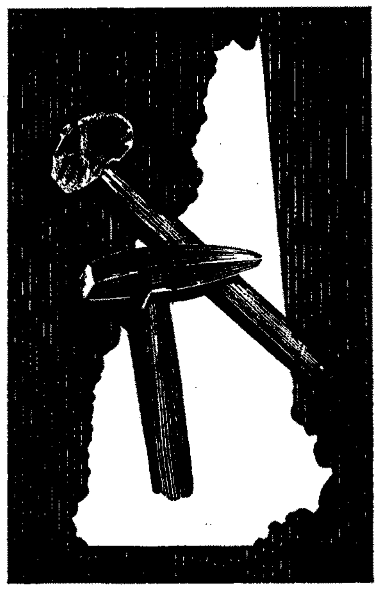
# The Bedrock Geology of the HOLDERNESS QUADRANGLE NEW HAMPSHIRE

By Evan John Englund



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# ABSTRACT

The bedrock of the Holderness Quadrangle consists primarily of sillimanite-grade metasediments of the Littleton Formation and rocks of the New Hampshire Plutonic Series. Two mappable subunits, the Clay Brook Member and the Roundtop Member, have been distinguished within the Littleton Formation. The New Hampshire Plutonic Series is represented by the Kinsman Quartz Monzonite and the Winnipesaukee Quartz Diorite, which comprise the Winnipesaukee Pluton, and by binary quartz monzonite stocks.

Analysis of minor structures shows that the metasediments were subjected to three deformations. The first produced isoclinical, recumbent folds with northwesterly-trending axes and a strong, approximately horizontal axial plane foliation parallel to bedding. This axial plane surface was refolded gently about northwest-trending axes and vertical axial planes. The final deformation about northeast-trending axes produced gentle folds in the southern part of the quadrangle and isoclinical folds in the northwest, all with vertical or steeply dipping axial planes.

Chemical analyses show that New Hampshire Plutonic Series rocks are of igneous origin and have characteristically higher $Na_2O$ contents than the metasediments. Composition and textural variations of the Kinsman Quartz Monzonite, some of which are suggestive of granitization of assimilation of metasediments, are found to be chemically more consistent with an origin by magmatic differentiation.

Interpretation of gravity anomalies indicates that the Winnipesaukee Pluton is nowhere thicker than 4 km. It was emplaced as a relatively thin sheet during the northwest-trending deformation, and was folded during the northeast-trending deformation. The late northeasterly folding resulted in an axial plane foliation within the pluton and in strong cataclasis. Emplacement of a number of such sheet-like plutons may have been the immediate source of the heat required for the regional metamorphism.

The geology of the Holderness Quadrangle is generally consistent with a plate tectonics model for the origin of the Appalachian-Caledonian system, although it is difficult to explain two periods of folding at right angles by a single deformational mechanism.

## TABLE OF CONTENTS

## LIST OF TABLES



## LIST OF FIGURES

## LIST OF PLATES

# INTRODUCTION

## Geography

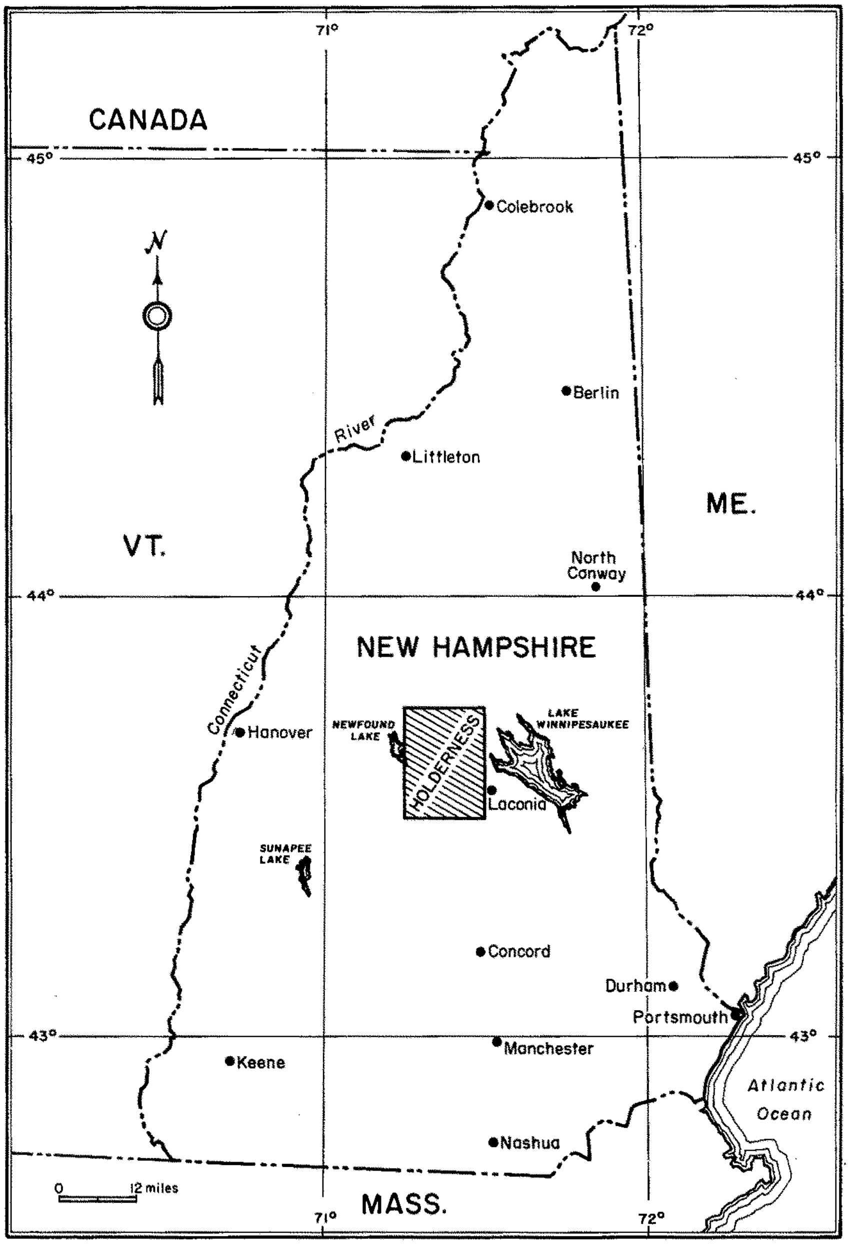
The Holderness Quadrangle lies in the geographic center of New Hampshire (Fig. 1). It has an area of approximately 220 square miles, most of it hilly and forest-covered, and is part of the New Hampshire "Lakes Region". Elevation ranges from about 320 feet at its lowest point on the Pemigewasset River, to 2187 feet on Plymouth Mountain.

