tectonic thickening or thinning are possible. The writer believes the Moody Brook formation is at least 600 feet thick.

Correlation.—The writer suggests a possible correlation with the upper part of the Waterville formation as defined by Osberg (in preparation) in the Waterville area.

**Correlation of the Southern Sequence**

The four units of the southern sequence have a lithologic correlation with at least part of the stratigraphic section found in the Waterville-Augusta areas. However, the specific correlation of individual units between the two areas remains in doubt for several reasons. (1) As seen on Fig. 1, over 45 miles separate the Bryant Pond area from the Waterville region. (2) Over much of the intervening area only reconnaissance mapping has been done in recent years. Somewhat more detailed mapping was done nearly twenty-five years ago, (Fisher, 1942) (Hanley, 1939) but has been largely superseded by more recent work. (3) The stratigraphic section in the Waterville-Augusta area is still somewhat tenuous, especially as it is extended westward. Moreover, recent mapping and reconnaissance (under the auspices of the Maine Geological Survey) in the region between Bryant Pond and Waterville indicates rather complex stratigraphic problems. (4) Finally, considerable structural complexity exists in the intervening region.

Hence, in the light of the preceding, it must be realized that even a tenuous correlation of specific units between the two areas may be premature. The following correlations may then be considered as suggestions; especially inasmuch as the most recent and important work in the Waterville area (Fig. 1), by Professor P. H. Osberg of the University of Maine, is of this date not yet completed.

Plate II shows the relationship of the southern sequence to the stratigraphy in nearby regions including the sections determined by Fisher (1942) in the Lewiston area and Hanley (1939) in the Poland quadrangle. The terminology used by these two workers is now largely obsolete but a rough correlation, based upon lithologic similarity, can be made with similar units in the Bryant Pond area.

In the Waterville-Augusta area, P. H. Osberg (oral communication) has described a number of units which are lithologically quite similar to those in the Bryant Pond area. This is especially true when these units are traced into more highly metamorphosed regions. Moreover, two of the units contain Silurian graptolites. Some of these units may be summarized as follows:

<table>
<thead>
<tr>
<th>Vassalboro formation</th>
<th>calcareous sandstone, quartz-biotite-actinolite granulite, calc-silicate granulite</th>
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<tbody>
<tr>
<td>Waterville formation</td>
<td>light-green slate and sandstone, mica schist and sillimanite schists</td>
</tr>
<tr>
<td>Benton Falls ls.</td>
<td>interbedded limestone and shale, and marble</td>
</tr>
<tr>
<td>Waterville* formation</td>
<td>light-green slate and thin sandstone</td>
</tr>
<tr>
<td>Mayflower Hill* formation</td>
<td>slate, feldspathic s.s., phyllite, calcareous sandstone</td>
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</table>

*Fossils present.

The Mayflower Hill formation and the lower part of the Waterville formation contain graptolites which indicate a Silurian age. The writer would suggest that the Patch Mountain formation is equivalent to the Mayflower Hill formation and that the Noyes Mountain, Berry Ledge, and Moody Brook formations are equivalent to the Waterville formation. It should also be noted that some of the other units tentatively defined by Osberg (oral communication) which are not described above, are lithologically similar to the Vassalboro, Waterville, and Mayflower Hill formations. As yet the exact relation of these two groups of strata is not fully known. Hence the possibility must be considered that the southern sequence of the Bryant Pond area
could correlate with these other units rather than with the Waterville formation etc. In conclusion, the only firm correlation which can be made is that the southern sequence of the Bryant Pond quadrangle correlates with some part of the Waterville section rather than with the well known stratigraphic section of Littleton-Moosilauke area of New Hampshire, Billings (1937).

The correlations suggested above are based upon several lines of evidence. The firm correlation made between the southern sequence and some part of the Waterville sequence is based upon material drawn from the literature (in particular Fisher (1942), Hanley (1939) and Barker (1961), see Fig. 1), personal communication with P. H. Osberg on his work in the Waterville area, and some reconnaissance by the writer. The most significant facts concerning this correlation are: (1) Large amounts of carbonate-rich strata or their metamorphosed equivalents are present throughout the region from the southern part of the Bryant Pond quadrangle to the Waterville-Augusta area. (2) Fisher (1942) and Hanley (1939) have demonstrated in at least reconnaissance fashion that the carbonate-rich strata of the Waterville area extend throughout the Lewiston and Poland areas to the southeast corner of the Bryant Pond quadrangle.

The more specific correlations suggested by the writer between the Waterville and Bryant Pond areas are based upon lithologic similarity (although in different metamorphic grades), similar sequence of lithologic types, and similar thicknesses for the correlated units.

The ages presented on Plate II for the units in the southern sequence are largely the result of the suggested correlations with the Mayflower Hill and Waterville formations of the Waterville area. Osberg (personal communication) has informed the writer that graptolites found in the Mayflower Hill formation indicate a Late Llandovery age and that graptolites in the Waterville formation also indicate a Silurian age.

NORTHERN SEQUENCE

Thompson Mountain formation

General Statement.—The Thompson Mountain formation is found only along the eastern border of the Bryant Pond quadrangle (Plate 1); good exposures are present only on the two hills north of the village of Redding.

The most extensive areas underlain by this formation are in the Buckfield quadrangle, and Mr. Jeffrey Warner has kindly informed the writer that very good exposures are present on Thompson Mountain; hence the name Thompson Mountain formation is used here for the first time. Due to the very limited and poor exposures of this unit in the Bryant Pond quadrangle it cannot be described in great detail here. In general it is a rusty-weathering migmatitic gneiss. The dark bands predominate and consist of coarse quartz-feldspar-two mica-sillimanite schist and in a few cases quartz-feldspar-biotite granulite. Fibrolitic sillimanite is common in the biotite flakes. The light-colored fraction is made up of lenses and bands of pegmatitic quartz and oligoclase. The light bands and lenses are 1 to 4 inches thick and occur at irregular intervals within the dominant dark fraction. Some blebs of coarse-grained microcline and quartz are present. Mr. Jeffrey Warner has determined that the Thompson Mountain formation lies conformably below the Shagg Pond formation in the nose of a northward-plunging anticline. Hence it is the lowest unit within the northern sequence in the Bryant Pond quadrangle. The Thompson Mountain formation is considered to be Siluro-Devonian in age. No reliable estimate can be made on the thickness of this unit in the Bryant Pond quadrangle. It was probably derived from shale and feldspathic sandstone deposited in a reducing environment. The Thompson Mountain formation is similar to the stratigraphically higher Billings Hill formation.

Shagg Pond formation

General Statement.—The Shagg Pond formation, which is composed largely of coarse-grained migmatitic, gray gneisses with varying amounts of inter-bedded quartz-feldspar-biotite granulite, is found mainly in the eastern part of the quadrangle as a broad northeast-trending belt (Plate I). The name Shagg Pond formation is proposed here for the first time. Good and readily accessible exposures are found at the type locality of Shagg Pond as well as southward along the road to Redding.

Lithology.—In general two distinct lithologic facies can be distinguished on the basis of grain size and mineralogy: north of
Moll Ockett Mountain, the Shagg Pond formation is coarse-grained, migmatitic and contains numerous megacrysts of muscovite. The northern facies of the Shagg Pond formation consists largely of gray coarse-grained migmatitic gneisses. The light-colored fraction is made of lenses and bands of pegmatitic quartz and plagioclase (An25) and in some cases quartz monzonite that are irregularly spaced and generally 1 to 4 inches thick. In most cases they are parallel to the foliation of the dark fraction of the rock and probably account for slightly more than one-third of the volume of a given outcrop. In some areas, however, the light-colored fraction is so abundant that the dark layers are merely inclusions in a groundmass of pegmatite or quartz monzonite. Coarse blebs of intergrown quartz and microcline are common also.

The dark fraction of the migmatitic gneiss is generally a coarse, well-foliated quartz-feldspar-two mica-sillimanite schist. Muscovite is generally in coarse megacrysts up to 5 cm long and in some cases lies at high angles to the foliation, which is determined mainly by the biotite. Fibrolitic swarms of sillimanite are common in the biotite and in a few cases in the muscovite. Granular quartz and plagioclase form the groundmass of the dark fraction of the gneiss and occur as lenses bounded by coarser micas. Some microcline is present. Garnets are common but not abundant in the coarse mica-rich fraction of the migmatitic gneiss. Averaged modes and assemblages of the coarse dark fraction of the gneiss are given in Table 3.

Many thin beds of granulite, consisting essentially of biotite, quartz, and plagioclase with minor amounts of garnet, microcline, and muscovite occur within the dark fraction of the gneiss. The modes are presented in Table 3. The granulite beds are generally one-half to three inches thick and constitute from one-sixth to one-fifth of the dark fraction of the Shagg Pond formation. The presence of these beds emphasizes the stratified nature of the unit and shows that the light-colored bands and foliation are parallel to bedding.

Bedding shows very well on weathered outcrops in areas where the migmatitic nature of the Shagg Pond formation is lacking or restricted. There the thin biotite granulite beds are gradational into the thicker beds of coarse two mica schist which is dominant in the dark fraction of the gneisses. In most cases the coarse mica schist beds are 4 to 10 inches thick. Good exposures, quite free of the migmatitic banding, are found at the road outcrops near the outlet of Shagg Pond and on top of Speckled Mountain. Most commonly migmatitic banding obscures the relation between the quartz-feldspar biotite granulite and coarse quartz-feldspar-two mica-sillimanite schist.

The southern facies of the Shagg Pond formation is seen in its most northerly occurrence on the north slopes of Moll Ockett Mountain. The change to the southern facies is gradational and is characterized by the disappearance of the megacrysts of muscovite, an increase in the amount of sillimanite and garnet, and an increase in the relative amount of quartz-feldspar-biotite granulite. Coarse quartz-feldspar-two mica-sillimanite schist accounts for about two-thirds to three-fourths of the dark fraction of the southern facies and is finer grained than that of the northern facies. Fibrolitic sillimanite commonly occurs as conspicuous platelets and lenses, and numerous 2 mm red garnets are scattered throughout the rock. Similar garnets are also common in the quartz-feldspar-biotite granulite. The schistose part of the gneiss is better foliated than in the northern facies. Good exposures are present on Moll Ockett Mountain and the west sides of the hills immediately to the west. Table 3 shows the estimated average modes of the Shagg Pond formation. Only small amounts of microcline have been observed in the groundmass of this unit, but coarse-grained microcline is commonly observed in the migmatitic lenses and sometimes as individual megacrysts.

Minor minerals within the dark fraction of the migmatitic gneiss include apatite, zircon (in biotite), chlorite (retrograde), tourmaline, and rutile. The opaque minerals are pyrrhotite and graphite.

A few 4 to 12 inches thick beds of quartz-plagioclase-diopside granulite are also present in the Shagg Pond formation and have been stretched into a series of ellipsoidal boudins. However, some calc-silicate nodules probably represent metamorphosed dolomitic concretions.

In the vicinity of Redding and on the west side of the hill northwest of South Woodstock there are calc-silicate granulite units 20 to 50 feet thick. Near Redding they are largely quartz, diopside, hornblende, and calcic plagioclase but they grade into im-
Table 3A
Average Estimated Modes of the Siluro-Devonian Formations of
the Northern Sequence of Units, Bryant Pond Quadrangle, Maine.

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Average An Content of Plagioclase 25 34 25 63 25 38 62

(1) Shagg Pond formation: quartz-feldspar-two mica-sillimanite schist fraction of the migmatitic gneiss (northern facies).
(2) Shagg Pond formation: quartz-feldspar-biotite granulite.
(3) Shagg Pond formation: quartz-feldspar-two mica-sillimanite schist fraction of the migmatitic gneiss (southern facies).
(4) Shagg Pond formation: calc-silicate granulite.
(5) Billings Hill formation: quartz-feldspar-two mica-sillimanite schist fraction of the migmatitic gneiss.
(6) Billings Hill formation: quartz-feldspar-biotite granulite.
(7) Billings Hill formation: calc-silicate granulite.

Table 3B
Assemblages in the Siluro-Devonian Formations of the Northern Sequence
(Numerals indicate number of specimens)

(1) Shagg Pond formation (Northern facies)
Quartz-feldspar-two mica-sillimanite-schist
qtz-plag-bio-musc-sill-garn (2)
qtz-plag-bio-musc-sill-micr (1)
qtz-plag-bio-musc-sill (2)

(2) Shagg Pond formation
Quartz-feldspar-biotite granulite
qtz-plag-bio-musc-sill-garn (2)
qtz-plag-bio-garn (4)
qtz-plag-bio-musc (1)
qtz-plag-bio-musc-sill (1)
qtz-plag-bio-musc-biotite (1)

(3) Shagg Pond formation (Southern facies)
Quartz-feldspar-two mica-sillimanite schist
qtz-plag-bio-musc-sill-garn-micr (5)
qtz-plag-bio-musc-sill (3)
qtz-plag-bio-musc-sill-garn (6)
qtz-plag-bio-musc-garn (1)

(4) Shagg Pond formation
Calc-silicate granulite
qtz-hbl-diop-scp-cal (1)
qtz-hbl-diop (1)
qtz-hbl-garn-bio (1)
qtz-plag-diop-garn-scp-cal (1)

(5) Billings Hill formation
Quartz-feldspar-two mica-sillimanite schist
qtz-plag-bio-musc-sill (9)

(6) Billings Hill formation
Quartz-feldspar-biotite granulite
qtz-plag-bio (4)
qtz-plag-bio-musc-sill-garn (1)
qtz-plag-bio-musc-garn-micr (1)
qtz-plag-bio-garn (1)
qtz-plag-bio-musc-sill (1)
qtz-plag-bio-musc-biotite (1)

(7) Billings Hill formation
Calc-silicate granulite
qtz-plag-hbl-bio (1)
qtz-plag-Fetrem-bio (1)
qtz-plag-diop-hbl (2)
qtz-plag-diop-bio (1)

pure marbles northwest of South Woodstock. The modes of some calc-silicate rocks and one impure marble are presented in Table 3.

The Shagg Pond formation is distinguished from the underlying Thompson Mountain formation by its gray-weathering surfaces. It is in fault contact with much of the Moody Brook formation of the southern sequence and distinguishable from it in several ways. As shown in the modes, (Tables 2 and 3) the Moody Brook formation is considerably less aluminous than the gneisses of the Shagg Pond formation, but, on the other hand, it is generally richer in garnet. On weathered surfaces the clots of sillimanite in the Shagg Pond formation produce a characteristic very knobby surface. This is a very restricted
feature on the weathered surfaces of the Moody Brook formation. Thin calc-silicate bands are much more common in the Moody Brook than in the Shagg Pond. The per cent of quartzfeldspar-biotite granulite or schist as opposed to muscovite and sillimanite-rich schists is much greater in the Moody Brook formation. The fibrolitic sillimanite in the northern sequence tends to occur as indistinct ragged tabular wisps. Sillimanite is less abundant throughout the southern sequence. A few exposures of the gneisses of the southern facies of the Shagg Pond formation have megacrysts of muscovite while none of the schists in the southern sequence have this feature. Bedding is better developed within the Moody Brook unit.

Thickness.—An estimate on maximum thickness of the Shagg Pond formation was obtained on Moll Ockett Mountain where both the top and bottom of the unit are exposed. The maximum thickness is found to be 4500 feet. Topographic correction has been neglected and, inasmuch as numerous isoclinal minor folds are visible in the Moll Ockett region, it is estimated that the maximum thickness is more nearly 3000 feet.

Origin.—The Shagg Pond formation is derived from shales and interbedded minor feldspathic, somewhat argillaceous sandstones. Such an origin is indicated by the micaceous nature of the present coarse-grained quartz-feldspar-two mica-sillimanite schist and the quartzose, feldspathic nature of the quartz-feldspar-biotite granulite. The interbedded nature of these granulites and schists in the dark fraction of the migmatitic gneisses seems to preclude a volcanic origin for the granulite. The thin calc-silicate beds probably represent dolomitic, argillaceous sandstones. The impure marble found on the hill to the northwest of South Woodstock is probably the metamorphosed equivalent of a siliceous dolomite.

Correlation.—The Shagg Pond formation lies below the Billings Hill formation and the Concord Pond member of the Littleton formation. It will be shown later that the Concord Pond member is probably equivalent to the lower gneiss unit of the Littleton formation mapped by Fisher (1952, 1962) in the Bethel quadrangle. Hence the Billings Hill formation and the Shagg Pond formation are both stratigraphically lower than any units in the Bethel area. Reasons are put forth in a later section which favor a “Siluro-Devonian” (?) age for the conformable sequence; Thompson Mountain formation, Shagg Pond formation, and Billings Hill formation.

Billings Hill formation

General Statement.—The Billings Hill formation consists mainly of migmatitic rusty-weathered, coarse-grained gneisses, but as much as one-fourth of the unit is medium-grained, dense rusty-weathered quartz-feldspar-biotite granulite. Minor calc-silicate beds and calc-silicate nodules are present also. The name Billings Hill formation is proposed here for the first time. This unit is present as a broad belt that covers much of Spruce Mountain and Billings Hill (Plate 1). It continues southward almost to South Woodstock and thence northeastward almost to the northeast corner of the quadrangle, where it then continues into the Buckfield quadrangle. Two isolated areas of the Billings Hill formation are found along the northern boundary of the quadrangle in the vicinity of Thurston Mountain and Little Zircon Mountain. In general, exposures of this unit are good, especially on the higher and steeper hills. The best and most accessible exposures are located along the brook that parallels Billings Hill Road on the west side of Billings Hill. Hence this brook and vicinity have been chosen as the type locality for the formation.

The Billings Hill formation is distinguished from the units above and below it on the basis of its rusty weathering. Inasmuch as some rusty lenses occur in the units above and below, the boundary of the Billings Hill formation is commonly difficult to define. However, the writer believes that it is a legitimate stratigraphic unit for the following reasons. In the vicinity of Chamberlain Mountain the structural data indicate a northwest-plunging syncline and this is also indicated by the outcrop patterns. Moreover in the vicinity of South Woodstock the structural data and outcrop patterns both indicate a north-plunging syncline. Inasmuch as these two lines of evidence both lead to the same structural interpretation (in two different localities), it would seem that the outcrop patterns are those of stratigraphic units. In the Bethel quadrangle Fisher (1952) also found that the outcrop pattern of rusty and gray-weathered migmatitic gneiss units lead to the same structural interpretation as that
derived from structural data such as strikes and dips, and minor structures. Another reason for accepting this method of distinguishing units is that the pyrrhotite (or at least the original sulfide) in the rusty units is sedimentary in origin, as shown by Fisher (1952).