Portions of Squam, Winnisquam, and Newfound Lakes lie within the quadrangle boundaries, as well as all of Little Squam, Waukewan, and a number of smaller lakes. Lake Winnipesaukee is just east of the area. The Pemigewasset River runs through the western half of the area from north to south, joining the Winnipesaukee River about five miles south of the quadrangle to become the Merrimack River.

The principal villages are Ashland, Bristol, and Meredith. The permanent population of the area is quite small, and the large numbers of summer homes and camps indicate the dominance of the tourist industry in the local economy. The importance of agriculture has been declining since the mid-1800's and is continuing to decline. The ubiquitous stone walls, cellar holes, etc., in the presently wooded areas indicate that at one time, most of the quadrangle, perhaps as much as 70 to 80% of the land area, was cleared for cultivation and pasture. By 1925, only 15% of the land area was cleared, while in 1955, this had decreased to about 5%. A number of areas shown as cleared land on 1955 maps and air photos have since been allowed to grow back. Present agricultural activity is essentially limited to small-scale dairy farming and to apple growing.

## Previous Work

Previous geological work in the Holderness Quadrangle has consisted primarily of reconnaissance mapping during three statewide surveys; those of Jackson (1844), Hitchcock (1874, 1877, and 1878), and Billings (1956). The primary purpose of the Jackson survey was the location of mineral deposits. No unusual or economically important minerals were reported by Jackson in the Holderness Quadrangle. The Hitchcock survey produced a geologic map of New Hampshire, which, in the Holderness Quadrangle, is quite accurate in showing the contacts between the schist, the "porphyritic gneiss" (Kinsman Quartz Monzonite), and the "lake gneiss" (Winnipesaukee Quartz Diorite). The Billings (1956) survey resulted in the modern geologic map of New


CANADA
71°
72°
45°
45°
Colebrook
Berlin
River
Littleton
ME.
VT.
North Conway
44°
44°
NEW HAMPSHIRE
Connecticut
Hanover
NEWFOUND LAKE
HOLDERNESS
LAKE WINNIPESAUKEE
Laconia
SUNAPEE LAKE
Concord
Durham
Portsmouth
43°
43°
Keene
Manchester
Atlantic Ocean
Nashua
0
12 miles
71°
MASS.
72°
Figure 1

# INTRODUCTION

## Geography

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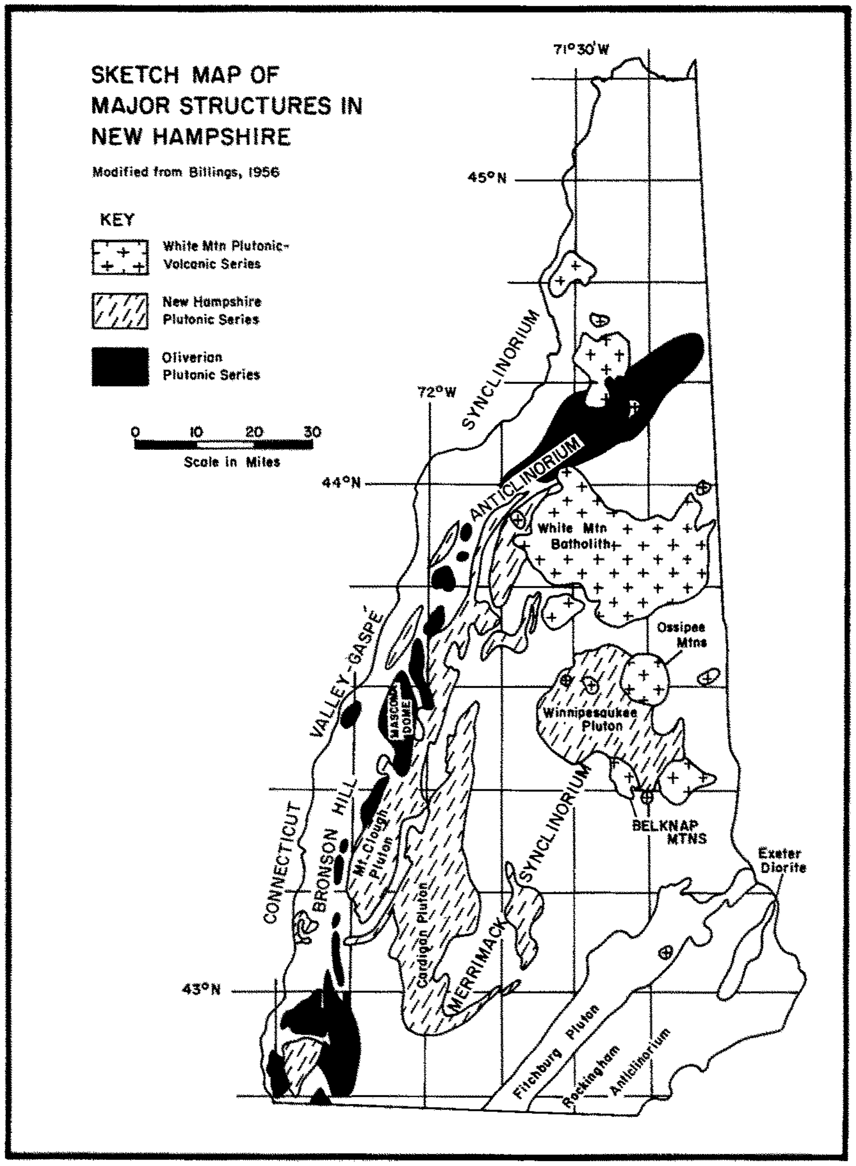
SKETCH MAP OF
MAJOR STRUCTURES IN
NEW HAMPSHIRE
Modified from Billings, 1956
KEY
White Mtn Plutonic-Volcanic Series
New Hampshire Plutonic Series
Oliverian Plutonic Series
0 10 20 30
Scale in Miles
71°30'W
45°N
72°W
44°N
43°N
SYNCLINORIUM
ANTICLINORIUM
White Mtn Batholith
Ossipee Mtns
Winnipesaukee Pluton
BELKNAP MTNS
Exeter Diorite
VALLEY-GASPÉ
CONNECTICUT
BRONSON HILL
MASCOMA DOME
Mt. Clough Pluton
Cardigan Pluton
MERRIMACK
SYNCLINORIUM
Fitchburg Pluton
Rockingham Anticlinorium

Figure 2

axes (Osberg, et. al., 1968). There the axis can be closely defined on the basis of mapping of well-dated fossiliferous metasediments. It enters New Hampshire at approximately latitude 44°00′ N, trending S 40° W. This trend would carry the axis somewhat just east of the Holderness Quadrangle.

The Rockingham anticlinorium (Fig. 2) in southeastern New Hampshire, is bounded on the southeast by the Atlantic coast and on the northwest by the Fitchburg Pluton. The stratigraphic succession of Ordovician (?) Kittery, and Silurian (?) Eliot and Berwick Formations defines the northwest flank of the anticlinorium, which suggests that the rocks in the eastern part of the Merrimack synclinorium lie stratigraphically above the Berwick Formation. However, the Fitchburg Pluton separates the Rockingham anticlinorium from the Merrimack synclinorium.

The age, origin, and structure of the Fitchburg Pluton are not well understood. Some of the rocks are petrographically identical to the Concord Granite of the New Hampshire Plutonic Series and thus would presumably be Devonian. R. S. Naylor (oral communication), however, has a preliminary Rb-Sr Precambrian date on some Fitchburg Pluton rocks. If this date is confirmed, it will require some major changes in the interpretation of the stratigraphy and structure of southwestern New Hampshire, particularly with respect of the Merrimack synclinorium.

The dominant north-northeasterly structural trend in New Hampshire is primarily a result of compressive folding during the Acadian orogeny in Middle Devonian time. In the Merrimack synclinorium in New Hampshire there is currently no known evidence from the regional structure or stratigraphy to suggest any deformational events prior to the Acadian orogeny.

## Purpose of Paper

The current study of the geology of the Holderness Quadrangle was undertaken for several reasons. First, in order to produce a geologic map of the quadrangle, subdividing the Littleton Formation into mappable subunits if possible. Second, to determine whether the stratigraphy and structural geology is consistent with the hypothesis that the area constitutes a root zone from which large-scale nappes may have originated. Third, to examine the structure and composition of the Winnipesaukee Pluton in order to determine whether it originated by granitization or magmatic process. Fourth, to locate the axis of the Merrimack synclinorium, if it passes through the Holderness Quadrangle, correlating the structure and stratigraphy with surrounding areas of New England.

A summary of the field methods used during this study is presented in the Appendix.

## Acknowledgements

The writer gratefully acknowledges financial support from the Geological Society of America, The Society of Sigma Xi, the

Department of Earth Sciences at Dartmouth College, and the New Hampshire Department of Resources and Economic Development.

The writer also wishes to thank all of the many people who provided assistance, advice, and encouragement when needed. Of these, special thanks are due Professor John B. Lyons, who suggested the project, and who always had time to talk about the geology of New Hampshire. Thanks are also due R. G. Clark, F. R. Koutz, and D. L. Nielson for assistance at various times in the field and in the laboratory.

The writer is indebted to Dr. and Mrs. James Townsend of Bridgewater, New Hampshire for their hospitality, and to his wife, Lillian, who assisted both in the field, and in the preparation of this report.

# STRATIGRAPHY

## General Statement

The bedrock of the Holderness Quadrangle consists primarily of igneous rocks of the New Hampshire Plutonic series, and of strongly metamorphosed sedimentary rocks. These igneous and metamorphic rocks are generally thought to be of Early to Middle Devonian age (Billings, 1956; Plate 1). Lamprophyre dikes of probable Late Jurassic age are frequently seen cutting the older rocks, but their total volume is relatively small.