Lithology.—The Billings Hill formation consists of two major rock types. Migmatitic gneiss in which the coarse quartz-feldspar-two mica-sillimanite schist makes up most of the dark fraction is the predominant rock type in the Billings Hill formation. Probably three-fourths of the unit is composed of this rock type. Possibly the light-colored fraction accounts for one-third to one-half of this migmatitic gneiss. The light-colored fraction is particularly abundant in the vicinity of Spruce Mountain where the dark fraction is commonly present only as inclusions in a groundmass of pegmatitic and quartz monzonite. Such areas of very large amounts of light component are indicated by an overprint pattern on Plate I. Migmatitic gneiss in which the dark fraction consists of quartz-feldspar-biotite granulite is the other major rock type in this unit. As in the Shagg Pond formation, scattered thin beds of this granulite are present in the migmatitic gneiss in which the dark fraction is mainly coarse quartz-feldspar-two mica-sillimanite schist. However in some outcrops the quartz-feldspar-biotite granulite is dominant over the coarse quartz-feldspar-two mica-sillimanite schist of the dark fraction and the rock grades into a migmatite consisting of quartz-feldspar-biotite granulite and pegmatitic quartz-feldspar bands. In some exposures where the light-colored bands are also of minor importance the rock is a massive-bedded quartz-feldspar-biotite granulite. It is possible to find as much as 100 feet of granulite which is relatively free of migmatitic banding. Exposures in the brook beside Billings Hill Road present good examples of this rock type.

The Billings Hill formation is very similar to the underlying Shagg Pond formation, except for its rather high content of iron sulfide. The rusty-weathered surfaces are deep reddish-brown to black and the stain commonly penetrates as much as two inches into the rock. Because of the iron oxide stains, it is often more difficult to discern bedding on weathered surfaces than it is on a corresponding outcrop in the Shagg Pond formation. The bedding and migmatitic banding are a little more contorted in this unit as compared to the gray units immediately above and below. On fresh surfaces this unit is very similar in appearance to the gray units except for a very soft greenish tinge to the normally gray or grayish-brown color. This greenish tinge is apparently due to finely disseminated sulfide. In many specimens, however, one can also see megascopic anhedral smeared-out grains of the sulfide. The sulfide is pyrrhotite as it is magnetic. Fisher (1962) reports that the pyrrhotite in similar gneisses in the Bethel quadrangle contain small amounts of exsolved pyrite.

Table 3 shows the modes of the Billings Hill formation. Comparison with those from the Concord Pond member or the Shagg Pond formation shows that these units have similar mineral contents except that the Billings Hill formation contains considerably greater amounts of opaques. Pyrrhotite is the main opaque mineral, but graphite is also present. It is also notable that garnet occurs in minor amounts in this unit and is totally absent from the gneissose fraction. No calc-silicates greater than one to one and one-half feet thick are present in this unit.

Thickness.—Estimates of thickness of the Billings Hill formation can be obtained from either limb of the north-plunging syncline immediately north of South Woodstock. By means of trigonometry a maximum thickness of 3000 feet is obtained but because of isoclinal minor folding, 2000 feet (two thirds of the maximum thickness) may be a more reasonable figure.

Origin.—The Billings Hill formation is derived mainly from shales and siltstones. The presence of considerable amounts of sedimentary sulfide and the presence of graphite indicate that the original sediments were deposited in a marine reducing environment. The biotite granulites represent feldspathic sandstones which were also deposited in a reducing environment. The minor beds of calc-silicate granulite are probably derived from dolomitic, feldspathic sandstones.

Correlation.—Inasmuch as the Concord Pond member seems to be equivalent to the lower gray gneiss unit in the Bethel quadrangle, it would seem that the Billings Hill formation is stratigraphically lower than anything in that area. The writer considers the Billings Hill formation to be at the top of a Siluro-
Devonian sequence immediately below the Devonian Littleton formation.

Littleton formation

General Statement.—The Littleton formation has been subdivided into three members in the Bryant Pond quadrangle, the Concord Pond member, the Wilbur Mountain member, and the Howard Pond member.

Concord Pond member

General Statement.—The Concord Pond member is mainly a gray-weathered, coarse-grained, quartz-feldspar-two mica-sillimanite, migmatitic gneiss interbedded with lesser amounts of gray to dull-brown quartz-feldspar-biotite granulite. Minor amounts of calc-silicate granulite are also present. The name Concord Pond member is proposed here for the first time. This unit is well exposed on many of the mountain tops in the north-central part of the area. Particularly good exposures are present on Kimball Hill, Mount Zircon, Bean Mountain, and Davis Mountain (Plate I). Very extensive exposures are also present at Concord Pond, the type locality.

Included within the Concord Pond member are one or more 50 to 75 foot thick horizons of calc-silicate granulite. As shown on Plate I, no attempt has been made to consider the outcrops of this rock type as belonging to a single horizon, although further detailed work may demonstrate this to be the case. Very good examples of such rocks are exposed in the southeast corner of Concord Pond.

In general, map pattern and structural data indicate that the Concord Pond member is conformable with the strata above and below it.

Lithology.—The Concord Pond member consists of about 75 per cent migmatitic gneiss, 20 to 25 per cent quartz-feldspar-biotite granulite, and minor amounts of quartz-plagioclase-diopside-granulite.

In most areas the Concord Pond member is made of highly migmatitic gneiss. Commonly as much as one-half of the rock consists of lenses and pods of pegmatitic quartz-feldspar or quartz monzonite. These bands are 1 to 4 inches thick and in most cases less than 10 feet long. In some of the smaller pods and lenses the feldspar is coarse, translucent microcline with a rather pearly luster. In some specimens this microcline occurs as single anhedral to subhedral crystals up to 2 inches long. In the larger pods and lenses the feldspar is plagioclase with composition near oligoclase. In many places the light-colored fraction of the rock increases and the rock grades into a quartz monzonite or pegmatite with schistose inclusions.

The dark fraction of the migmatitic gneiss of the Concord Pond member is predominantly coarse-grained quartz-feldspar-two mica-sillimanite schist. Lesser amounts of quartz-feldspar-biotite granulite are present in the dark fraction. Because of the thin beds of granulite, one can usually find bedding in any outcrop. In the vast majority of cases, the bedding foliation and migmatitic banding are parallel. In some areas where the amount of the light fraction is reduced, it is possible on weathered surfaces to see that the thin granulite beds are gradational into the thicker coarse quartz-feldspar-two mica-sillimanite schist beds. This is especially true in the northeastern part of the quadrangle. In these exposures the Concord Pond member is nearly identical with the Shagg Pond formation.

In hand specimen, attention is focused mainly on the coarse schistose fraction of the gneiss. It is a coarse quartzose two mica rock. Muscovite forms poikilitic megacrysts up to 2 inches long. Usually they are parallel to the foliation but not uncommonly lie at high angles to the foliation. Granular quartz and plagioclase form wisps and lenticles that fill in the spaces among the larger mica flakes. By a decrease in the micaceous nature and an increase in quartz and feldspar the coarse two mica fraction of the gneiss grades into the quartz-feldspar-biotite granulite. Sillimanite is present as felted masses of fibrolite intergrown with biotite. On weathered surfaces these biotite-sillimanite intergrowths produce a rather ribbed knobby appearance (Photo 3). Small red garnets (1 to 2 mm) are common in the coarse schistose fraction of the gneiss.

Microscopically, muscovite contains long needles of fibrolite in a few specimens. A few thin sections also show that microcline occurs sparingly in the quartz-feldspar groundmass of the coarse
schistose fraction of the gneiss. Granular lenses of quartz and plagioclase lie between the mica plates or enclose them if the amount of mica is small. The plagioclase ranges in composition from An$_{30}$ to An$_{50}$. Biotite is usually more abundant than muscovite. Accessories are opaques, tourmaline, apatite, zircon (in the biotite) and chlorite. Pyrrhotite is the major opaque mineral in the rusty-weathering lenses within the ordinarily gray Concord Pond member, but some graphite is also present.

The quartz-feldspar-biotite granulite occurs as thin, discontinuous beds within the gneiss although in some instances they are rare. Commonly the beds are one-half to two inches thick.

### Table 4A

Average Estimated Modes of the Members of the Littleton Formation, Bryant Pond Quadrangle, Maine.

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| Average An Content | 23 | 26 | 70 | 73 | 24 | 29 | 74 | 29 | 68 |

(1) Concord Pond member: quartz-feldspar-two mica-sillimanite schist.
(2) Concord Pond member: quartz-feldspar-biotite granulite.
(3) Concord Pond member: Thin calc-silicate granulite lenses and concretionary nodules.
(4) Concord Pond member: mappable calc-silicate horizons.
(5) Wilbur Mountain member: quartz-feldspar-two mica-sillimanite schist.
(6) Wilbur Mountain member: quartz-feldspar-biotite granulite.
(7) Howard Pond member: quartzose calc-silicates of the upper and lower units.
(8) Howard Pond member: micaeous schist of the upper and lower units.
(9) Howard Pond member: calc-silicate and biotite granulite of the middle unit.

### Table 4B

Assemblages in the members of the Littleton formation (Numerals indicate number of specimens)

(1) Concord Pond member

| qtz-plag-bio-musc-sill-garn-micr | 2 |
| qtz-plag-bio-musc-sill | 14 |
| qtz-plag-bio-musc-sill-garn | 7 |
| qtz-bio-musc | 1 |
| qtz-plag-bio-musc-sill-micr | 3 |

(2) Concord Pond member

| qtz-plag-bio-musc-sill-garn | 4 |
| qtz-plag-bio-musc | 6 |
| qtz-plag-bio-musc-sill | 2 |
| qtz-plag-bio-garn-mier | 1 |
| qtz-plag-bio-garn | 2 |
| qtz-plag-bio-garn-micr | 2 |
| qtz-plag-bio-musc-micr | 2 |
| qtz-plag-bio-musc-sill-micr | 2 |
| qtz-plag-bio-musc-sill | 1 |

(3) Concord Pond member

| qtz-plag-diop-calc | 1 |
| qtz-plag-diop-garn | 1 |
| qtz-plag-diop-trem-bio | 1 |
| qtz-plag-diop-alc | 1 |
| qtz-plag-diop-garn-calc | 1 |

(4) Concord Pond member

| qtz-plag-diop-bio | 3 |
| qtz-plag-diop-garn-calc | 2 |
| plag-diop-scap-mier | 1 |
| qtz-plag-diop-bio | 1 |
| qtz-plag-garn | 1 |
| qtz-plag-bio | 1 |
| qtz-plag-diop-scab-mier | 1 |

(5) Wilbur Mountain member

| qtz-plag-bio-musc-sill | 3 |
| qtz-plag-bio-musc-sill-garn | 2 |
| qtz-plag-bio-musc-sill-micr | 1 |
(6) Wilbur Mountain member
Quartz-feldspar-biotite granulite
- qtz-plag-bio-garn (1)
- qtz-plag-bio-musc-garn-micr (2)
- qtz-plag-bio-musc-micr (2)

(7) Howard Pond member
Quartzose calc-silicate of the upper and lower units
- qtz-plag-bio (3)
- qtz-plag-bio-garn-trem (1)
- qtz-plag-trem (1)
- qtz-plag-hnbl-diop (1)
- qtz-plag-trem-diop (1)
- qtz-plag-bio-trem-diop (1)
- qtz-plag-trem-phlog (1)

(8) Howard Pond member
Micaceous schist fraction of the upper and lower units
- qtz-plag-musc-bio (1)
- qtz-plag-musc-bio-garn (1)
- qtz-plag-musc-bio-micr (2)
- qtz-plag-musc-bio-sill (1)
- qtz-plag-musc-bio-sill-micr (1)

(9) Howard Pond member
Calc-silicate and biotite granulite of the middle unit.
- qtz-diop-trem-bio-micr (1)
- qtz-plag-diop-trem (5)
- qtz-plag-bio (1)
- qtz-plag-diop-trem-garn-calc (1)
- qtz-plag-trem-garn-bio (1)

but in some places such as Bryant Mountain, Chamberlain Mountain, and near Abbotts Mills, the granulite occurs as massive beds more than 100 feet thick. On stream and road outcrops the quartz-feldspar-biotite granulite beds are not immediately obvious, partly because of the coarse gneissose nature of the rest of the rock. But on well exposed and weathered outcrops the granulite is well defined because it weathers out preferentially to form depressions. This is particularly evident in areas where the migmatitic component is subordinate. In such areas there is a cyclical interbedding of the granulite and the very coarse-grained quartz-feldspar-two mica-sillimanite schist. The coarse-grained fraction is vastly predominant and makes up about 80 per cent of the rock.

In hand specimen, the quartz-feldspar-biotite granulite is very similar to that in the Billings Hill formation. In some cases biotite occurs as elongate megacrysts up to one centimeter in length and in a few specimens small pebbles of rounded quartz and feldspar, one-sixteenth to one-fourth inch across, are present (especially in the thicker beds of biotite granulite). In some exposures of thick-bedded granulite, bedding is difficult to discern and in other exposures, where the grain size is 2 to 4 mm or greater, the rock is massive. In such outcrops, one might mistake it for a sill-like intrusive. Garnet commonly occurs as small anhedral grains but also occurs as irregular, poikilitic, megacrysts up to one-half inch in diameter.

In thin section, the quartz-feldspar-biotite granulite is again similar to that in the Billings Hill formation. The modes of the granulite (Table 4) show that in some rocks microcline is present in addition to plagioclase.

Calc-silicate lenses and beds from 1 to 15 inches thick have been found occasionally throughout the Concord Pond member. In the quartz-feldspar-biotite granulite they occur as light-colored, even bands 1/2 to 1 inch thick with up to 10 such bands grouped within a 1 to 11/2 foot zone. Usually these bands are rich in calcic plagioclase and quartz with green hornblende and biotite as the major dark minerals. In the thicker calc-silicate bands, especially within the gneissose parts of the member, the calc-silicate grades into a more or less quartz-plagioclase-diopside granulite. Boudined calc-silicate beds are present also along with true calc-silicate nodules showing well-defined concentric structures. Table 4 presents modes of these various calc-silicate types. Despite the wide distribution of the calc-silicate rocks mentioned above, they are insignificant in total amount.

Also present within the Concord Pond member are one or more 50 to 75 foot thick horizons of quartz-plagioclase-diopside granulite. In the field this rock is a grayish-green color and is flaggy to massive bedded. Texturally it is a fine-grained dense granular rock. The plagioclase compositions are near An55. The diopside grains are usually very poikilitic. Table 4 shows the estimated modes of this rock type.

The Concord Pond member is distinguished from the overlying Wilbur Mountain member and the underlying Billings Hill formation mainly by its lack of rusty-weathering outcrops. Except for the rusty-weathering surfaces, the units above and below the Concord Pond member are lithologically very similar to it. Moreover, in a number of cases the Concord Pond member has rusty-weathered patches. Indeed, the large "U" shaped outcrop pattern on Chamberlain Mountain is a lense of rusty gneiss within
the Concord Pond member. In general, the limits of this unit have been set rather arbitrarily to exclude outcrops that are extensive and very rusty. In many cases it was quite difficult to decide whether to include an outcrop or series of outcrops in the Concord Pond member or in one of the rusty units.

Thickness.—In the syncline that plunges northward from South Woodstock it is possible to consider half of the outcrop pattern and by trigonometry obtain a minimum thickness of 1700 feet for the unit. This estimate neglects the possibility of repetition of beds by minor folding. Another estimate of thickness can be obtained from the northwest-plunging syncline in the vicinity of Chamberlain Mountain. Here a thickness of 3400 feet is obtained with no correction made for repetition by minor folding. Assuming a one-third repetition of bedding due to folding, a thickness of 2500 feet is a reasonable estimate. The very broad outcrop area of the Concord Pond member in the north-central part of the quadrangle is due to rather open folds plunging to the north.

Origin.—The protolith of the Concord Pond member is certainly sedimentary. The coarse-grained micaceous part is derived from shale and siltstone while the quartz-feldspar-biotite granulite originated from impure feldspathic sandstone. Gradations from the original sandstone to the shale probably appeared as large-scale graded bedding in the original sediments, but this has largely been obliterated by the intense metamorphism to which the region has been subjected. The gneissose migmatitic nature of most of the unit is probably due to metamorphism and will be treated further under the section on metamorphism. Fisher (1952, 1962) has shown by chemical analyses that essentially the same gneisses in the Bethel quadrangle have formed isochemically from shales.

A possible volcanic protolith for the quartz-feldspar-biotite granulite (especially the thicker beds) has been rejected for two reasons. One is that the estimated modes indicate that it is often too quartzose to have an igneous origin. Second, the presence of some well-rounded quartz and feldspar clasts suggests a sedimentary origin.

The quartzose calc-silicate granulite is very likely derived from dolomitic feldspathic sandstones. Those with larger amounts of hornblende and biotite were probably somewhat shaly dolomitic sandstones.

Correlation.—The Concord Pond member is correlated with part of the Lower Devonian Littleton gneiss in the Mount Washington quadrangle. Fisher (1952, 1962), in the Bethel quadrangle, has shown that these gneisses can be traced into the Mount Washington area.

Wilbur Mountain member

General Statement.—The Wilbur Mountain member consists largely of rusty-weathered, coarse-grained migmatitic gneiss and quartz-feldspar-biotite granulite. The whole unit lies within the sillimanite plus potassium feldspar zone of metamorphism (Plate I). The name Wilbur Mountain member is proposed here for the first time. It is well exposed on the northwest and northern slopes of the Wilbur Mountain area and on the two hills to the south-southwest of Rumford Corner. The best outcrops of the Wilbur Mountain member are in the brook one mile south of East Bethel.

Although this writer considers this unit to be a member in the Littleton formation, it could also be interpreted as a lens of
rusty gneiss within the gray-weathered Concord Pond member. It would then be similar to the lens of rusty gneiss in the Concord Pond member on Chamberlain Mountain. Because of the poor exposures along the Androscoggin River, there is no way of demonstrating the validity of calling this rusty gneiss a stratigraphic unit.

Lithology.—The Wilbur Mountain member is composed dominantly of rusty-weathered migmatitic gneiss, very similar to that of the Billings Hill formation. The light-colored bands are mainly pegmatitic quartz and feldspar and medium-grained quartz monzonite. In general, the migmatitic bands grade from one-inch bands to a scale where the rock is essentially a pegmatite or quartz monzonite with only inclusions of schist. In most areas the light-colored fraction probably accounts for one-third to one-half of a given outcrop.

The dark fraction consists predominantly of coarse quartz-feldspar-two mica-sillimanite schist with lesser amounts of quartz-feldspar-biotite granulite. Bedding can usually be determined in the dark fraction of the migmatitic gneisses by the interbedded granulite, and it is generally found that the foliation in the dark fraction, bedding, and the migmatitic bands are parallel. The granulite accounts for less than 10 per cent of the Wilbur Mountain member. However, in a few places the quartz-feldspar-biotite granulite becomes dominant and may be present as massive beds up to 15 feet thick with little or no migmatitic banding present. Inasmuch as the Wilbur Mountain member is very similar to the Billings Hill formation no further description is necessary. The modes and assemblages of this member are given in Table 4.