Billings (1937), working in the less intensely metamorphosed rocks of the Littleton-Mooselaukee area where some fossils remain, established the first definitive stratigraphic succession for New Hampshire (Fig. 3), and showed that the metamorphic rocks range in age from Middle Ordovician (?) to Lower Devonian. These formations have been traced southward along strike in a 20 to 30 mile-wide belt, through most of western New Hampshire.

Moke (1946) showed that the rocks of the Littleton Formation in the Moosilaukee area could be traced continuously into the Plymouth Quadrangle, and from there into the Holderness Quadrangle. This implies that the metasediments of the Holderness Quadrangle are Lower Devonian in age. No new fossil evidence has been found to refute this interpretation, but several lines of reasoning suggest that the actual situation may be much more complex.

Thompson, et. al (1968), now interpret the Middle Ordovician (?) Orfordville Formation of Hadley (1939) as an inverted succession of the Ammonoosuc and Partridge (Ordovician), Clough and Fitch (Silurian), and Littleton (Devonian) Formations. They have also reassigned some quartzites and "rusty schists" of the Littleton Formation to the Clough and Partridge, respectively. This simplifies the stratigraphy while complicating the structure, as it requires the invocation of large scale nappes to account for the inverted stratigraphy. A sequence of three such nappes, folded east over west, has been delineated in the Connecticut Valley area, where they are now seen to be draped over the Oliverian gneiss domes of the Bronson Hill anticlinorium.

The nappe structures involve rocks ranging from Ordovician to Devonian in age, and when unfolded, these formations would have originally extended far to the east into the Merrimack synclinorium. It is difficult to imagine a mechanism whereby these nappes could have been emplaced from the Merrimack synclinorium area, while leaving behind only the uppermost of the metasedimentary formations.

| SYSTEM | SERIES | FORMATION | COLUMNAR SECTION | LITHOLOGY WHERE LEAST METAMORPHOSED | THICKNESS IN FEET |
|---|---|---|---|---|---|
| DEVONIAN | Lower | Littleton formation | Dlv | Dark-gray slate and dark-gray sandstone. Dlv = volcanic rocks, including buff to white soda-rhyolite, soda rhyolite volcanic conglomerate, dark-green chlorite schist and lighter green chlorite-actinolite schist. | 15,000± |
| SILURIAN | Middle | Fitch formation | | Blue-gray limestone, white marble, buff slaty dolomite, gray calcareous slate, gray arenaceous limestone, white calcareous sandstone, gray slate, white arkose, and light-gray quartz conglomerate. | 0-769 |
| | Lower or Middle | Clough quartzite | | Gray to white quartzite and quartz conglomerate. | 0-1200 |
| | | Unconformity | | | |
| PRE-SILURIAN Probably Ordovician | Upper Ordovician? | Partridge formation | | Chiefly dark-gray slate. | 0-2000 |
| | | Ammonoosuc volcanics | | Buff to white schistose soda-rhyolite tuff, breccia, and volcanic conglomerate; dark-green chlorite schist; dark-green chlorite-epidote schist; gray slate; gray impure quartzite. | 2000-5000 |
| | | Albee formation | | Light-green slate and phyllite, light-green quartzose phyllite, and light-green to white quartzite. | 5000 |
| | Middle Ordovician? | Orfordville formation | Oos<br>Oop<br>Ooh | Dark-gray slate and grey arenaceous slate. Oos = Sunday Mountain volcanic member, chiefly slaty soda-rhyolite tuff. Oop = Post Pond volcanic member, chiefly chlorite schist. Ooh = Hardy Hill quartzite, gray to white quartzite and quartz conglomerate. | 3500-4000 |

**Figure 3**. Stratigraphic sequence, Ordovician through Devonian, in western, central, and northern New Hampshire (after Billings, 1956).

| | | |
|---|---|---|
| Lower Devonian and Lower Devonian (?) | SEBOOMOOK FORMATION | Cyclically bedded metashale and metasandstone. |
| | MADRID FORMATION | Calcareous metasandstone and metapelite. |
| Lower Silurian and Lower Silurian (?) | SMALLS FALLS FORMATION | Rusty weathering, sulfidic metashale, metasandstone, and metaconglomerate. Upper few hundred feet is clacareous. |
| | PERRY MOUNTAIN FORMATION | Bedded metasandstone, metashale, and metaconglomerate. |
| | RANGELEY FORMATION | Metashale, metasandstone, and metapelite in Merrimack synclinorium, vitreous quartzite and conglomerate in Boundary Mountains anitclinorium. |
| Upper Ordovician (?) | QUIMBY AND GREENVALE COVE FORMATIONS | Metagraywacke, felsic metavolcanic rock, metapelite, and metasandstone. |

**Fig. 4,** Stratigraphic Section in the Merrimack Synclinorium of western Maine (after Boone, et. al., 1970, p. 3).

The stratigraphy of the west flank of the Merrimack synclinorium along strike in Maine (Fig. 4) provides a further indication that the metasediments in central New Hampshire are probably not exclusively Devonian. Although fossil dating has not been conclusive (Boone, et. al., 1970), a thick Silurian section of up to 15,000 feet of metasediments, as well as some rocks of probable Ordovician age, lie below the Devonian in that area. Unfortunately, direct correlation of any of these units southwestward into New Hampshire is prevented by the intervening Sebago Pluton of western Maine.