Thickness.—Estimates for the thickness of the Wilbur Mountain member are hard to give as only the bottom is well exposed. Moreover its structural position and the lack of good exposures make it impossible to arrive at any definite figure for the thickness of the unit. Probably it is on the order of 300 to 500 (?) feet thick.

Origin.—The pelitic nature of the rusty gneisses indicates that the original sediment was a shale. The presence of pyrrhotite indicates that the shale probably formed under reducing conditions. Fisher (1962, p. 1400) has given reasons for believing that the pyrrhotite (or at least the original sulfide) is sedimentary in origin.

The protolith of the quartz-feldspar-biotite granulite was probably a feldspathic, somewhat argillaceous sandstone. The high quartz content would seem to rule against a purely volcanic origin although metamorphism of an impure, reworked volcanic could possibly result in rock similar to the granulite.

The thin calc-silicate granulite beds are derived from originally dolomitic feldspathic sandstones and the calc-silicate nodules were probably dolomitic concretions.

Correlation.—The writer believes that the Wilbur Mountain member lies within the Devonian Littleton formation and probably correlates with the pyrrhotite gneiss mapped in the Bethel quadrangle by Fisher (1962). In particular, it probably lies near or at the top of what Billings et al. (1946) has called the gneiss of the Littleton formation in the Mount Washington area.

Howard Pond member

General Statement.—The Howard Pond member is made up of three units. The upper unit consists of interbedded rusty-weathering quartz-mica schist and dark quartzites that contain some calc-silicate minerals. The quartzite is free of calc-silicate minerals in the higher parts. The middle unit is a very quartzose calc-silicate granulite and quartz-feldspar-biotite granulite, whereas the lower unit seems to be similar to the upper one but is very poorly exposed. The Howard Pond member is present on Howard Mountain in the northwest corner of the quadrangle with small patches of the lower unit also present on the summit and northwestern slopes of Wilbur Mountain. The name Howard Pond member is proposed here for the first time. The upper unit is best exposed in the Rumford quadrangle at the bridge just below the outlet of Howard Pond. The calc-silicate and quartz-feldspar-biotite granulite unit are well exposed along the steep part of the south slope of Howard Mountain.

Lithology.—The upper and lower units of the Howard Pond member consist of alternating beds of quartz-mica schist and thin dark calc-silicate bearing quartzites in one to six inch intervals, but in places the two rock types are interbedded on a scale one-eight to one-fourth of an inch thick. Both units con-
tain considerable pyrrhotite and thus have rusty-weathered surfaces. In the Bryant Pond quadrangle the calc-silicate bearing quartzite makes up at least 40 per cent of the upper and lower units of the Howard Pond member. It is a very dark gray to black and the calc-silicate nature is evidenced in the field only by the common occurrence of an iron-bearing tremolite. This amphibole is characteristic of the unit and has been observed by the writer in parts of the Rumford quadrangle. It occurs as one-eighth to one-fourth inch needles which flash brightly in reflected light, whereas in this section this amphibole is seen to be almost perfectly euhedral though sometimes enclosing quartz poikilitically. The modes in Table 4 show the mineralogy of this rock type. It is a very fine-grained, dense rock except for the megascopic amphibole. It contains about 10 per cent of opaque minerals, most of which is pyrrhotite, but some graphite is also present.

The micaceous fraction in the upper and lower units of the Howard Pond member is a fine-grained, well-foliated micaceous rock with a silvery blue-gray color on fresh surfaces. On a weathered surface it is rather punky and much stained by iron oxides. In a few cases fresh specimens of the mica schist show that the foliation is crinkled. In general, however, the foliation parallels the bedding of the rock. The modes of this rock type (Table 4) show that it is rather quartzose and that muscovite predominates over biotite. Opaques are abundant and consist mainly of pyrrhotite. Some specimens have abundant fibrolitic sillimanite.

The middle unit of the Howard Mountain member is interbedded fine-grained, light-colored calc-silicate rock and dull-gray to purplish quartz-feldspar-biotite granulite. The interbedding generally occurs on a scale ranging from 1 to 4 inches. In general the quartz-feldspar-biotite granulite predominates and probably accounts for up to 60 per cent of the middle unit. It is mainly a quartz-biotite-feldspar rock in which quartz predominates slightly. In some outcrops this granulite is interbedded with one-sixteenth to one-half inch beds of a very light-colored quartzose calc-silicate rock (usually these bands contain megacrysts of hornblende).

The calc-silicate rock is light-gray to green and quartzose as seen in the modes (Table 4). The light-green color is the result of diopside, slightly iron-bearing tremolite, and in a few specimens hornblende instead of tremolite. The amphiboles are usually euhedral in thin section and are seen to be one-eighth to one-fourth inch needle-like laths in hand specimen. In a few cases the amphiboles become so abundant that the rock can be called an amphibolite. In some instances the calc-silicate granulite was found to contain microcline.

In a general way this unit grades into the other two units by an increase in pyrrhotite and micas.

Thickness.—It is difficult to obtain figures for the thicknesses of the upper and lower units of the Howard Mountain member. The major difficulties lie in the fact that the top of the upper unit and bottom of the lower unit cannot be located in the Bryant Pond quadrangle. The middle unit is probably on the order of 200 feet thick. Forsythe (1955) shows it with a considerable width in outcrop pattern but does not mention any figures on the thickness. Possibly it is thicker in the Rumford quadrangle or repeated by folding. The writer estimates it to be 600 feet thick.

Correlation.—Inasmuch as Billings (unpublished work) and Milton (1961) have extended the Boott member of the Littleton formation through the Gorham and southern part of the Old Speck Mountain quadrangles, respectively, the writer correlates the Howard Pond member with the Boott member. This correlation is based upon lithologic continuity and lithologic similarity with regard to the calc-silicate rocks.

Correlation of Northern Sequence

As shown on Plate II, the age designations for the northern sequence are derived from correlations with the strata underlying much of New Hampshire. Fossils indicate a Lower Devonian age for the Littleton formation (first defined by Ross, 1923) in the Littleton-Moosilauke area, Billings (1937), Billings and Cleaves (1934). Mapping by Billings et al. (1946) in the Mount Washington area, by Billings and Fowler-Billings (unpublished) in the Gorham quadrangle, Milton (1961) in the Old Speck Mountain quadrangle and Fisher (1952, 1962) in the Bethel quadrangle (see Fig. 1) has extended the Littleton formation eastward to the Bryant Pond area. As seen on Plate II the workers in the
Gorham, Old Speck Mountain and Bethel areas have, following Billings et al. (1946), divided the Littleton formation into three members.

In the Bryant Pond area, the writer correlates the Howard Pond member with the Boott member and the Concord Pond and Wilbur Mountain members with the lower Littleton. It should be noted that Fisher (1962) has further refined the lower Littleton as exposed in the Bethel quadrangle by the addition of two units of interbedded quartz-mica schist and sillimanite schist. The writer has not been able to define any units in the Bryant Pond area which coincide with these units. A possible suggestion is that they may be equivalent to some of the zones in the Concord Pond member which contain larger amounts of biotite schist.

In a northward direction the Concord Pond member can be traced into the undifferentiated gneisses and schists mapped by Forsythe (1955) in the eastern and southern parts of the Rumford quadrangle. The Howard Pond member can also be traced northward through the Rumford area where it has been mapped by Forsythe as a band of calc-silicate granulite with interbedded quartzite and micaceous, sulfide-rich schist on either side.

The upper Littleton is not exposed in the Bryant Pond area but reconnaissance in the Rumford area by Billings, J. Warner, and the writer indicates that the strata to the northwest of the interbedded quartzites and rusty schists and the calc-silicate unit are lithologically similar to the upper Littleton. Moreover they may be traced into strata in the Old Speck Mountain area, which Milton (1961) correlates with the upper Littleton. Hence these strata in the Rumford area may tentatively be correlated with the upper Littleton of the Mount Washington area.

Inasmuch as the Concord Pond member seems to be equivalent to the lower part of the gneisses mapped by Fisher (1952) in the Bethel quadrangle, it would seem that the Billings Hill formation is stratigraphically lower than anything in the Bethel quadrangle. In view of this fact, one might become concerned about how far down in the section to extend the name “Littleton” and whether to consider the lower units to be Lower Devonian in age. If one extends the term “Littleton formation” and the implied age to all of the gneisses (no matter how far down in the section), one would have to lump a tremendous thickness of sediments within one formation and with little definite evidence to support such an approach.

Indeed Billings et al. (1946) have pointed out that the fossil horizon within the Littleton formation lies 2500 feet above the base of the unit. Hence even the lower gneisses in the Mount Washington area could conceivably be older than Lower Devonian. Such facts indicate that it is probably more reasonable to impose an arbitrary lower limit to what is included within the Littleton formation. In order to avoid this problem the writer has decided to consider the base of the Concord Pond member as the base of the Lower Devonian Littleton gneisses. The three lower units within the northern sequence will be considered as three separate formations of “Siluro-Devonian” (?) age.

The basis for the preceding correlations are: lithologic similarity, similar sequence of three units, and lithologic continuity. Only the Boott member is found to be discontinuous as it is traced from the Mount Washington area eastward to the Rumford and Bryant Pond quadrangles. Local names have been used for the Littleton formation in the Bryant Pond area because the writer has subdivided the lower gneiss member and because the Boott member is not completely lithologically continuous with the calc-silicate horizon within the Howard Pond unit.

It should be noted here that the preceding age designations for the northern sequence are primarily dependent upon the age designations presented by Billings et al. (1946) for the three units in the Mount Washington area. If the earlier interpretation presented by Billings (1941) were the acceptable one, then the age designations in the Bryant Pond area would of course follow. In that case the gneisses below the Howard Pond member would be equivalent to the Ordovician Partridge formation. The Howard Pond member would then be equivalent to the Fitch formation of Silurian age.

Further mapping by Dr. K. Pankiwskyj and Mr. J. Warner in the Dixfield and Buckfield quadrangles respectively and the areas to the east as far as the Waterville area should give further evidence bearing on the interpretation for the Mount Washington area which is accepted in this report. For example, if the Howard Pond member as extended through the Dixfield quadrangle by
Pankiwskyj were shown to be equivalent to one of the Silurian “ribbon limestones” in the Waterville area, then it would be necessary to reconsider the possibility that the gneisses below the Howard Pond member could be equivalent to the Ordovician Partridge formation.

**INTRUSIVE ROCKS**

**General Statement**

Five groups of intrusive rocks can be distinguished in the Bryant Pond quadrangle. They are the Songo granodiorite, Speckled Mountain quartz monzonite, pegmatite, binary granite, and post-metamorphic trap dikes. The first two are extensively developed in the Bethel quadrangle where Fisher (1962) has defined their type localities and demonstrated that they belong in the New Hampshire magma series (Billings, 1956). The pegmatites and binary granite are also included within the New Hampshire magma series. The post-metamorphic dikes are much later than the rocks of the New Hampshire series and are not abundant.

**Songo Granodiorite**

The Songo granodiorite (Fisher, 1962) underlies about 70 square miles in the southeastern third of the Bethel quadrangle and Plate 1 shows that it also underlies much of the western and central portions of the Bryant Pond quadrangle. Reconnaissance by the writer indicates that the Songo granodiorite extends into the northeast corner of the Fryeburg quadrangle and the northwestern corner of the Norway quadrangle but that most of these two quadrangles is underlain by a binary granite to be described in a later section.

The Songo granodiorite is a non-resistant rock inasmuch as few major hills are underlain by it. It is a rather coarse-grained, dark-gray, massive biotite-rich rock. Plagioclase and quartz are the major light-colored minerals. In the interior parts of the body poor alignment of biotite flakes results only in a weak foliation, but as the contacts are approached the foliation improves. At some contacts the foliation is excellent and is greatly emphasized by inclusions of schist. In a few places long layers of schist and granodiorite alternate on a 1 to 2 inch scale in outcrops up to 30 feet thick. Plate III shows the foliations measured and emphasizes the fact that the body is generally discordant but locally concordant. Along much of the southern boundary the granodiorite is in fault contact with the southern sequence of metasediments. In general, it seems to have the shape of a large locally concordant dome. This is substantiated by the excellent foliation which is parallel to the contacts near the borders. In general, the amounts of pegmatite and quartz monzonite in the metamorphic rocks is greater immediately surrounding the Songo granodiorite, resulting in a poorly-defined aureole. The large north-south trending belt of pegmatite and quartz monzonite shown on Plate I may be part of this aureole which has not yet been eroded from the top of the major pluton. It can be seen that this complex of pegmatite and quartz monzonite underlies a series of higher hills as compared to the adjacent areas underlain by granodiorite. This would seem to indicate that only the top of this pluton has been breached by erosion.

Exfoliation is well-developed at several localities within the Songo granodiorite. It is best exposed at the big cliff on Bucks Ledge, just east of North Pond. The exfoliation slabs seem to be curved and define a number of small exfoliation domes.

In thin section the granodiorite is seen to consist mainly of biotite, quartz, and plagioclase (An50) (Table 5). This biotite is pleochroic in shades of green and brown. Hornblende and microcline are present in minor amounts and apatite, zircon, sphene, and opaques are the common accessory minerals. In thin section the rock has a coarse hypidiomorphic texture. Most of the specimens collected by the writer are actually quartz diorites rather than granodiorites. However, inasmuch as Fisher (1962) called this rock a granodiorite and noted that it grades into a quartz diorite in places, the writer prefers to retain the name Songo granodiorite for the whole pluton. Chlorite is a retrograde product after biotite and is particularly common in the vicinity of the major fault shown on Plate I.

Fisher (1952) has demonstrated that the Songo granodiorite is a member of the New Hampshire magma series (Quinn, 1944; Billings, 1956).

Green (1960) has discussed some of the recent radioactive dating of rocks in northern New Hampshire and particularly for rocks of the New Hampshire magma series in the Errol quad-
rangle. He presents six Potassium-Argon ages from micas (obtained from Hurley et al. (1958) which range from 252 to 340 million years. Green (1960, p. 125) also states, “A single determination by the U. S. Geological Survey on zircon from the Umbagog granodiorite—gave a lead-alpha age of 360 ± 40 m.y.—”.

According to the most recent time scale by Holmes (1960) the figures mentioned above would indicate ages ranging from Upper Devonian to Middle Carboniferous. Boucot (1954) has presented geological evidence that indicates that in Maine the intrusives belonging to the New Hampshire magma series are Middle Devonian or younger. In the Bryant Pond area the only evidence bearing on this problem is that the intrusives of the New Hampshire magma series cut the Lower Devonian Littleton formation. Hence one can only say that these intrusives are post-Early Devonian.

**Speckled Mountain Quartz Monzonite**

The Speckled Mountain quartz monzonite was defined by Fisher (1962) at its type locality at Speckled Mountain in the southwest corner of the Bethel quadrangle. In the Bryant Pond quadrangle it is present mainly in small bodies throughout the metamorphic rocks. However, one large body is present in the vicinity of Rumford Corner, just south of the Androscoggin River.

It occurs as small concordant and discordant lenses and irregular masses and dikes. These bodies range from a few feet across to hundreds of square feet in area. Only one large body is shown on Plate 1, but many of the stipled areas contain large amounts of the quartz monzonite as well as pegmatite. In general, it is more common around the borders of the Songo granodiorite.

The quartz monzonite commonly grades into pegmatites or into migmatitic gneiss by an increase in schist inclusions. Schist inclusions are common and in some cases become so drawn out that they merely grade into wisps of biotite flakes. Such drawn-out schist inclusions commonly have more garnets than the surrounding quartz monzonite. Foliation is prominent in the quartz monzonite and consists of biotite folia and some thin concentrations of small muscovite grains. The foliation is almost always parallel to the contacts of the body.

The Speckled Mountain quartz monzonite is a light-colored medium-grained rock. The major minerals are quartz, plagioclase, microcline, biotite, and muscovite. The composition of the plagioclase ranges from An₂₀ to An₈₅. Common accessories are sillimanite, apatite, zircon, opaques, and garnet. Micropegmatic intergrowths of quartz and plagioclase are common in minor amounts. Chlorite is a retrograde product of biotite. Some clots of muscovite with intergrown fibrolitic sillimanite are occasionally seen in hand specimen and are probably remnants of digested schist. Table 5 presents the modes of the Speckled Mountain quartz monzonite in the Bryant Pond quadrangle.

Fisher (1962, p. 1405) notes, “The rock is similar, except for its banding, to the Concord (Fowler-Billings, 1949, p. 1268) and Bickford (Billings, 1941, p. 896-898) granites of the New Hampshire magma series.”

As pointed out by Fisher, the quartz monzonite dikes cut pegmatite and vice-versa and pegmatite dikes cut the Songo granodiorite. Hence, although the quartz monzonite has not been observed to cut the granodiorite, it is probably contemporaneous with the pegmatites and thus younger than the granodiorite. In addition the greater concentration of quartz monzonite and pegmatite in the periphery of the granodiorite would seem to

<table>
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<th>Table 5</th>
<th>Average Estimated Modes of the Intrusive Rocks in the Bryant Pond Quadrangle, Maine.</th>
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<td>No. Averaged</td>
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<td>Hornblende</td>
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<td>Chlorite</td>
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<td>An Content of Plagioclase</td>
<td>36</td>
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(1) Songo granodiorite.
(2) Speckled Mountain quartz monzonite.
indicate that they were emplaced sometime after the granodiorite body.

The large body of pegmatite and quartz monzonite shown in the west and southwest parts of the Bryant Pond quadrangle is at least 40 per cent quartz monzonite. In that complex of intrusive rocks the quartz monzonite is commonly found grading into pegmatite by an increase in grain size. However, the two rock types commonly have sharp contacts in nearby outcrops and even cross cut each other in sharp dikes. It would seem that the two rock types formed at the same time and probably have a similar origin as late magmatic differentiates from the Songo granodiorite. Numerous schist inclusions are also present in this area of pegmatite and quartz monzonite. As mentioned earlier, the writer believes that this area represents the remnant of an aureole of pegmatite and quartz monzonite which once enclosed the Songo granodiorite.

**Pegmatites**

Pegmatites are exceedingly common throughout the Bryant Pond quadrangle but are especially abundant in the southern third, near the Songo granodiorite, and in the zone mapped as the pegmatite and quartz monzonite in the western part of the area. The pegmatites, which range in size from the 1 to 4 inch pegmatitic bands in the schist to bodies that underlie several square miles, form sills, discordant lenses, dikes, and irregularly shaped bodies. In the schists and gneisses the bodies are mostly concordant lenses of various dimensions and constitute much of the migmatitic banding of the gneiss. In many places the schistose parts become subordinate and are present only as wisps and blebs. In general the boundaries of the pegmatites in the schistose rocks are rather indistinct.