The only general statement that can be made about the stratigraphy of central New Hampshire is that no definitive fossil age evidence is yet available, and that the rocks may well represent all, or any part(s) of, the Ordovician, Silurian, and Lower-Middle Devonian rocks.

## Littleton Formation

The Littleton Formation in the Holderness Quadrangle contains a variety of metasedimentary rocks, including quartzites, pelitic schists, and calc-silicate granofels. No metavolcanic rocks have been recognized in the area to date. Two mappable sub-units have been distinguished within the Littleton Formation, and have been informally designated the Clay Brook Member and the Roundtop Quartzite Member. These will be discussed separately.

The major portion of the Littleton Formation is a thick sequence of mica schists, often grading into or alternating with beds of impure quartzite (quartz-feldspar-mica granofels). Somewhat less abundant is calc-silicate granofels, which invariably occur in isolated boudins. A typical calc-silicate boudin would be on the order of 4 inches by 10 inches, and may be separated from other boudins along strike by as much as several meters, which indicates that the original beds have undergone extreme extension.

Compositional layering which can be identified as bedding generally occurs as either thin, alternating beds of mica schist and granofels with no distinct graded bedding (Figs. 23, 24) or as thicker, less well-defined bedding in predominantly micaceous schists (Fig. 20). Unfortunately the effects of metamorphism and deformation frequently make it difficult to determine the nature of the original bedding (Fig. 22).

Quartz and biotite are the dominant and ubiquitous minerals in the non-calcareous rocks. Oligoclase is generally present in small amounts, and garnet, sillimanite and muscovite are common. The mineral assemblages typically found in the Littleton area, in addition to quartz: biotite-sillimanite-(muscovite)-(plagioclase); biotite-garnet-(muscovite)-(plagioclase); and biotite-sillimanite-garnet-(muscovite)-(plagioclase). Biotite and garnet are often partially altered to chlorite. Accessory minerals are apatite, ilmenite, pyrite, pyrrhotite, graphite, sphene and tourmaline. An unusual chloritoid-bearing rock was observed in one location. Some estimated modes are given in Table 1.

The calc-silicate boudins are usually concentrically zoned. The cores typically contain a pale, salmon-colored assemblage of quartz-diopside-grossularite. Surrounding this is an intermediate zone of quartz-diopside-grossularite-clinozoisite, and an outer, gray-green zone of quartz-plagioclase-antinolite. Sphene is a common accessory.

Weathered surfaces of the Littleton Formation are generally light to dark gray, as are fresh surfaces. Rusty weathering is not uncommon, however, and may sometimes be due to sulfide-rich beds, but is often related to sulfide-bearing veins or fracture fillings.

## CLAY BROOK MEMBER

### Lithology

The Clay Brook Member (informal name) of the Littleton Formation is characterized by very rusty weathering outcrops often with a limonitic crust on outcrop surfaces, and by an abundance of sulfides and graphite. The type locality is a series of outcrops along the bed of Clay Brook, about 2 miles southeast of Plymouth Mountain (Plate 1). Good outcrops are readily accessible along Highway 3 and 25 just south of Plymouth, from the point where it enters the quadrangle southward to the intersection with River Road.

The most diagnostic rock type is a very hard, gray, fine-grained granofels with abundant, finely disseminated pyrrhotite. In the field, this rock was called a sulfidic quartzite, but microscopic examination revealed it to be a calc-silicate granofels. The rock contains quartz, pyrrhotite, sphene, zoisite, tremolite, diopside, plagioclase, and graphite. A second rock type characteristic of the Clay Brook Member is a graphite-rich schist containing muscovite, quartz, a very fine-grained white mica (paragonite?), plagioclase, graphite, pyrrhotite, and a small amount of biotite. The rock is usually quite friable, and is stained a greenish yellow on weathered surfaces. This rock and the gray granofels are locally interbedded with biotite schist. Garnet and sillimanite are very rare in the Clay Brook Member. Some estimated modes are presented in Table 1.

### Distribution

The Clay Brook Member crops out in a band approximately one mile wide and seven miles long, in the northwest quarter of the Holderness Quadrangle (Plate 1). The bedrock immediately north in the Plymouth Quadrangle is covered by a thick glacial deposit; but two outcrops of the Clay Brook Member were located along strike, indicating that the unit can be extended at least four miles into the Plymouth Quadrangle.

Rusty, graphitic rocks apparently belonging to the Clay Brook Member were observed in isolated outcrops in several other areas; notably within the contact zone (Dlk) one mile southeast of the village of Ashland; near the Kinsman-Littleton contact at Saddle Hill, southeast of New Hampton village; one-half mile northeast of the village of Bristol; and near the base of Hershey Mountain in Prescott Brook, southeast of Bristol village.

## Thickness

The outcrop of the Clay Brook Member attains a maximum width of approximately 5,600 feet at the northern boundary of the quadrangle. Bedding and foliation in this area dip very steeply or vertically (Fig. 18b), so that 5,600 feet would be the maximum true thickness for the unit. However, this area is also characterized by upright isoclinal folding which could have at least doubled the unit, making the maximum thickness 2,800 feet. An earlier period of isoclinal folding has also been recognized in the Holderness Quadrangle, which greatly complicates the interpretation of stratigraphic thickness, so that a further reduction of the Clay Brook Member's thickness to a maximum of 1400 feet is not unlikely. In any event, some repetition is probably present, so that the maximum true thickness is probably considerably less than 5,600 feet, and possibly only a quarter of this amount.