The migmatitic pegmatites in the calc-silicate rocks are restricted. In most cases the pegmatite bodies in the calc-silicate units are dikes or large irregular-shaped bodies with rather sharp contacts. It is a notable fact that the majority of the large pegmatite quarries are located in the calc-silicate units.

The pegmatite bodies located within the Songo granodiorite are either very irregular and diffuse in shape or occur as sharp cross-cutting dikes.

Most of the pegmatites are simple quartz-feldspar-mica rocks, but in the southern third of the quadrangle they are commonly more complex and contain many of the more exotic minerals, such as beryl, tourmaline, lepidolite, and apatite. In general, the pegmatites show little obvious internal structure and commonly they are gradational into quartz monzonite. On Plate 1, only areas with more than 75 per cent pegmatite are mapped as pegmatite. The stippled areas contain 50 to 75 per cent pegmatite and/or quartz monzonite.

**Binary Granite**

Binary granite underlies a small area in the western part of the town of Norway, just south of Pierce School. It is a medium-grained to coarse-grained equigranular massive binary granite. Muscovite and biotite are present in about equal amounts and constitute approximately 15 per cent of the rock. No thin sections of this rock type were studied because the exposures are so weathered.

Reconnaissance by the writer indicates that this binary granite underlies much of the Norway and Fryeburg quadrangles and is probably continuous with what Hanley (1939) has termed the Sebago batholith. It probably belongs to the New Hampshire magma series. If it is continuous with the granite at Sebago Lake, the body constitutes an immense batholith.

The part of the Sebago batholith described by Hanley (1939) in the Poland quadrangle is composed of binary granite and apparently is the same as the granite described above. The relation of this granite to the Songo granodiorite is at present unknown. Possibly the two rock types are just phases of one large body or the large body may consist of multiple intrusions.

**Post Metamorphic Dikes**

A total of sixty-five post metamorphic dikes has been observed in the Bryant Pond quadrangle. Fig. 2 shows the distribution and attitude of those dikes on which measurements could be obtained. The dikes, 6 inches to 18 inches wide, are generally diabasic in appearance and have dark-gray to brown-weathering surfaces. Commonly they are preferentially weathered, and in some cases on exposed hilltops furrows have formed that are
several inches deep and as wide as the dike. In most cases the
dikes have very sharp, well-defined contacts and are distinctly
-tabular. The majority of them strike northeasterly, dip steeply
and are thus essentially parallel to the major joint system in
the area.

No thin sections of the dikes were studied by the writer. How­
ever, Milton (1961) in the Old Speck Mountain quadrangle, Fisher
(1952) in the Bethel quadrangle, and Hanley (1939) in the Poland
quadrangle have done some work on the petrography of similar
dikes and found that most of them are diabase and basalt, with
a few camptonite, kersanite, and bostonite dikes also present.
Fowler-Billings (1944) has described similar dikes in the
Mount Washington area and concluded that they are co-magmatic
with the White Mountain magma series. As pointed out by
Milton (1961), recent radioactive ages for the White Mountain
magma series, e.g. Lyons et al. (1957) or Aldrich et al. (1958),
indicate that it is Triassic in age. Hence, if the dikes are co­
magmatic with the White Mountain magma series, then they
are probably Triassic in age also.

The recent ages determined for the White Mountain magma
series average near 186 million years old (Lyons et al., 1957),
which according to Marble (1950) would be Late Permian to
Early Triassic in age. However, by the more recent Holmes
(1960) time scale this figure would be Late Triassic in age,
which would suggest a correlation with the Late Triassic basals
associated with the Newark series in Connecticut and Massachu­
setts.

STRUCTURE

General Statement

The strata in the Bryant Pond quadrangle have been deformed
into complex tight folds, many of which have been overturned.
Intrusive rocks are common and may be divided into five groups,
four of which belong to the New Hampshire magma series. The
Songo granodiorite and binary granite form distinct large, mapp­
able bodies, whereas the pegmatites and Speckled Mountain
quartz monzonite are commonly intimately mixed with metasedi­
ments. In most cases the inclusions in pegmatites and Speckled
Mountain quartz monzonite have attitudes similar to those of
the metasediments enclosing the intrusive body. Hence one
 can map stratigraphy through these relatively small intrusive
bodies. The second group of intrusives consists of postmetamor­
phic trap dikes. Faulting has been important and is most evi­
denced by the large normal fault (Moll Ockett fault) shown on
Plate I.

Several related methods were employed to determine the geo­
logical structure in the Bryant Pond quadrangle. Outcrop pat-
terns of the various distinguishable stratigraphic units were plotted on a topographic base as field work proceeded. In conjunction with determining the map pattern the writer recorded and plotted strikes and dips of the bedding wherever possible. The various minor structures have been treated in a similar fashion. The minor structures observed include crinkles, mineral lineations, minor folds, boudins, and ripples or broad rolls of bedding planes. Plate III shows the strikes and dips plotted and their relation to the map pattern. As noted earlier, bedding and foliation are parallel; thus Plate III is also a plot of foliation. Plate IV presents a plot of the minor structures with the various stratigraphic units also shown. As will be noted later, some use has also been made of several methods of plotting the data on minor structures on a Schmidt equal area net.

Little use has been made of primary evidence for tops in the present work, mainly because such evidence is lacking. Either such evidence never existed in the strata or else it has been destroyed. The writer favors the second possibility. Graded bedding is the only primary evidence for tops that has been observed by the writer, but in only three localities were they "readable." Inasmuch as the beds are isoclinally folded at these localities, no true indications of tops could be obtained. Since cleavage is mainly a foliation parallel to bedding, it is impossible to use cleavage-bedding relations to determine the tops of stratigraphic units. To some extent the writer has relied upon stratigraphic sequences determined in the surrounding areas to tell which units are the younger ones in the section.

In general the writer has found it convenient to consider the areas north and south of the Moll Ockett fault as two distinct structural units. Each area displays its own style of folding. Each of these two structural units will be described and then compared and contrasted. The forces that caused the observed structures are discussed at the end of this section.

Structure South of the Moll Ockett Fault

General Statement.—The structural features south of the Moll Ockett fault have several distinctive general characteristics. All of the major folds trend about N. 30° W. and most measurements on bedding and minor structures have similar strikes. The distinct topographic lineament south of the Moll Ockett fault is also parallel to these structural trends. In large areas the bedding dips at low angles to the east. The major northwest-trending folds appear to be strongly overturned to the southwest.

The structure south of the Moll Ockett fault has been deduced largely by means of the Berry Ledge and Patch Mountain formations, both of which are excellent mappable units. In very few cases has the writer found it difficult to distinguish these two units from each other or from the other units in the southern sequence.

Minor Structures and Attitudes of Bedding.—Plate IV shows the types of minor structural features observed in the Bryant Pond quadrangle and the attitudes recorded. Crinkles, small folds (i.e. less than one foot wave length) and large folds (up to outcrop scale) are the most abundant structural features. Most of them have northwest strikes and variable plunges ranging from 5° to 35° to the north and south. Another prominent type of minor structure consists of broad open rolls on the bedding planes. Such “rolls” range from a few inches in wave length to 100 feet. The axial traces of most of them have the characteristic northwest or southeast strikes but a few strike northeasterly, approximately perpendicular to the usual trend. These cross structures are presumably later than the other minor structures and are the cause of the rather random fashion in which the majority of the minor structures plunge to the northwest and southeast.

As shown in Plate IV, many of the minor folds, on which the axial plane could be measured, are overturned toward the southwest. Of the 38 folds on which the axial plane could be measured only five have axial planes that are vertical or dip to the southwest. Probably a dip of 45° to the northeast is the most common attitude. Many of these folds are isoclinal or nearly so.

Minor structural features are particularly common in parts of the Berry Ledge formation, some folds being very complex and contorted (Photo 4). Possibly the thin marble bands were exceptionally plastic, thus making the unit relatively incompetent. The Noyes Mountain and Moody Brook formations display much more crinkling than the Patch Mountain or Berry Ledge formations but also have numerous distinct minor folds ranging up to 40 feet across. Photo 5 shows such a fold in the Noyes Mountain formation. Most of these folds are simpler in shape,
even if isoclinal, than the folds in the Berry Ledge and Patch Mountain formations.

Plate III shows a general plot of the attitudes of bedding measured in the southern and northern sequences. South of the Moll Ockett fault the strikes are mainly toward the northwest. East of a northwest-trending line through Rock Dundee the beds generally dip at low angles to the east. West of a similar line passing through Mud Pond the bedding generally dips eastward at moderate angles. Between these two lines the dips are commonly vertical or very steep to the northeast or southwest.

Major Structures.—The major structural features south of the Moll Ockett fault (Fig. 3) are the Norway anticlinorium, the Pierce School syncline, the Stearns Hill syncline, and the Moose Pond anticline. These structural features are not very well shown by the strikes and dips of the bedding because they are very tight to isoclinal folds. The validity of these structures rests largely upon the map pattern and the assumption that the stratigraphic sequence presented earlier is correct.

The Norway anticlinorium, the dominant structure, extends in an east-west direction across much of the southern border of the Bryant Pond quadrangle. The core of this anticline consists of the Patch Mountain formation, which is the lowest unit in the southern sequence. The map pattern of the Patch Mountain formation and the overlying units indicates that this anticline consists of several smaller folds, which include the Patch Mountain anticline, Nobles Corner syncline, and West Paris anticline. In the area between the Nobles Corner syncline and the West Paris anticline there is a rather poorly defined anticline and the Eastman Hill syncline. Fig. 3 shows the axial traces of these folds. Plate I shows cross sections FF' and EE' through the Norway anticlinorium and the several folds that constitute this major fold. The central portion of the Norway anticlinorium is more complex than the outer parts and is occupied by a zone of steep and variable dips. On the east and west margins of the anticlinorium the folds are tighter but simpler and are more severely overturned to the southwest.

The overturned nature of the major fold is derived from the fact that in the vast majority of cases the bedding dips to the
northeast and from the fact that many of the minor folds observed indicate overturning to the southwest. The map pattern indicates that the Norway anticlinorium plunges about 10° to 15° northwest. The minor structural features show variable plunges to the northwest and southeast.

Only the eastern half of the Pierce School syncline is shown on Plate I. The remainder of this fold has been cut out by intrusives belonging to the New Hampshire magma series.

The Stearns Hill syncline is a long narrow structural feature that lies immediately east of the Norway anticlinorium. It is defined by two parallel bands of the Berry Ledge formation and the fact that the central part consists of the Moody Brook formation. Cross sections EE' and FF' shows that this syncline is nearly recumbent. This interpretation is held inasmuch as both limbs dip at rather low angles to the northeast. The evidence from minor structural features also indicates that the fold is overturned, but the axial planes of the minor folds do not dip at angles as low as that shown on the cross sections for the major fold.

The Moose Pond anticline is east of the Stearns Hill syncline. It is shown as strongly overturned to the southwest for the same reasons advanced for the Stearns Hill syncline. The core is composed of the Noyes Mountain formation and is largely defined by the two bands of the Berry Ledge formation on each side of Moose Pond. The northern end of this anticline is cut off by the Moll Ockett fault. Reconnaissance by the writer and personal communication from Mr. J. Warner indicates that the Moose Pond anticline closes in the western part of the Buckfield quadrangle. Hence the structure appears to be plunging into the southeast. A subsidiary syncline about two miles north of Moose Pond seems to be plunging to the northwest where it is cut off by the Moll Ockett fault.

**Structure North of the Moll Ockett Fault**

General Statement.—The structure north of the Moll Ockett fault consists of large folds, some of which are open and some of which are isoclinal and slightly overturned toward the west. Attitudes measured on bedding are generally steep. The structure is deduced from outcrop pattern, strikes and dips, and minor structural features. The map pattern is mainly based upon units that are distinguished by the presence or absence of notable amounts of pyrrhotite. Those rocks with significant amounts of pyrrhotite weather to a distinctive rusty color.
Minor Structures and Attitudes of Bedding.—Plate IV shows the distribution and attitudes of the various minor structural features north of the Moll Ockett fault. The minor structural features are mainly crinkles, minor folds, and mineralogical lineations and are equally well-developed in all of the major units of the northern sequence. The mineralogical lineations are largely sills of sillimanite intergrown with biotite and in some cases muscovite. Inasmuch as these mineralogical lineations are parallel to the axes of the minor folds, they are probably b lineations also. The minor folds are commonly isoclinal, or nearly so. From Plate IV it is obvious that most of the minor structural features in the area north of the Moll Ockett fault fall into two general groups, one with plunges at moderate angles to the northwest and the other with plunges at moderate angles to the northeast. In general the minor structural features are closely related to the major folds and the writer has found it convenient to employ diagrams which emphasize the relations between the major and minor folds.

Plate III shows a plot of bedding attitudes observed in the structural features north of the Moll Ockett fault. Several folds are clearly outlined by the attitudes of bedding. In addition, it is clear that there is a distinct northwest structural trend near the Songo granodiorite but that to the east and northeast the structural trends swing into a distinct northeast orientation.

Major Structural Features.—Three major folds dominate the geology north of the Moll Ockett fault. These are the Chamberlain Mountain-Wilbur Mountain syncline, the Spruce Mountain anticline, and the South Woodstock syncline. In the northeast corner of the area there are three smaller folds: the Thurston Mountain synform, the Little Zircon Mountain synform, and the Spears Stream synform. Fig. 3 shows the locations of the axial traces of these folds. Sections AA', BB', CC'B', and DD' on Plate I depict some of the features of these folds.

Chamberlain Mountain-Wilbur Mountain syncline occupies a broad zone stretching from the northwest slopes of Spruce Mountain to the northwest corner of the Bryant Pond quadrangle. The southwest limb is subparallel with the contact of the Songo granodiorite but is also partly cut off by this intrusive. In the northern part of the township of Woodstock it is a rather broad open fold, but in the township of Bethel it is a rather tight fold overturned to the west.

The lowest unit involved in this fold is the Shagg Pond formation and the highest is the Howard Pond member of the Littleton formation. The southeast end of the Chamberlain Mountain-Wilbur Mountain syncline is clearly outlined by the outcrop pattern and the attitudes of the bedding (Plate III). The outcrop pattern indicates that the fold opens in a northwesterly direction. On the other hand, Fig. 4, which is a plot of perpendiculars to bedding planes, and Fig. 5, which is a plot of linear features, indicate that the plunge is more nearly to the north. The geometry of this fold may be the result of forces associated with the syntectonic intrusion of the Songo granodiorite. This will be considered further in a later section. Fig. 6 shows a plot of the lineations from the Wilbur Mountain section of the syncline, and it can be seen that the plunge is slightly west of north in that area. In the vicinity of Howard Mountain the open nature of the syncline and its north-northwesterly plunge are clearly indicated by the map pattern and attitudes of bedding.

The Spruce Mountain anticline is in the north-central portion of the quadrangle and occupies much of Spruce Mountain. This anticline is rather poorly defined due to rather poor exposures in the vicinity of the nose of the fold. The north-plunging nose is located about one-half mile south of the Milton School. Northward it merges with one of the indistinct smaller folds within the broad, shallow syncline that occupies much of the towns of Milton and Rumford. This anticline is defined largely by its outcrop pattern and the fact that it lies between two distinct, large synclines. The lowest unit exposed in the core is a small patch of the Shagg Pond formation exposed on the south slopes of Spruce Mountain. Cross sections CC'-B' and DD' illustrate the subsurface form inferred for this fold.

The South Woodstock syncline is a tight fold, the southern end of which begins about one-quarter of a mile north of South Woodstock and continues northward to the vicinity of Concord Pond where it broadens rapidly into an extensive, shallow synclinorium. In this synclinorium the folding is tight but with low amplitudes as indicated by the fact that the bedding attitudes are steep but only one unit is exposed. The syncline west of Moll Ockett Moun-
tain trends in a northerly direction to the vicinity of Concord Pond. North of Concord Pond no single axis defines the fold. Several poorly defined folds continue north-northwest for a few miles and then swing into the northeasterly trends which prevail throughout most of northwestern Maine. Cross sections CC'BB' and DD' show how this fold changes from a tight syncline into a more open synclinorium.

The center of the South Woodstock syncline is occupied by the Concord Pond member of the Littleton formation. The western limb consists of the Billings Hill and Shagg Pond formations and is partly cut off by the Songo granodiorite. The eastern limb consists of the Billings Hill, Shagg Pond and Thompson Mountain formations and appears to be somewhat overturned at least near the center of the fold. Mr. J. Warner (personal communication) has shown that the next major structure to the east in the Buckfield quadrangle is a broad anticline that plunges northward.

The South Woodstock syncline is very well defined at its southern end by bedding attitudes and outcrop pattern (Plates III and I). From Plate III it is seen that the bedding attitudes are very steep in this area and the minor structures (Plate IV) show that the fold plunges in a northerly direction. Fig. 7 shows a plot of the lineations in the vicinity of the southern end of the fold. It emphasizes the fact of a northerly plunge and demonstrates that the average plunge is about 25° N.

The Thurston Mountain, Little Zircon Mountain, and Spears Stream synforms are three small folds located in the northeast corner of the Bryant Pond quadrangle and are indicated mainly by the presence of the Billings Hill formation in the cores. From the map pattern these three northeasterly-trending folds appear to be subsidiary anticlines, within the South Woodstock syncline, that have sufficient amplitude to bring the Billings Hill formation above the present erosion surface. The noses at the northern ends of these folds are not exposed within the Bryant Pond area and presumably lie a few miles north in the Rumford quadrangle. Section BB' presents the internal structure of these folds.

As seen on Plate IV, the lineations near the southern extremities of the cores of these three folds plunge to the northeast. This is inconsistent with an anticlinal interpretation deduced from the map pattern alone. Moreover, Fig. 8, which is a plot
of the intersections of bedding planes for the southern end of the Little Zircon anticline, also indicates a northeasterly plunge for the folds. These three areas of rusty-weathering rocks could be interpreted as anticlinal cores that rose to such heights that the plunge was inverted. A second interpretation is that the three patches of rusty-weathering gneiss are synclines superimposed on the inverted limb of a large recumbent fold of which the Billings Hill formation is the core (Fig. 9). This would result in structural features which would be consistent with the plunges of the minor structures. To conclusively prove this hypothesis, much further detailed field work would be required in the Bryant Pond and Rumford quadrangles, and the work in progress in the Buckfield and Dixfield quadrangles would have to be completed. Nonetheless, in order to have an interpretation consistent with the available data, the writer prefers this second hypothesis, mainly on the grounds of simplicity. However, as a third hypothesis it is also possible that the rusty-weathering gneisses at Thurston Mountain and Little Zircon Mountain are merely large lenses within the Concord Pond member. This would eliminate both of the preceding hypotheses.
IGNEOUS ROCKS

General Statement.—The shapes, sizes and distributions of the intrusive bodies in the Bryant Pond area have been described in a preceding section. The Songo granodiorite will be emphasized the most inasmuch as it is the most important for an understanding of the structure of the Bryant Pond quadrangle.

Songo Granodiorite.—The Songo granodiorite is a rather coarse-grained rock that is massive to poorly foliated away from the main contacts. Moreover, the foliation has no consistent trends from outcrop to outcrop. The foliation is due to alignment of biotite flakes and some concentration of the flakes in layers in an otherwise homogeneous, equigranular rock. Near the borders of the body, however, the foliation is extremely well developed and along the northeastern border of the pluton it is the most prominent feature of the rock. As seen in Plate III, the foliation along the northeastern border of the body generally parallels the contacts and is subparallel with the bedding in the adjacent metamorphic rocks. Along the south-central border of the Songo granodiorite, where it is in fault contact with the southern sequence, the foliation tends to dip inward from the contacts of the body and is paralleled to a minor extent by the bedding in the metasediments immediately south of the Moll Ockett fault. This has probably been caused by the movement along the fault.

Inclusions are rare in the interior of the pluton but common along the borders. Along the borders the inclusions are parallel to the foliation. In many cases the inclusions are simple tapering slabs that may be up to a few feet thick and 30 feet long. In some exposures, however, as described on p. 50, the granodiorite and metasediments may be intimately interlayered on a 1 to 2 inch scale over a thickness of 30 feet or more.

In many places the contact between the metasediments and granodiorite is obscured by a thick zone of Speckled Mountain quartz monzonite and pegmatite. This is particularly common southwest of East Bethel, south of Days Ridge, north of Moody Mountain and near Curtis Hill. Fisher (1952, p. 89) notes a similar zone between the granodiorite and metasediments in the Bethel quadrangle. The writer has noted in an earlier section that the large amounts of pegmatite and quartz monzonite in the metasediments immediately surrounding the granodiorite body form a modified aureole. There seems to be no contact metamorphism of the metasediments enclosing the pluton.

The Songo granodiorite is a locally conformable but regionally disconformable intrusive body (Plate III). Fisher (1952, 1962) noted the same relationship in the Bethel quadrangle. In general the contacts are sharp, especially where the zone of Speckled Mountain quartz monzonite and pegmatite is absent. As shown on Plate III, the bedding in the metasediments along the northeastern portion of the granodiorite is conformable to the border of the pluton as much as a mile away from the contact. Yet in the area immediately north of Curtis Hill the bedding in the metasediments appears to be cut off abruptly by the intrusive body.

From the preceding facts, the writer infers that the Songo granodiorite is a dome-shaped body, except where cut out by the later Moll Ockett fault and in the vicinity of Curtis Hill. Along part of the southern border of the body the contact is believed to dip steeply to the north. The exact relations are largely obscured by the Moll Ockett fault. The small patches of Songo granodiorite south of the fault are probably slices of the pluton which were carried up from depth (Fig. 10). Fisher
(1952, 1962) also believes the granodiorite body to be domical in shape. Moreover, the writer believes that it was forcibly intruded and possibly moved northeastward at the same time that it moved upward.

Another notable structural feature of the Songo granodiorite is the presence of exfoliation domes up to one-half mile or more in diameter. The best example is displayed on Bucks Ledge and can be clearly seen on the east side of North Pond from Maine State Route 25. Other exfoliation domes are present on Goss Ledge, Walkers Mountain and the two hills immediately north of it.

Speckled Mountain Quartz Monzonite.—The Speckled Mountain quartz monzonite has been described in a previous section. Structurally it is a rather variable rock type in that it occurs as sills, dikes, and irregular bodies. Probably sills are the most common structural form in which this rock type occurs. Most bodies of the quartz monzonite are probably on the order of a few hundred feet long and 50 to 100 feet thick; but some are much larger and Plate I shows a large body near Rumford Corner that has been mapped separately. Many areas of metamorphic rocks are so highly intruded by this rock type that the metasediments are merely inclusions. The fact that the inclusions maintain attitudes consistent with the metasediments outside of the highly intruded areas would seem to indicate that the quartz monzonite came in as a series of small intrusions rather than one or two relatively large pulses. As mentioned earlier, some of the migmatitic banding in the gneissose units is a result of innumerable 1 to 4 inch bands of quartz monzonite along with pegmatite bands.

Foliation is well developed in the Speckled Mountain quartz monzonite and is usually parallel to the contacts of the body. The well-developed foliation and commonly intricate shapes of the quartz monzonite bodies would seem to indicate that it was very fluid at the time of intrusion.

Pegmatites.—The pegmatite bodies have been described previously on p. 70. Many workers have published papers on the pegmatites in Maine and have presented much detailed structural information on a number of the pegmatite bodies in the Bryant Pond quadrangle. Among the more important works are those by Bastin (1911), Cameron et al. (1945), Cummings (1955), and Landes (1925).

Post-Metamorphic Dikes.—A total of 65 post-metamorphic basalt dikes have been recorded by the writer in the Bryant Pond quadrangle. The dikes are unmetamorphosed and cut all of the
other rock types in the area at high angles. Figure 2 shows that most of the dikes lie to the south of the Moll Ockett fault. Most of them occur as single dikes but in a few areas as many as six parallel dikes up to 2 feet thick are present in a zone 20 to 30 feet thick.

The dikes have very sharp contacts with the country rock, and some of the larger dikes have thin, chill borders. Only a very few post-metamorphic sills have been observed and may be only offsets of the dikes which parallel the bedding in the enclosing rock. In the vast majority of cases the dikes are 8 inches to 18 inches thick. However, a few dikes up to 30 feet thick were found on the south side of the hills immediately south of Redding. A common feature in the thin dikes is the presence of numerous joints perpendicular to the contacts. These joints are probably tension fractures formed when the dike cooled.

Fig. 2 shows that most of the dikes trend in a northeasterly direction. Fig. 11, which is a plot of the poles perpendicular to the dikes, emphasizes this same trend. It will be seen in a following section that the major joint pattern in the area has the same orientation. Hence it would seem that the intrusion of the dikes is related to planes of weakness within the country rock. Possibly the dikes and joints are genetically related. This point will be considered further after consideration of the joints and faults present in the Bryant Pond quadrangle.

**FAULTS**

One major and one minor fault are shown on Plate I. The Moll Ockett fault lies at the base of the two prominent spurs of Moll Ockett Mountain. This fault is located in the south-central part of the Bryant Pond quadrangle and strikes approximately N. 55° E. It lies at a conspicuous topographic break between the more hilly, irregular topography on the north and the more gentle northwest oriented hills on the south side of the fault. The Moll Ockett fault has considerable stratigraphic significance inasmuch as it brings the southern sequence of Ordovician to Silurian age in contact with the northern sequence of Silurian to Lower Devonian age. However, the fault also prevents the determination of the true stratigraphic relations between the two sequences in the Bryant Pond quadrangle.

The Moll Ockett fault is indicated by several lines of evidence.
1) The sharp topographic break and change in pattern of topography between the area north of the fault and that to the south (Plate I). North of the fault the topography has no particular trends or relation to the structure. South of the fault the topography has a distinct north-westerly trend that reflects the underlying structural trends. 2) There is a sharp stratigraphic and lithologic break between the areas on each side of the fault. This is reflected by the topography and map patterns shown on Plate I. In no case has the writer been able to find float of the distinctive calc-silicate rocks of the southern sequence on the north side of the Moll Ockett fault. 3) There is a break in the type of folding and structural pattern. North of the fault the strata strike northwest to northeast and dip at angles greater than 45° and commonly at angles of 70° to vertical. To the south the bedding usually strikes in a northwesterly direction and dips at angles less than 35° and in large areas less than 20°. This comparison applies particularly to the zone within one mile north and south of the fault. Moreover the folds north of the fault have steep axial planes whereas many of the folds south of the fault are strongly overturned toward the west. 4) Breccia is found in several places along the fault zone (Plate I). At the eastern-most outcrop the breccia is marked by a silicified zone 20 feet thick of brecciated pegmatite and quartz monzonite with numerous cavities filled with drusy quartz. Some pyrite is present also. 5) Retrograde effects are common in a zone about one-half to one mile wide along the whole length of the fault zone. These effects are indicated mainly by chlorite replacing biotite and the feldspars altering to sericite. In some instances the biotite schists have been changed into chlorite schists. As noted earlier, retrograde effects are common throughout the southern structural unit, but only along faults have true chlorite schists been formed.

The Moll Ockett fault appears to be a high-angle normal fault down-dropped to the north. The high-angle nature is attested by the fact that the fault has a relatively straight trace. Topography seems to have very little effect on the fault trace. Moreover, at the breccia zone located one-half mile south of B.M. 577, one can observe slickensides along planes which dip 60° to 80 N. Also, it can be seen on Plate III that the foliation in the meta-sediments just south of the above mention breccia locality has
been dragged into near parallelism with the fault and dips steeply to the northwest. Hence it would seem that the fault zone dips very steeply to the northwest.

Two lines of evidence support the conclusions that it is a normal fault downthrown on the north side. First, it is a steep-dipping structure and is paralleled by the major joint system of the area. It is to be noted that jointing is most prominent in the area south of the fault and is accompanied by several small, parallel faults. These features taken together would seem to indicate that decreased horizontal compressive stresses controlled the formation of the faults and joints. Thus it is probably more reasonable to think in terms of a normal fault when considering the Moll Ockett fault. Second, the southern sequence is older than the northern sequence and presumably was below it at one time. Hence, with the steep, northwest-dipping fault zone, it is most reasonable to imagine the older southern sequence being moved upward as the southern block rose and the younger northern sequence moved downward as the northern block moved downward. No estimates can be made from the evidence in the Bryant Pond quadrangle on the total displacement that has taken place on the Moll Ockett fault. It is interesting to note that a similar big northeast-trending fault has been mapped by Billings et al. (1946) in the Mount Washington area and by Moench (1963) in the Phillips quadrangle. The Moll Ockett fault will be related to joints and post-metamorphic dikes in a later section. Suffice it to say that it is definitely post-metamorphic in age as it cuts and causes retrograde metamorphism of all of the rock types except the post-metamorphic dikes.

One small fault also is shown on Plate I in the vicinity of Andrews Brook about one mile north of Trap Corner. It is suggested by breccia and a zone about 30 feet wide in which jointing is very close-spaced. Commonly the joints in this fault zone are very well defined with a one-half to one inch spacing. The Berry Ledge formation has been offset a small amount in a direction indicating that the south side moved upward. Several other very minor fault zones were observed in the Bryant Pond quadrangle, but inasmuch as no significant displacement could be determined, they are not shown on Plate I.

One major difficulty arises with the preceding interpretation of normal faults with the major downthrow on the north side.

The Songo granodiorite has been described as a broad dome that dips out under the metasediments. Hence, one might expect that upthrow of the area south of the Moll Ockett fault would bring much more of the Songo granodiorite to the surface. This does not seem to have happened. One possible explanation, which was suggested on p. 71, is that the Songo granodiorite may not have been dome-shaped along its southern margin. The contact may have dipped toward the north instead of toward the southeast out under the metasediments.

**Joints**

A total of 107 joints have been measured in the Bryant Pond area. They have been observed and measured in all of the rocks in the area. Most of the measured joints are distinct and sharp, and in many places they are in groups with a 6 to 12 inch spacing between each joint. In the vicinity of the Moll Ockett fault and throughout the southern structural unit, it is common to find innumerable very thin seams or joints that have no visible separation. They are made evident by the fact that the country rock is strongly chloritized for one-half to one inch on each side of the seam. These thin seams are parallel to the more obvious joints along which there is usually much chloritization. It is a notable feature that the modes of the units in the southern sequence contain considerably more chlorite than the modes of the northern sequence. Similar thin joints and seams in the Songo granodiorite are notable in that epidote has been introduced or formed along them and in the immediately adjacent granodiorite. Thus on well-exposed surfaces they form long rib-like bands up to one and one-half inches wide which look like great welts stretching across the outcrop.

Fig. 12 is a plot of poles perpendicular to joint surfaces and shows that the majority of the joints have a distinct northeast trend and steep dips. Two other minor steep-dipping joint sets strike east-west and north-south. Fig. 13 is a similar plot for the joints south of the Moll Ockett fault and emphasizes the smaller degree of variation of the attitudes of jointing in that area as compared to the whole quadrangle. Comparison with the attitudes of the post-metamorphic dikes plotted on Fig. 11 and the faults plotted on Plate I demonstrates that all three features are essentially parallel to each other. It is interesting to
note that Hanley (1939) in the Poland quadrangle, L. W. Fisher (1942) in the Lewiston area, and I. S. Fisher (1952) in the Bethel quadrangle have found that the major joint set strikes north-easterly and dips steeply, and that the vast majority of post-metamorphic dikes in those areas have this same orientation.

Structural Synthesis

Relationship of the Structural Units North and South of the Moll Ockett fault:

From the preceding descriptive treatment of the structure it is apparent that there are two distinct structural units within the Bryant Pond quadrangle, one on the north side of the Moll Ockett fault and the other on the south side. Indeed, this is some of the most convincing evidence for the existence of the Moll Ockett fault. Several similarities and several marked differences between the two units are to be noted.

The similarities are as follows: 1) In both units, bedding has a distinct north-westerly trend, at least in part. (2) The major folds plunge toward the north. 3) Both units are in the same metamorphic grade and are cut by the same intrusive rocks. 4) Very tight to isoclinal minor folds are common in both areas. 5) The axial planes of folds in both areas are overturned toward the Songo granodiorite, but to a much lesser degree in the northern unit.

The major contrasts between the two structural units are the following: 1) In the southern unit all of the structural trends are in a north-westerly direction whereas the north-westerly structures in the northern unit show a distinct swing into northeasterly trends as the distance from the Songo granodiorite increases. 2) Most of the folds in the southern unit are very strongly overturned toward the southwest but the folds north of the Moll Ockett fault are only slightly overturned toward the west and southwest. 3) Large areas in the south have low easterly dips due to the overturning of folds. In the northern unit low dips are rare. 4) Different groups of strata are involved in the folds of each area. 5) The minor structural features in the northern structural unit generally strike and plunge in a northwest to northeast direction at moderate angles. As a result of later broad cross folds, the minor structural features in the south
plunge northwest and southeast at low to moderate angles. Apparently the later broad cross folds are absent in the northern unit. 6) Joints are better developed in the south and the number of post-metamorphic dikes is greater.

Time Relations Between the Two Structural Units:

Several lines of evidence indicate that the structural features north and south of the Moll Ockett fault formed at the same time and under the same general stresses. In the first place, all of the strata are at the same grade of metamorphism and are cut by the same intrusive rocks. Secondly, both structural units have, at least in part, the same northwesterly trends. A third point of evidence is derived from a consideration of the regional geology. In most of the surrounding areas the northern sequence and southern sequence are folded into northeast-trending folds. Only in a few restricted areas are the folds oriented in a north-westerly direction.

The fact that the structural pattern is much different on each side of the Moll Ockett fault is probably best explained by differences in lithology. The southern sequence contains much calc-silicate rock which is interbedded with some impure marble and has probably been relatively incompetent with respect to external forces. This may explain the more intense deformation and overturning in the southern structural unit.

Spatial Relations of the Structure in the Metamorphic Rocks to the Songo Granodiorite and Sebago Granite:

As indicated above, the regional structure throughout most of western Maine and northern New Hampshire displays a distinct northeasterly trend. However, in parts of the Poland, Buckfield, and Bryant Pond quadrangles the map patterns show a distinct swing into a northwest orientation. Moreover, these northwest-trending structural features are obviously subparallel to the eastern and northeastern borders of the Sebago granite and Songo granodiorite and within a few miles of the plutons. Hanley (1939) has found that the northwest-trending folds in the Poland quadrangle are intensely deformed and overturned toward the southwest. Mr. J. Warner (personal communication) believes that the same situation probably exists in the western part of the Buckfield quadrangle south of the Moll Ockett fault.

On the western side of the Songo granodiorite the major folds appear to strike consistently toward the northeast. However, some very broad northwest-trending open folds are also present. Billings (personal communication) has observed some small northwest-trending folds in the Gorham quadrangle and Milton (1961) has described similar folds near the southern border of the Old Speck Mountain quadrangle. It is also evident from a consideration of the mapping by Billings (personal communication) and Milton (1961) that just southwest of the mutual corner of the Milan, Old Speck Mountain, Bethel, and Gorham quadrangles a broad northwest-trending culmination is at right angles to a large northeast-trending syncline. This is shown by the fact that the Boott member crosses the Mahoosuc Range.

Time and Sequence of Folding:

The youngest strata involved in folding in western Maine and north-central New Hampshire belong to the Lower Devonian Littleton formation. These strata have been extensively intruded by the Middle to Late Devonian New Hampshire magma series. Hence the major orogeny which has affected the rocks of the Bryant Pond area is probably Middle to Late Devonian in age. This would correspond to the well recognized Acadian Orogeny.

The folding probably began before the intrusions of New Hampshire magma started. This is shown by the fact that some of the folds, such as in the vicinity of Curtis Hill, are sharply truncated by relatively massive igneous bodies. This is even more striking in nearby areas of lower metamorphic grade, such as the area just north of Augusta (Barker, 1961). However, in the Bryant Pond area folding probably continued for some time after the main intrusions. This interpretation is supported by two facts: 1) The folds are overturned toward the major intrusives and more intensely deformed in the same direction. This is particularly true in the southern structural unit and would seem to imply that the plutons were already emplaced at the time the folds were forming. 2) Some of the pegmatite and quartz monzonite sills have been folded and boudined. Most of the larger pegmatites and all of the Speckled Mountain quartz monzonite are comagmatic with the Songo granodiorite and Sebago granite.
On the other hand, two facts indicate that at least in the Bryant Pond area the intrusive aspects of the Acadian Orgeny outlasted the folding. First, there are dikes of pegmatite and quartz monzonite that are totally unaffected by deformation. These dikes have very sharp contacts and are texturally massive. This would not be expected if they had been subjected to deformative stresses. The second fact is that the texture of the internal parts of the Songo granodiorite shows no evidence of cataclasis or strain.