## Age and Correlation

An examination of various stratigraphic sections in northern New England reveals that calcareous rocks are apparently characteristic of Silurian (Fig. 4). Calcareous Silurian units include the Fitch Formation of western New Hampshire, the Eliot and Berwick Formations and the Merrimack Group of southeastern New Hampshire (Billings, 1956) and the Madrid and Smalls Falls Formations of western Maine (Boone, et. al., 1970, p. 15). The Smalls Falls Formation in particular resembles the Clay Brook Member in that it contains both calcareous rocks and sulfidic metasandstones. In the absence of any better stratigraphic criteria, the Clay Brook Member is tentatively placed at the base of the Littleton Formation, and considered to be of possible Silurian age.

The Francestown Member of the Littleton Formation in the Peterborough Quadrangle (Greene, 1970) closely resembles the Clay Brook Member. Estimated modes of the Francestown Member (after Greene, 1970) are included in Table 1 for comparison. Rocks which are lithologically similar to the Clay Brook Member have recently been observed in the northeast part of the Hillsboro Quadrangle (D. L. Nielson, personal communication) and in the Mt. Kearsarge and Penacook Quadrangles (J. B. Lyons, personal communication). These may eventually prove to connect the Clay Brook and Francestown Members. Greene places this unit in the upper Littleton Formation, primarily because the rocks are interpreted as becoming younger toward the axis of the Merrimack Synclinorium. However, structural evidence in the Holderness Quadrangle, to be discussed in detail later, shows that the actual structure is much more complex, so that even if the Clay Brook Member can be positively correlated with the Francestown Member, a Devonian age is not assured.

# ROUNDTOP QUARTZITE MEMBER

## Lithology

The Roundtop Quartzite Member (informal name) of the Littleton

Formation was identified as a thin mappable unit in the southwestern part of the Holderness Quadrangle in 1971, and the geologic map (Plate 1) shows its presently known distribution. Further detailed mapping conceivably could extend the unit to the northeast or the southeast of the area between Bristol Peak and Peaked Hill, where it was last observed. The type locality is at the top of Roundtop Hill, 1 mile south of the village of Bristol. The unit at this point is a thickly bedded, massive, impure quartzite which is generally light brown on fresh surfaces and often white or light gray on weathered surfaces. Two other rock types have been included in the Roundtop Quartzite Member. The first, which can be seen about 500 feet southeast of the southeast bay of Newfound Lake (Plate 1), is a quartzite exhibiting good graded bedding (Plate 2) in layers 4 to 6 inches thick. The second type, observed on Sugar Hill, 1 mile north of the village of Bristol where the hill is crossed by a power line, consists of irregular, sometimes discontinuous bands of impure quartzite up to 1 inch thick, separated by very thin (less than 1/8 inch) micaceous layers. This "pin-striped" rock may be merely a sheared version of the quartzite with graded bedding. The mineralogy of these rocks is dominated by quartz, with either garnet, or biotite plus muscovite as the secondary minerals. Plagioclase may be present. Some estimated modes are presented in Table 1.

**Distribution**

The Roundtop Quartzite Member is exposed in a relatively thin belt along the top of Periwig Mountain and Roundtop Hill, south of the village of Bristol. North of Bristol, it appears along the top of Sugar Hill and intermittently along the top of a ridge running southwest from Bristol Peak toward the southeast end of Newfound Lake. An additional outcrop has been tentatively mapped between Bristol Peak and Peaked Hill, but it is not certain that this belongs to the Roundtop Quartzite Member.

South of Bristol, the quartzite is relatively flat-lying, in a series of gentle folds whose axes plunge shallowly to the northwest. In contrast, the quartzite north of Bristol occurs in isoclinal folds with vertical axial planes and nearly horizontal, north to north-northeast trending axes.

**Thickness**

The same structural difficulties encountered in determining the thickness of the Clay Brook Member also apply to the Roundtop Quartzite Member. In the outcrop near Newfound Lake in which good graded bedding is present, for example, the graded bedding appears to show an isoclinal syncline adjacent to an overturned anticline, which would require a recumbent fold with both upright and inverted limbs, subsequently refolded isoclinally. Nevertheless, the relatively restricted occurrence of the unit indicates that it is quite thin, with a probable maximum thickness of 150 feet or less.

**Age and Correlation**

Although the Roundtop Quartzite Member has not been definitely

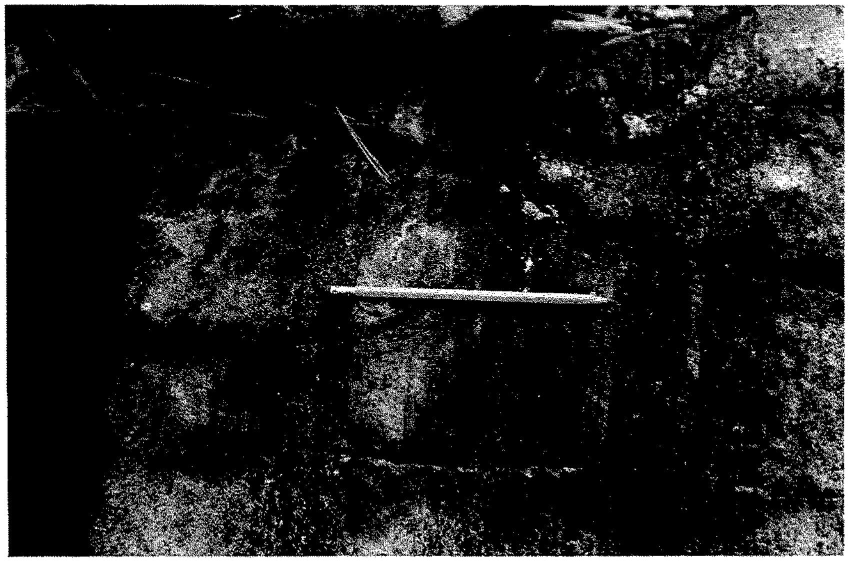

**Plate 2.** Graded bedding in the Roundtop Quartzite member near the southeast end of Newfound Lake. Beds dip vertically, and top to the right. Rocks several feet to the right of the photo top to the left, indicating an isoclinal syncline.

located north of Bristol Peak, projection a short distance to the north would bring it into close proximity with the Clay Brook Member. The most logical correlation for the Roundtop Quartzite Member on a regional basis would be with the Silurian Clough Quartzite of western New Hampshire. The Clough, as remapped by Thompson, Robinson, and others (Thompson, et. al., 1968) now includes some quartzites formerly thought to be in the Littleton Formation. Either the Clay Brook Member or the schists which contain abundant calc-silicate boudins might then be correlated with the calcareous Silurian Fitch Formation, which overlies the Clough.

Table 1, Estimated Modes of Some Metamorphic Rocks

| Sample Number | Rock Type and Location |
|---|---|
| 30-69 | Biotite-garnet-sillimanite-muscovite schist, Dl, surrounding calc-silicate boudins (#30a-69); Interstate 93 at Meredith-Sanbornton town line. |
| 35-69 | Calc-silicate boudin, Dl, Interstate 93 due west of North Sanbornton. |
| 12-71 | Chloritoid schist, Dl; near contact with bqm, north flank of Plymouth Mountain. |
| 15-71 | Thin-bedded biotite-garnet-muscovite schist, Dl; Highway 3b, 3 miles north of Highway 104. |
| 21-71 | Fine-grained schist, Dl; Highway 175, 1 mile north-northwest of Church Hill. |
| 8-71 | "Pin-striped" quartzite, Dlq; north end of Sugar Hill, north of Bristol Village. |
| 9-71 | Massive impure quartzite, Dlq; north end of Sugar Hill, North of Bristol Village. |
| 31-69 | Graphitic mica schist, Clay Brook Member (Dlc); Highways 3 and 25, ½ mile northeast of junction with River Road. |
| 11-71 | Sulfidic calc-silicate granofels, Clay Brook Member (Dlc); Highways 3 and 25, ½ mile northeast of junction with River Road. |
| 16-71 | Graphitic muscovite schist, Dlc; Clay Brook, southeast of Plymouth Mountain. |

| | Littleton Formation | | | | | Roundtop Quartzite Member | | Clay Brook Member | | | Francestown Member (after Greene, 1970) | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 30-69* | 35-69* | 12-71* | 15-71* | 21-71* | 8-71 | 9-71 | 31-69* | 11-71* | 16-71 | 1 | 2 | 3 |
| Quartz | 72 | 50 | 25 | 23 | 50 | 70 | 80 | 35 | 40 | 20 | 78 | 45 | 75 |
| Biotite | 20 | --[1] | -- | -- | 15 | 12 | -- | 10 | -- | -- | -- | -- | -- |
| Muscovite | 2 | -- | 10 | 10 | 5 | 12 | tr[2] | 20 | -- | 35 | -- | -- | 15 |
| Garnet | 2 | 3 | 5 | 5 | -- | 1 | 15 | -- | -- | -- | -- | -- | -- |
| Sillimanite | 4 | -- | -- | -- | -- | -- | -- | -- | -- | -- | -- | -- | -- |
| Plagioclase | -- | 25 | -- | -- | 25 | -- | 5 | 30 | 40 | 10 | 12 | 30 | 10 |
| Opaque Oxides | tr | tr | 5 | 2 | tr | 2 | tr | -- | -- | -- | tr | m[3] | tr |
| Epidote | -- | 2 | -- | -- | -- | -- | -- | -- | tr | -- | 3 | 4 | -- |
| Actinolite | -- | 12 | -- | -- | -- | -- | tr | -- | 5 | -- | 4 | 10 | -- |
| Diopside | -- | 8 | -- | -- | -- | -- | tr | -- | 10 | -- | tr? | -- | -- |
| Chloritoid | -- | -- | 20 | -- | -- | -- | -- | -- | -- | -- | -- | -- | -- |
| Sericite | -- | -- | 35 | 35 | -- | 3 | -- | -- | -- | 30 | -- | -- | -- |
| Staurolite | -- | -- | tr? | -- | -- | -- | -- | -- | -- | -- | -- | -- | -- |
| Chlorite | -- | -- | tr | 25 | 5 | -- | -- | -- | -- | -- | -- | -- | -- |
| Sphene | -- | tr | -- | -- | tr | -- | tr | -- | tr | -- | m | 2 | -- |
| Graphite | -- | -- | -- | -- | -- | -- | -- | 5 | tr | 5 | m | 2 | -- |
| Sulfides | -- | -- | -- | -- | -- | -- | -- | -- | 5 | -- | 4 | 7 | m |

* See Table 4 for chemical analysis.
1. -- - not present
2. tr - trace
3. m - minor

**Table 1**

# PLUTONIC ROCKS

## New Hampshire Plutonic Series

The New Hampshire Plutonic Series is represented in the Holderness Quadrangle by the Meredith phase of the Kinsman Quartz Monzonite, the Winnipesaukee Quartz diorite, and the binary quartz monzonite. The Kinsman Quartz Monzonite and Winnipesaukee Quartz Diorite together form the Winnipesaukee Pluton, which is exposed in a roughly circular area with a diameter of 30 to 35 kilometers. The pluton is exposed in the northeast part of the Holderness Quadrangle covering about one third of the quadrangle area (Plate 1). The largest portion of the Kinsman phase of the pluton lies within the Holderness Quadrangle.