Some idea of the time relations between the northeast and northwest-trending folds can be obtained from the Bryant Pond quadrangle. In general it would seem that the northeast-trending folds in the northeast corner of the Bryant Pond quadrangle and elsewhere in western Maine were formed somewhat before the northwest-trending structures. Evidence for this view can be seen on Plate IV, which shows the minor structures observed in the Bryant Pond quadrangle. In the vicinity of the Chamberlain Mountain-Wilbur Mountain syncline it can be seen that the minor structures (which are mainly b lineations) plunge into the northwest, but some also plunge into the north and northeast. Moreover, as the distance away from the Songo granodiorite increases, all of the lineations rapidly swing into a constant northeasterly plunge with little or no evidence of any remaining northwest-plunging lineations. This is strong evidence that the northeast-plunging lineations distinctly pre-date the northwest-plunging lineations and likewise with the respective fold structures. In addition, if the conclusion that the northeast-trending structures pre-date the major intrusive action is correct, then the northwest-trending folds must be somewhat younger folds. This follows inasmuch as the clear cut spatial and structural relations (e.g. overturning toward the major plutons) between the plutons and northwest-trending folds indicates that they must be intimately related in time. Indeed, this fact can be used to support the idea that the northeast-trending folds predate the major intrusive activity.

Nature of Fold-Producing Stresses

Northeast-Trending Structural Features:

A consideration of the regional geology shows that the folds have a distinct northeast trend throughout much of northwest-

ern Maine and northern New Hampshire. The same northeasterly trends prevail throughout the state of New Hampshire as can be seen on the geologic map of that state (Billings, 1956). A comprehensive stress directed in a southeast-northwest direction is the most obvious and simple interpretation to explain the northeast-trending structures in the Bryant Pond area and the surrounding parts of New England.

In some cases it is possible to determine the direction in which the active compressive stress was oriented. Billings (1954, p. 226) states, “In most cases the active force, analogous to a moving piston, operates on one side of the folded belt, whereas a resisting force is induced by a stationary block on the other side.”

In order to determine the direction of the active forces which caused the folds under consideration, it is necessary to make the assumption that there has been only one major period of folding. Thus the northwest-trending folds would be formed sometime during the period in which the northeast-trending folds were forming in the surrounding areas.

Two lines of evidence in the area immediately around the Songo granodiorite and Sebago granite indicate that at least in the Bryant Pond quadrangle the compressive stress was actively directed from the southeast to northwest. In the first place, the folds on the east side of the Sebago granite and Songo granodiorite, although trending toward the northwest, are strongly overturned toward the west. This strongly suggests that the stress was directed toward the west. The northwest trend of the folds will be explained below. Secondly, in the Bethel area Fisher (1952, 1962) finds no evidence that the northeast-trending folds are overturned toward the southeast, i.e. in the direction of the Songo granodiorite which was emplaced before all of the fold-producing stresses had ceased. This would imply that after the formation of the northeast-trending folds in the Bethel quadrangle, they were not subject to any active compressive force from the northwest which would have overturned the folds toward the Songo granodiorite.

Northwest-Trending Folds:

Three different models can be proposed and considered in an attempt to explain the northwest-trending folds which, along the eastern borders of the Sebago granite and Songo granodiorite, were superimposed on the somewhat earlier northeast-trending
structures. The three models are as follows: 1) The plutonic bodies are much younger than the northeast-trending folds and have caused the northwest-trending folds. 2) The intrusions are older than the folding and the northwest-trending folds were caused by the same forces that caused the northeast-trending folds. 3) The intrusions and folding are syntectonic.

In the light of the evidence presented on the preceding pages the writer favors the third model. The major regional stress can be envisioned as oriented from southeast to northwest with the active stress being applied from the southeast. At some time after this stress was in operation and the northeast-trending folds were formed, it can be assumed that the large plutonic bodies started to intrude upward and possibly northeastward into the metasedimentary strata. It can then be postulated that the regional southeast-to-northwest stress pattern would be locally deflected as a result of the new stresses resulting from the intruding plutonic bodies. The deflected forces could be oriented more nearly toward the west or slightly south of west and cause the northwest-trending folds which are overturned toward the large intrusive bodies.

Post-Metamorphic Structural Synthesis.

The post-metamorphic dikes, joints, and faults found in the Bryant Pond quadrangle have been described in preceding sections. It was pointed out that all three of these post-metamorphic structural features are steeply dipping and have northerly trends. Inasmuch as joints and normal faults probably indicate at least local stretching of the Earth's crust, it is reasonable, in the light of the sand-box experiments by Hubert (1951), to consider those in the Bryant Pond area to be the result of a decrease in the horizontal compressive stresses or forces. In the present case, one can postulate that the greatest principal stress axis was vertical and the least principal stress axis horizontal in a northerly direction. Hence it would seem reasonable to consider them to be genetically related in time and cause. The dikes could also be genetically related but no evidence is immediately obvious in the Bryant Pond quadrangle. It is possible that the dikes were intruded at a considerably later time along the major joint set which is an obvious plane of weakness.

It was noted earlier that in the surrounding parts of Maine the major joint set is steeply dipping and strikes northerly and is paralleled by the vast majority of the post-metamorphic dikes. Fowler-Billings (1944, p. 1262) has emphasized the fact that northeast trends and steep dips are the dominant attitudes for post-metamorphic dikes in much of northern New Hampshire and notes also that this is the general attitude of the large, post-metamorphic, normal fault (Pine Mountain fault) in the Mount Washington area. It is notable that in the areas of northwest-trending folds the joints and dikes are almost perpendicular to the fold axes. Hence they are unrelated to the fold trends.

Thus it becomes evident that at some time after the Acadian Orogeny much of northwestern Maine and northern New Hampshire may have been subjected to a decrease in the northwest-southeast compressive forces with a concomitant development of northeast-trending joints and a few parallel normal faults. Possibly a regional doming may have resulted in a vertical orientation for the greatest principal stress axis. Post-metamorphic trap dikes were intruded along some of these tension fractures either as they formed or some time after. Fowler-Billings (1944) has shown these dikes to be co-magmatic with the White Mountain magma series, which, as noted earlier, has within recent years been dated as Late Triassic by radioactive dating methods. Inasmuch as there is ample evidence of Late Triassic normal faults associated with trap rock intrusion in southern New England and Nova Scotia, the writer believes that this same relationship should be considered as a possibility to explain the origin of the faults, joints, and post-metamorphic dikes in the Bryant Pond quadrangle.

One other interesting aspect of the postulated stress distribution is that they could possibly explain the broad, low amplitude cross folds which are present within the southern structural unit. If these cross warps formed at a time well after the main period of folding, they would explain why the minor structures south of the Moll Ockett fault do not have plunges that are consistent with the plunge directions of the major folds as indicated by the outcrop patterns.

**PETROLOGY OF THE METAMORPHIC ROCKS**

**General Statement**

The petrology of the metamorphosed strata in the Bryant Pond quadrangle can be considered in several separate cate-
of the pelitic rocks and the origin of the migmatitic banding, sillimanite-microcline-muscovite-plagioclase (with a composi-
tion near An$_{20}$ to An$_{30}$) from those in which these two minerals in the four-phase assemblage sillimanite microcline-muscovite-plagioclase cannot co-exist with a plagioclase more calcic than about An$_{20}$. As can be seen on Plate 1, this isograd cuts formational boundaries at high angles.

In order to make meaningful deductions from an observed mineral assemblage, it is necessary to consider whether or not the assemblage represents equilibrium. Several criteria have been used by the writer to decide whether or not a particular mineral is in equilibrium with the surrounding mineral grains. (1) If the grain boundaries are sharp and distinct, a mineral is considered to be in equilibrium with the surrounding grains. (2) In an equilibrium assemblage the grains of a particular mineral should all be approximately the same size if they are surrounded by similar grains. (3) The number of minerals present should satisfy the mineralogical phase rule. In the light of such criteria the writer believes that the assemblages listed in the following sections reached chemical equilibrium during metamorphism.

Mineralogy

General Statement.—From the viewpoint of metamorphic petrology the mineralogical variation within the metamorphosed strata in the Bryant Pond quadrangle is restricted to a relatively small number of significant minerals, because all the sedimentary units have been subjected to essentially the same degree of metamorphism. The significant minerals in the mica schists are biotite, muscovite, plagioclase, quartz, sillimanite, garnet and microcline. In the calc-silicate rocks the significant minerals are plagioclase, diopside, calcite, scapolite, microcline, hornblende, tremolite, garnet, biotite, quartz, and sphene. Minerals resulting from retrograde metamorphism include chlorite, calcite, sericite and zoisite.

Plagioclase.—Various members of the plagioclase series are abundant in the metasediments throughout the Bryant Pond quadrangle. In the calc-silicate rocks the plagioclase occurs as well-twinned (albite-law) equant grains in the groundmass. The An content ranges from about An$_{40}$ to An$_{50}$. In the schist and granulite facies of the various gneiss units the plagioclase is much less calcic. In the quartz-feldspar-biotite schist and granulite the plagioclase ranges from An$_{25}$ to An$_{30}$, while in the quartz-feldspar-two mica-sillimanite schists the plagioclase ranges from An$_{15}$ to An$_{25}$, but in most specimens the composition is between An$_{25}$ and An$_{30}$. The plagioclase in the pegmatitic quartz-feldspar bands of the gneiss units is generally oligoclase.

Garnets.—Garnets are common but not abundant in most of the metamorphosed strata in the area. In the micaceous gneiss they commonly occur as distinct red to reddish-purple anhedral, round grains 1 to 2 mm in diameter. The garnets in some of the migmatic quartz-feldspar-biotite schists and granulites are similar to those in the mica gneisses, but in most they are very irregular, anhedral, highly poikilitic grains. Garnets in the calc-silicate rocks are also irregular, very poikilitic and some are slightly birefringent when viewed between crossed nicols. Garnets are rare in the pyrrhotite-rich gneisses. None of the specimens listed in the modes of the Billings Hill formation have garnets and garnet is rare in the Wilbur Mountain member of the Littleton formation. Heald (1948) in the Lovewell Mountain quadrangle of New Hampshire and Fisher (1952, 1962) in the Bethel quadrangle report similar sulfide rich gneisses in the same grade of metamorphism which contain no garnets. Heald (1948, p. 38) suggests, "Apparently insufficient iron was available for garnet after the formation of pyrrite in these gneisses."

The writer has determined the refractive indices of 17 garnets from the migmatitic quartz-feldspar-two mica-sillimanite gneiss, and calc-silicate rocks. The refractive indices of the garnets from the calc-silicate rocks range from 1.747 to 1.794 and are noticeably lower than those in the pelitic rocks which range from 1.800 to 1.813. This difference is expectable inasmuch as the garnets in the calc-silicates are probably composed mainly of grossularite. The garnets in the pelitic rocks are most likely...
almandite but may contain minor amounts of the Ca$_3$Al$_2$(SiO$_4$)$_3$ component. Moreover, Fisher (1952, 1962) states that the garnets in the gneisses of the Bethel quadrangle have a refractive index near 1.804 and a specific gravity of 4.15 and believes that only minor amounts of Ca$_3$Al$_2$(SiO$_4$)$_3$ and Mn$_3$Al$_2$(SiO$_4$)$_3$ are present in them.

Heald (1948, p. 38) has determined the following data for a garnet from the Kinsman quartz monzomite:

- **Refractive index** = 1.803
- **Specific gravity** = 4.10
- **Size of unit cell**
  
  \[ a = 11.50 \text{ A} \]

The chemical formula of the garnet is:

\[(\text{Fe}^{2+}\cdot\text{Mg}^{2+}\cdot\text{Ca}^{2+}\cdot\text{Mn}^{2+})_3\text{Al}_2(\text{SiO}_4)_3\]

The molecular composition is:

- **Almandite** 81%
- **Pyrope** 16%
- **Spessartite** 2%
- **Grossularite** 1%

Heald also found that the refractive indices of the garnets in the metamorphic rocks ranged from 1.801 to 1.808 and averaged at 1.805. Moreover, the garnet from one biotite gneiss had a specific gravity of 4.12 and the length of the cell edge was 11.50 A. Hence he concluded that the garnets in the metamorphic rocks are similar in composition with those in the Kinsman quartz monzomite.

The garnet compositions suggested by Heald (1948) and Fisher (1952) are of interest in that both workers were considering rocks similar in composition to the gneisses of the Bryant Pond area and at about the same grade of metamorphism. Thus the writer believes that only minor amounts of grossularite and spessartite are present in the garnets from the migmatitic gneisses of the Bryant Pond quadrangle.

Iron Sulfides.—Pyrrhotite is the common sulfide in the metamorphosed strata throughout the Bryant Pond quadrangle. It appears as small anhedral blebs, wisps, and thin irregular stringers. In some specimens it can be seen with the naked eye. It is light yellow and moderately magnetic.

Pyrite is restricted to shear zones and is associated with much retrograde chlorite, sericite, and in some cases calcite. Commonly it forms distinct euhedral, striated cubes 1 to 3 mm in diameter.

Fisher (1952) reports a similar occurrence of pyrrhotite in some of the gneisses of the Bethel quadrangle. Moreover, he mentions (p. 76), “These grains are cut by a network of tiny veinlets of another yellow metallic mineral, probably pyrite.” However, Fisher suggests that these pyrite veinlets may be of retrograde origin.

According to Fisher, (1962, p. 1400):

“...The evidence for a sedimentary origin for the components of pyrrhotite is: (1) the pyrrhotitic gneisses represent a stratigraphic unit; (2) the pyrrhotite is well disseminated through the pyrrhotitic gneiss and shows no systematic concentrations; (3) the iron content of the pyrrhotitic gneiss is similar to that of the other metasedimentary rocks; (4) the reducing conditions of deposition favoring the preservation of sulfides would also favor preservation of organic matter, and the pyrrhotitic gneiss is the only rock in the quadrangle in which graphite can commonly be recognized; (5) silicified fault zones are the only evidence of hydrothermal activity, and none of the silicification is spatially related to the pyrrhotitic gneiss; and (6) pyrrhotite is known to occur in metamorphosed argillaceous sediments, especially those of high carbon content.”

Reasons 1, 2, and 5 certainly hold true in the Bryant Pond quadrangle and thus the writer is in agreement with Fisher’s view on a sedimentary origin of the iron and sulfur components of the pyrrhotite.

Amphiboles.—Common hornblende is present in many of the calc-silicate rocks of the Bryant Pond quadrangle. In most rocks it is associated with diopside and/or biotite. Tremolite is abundant only in the quartz-rich calc-silicate rocks of the Howard Pond member of the Littleton formation, where it forms distinct euhedral laths about 3 to 4 mm long and 1 mm thick. Commonly
it is associated with diopside, calcic plagioclase, quartz, phlogopitic biotite and in some specimens garnet. Specimen (3) 6/21/59 in a rock containing quartz, diopside and plagioclase (An77) has the following properties:

\[
\begin{align*}
1.625 < n_x < 1.628 & \quad \text{optically negative} \\
1.638 & \approx n_z \\
\end{align*}
\]

z = light yellow-green
x = light yellow-brown
y = light yellow to colorless

From the tables in Winchell (1951, Fig. 323, p. 433) this amphibole is tremolite with 15 to 20 per cent of the CaFe2Si2O6(OH)2 end member.

Common hornblende is the amphibole in most of the calc-silicate rocks of the northern and southern sequence of meta-sediments and in the calc-silicate nodules of the northern sequence. The hornblende, especially in the southern sequence, commonly appears as rather shabby irregular grains, which in a few specimens appear to be altering to diopside. Below are listed the optical properties of two hornblendes. The optics of the co-existing diopsides will be presented in the following section.

(17) 9/20/61: co-existing with quartz, diopside, scapolite, and calcite; from the Shagg Pond formation.

\[
\begin{align*}
1.635 < n_x < 1.637 & \quad z \wedge c = 20^\circ \quad z = \text{dark blue-green} \\
1.657 < n_x < 1.659 & \quad 2V = 70^\circ \quad x = \text{light yellow-brown} \\
1.669 < n_x < 1.671 & \quad \text{optically (-)} \quad y = \text{light green} \\
\end{align*}
\]

From Tröger (1959, Fig. 187):

\[
\begin{align*}
\text{Mg}/\text{Fe}^{++}, \text{Mn}, \text{Ti} & = .75 \text{ to } .85. \\
\end{align*}
\]

(2) 6/14/61: co-existing with diopside, calcite, quartz, calcic plagioclase; from the Patch Mountain formation.

\[
\begin{align*}
1.639 \approx n_x & \quad z \wedge c = 20^\circ \quad z = \text{dark blue-green} \\
1.652 \approx n_x & \quad 2V = 70^\circ \quad x = \text{light yellow-brown} \\
1.663 \approx n_x & \quad \text{optically (-)} \quad y = \text{green} \\
\end{align*}
\]

From Tröger (1959, Fig. 187):

\[
\begin{align*}
\text{Mg}/\text{Fe}, \text{Mn}, \text{Ti} & = .80. \\
\end{align*}
\]

Pyroxenes.—Diopside is the only pyroxene that the writer has observed in the Bryant Pond quadrangle. It is the characteristic mineral of most of the calc-silicate rocks, especially in the southern sequence. Generally it is green and forms equant poikilitic grains in the groundmass. In most specimens it is so abundant that the whole rock is green. Table 7, p. 103 & 104 shows the assemblages in which diopside is present in the Bryant Pond quadrangle. Below are presented the data on two specimens of diopside from the same rocks from which were obtained the two hornblendes previously described.

(17) 9/20/61: from impure marble lens in the Shagg Pond formation.

\[
\begin{align*}
1.723 < n_x < 1.726 & \quad 2V = 65^\circ \quad z = \text{green} \\
1.692 < n_x < 1.694 & \quad \text{optically (+)} \quad x = \text{colorless} \\
\end{align*}
\]

\[
\begin{align*}
z \wedge c & = 42^\circ \quad y = \text{light yellow-green} \\
\end{align*}
\]

(2) 6/14/61: from diopside granulite in the Patch Mountain formation.

\[
\begin{align*}
1.683 < n_x < 1.687 & \quad 2V = 60^\circ \quad z = \text{light green} \\
1.714 < n_x < 1.717 & \quad c = 43^\circ \quad x = \text{colorless} \\
\text{optically (+)} \quad y & = \text{very light yellow-green} \\
\end{align*}
\]

By means of tables in Winchell (1951) it is possible to determine that specimen (17) 9/20/61 contains 45 mole per cent of CaFeSi2O6 and that specimen (2) 6/14/61 contains 35 mole per cent. Inasmuch as the diopside-bearing rocks generally contain excess quartz, it can be assumed that small amounts of alumina will substitute into the Si positions in the diopside. Hence one can probably consider the Al content in the diopside to be negligible.

Microcline.—Microcline is the only potassic feldspar that the writer has observed in the Bryant Pond area. It occurs commonly as anhedral grains in the groundmass of the calc-silicate rocks, and in some of the mica schist and biotite schist fractions of the migmatitic gneisses. In some of the latter the microcline is microperthitic. Potassic feldspar is also in scattered blebs and stringers of coarse-grained quartz and microcline. Such blebs and stringers are commonly 3 to 6 inches long and 1 to 2
inches thick. In some respects they look like flattened or elongated megacrysts of microcline with intergrown quartz. Indeed, some quartz-microcline blebs are not flattened and are more aptly described as anhedral megacrysts. In a very few places the megacrysts are subhedral. The quartz in the blebs, stringers, and megacrysts of microcline has distinct outlines that suggest euhedral quartz crystals. Under high magnification small grains of the microcline show the characteristic grid pattern twinning and thin bands of plagioclase demonstrate the perthitic nature of the microcline.

Fisher (1962) describes similar blebs and megacrysts of potassic feldspar in the gneisses of the Bethel quadrangle but says that they are microperthitic orthoclase rather than microcline. Heal (1948) reports similar microperthitic megacrysts of orthoclase in the gneisses of the Lovewell Mountain quadrangle in New Hampshire. The writer has no explanation as to why these other workers have found the megacrysts to be orthoclase rather than microcline.

In general, the microcline is widespread but not abundant in rocks that also contain sillimanite. Moreover, this association is absent in the northeast corner of the Bryant Pond quadrangle. This feature will be considered later.

The refractive indices of 13 specimens of the perthitic megacrysts of microcline have been determined by the writer. Because the megacrysts are perthitic, it is possible to obtain only composite indices of the two feldspars. The average indices determined are:

\[ n_r \approx 1.525 \]
\[ n_o \approx 1.521 \]

It can be seen that these values are actually closer to those of orthoclase than those of microcline. Indeed, some specimens have refractive indices which are identical with those given in Winchell (1951) for orthoclase. However, the presence of microcline twinning shows the feldspar to be triclinic; hence it is microcline.

In small cleavage fragments the microcline is seen to be nearly transparent. In this way the microcline can readily be identified in the field as the plagioclase is usually dull white to gray and non-transparent.

Muscovite.—Muscovite is present in most of the pelitic rocks of the Bryant Pond quadrangle. Table 6 on p. 97 presents a list of the typical assemblages which contain muscovite. In the micaceous southern units the muscovite flakes are parallel to the biotite flakes and thus coincide with the foliation of the rock. In the gneisses of the northern units the muscovite forms large poikilitic flakes, which are as much as 4 cm long but are more commonly about 1 cm in length. In many rocks these megacrysts lie at high angles to the foliation of the rock, which is largely defined by the parallelism of the smaller biotite flakes. Most commonly, however, the muscovite plates are parallel to the foliation. In some specimens swarms and felted masses of fibrolitic sillimanite are present in the muscovite.

The writer has determined the basal spacings of three samples of muscovite from the gneisses of the northern sequence. In all three specimens the assemblages are muscovite-quartz-biotite-microcline-plagioclase \((\text{An}_{25})\)-sillimanite. The basal spacings were determined by means of a Norelco X-Ray diffractometer scanning at 1/4 minute with thorium oxide as an internal standard and using CuKα radiation. Measurements of 2θ for 005 plane gives the following results:

<table>
<thead>
<tr>
<th>Sample number</th>
<th>2θ (°)</th>
<th>d(001)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(6) 6.21/61</td>
<td>45.39</td>
<td>9.982 A</td>
</tr>
<tr>
<td>(1) 8.21/61</td>
<td>45.38</td>
<td>9.984 A</td>
</tr>
<tr>
<td>(2) 8.25/61</td>
<td>45.39</td>
<td>9.982 A</td>
</tr>
</tbody>
</table>

These values can be compared with those determined by Milton (1961) from rocks in the sillimanite and staurolite zone of metamorphism. Milton (1961, p. 154) obtained the following values:

<table>
<thead>
<tr>
<th>Sample</th>
<th>2θ (°)</th>
<th>d(001)</th>
</tr>
</thead>
<tbody>
<tr>
<td>OSM 25</td>
<td>9.914 A</td>
<td>OSM 32 9.950</td>
</tr>
<tr>
<td>OSM 431</td>
<td>9.925 A</td>
<td>OSM 104 9.956</td>
</tr>
<tr>
<td>OSM 382</td>
<td>9.942 A</td>
<td>OSM 395 9.964</td>
</tr>
<tr>
<td>OSM 473</td>
<td>9.947 A</td>
<td>OSM 264 9.966</td>
</tr>
<tr>
<td>OSM 505</td>
<td>9.948 A</td>
<td>OSM 21  9.978</td>
</tr>
</tbody>
</table>

Rosenfeld, Thompson and Zen (1958) have determined the basal spacings of two muscovites from Vermont, one of which has been analysed. They obtained the following values:

9.918 A - \( \text{K}_{0.06}\text{Na}_{3.88}\text{Ca}_{0.8} \) (mole %), Gassetts, Vermont.

9.981 A - presumably nearly pure \( \text{KAl}_3\text{Si}_3\text{O}_{12}(\text{OH})_2 \) from the biotite zone, Danby, Vermont.
It is known that the basal spacings in muscovite increase with a decrease in the amount of Na present in the mineral. The presence of Ca probably also affects the basal spacing in a similar manner but as yet the amount is unknown. Moreover, it is not known if the effect on the basal spacing due to variation in the K/Na ratio is linear. With these reservations in mind, one can tentatively state that the muscovite in the Bryant Pond area is of the various migmatitic gneiss units. Generally it makes up in the sillimanite plus potassium feldspar zone of metamorphism.

Inasmuch as the assemblages above indicate that alumina is in excess, it is probably correct to follow Crosby (1960) and Green (1960) and use the following formula for biotite:

\[ \text{KFe}_2 \gamma_3 \text{Al}_1 \gamma_2 (\text{Al}_1 \gamma_3 \text{Si}_2 \gamma_4 \text{O}_{10}) (\text{OH})_2 \].

This formula corresponds to the one commonly found by analyzing biotite from aluminous rocks. By means of the diagram on p. 374 of Winchell and Winchell (1951) the Fe++/Mg++ ratio indicated for each biotite specimen was obtained.

Another biotite composition was studied optically with the following results:

(A) 6-9/10/60 ............... in the assemblage biotite-sillimanite-muscovite-quartz-plagioclase
1.634 < \varnothing < 1.636 \quad \frac{\text{Fe}}{\text{Mg}} = .42

—from biotite schist interbedded with calc-silicate granulite in the Patch Mountain formation.

In this case it is assumed that this biotite lies between KFe$_2$ (AlSi$_3$O$_{10}$) (OH)$_2$ and KMg$_2$ (AlSi$_3$O$_{10}$) (OH)$_2$, and no KFe$_2$ (Al$_{1/2}$Si$_{1/2}$O$_{10}$) (OH)$_2$ or KMg$_2$ (Al$_{1/2}$Si$_{1/2}$O$_{10}$) (OH)$_2$ is present.

Pelitic Rocks

Pelitic schists are very abundant as the dark fraction of the gneisses that constitute much of the northern and southern sequence in the Bryant Pond quadrangle. Most of them are in the sillimanite plus postassium feldspar zone. The mineralogical variation in these rocks is limited and includes the following minerals important with regard to the metamorphic petrology: biotite, muscovite, quartz, plagioclase, microcline, sillimanite, and garnet. Quartz is present in all assemblages and muscovite in most of them. Table 6 lists all of the assemblages found in the quartz-feldspar-two mica-sillimanite schists and quartz-feld-
spar-biotite granulites that occur in the dark fraction of the migmatitic gneisses. Assemblages with microcline and sillimanite represent rocks in which microcline is present in the groundmass. Many other instances of co-existing microcline and sillimanite were found in which the microcline is present as megacrysts. In such rocks the microcline was identified by crushed fragments in an immersion mount; hence no modes were obtained for these specimens.

The assemblages listed in Table 6 can readily be represented in the system $\text{SiO}_2$ - $\text{MgO}$ - $\text{Al}_2\text{O}_3$ - $\text{FeO}$ - $\text{K}_2\text{O}$ - $\text{H}_2\text{O}$. Thompson (1957) has pointed out how such a system can be used to describe the metamorphism of pelitic rocks. Moreover, he has described a procedure by which the minerals found in this system can be projected from muscovite and thus graphically represented in a two dimensional diagram that resembles the common triangular diagrams used to describe ternary systems. Some of the conditions which must be met to successfully consider Thompson's graphical method of treating pelitic rocks are: 1) Muscovite must be present in any assemblage to be plotted. 2) Quartz must be present. 3) $\text{H}_2\text{O}$ must act as a perfectly mobile component. Conditions (1) and (2) are met by most of the pelitic rocks in the area, and Thompson (1957) (1959) has presented reasons why $\text{H}_2\text{O}$ can be considered to be a mobile component and its chemical potential or activity regarded as an externally controlled variable much like pressure and temperature.

Fig. 14 shows the basic projection proposed by Thompson (1957) and Fig. 15 shows how the assemblages of pelitic schists in the Bryant Pond area can be plotted on this projection. Obviously the plagioclase in these rocks must be ignored as it is a calcium-sodium mineral which therefore does not lie in the system we are considering. Moreover, strictly speaking, microcline, especially in the mica-sillimanite rocks, does not lie in the projection because it contains significant amounts of sodium.

The maximum number of minerals in any assemblage to be represented on Thompson's projection should in general be 3 or less. Of course quartz and muscovite are understood to be present also, but they are not shown specifically on the projection. Thompson (1957) has pointed out that some apparent violations of this rule are resolved by the fact that some one of the minerals under consideration may contain significant amounts of some component which is not in the system $\text{SiO}_2$ - $\text{FeO}$ - $\text{Al}_2\text{O}_3$ - $\text{K}_2\text{O}$ - $\text{H}_2\text{O}$. This will tend to make this mineral occur in assemblages in which the pure phase or mineral is unstable. Thompson (1957), Green (1960), Crosby (1960) and Milton (1961) have noted that $\text{Ca}^{++}$ in the garnet is the most likely troublemaker. In the present rocks, if we ignore the microcline in the sillimanite-bearing assemblages where it is perthitic and thus at least initially sodic, then it is found that the assemblages in the pelitic rocks of the Bryant Pond area follow the rule that a single assemblage should never have in it more than three of the projected minerals. Actually this is an indication that the garnets are relatively free of $\text{CaO}$ and $\text{MnO}$.

### TABLE 6

<table>
<thead>
<tr>
<th>Assemblages with muscovite and quartz:</th>
<th>No. of Specimens</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\text{qtz}$-$\text{plag}$-$\text{mirc}$-$\text{bio}$-$\text{musc}$-$\text{sill}$-$\text{garn}$</td>
<td>9</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{plag}$-$\text{mirc}$-$\text{bio}$-$\text{musc}$-$\text{sill}$</td>
<td>8</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{plag}$-$\text{mirc}$-$\text{bio}$-$\text{musc}$-$\text{garn}$</td>
<td>6</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{plag}$-$\text{bio}$-$\text{musc}$-$\text{sill}$-$\text{garn}$</td>
<td>3</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{plag}$-$\text{bio}$-$\text{musc}$-$\text{sill}$</td>
<td>32</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{plag}$-$\text{bio}$-$\text{musc}$</td>
<td>36</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{plag}$-$\text{bio}$</td>
<td>7</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{bio}$-$\text{musc}$-$\text{sill}$-$\text{garn}$</td>
<td>8</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{bio}$-$\text{musc}$-$\text{sill}$</td>
<td>2</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{bio}$-$\text{musc}$</td>
<td>2</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Assemblages without muscovite:</th>
<th>No. of Specimens</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\text{qtz}$-$\text{plag}$-$\text{mirc}$-$\text{bio}$-$\text{sill}$-$\text{garn}$</td>
<td>2</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{plag}$-$\text{mirc}$-$\text{bio}$-$\text{sill}$</td>
<td>5</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{plag}$-$\text{mirc}$-$\text{bio}$</td>
<td>2</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{plag}$-$\text{bio}$-$\text{sill}$-$\text{garn}$</td>
<td>6</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{plag}$-$\text{bio}$-$\text{sill}$</td>
<td>1</td>
</tr>
<tr>
<td>$\text{qtz}$-$\text{plag}$-$\text{bio}$</td>
<td>20</td>
</tr>
</tbody>
</table>

**Explanation**

- $\text{qtz}$ - quartz
- $\text{plag}$ - plagioclase
- $\text{mirc}$ - microcline
- $\text{musc}$ - muscovite
- $\text{sill}$ - sillimanite
- $\text{garn}$ - garnet
- $\text{bio}$ - biotite
It can be seen that Thompson's projection in the present study provides a graphical way of representing the assemblages listed in Table 6 and demonstrates the general composition ranges of the pelitic rocks of the area with respect to the system FeO - SiO$_2$ - MgO - Al$_2$O$_3$ - K$_2$O. It also provides a ready way to suggest an interpretation of the common occurrence of sillimanite intergrown with biotite. If we consider an original bulk composition in the garnet-sillimanite-biotite field and which lies near the biotite-sillimanite tie line, then the possibility arises that with a change in the metamorphic grade the biotite-sillimanite tie line may move in toward the center of the triangle (i.e. by a shift in the Fe/Fe$^+$ + Mg ratio). This would cause a decrease in the original amount of garnet, and if the original composition were eclipsed, a disappearance of garnet. The reaction would be:

$$\text{garnet} + \text{muscovite} \leftrightarrow \text{biotite} + \text{sillimanite} + \text{quartz.}$$

However, it does not provide a way of showing the change from rocks in which sillimanite is incompatible with microcline to rocks in which the two minerals are compatible. Nor does it provide a means of showing the sodic nature of the microcline that co-exists with sillimanite.

In order to describe more fully the metamorphism of the pelitic rocks in the Bryant Pond quadrangle it is possible to consider...
them in terms of the system $\text{SiO}_2 - \text{Al}_2\text{O}_3 - \text{CaO} - K_2\text{O} - Na_2\text{O} - H_2O$. This approach has been considered in detail elsewhere (Guidotti, 1963). The above system would include the minerals sillimanite, quartz, muscovite, potash feldspar, and plagioclase. As noted earlier in Table 6 all combinations of these five minerals occur in the pelitic rocks of the Bryant Pond quadrangle. However, one restriction exists inasmuch as the association potash feldspar-sillimanite $\pm$ other phases of the system $\text{SiO}_2 - \text{Al}_2\text{O}_3 - \text{CaO} - K_2\text{O} - Na_2\text{O} - H_2O$ can not be found in the northeast part of the area. It has been shown by the writer (Guidotti, 1963) that the change from those pelitic rocks without the association potassium feldspar-sillimanite to those with this association involves the reaction:

1. Muscovite + sodic plagioclase + quartz $\leftrightarrow$ sillimanite + sodic biotite + less sodic plagioclase + water.

This reaction involves an increase in metamorphic grade and is dependent upon pressure, temperature, activity of $H_2O$, and the $An$ content of the plagioclase. The northwest-trending hachured line in the N. E. corner of Plate I represents the zone in which the metamorphic conditions were sufficiently intense to effect the reaction indicated by equation 1. Thus, this line represents the sillimanite plus potassium feldspar isograd. This isograd has been discussed by numerous people working in high-grade metamorphic rocks; Heald (1948, 1950), Chapman (1952), Lundgren (1962), Snyder (1961) and Guidotti (1963). Snyder (1961) called it the sillimanite and potassium feldspar isograd and Guidotti (1963) called it the sillimanite-potash feldspar isograd. However, in the interests of euphony when listing a series of mineral isograds it would seem desirable to call this isograd the sillimanite plus potassium feldspar isograd.

As discussed in Guidotti (1963) the $An$-content of the plagioclase in the rocks probably has a strong effect on the reaction indicated in equation 1. In pelitic rocks where the initial plagioclase is very sodic the reaction will probably take place at lower temperatures than if the initial plagioclase were less sodic (assuming that the pressure and activity of water were the same in both cases). In the present case the sillimanite plus potassium feldspar isograd is drawn for rocks in which the initial plagioclase composition is near $An_{25}$. This results from the fact that in the Bryant Pond area the plagioclase in the rocks on both sides of the isograd is approximately $An_{25}$.

**Origin of the Migmatitic Gneisses**

The migmatitic nature of the gneisses in the Bryant Pond area is similar to what has been ascribed to granitization or metasomatism in many areas of high-grade metamorphism. In "granitization" the writer is including all of the various processes which have been proposed whereby material is added to an essentially solid, non-granitic rock to make it more like a "granite." Milton (1961, p. 165) has proposed that $K_2O$ was mobile or partially mobile during metamorphism and that large-scale potash metasomatism may have occurred in parts of the Old Speck Mountain quadrangle. This proposal is based entirely upon textural evidence. The writer is skeptical of Milton's interpretation inasmuch as it is extremely difficult to prove that textural features are not merely the result of local rearrangement of the initial components in a rock.

Fisher (1952, 1962) has also considered the possibility that "granitization" may have occurred in the gneisses of the Bethel quadrangle. These rocks are very similar to the northern sequence of the Bryant Pond quadrangle and are in the same metamorphic zone. Fisher has demonstrated by means of chemical analyses of the migmatitic high-grade metamorphic rocks from the Littleton formation in the Bethel area and sedimentary rocks from the Littleton formation in the Littleton area that little or no metasomatic composition changes have taken place in the Bethel area. Moreover, he has demonstrated that the compositions of the metamorphic rocks do not vary significantly from the average of 51 Paleozoic shale analyses compiled by Clarke (1924, p. 552). Fisher has suggested that the migmatitic banding or small lenses of conformable pegmatite has probably originated by partial fusion or metamorphic segregation of the light-colored minerals. He has pointed out how the schists immediately adjacent to the light-colored bands are commonly enriched in biotite and garnet. On the other hand, he has pointed out that the larger pegmatite bodies are a late intrusive phase of the Songo granodiorite. Evidence from the Bryant Pond area which supports this view was presented in an earlier section of this report. In general, the writer agrees with Fisher on the lack of metasomatism in these rocks and the importance of partial melt-
ing in the origin of the migmatitic banding. Further detailed work on the mineralogy (e.g., compositions) and relative proportions of quartz, plagioclase, and microcline in the light-colored bands of the migmatites will probably help to determine how significant partial melting may have been in forming the migmatitic gneisses. The possibility of partial melting will be mentioned further in a later section when the general conditions of metamorphism are considered.

**Petrology of the Calc-Silicate Rocks**

The mineralogy of the calc-silicate rocks throughout the Bryant Pond quadrangle has little variation. Only eleven minerals important for the analysis of the metamorphism are present. They are quartz calcite, calcic plagioclase, diopside, hornblende, microcline, biotite, scapolite, tremolite, and minor amounts of garnet and phlogopite. The garnets are probably members of the almandine pyrope series with large amounts of the Ca$_3$Al$_2$Si$_3$O$_{12}$ component. Tables 7a and b list the assemblages found in the calc-silicate rocks of the northern sequence and those in the southern sequence. It is a notable feature that all of these assemblages have quartz. The tabulated assemblages also emphasize the difference between the calc-silicate rocks of the northern and southern sequences. The most obvious difference is that calcite is present in many more assemblages in the southern calc-silicate rocks than in the northern ones. On the other hand, tremolite and garnet are more common in the northern calc-silicate rocks. Such differences are probably due to compositional variation alone, rather than reflecting a difference in metamorphic grade. The rocks with tremolite are probably poorer in the Na$_2$O than those with hornblende.

The writer believes that the zoisite commonly present in these rocks is a retrograde product after calcic plagioclase because clean, distinct grains of zoisite are rare. They are usually very ragged and they appear to be replacing calcic plagioclase. Moreover, zoisite has been observed in some of the distinctly altered rocks from shear zones. Of the minerals listed in the assemblages of Tables 7a and b only biotite, diopside, hornblende and plagioclase show signs of significant alteration. Biotite, and diopside to a very minor extent, display some alteration to chlorite. This is probably a late retrograde metamorphic effect. The relationship of hornblende to diopside is not so clear. Particularly in the calc-silicate rocks of the southern sequence, the hornblende grains are rather ragged and altered. In some cases, the hornblende even has the appearance of altering to diopside. Ragged grains of hornblende can be seen with several smaller grains of fresh-appearing diopside scattered in them. It is a rather notable fact that according to the modes presented by Barker (1961), Fisher (1942) and Hanley (1939), the stratigraphic equivalents of the Berry Ledge formation and the Patch Mountain formation in the Poland, Lewiston and Augusta areas contain considerably greater amounts of hornblende or actinolite (in the Augusta area) than in the Bryant Pond area. Unless this variation in the modes can be explained by a compositional change, it leads to the suggestion that the grade of metamorphism in the Bryant Pond area is significantly higher than in the Lewiston area ( sillimanite zone), and hence amphiboles may be less stable than diopside.

In discussing the correlation of the Berry Ledge formation with the Benton Falls limestone of the Waterville area it was noted that the latter seemed to have significantly more carbonate present than the former. It was indicated that metamorphic effects may be responsible for the decreased carbonate content in this unit from evidence seen in slabs cut perpendicular to the bedding. In these slabs it can be seen that reaction zones up to 1/4" thick have formed between the carbonate and biotite schist beds. The reaction zones are composed largely of diopside and are evidenced by distinct light, green bands separating the light grey carbonate-rich beds from the black, biotite schist beds. Hence an unmetamorphosed carbonate bed originally 1 inch thick can potentially be reduced to a bed 1/2 inch thick.

An interesting calculation concerning the massive calc-silicate rocks is the amount of dolomite necessary in the protolith to

<table>
<thead>
<tr>
<th>Assemblages in calc-silicate rocks from the southern sequence:</th>
<th>No. of specimens</th>
</tr>
</thead>
<tbody>
<tr>
<td>qtz - cale - plag - micr - diop - hnbl - bio</td>
<td>3</td>
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<tr>
<td>qtz - cale - plag - micr - diop - hnbl</td>
<td>6</td>
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<tr>
<td>qtz - cale - plag - micr - diop - hnbl - trem - bio</td>
<td>1</td>
</tr>
<tr>
<td>qtz - cale - plag - micr - diop</td>
<td>2</td>
</tr>
<tr>
<td>qtz - cale - micr - diop</td>
<td>1</td>
</tr>
<tr>
<td>qtz - calcite - micr - diop</td>
<td>1</td>
</tr>
</tbody>
</table>
Assemblages in the calc-silicate rocks from the northern sequence:

<table>
<thead>
<tr>
<th>Assemblage</th>
<th>No. of specimens</th>
</tr>
</thead>
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<td>qtz - calc - diop - hnb - scap</td>
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</tr>
<tr>
<td>qtz - calc - plag - diop - hnb - scap</td>
<td>4</td>
</tr>
<tr>
<td>qtz - calc - plag - diop - hnb - scap - gross?</td>
<td>1</td>
</tr>
<tr>
<td>qtz - calc - plag - diop - trem</td>
<td>9</td>
</tr>
<tr>
<td>qtz - calc - plag - diop - hnb</td>
<td>3</td>
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<tr>
<td>qtz - plag - micr - diop - hnb</td>
<td>1</td>
</tr>
<tr>
<td>qtz - plag - micr - diop - hnb - bio</td>
<td>1</td>
</tr>
<tr>
<td>qtz - plag - diop - hnb - scap</td>
<td>4</td>
</tr>
<tr>
<td>qtz - plag - diop - hnb - bio</td>
<td>2</td>
</tr>
<tr>
<td>qtz - plag - hnb - bio</td>
<td>1</td>
</tr>
<tr>
<td>calc - scap - diop</td>
<td>1</td>
</tr>
<tr>
<td>qtz - quartz</td>
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</tr>
<tr>
<td>calc - calcite</td>
<td>1</td>
</tr>
<tr>
<td>plag - plagioclase</td>
<td>1</td>
</tr>
<tr>
<td>micr - microcline</td>
<td>1</td>
</tr>
<tr>
<td>diop - diopside</td>
<td>1</td>
</tr>
<tr>
<td>hnb - hornblende</td>
<td>1</td>
</tr>
<tr>
<td>bio - biotite</td>
<td>1</td>
</tr>
<tr>
<td>trem - tremolite</td>
<td>1</td>
</tr>
<tr>
<td>scap - scapolite</td>
<td>1</td>
</tr>
<tr>
<td>gross - grossularite</td>
<td>1</td>
</tr>
</tbody>
</table>

that the protolith had to have 24.50 modal per cent dolomite and 75.40 per cent quartz. Moreover, it can also be calculated that a rock with 30 per cent diopside and 70 per cent quartz represents a 16.6 per cent decrease in volume from the above mentioned protolith.

From the preceding it is seen that in a rough way the original dolomite in the protolith of a diopside-bearing rock was probably a few modal per cent less than the observed diopside per cent. Actually the original dolomite may have been more nearly equal to the observed diopside modal per cent because the remainder of the rock may not be quartz alone but may include large amounts of calcic plagioclase, scapolite, microcline, etc. Moreover, to obtain more exact figures one would have to take into account Ca++ and Mg++ that are in minerals such as tremolite.

Retrograde Metamorphism

Retrograde metamorphism seems to be mainly associated with the post-metamorphic faulting and jointing. This is shown by the fact that the most intense and widespread retrograde effects are located along post-metamorphic faults and the associated joint pattern. It was pointed out earlier that many of the northeast-trending joints have a chloritized zone up to one inch wide on each side of the joint. The most extensive and thoroughly retrograded area is a zone up to one mile wide along the western half of the line on Plate 1 designating the Moll Ockett fault. In this zone much of the biotite in the migmatitic gneisses and the Songo granodiorite has been altered to chlorite.

Biotite partly altered to chlorite is commonly observed in the Bryant Pond quadrangle. Others are:

- Feldspars $\rightarrow$ sericite or saussurite
- Ca-Plagioclase $\rightarrow$ saussurite + calcite
- Ca-Plagioclase $\rightarrow$ zoisite
- diopside $\rightarrow$ chlorite.

Table 8 presents the modes of some specimens from localities where retrograde metamorphism was especially intense. All of the chlorite listed in the modes of the various stratigraphic units and intrusive rocks is believed to be retrograde in origin.
Conditions of Metamorphism

In order to obtain some idea of the conditions under which some particular metamorphic rocks formed, it is necessary to consider the mineralogy and mineral assemblages of the rocks and then compare these with experimental data on the same minerals and assemblages. In this way, depending upon the reliability of the experimental work, at least maximum and minimum limits can be placed upon the metamorphic conditions. However, many workers have pointed out that such comparison must always be received with much skepticism. It is commonly noted that (1) equilibrium is not always achieved in laboratory studies of minerals (and in some cases not in rocks either), (2) experimental studies commonly deal with some polymorph of the minerals observed in rocks, (3) experimental studies commonly use starting materials that are very different from the minerals involved in reactions taking place in the rocks, (4) the minerals formed in experimental studies are usually simpler in composition than in rocks, and (5) experimental work is commonly carried out at activities of H₂O and CO₂ that are unrealistic in rocks.

Nevertheless, lacking any better method, we must compare experimental curves on minerals and mineral assemblages with minerals and mineral assemblages observed in rocks to obtain any idea of the metamorphic conditions which might have prevailed. Fig. 16 presents a number of experimental curves from the literature.

The curves for the relations among kyanite, sillimanite, andalusite, and mullite are obtained from Clark et al. (1957) and Clark (1961). Only that segment of the kyanite-sillimanite curve between 1000 Centigrade, 18.2 kilobars and 1300 Centigrade, 21 kilobars was determined experimentally by these workers. However, drawing upon published work on other parts of the system Al₂SiO₅, field occurrences, and a knowledge of the densities of the minerals involved; Clark et al. (1957) and Clark (1961) purpose the curves shown in Fig. 16 for the minerals kyanite, sillimanite, andalusite, and mullite. These curves are qualitatively similar to those proposed by Miyashiro (1949) and Thompson (1955).

Assuming that these curves have some validity and that the sillimanite of the Bryant Pond area formed under equilibrium conditions (the latter assumption seems beyond doubt), then one can state with some assurance that the P-T conditions to which the rocks of the Bryant Pond area were subjected lie somewhere in the area outlined by the sillimanite field. In order to narrow the limits of possible pressures and temperatures, it is necessary to consider other experimental data.

Curve EF, taken from Danielsson (1950), represents the univariant reaction (in the presence of CO₂ gas):

Calcite + quartz ⇌ wollastonite + carbon dioxide.

Ellis and Fyfe (1956) have derived a similar curve. It is extremely unlikely that pure CO₂ gas was present during the metamorphism of the calc-silicate rocks. Hence this curve should probably be drawn at lower temperatures. Nevertheless, the association of quartz and calcite in some of the rocks of the Bryant Pond area clearly indicates that the temperature conditions shown by curve EF are the absolute maxima that could have occurred during metamorphism.

Curve GH represents experimental work on the stability of muscovite by Segnit and Kennedy (1961). The segment below 5000 bars was run in a steam-saturated system (Fig. 2 of Segnit and Kennedy, 1961) and represents the reaction:

muscovite + quartz ⇌ sanidine + sillimanite + quartz + water.

In the segment above 5000 bars (Fig. 3 of Segnit and Kennedy, 1961) the only H₂O in the system is that released upon the breakdown of muscovite. The reaction is:

muscovite + quartz ⇌ sillimanite + glass.
Segnit and Kennedy note that melting phenomena obscure the formation of sandine. Yoder and Eugster (1955) have presented a somewhat similar curve for the breakdown of muscovite, but, inasmuch as their system was quartz-free, it is obvious that their curve represents the maximum stability conditions of muscovite and has no direct bearing on quartz-bearing rocks.

The effect of lowering the activity of H₂O would clearly move that segment of the Segnit-Kennedy curve below 5000 bars to lower temperatures. That segment above 5000 bars would probably not be greatly affected.

The pelitic rocks of the Bryant Pond area contain abundant co-existing quartz and muscovite which is probably almost pure potassium mica as indicated by the d-spacings. Hence we can say with some assurance that the Segnit-Kennedy curve places a lower maximum temperature of formation limits than does the quartz-wollastonite curve. The fact that the sillimanite plus potassium feldspar isograd has been achieved and that the work by Heald (1950) indicates that muscovite is then close to being unstable would seem to indicate that temperatures during metamorphism in the Bryant Pond area could be close to those indicated by the Segnit-Kennedy curve (CH). However, a very low activity of H₂O could cause the muscovite to break down at lower temperatures.

From the preceding it would seem that the metamorphism in the Bryant Pond area took place under the conditions outlined by the curves sillimanite-andalusite, kyanite-sillimanite, GH, and sillimanite-mullite. Of course, this speculation is dependent upon the validity of curves discussed.

Another curve (IJ) shown in Fig. 16 is that determined by Tuttle and Bowen (1958) for the melting of Westerly and Quincy granite in the presence of steam. That segment above 4000 bars represents extrapolation of the original curve and may be somewhat misleading. Nevertheless, it is evident from curve IJ that the metamorphic conditions proposed above would support the contention that at least part of the migmatitic banding in the gneisses of the Bryant Pond area could be due to partial melting. Indeed, the abundance of quartz and feldspar in these rocks should favor appreciable partial melting at what Tuttle and Bowen (1958) have called the “Granite minimum.” This could be further substantiated by a detailed study of the modes in the light-colored bands of the migmatitic gneisses. Obviously if the activity of water were very low, then the curve IJ would be displaced to higher temperatures. As shown in Fig. 16, this curve could be displaced 100 °C. higher and still pass through the area outlined by curves sillimanite-andalusite, sillimanite-mullite, GH, and kyanite-sillimanite.

The curve KL represents the curve for the transition from calcite to aragonite as proposed by Jamieson (1953). Its main purpose is that it serves to corroborate the kyanite-sillimanite curve. No kyanite or aragonite have been observed in the Bryant Pond area.

The right-hand side of Fig. 16 shows the depths which would correspond to the pressures on the left-hand axis (taken from
Fig. 3 of Clark, 1961). These figures suggest values for depth of burial which some people find difficult to accept. Birch (1955) has suggested a geothermal gradient which would also require extreme values for depth of burial to attain the temperatures apparently required for the minerals and mineral assemblages of high-grade metamorphic rocks. To alleviate the temperature aspect of these problems, it has commonly been suggested by many geologists that the geothermal gradient may be higher in tectonically active areas. Clark (1961), on the other hand, has considered an "over pressure" phenomenon by which significant increases in total pressure could be ascribed to tectonic stresses. Although these are not solutions for the problems of burial depth, they at least suggest ways of circumventing or even solving the problems.

SUMMARY OF THE GEOLOGIC HISTORY

The deposition of the southern sequence is the first portion of the geologic record which is discernible in the Bryant Pond area. It is difficult to say how far to the west and northwest the strata of the southern sequence might have been deposited. The southern sequence is now cut off abruptly by the Moll Ockett fault and there is no evidence of these units anywhere to the northwest. Deposition of the southern sequence possibly began sometime in Pre-Middle Silurian with the Patch Mountain formation.

The Patch Mountain formation represents the deposition of at least 2000 feet of dolomitic shales and sandstones. This was followed by 800 feet of shale and argillaceous-feldspathic sandstones and then 150-250 feet of interbedded limestone and shale. These would be the protoliths of the Noyes Mountain and Berry Ledge formations respectively and probably were deposited in Silurian time.

Also in Silurian time, the Moody Brook formation was deposited as 600+ feet of argillaceous-feldspathic sandstones and shales. The record is then blank until at least Late Silurian time when the lower gneisses of the northern sequence were deformed.

The Thompson Mountain, Shagg Pond, and Billings Hill formations represent at least 5500 feet of shales and argillaceous-feldspathic sandstones with very minor amounts of dolomitic beds, which were deposited in Silurian to Early Devonian time. The Thompson Mountain and Billings Hill formations were deposited under reducing conditions with iron sulfide and free carbon resulting. Continued deposition of 3000 feet of similar rocks in the Early Devonian resulted in the Concord Pond member and the Wilbur Mountain member of the lower part of the Littleton formation. Some thin dolomitic sandstones deposited during this time have given rise to the thin calc-silicate granulite horizons in the Concord Pond member.

Following the deposition of the Wilbur Mountain member, the Howard Pond member was deposited. It represents a change to deposition of nearly equal amounts of interbedded shales and sandstones in a reducing environment followed by about 200 feet of dolomitic sandstones and feldspathic sandstones. More interbedded shale and sandstones then were deposited in a reducing environment. The final strata to be deposited and of which there is still a record are the interbedded shales and feldspathic sandstones which are now represented by the interbedded schist and quartzites of the upper Littleton formation. These strata were also deposited in Early Devonian time. Although they are nowhere exposed in the Bryant Pond quadrangle, their close proximity in the Rumford quadrangle and structural relations make it almost certain that they once covered much of at least the northern half of the Bryant Pond quadrangle.

In Middle Devonian time the strata in the Bryant Pond area started to be folded into the northeast-trending folds which are one of the dominant structural features throughout much of northern New England. Compressive stresses oriented in a northwest-southeast direction probably caused these folds. During the period of deformation the Songo granodiorite and Sebago granite were intruded, followed by the pegmatites and the Speckled Mountain quartz monzonite as late magmatic differentiates. The interaction of the intruding Songo granodiorite and Sebago granite with the northwest-southeast compressive forces probably caused the northwest-trending folds in the Bryant Pond area. The final consolidation of the Songo granodiorite and Sebago granite probably occurred shortly after the folding stresses ended. The final intrusions of pegmatite and Speckled Mountain quartz monzonite probably took place at this time also.

It would seem very likely that the major period of metamorphism which affected the rocks in the Bryant Pond area occurred
during the same period in which deformation and intrusion were taking place. The migmatitic banding which is prevalent in most of the strata originated at the same time. Much of it may represent partial fusion and then consolidation of the quartz-feldspar portion of the original shales. Some of the migmatitic banding near larger pegmatite bodies may be the result of lit-par-lit injection.

A long period of erosion followed the Middle to Late Devonian orogeny. All of the Littleton formation above the Boot member was eroded off and the nearest remaining outcrops are found about one-half mile north of Howard Mountain.

In Late Triassic time (?) the whole area was subjected to decreased horizontal stresses during which one major and several minor faults formed. Numerous joints largely parallel to the faults formed during this same period, and post-metamorphic diabase dikes were intruded along some of the planes of weakness indicated by the joints. Retrograde metamorphism occurred along many of the fault zones and to a lesser extent along joint surfaces. It is suggested that this period of decreased horizontal stress coincided with intrusion of the White Mountain magma series. The next recorded aspect of the geologic history concerns Pleistocene glaciation. Only the effects of the Wisconsin glaciation can be observed and include striations on bedrock surfaces, ground morain, ice-contact stratified drift, and outwash deposits.

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<table>
<thead>
<tr>
<th>Plate II</th>
<th>Correlation of Sections from Littleton-Moosilauke Area, New Hampshire, to Lewiston, Maine</th>
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<tbody>
<tr>
<td>North America</td>
<td>Europe</td>
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<tr>
<td>Formation or Member</td>
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</tr>
<tr>
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<tr>
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<td>Burnt Member</td>
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<td>Lower</td>
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<tr>
<td>Cuon</td>
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</tr>
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</tr>
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