The binary quartz monzonite occurs most commonly as smaller, apparently isolated bodies throughout the area, although that unit shown directly north of Newfound Lake on Plate 1 is part of a much larger body which is a prominent feature in the Cardigan, Rumney, and Plymouth Quadrangles.

### Kinsman Quartz Monzonite

The Meredith phase (Billings, 1956) of the Kinsman Quartz Monzonite is that part of the Kinsman associated with the Winnipesaukee Pluton. All of the Kinsman Quartz Monzonite bodies in the Holderness Quadrangle belong to the Meredith phase, and will hereinafter be referred to simply as "Kinsman", or "Kinsman Quartz Monzonite", except when comparisons are being made to rocks from other plutons. Varieties of the Kinsman are shown in Plates 3-7.

The Kinsman Quartz Monzonite contains white crystals of feldspar as large as 10 cm. long in a finer-grained, gray to black matrix of biotite, quartz, plagioclase, and K-feldspar. Garnets are not uncommon in the Kinsman, and where present, invariably show reaction rims of biotite, suggesting that garnets crystallized directly from the magma, and subsequently reacted with the magma to form biotite. Occassionally, where the Kinsman is not strongly sheared, biotite occurs in small, round clusters which may be relicts of completely altered garnets.

The feldspar megacrysts in the Kinsman Quartz Monzonite include both microcline and plagioclase ($An_{20}$-$An_{35}$). Microcline generally forms the larger crystals. Most of the megacrysts show Carlsbad twinning and range from euhedral to strongly sheared. Most samples of Kinsman show evidence of moderate to strong shearing, including

**Plate 3.** Kinsman Quartz Monzonite from an outcrop on Highway 104 about three miles northeast of the village of New Hampton. The quarter indicates scale. The K-feldspar megacrysts have been sericitized, which emphasizes the surrounding myrmekitic rims. The myrmekite is seen to be a late feature, as it is developed around sheared crystals.

**Plate 4.** Kinsman Quartz Monzonite from an outcrop on Highway 104 north of Wicwas Lake. The quarter indicates scale. This is a zone of intense shearing, indicated by the stretched and tapered megacrysts. Note that some crystals are euhedral, suggesting a second period of K-feldspar growth.

**Plate 5**. Kinsman Quartz Monzonite from the same outcrop as Plate 4. Some of the K-feldspar megacrysts have been completely replaced by quartz, leaving the myrmekite rims unaltered. This replacement phenomenon was not observed elsewhere. Note that this rock before replacement would have had an abnormally high $K_2O$ content. Texture suggests mechanically concentration of K-feldspar megacrysts.

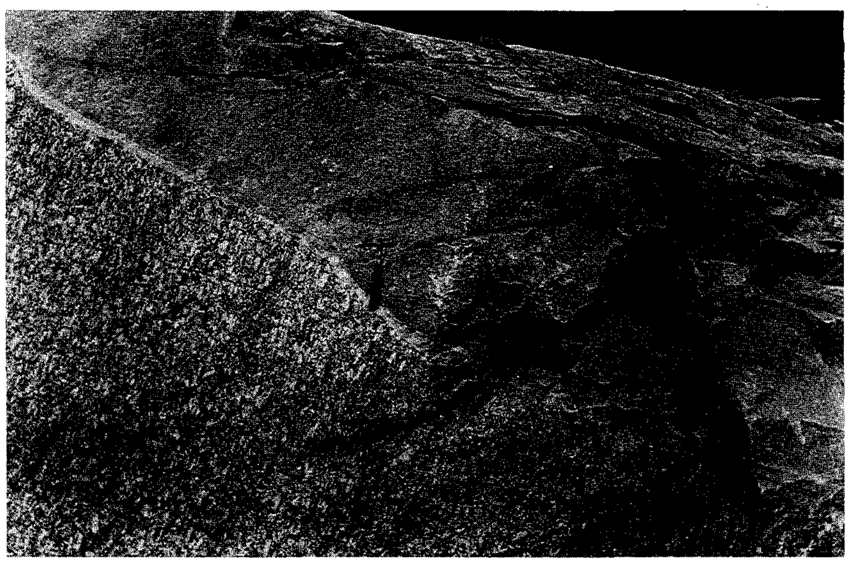

**Plate 6.** Well-foliated Quartz Monzonite cut by binary quartz monzonite, Highway 104, north of Wicwas Lake. The relative homogeneity of the Kinsman here contrasts sharply with Plate 7.

**Plate 7.** Kinsman Quartz Monzonite cut by dikes of nonfoliated binary quartz monzonite, from the same outcrop as Plate 5. The folded layered structure of the Kinsman is interpreted as being due to a deformation during the final stages of crystallization of the magma. The sheared megacrysts are always parallel to the layering, while the euhedral megacrysts are more randomly oriented, indicating post-shearing growth of some K-feldspar megacrysts. Note that a small amount of offset occurred during intrusion of the larger dike of binary quartz monzonite.

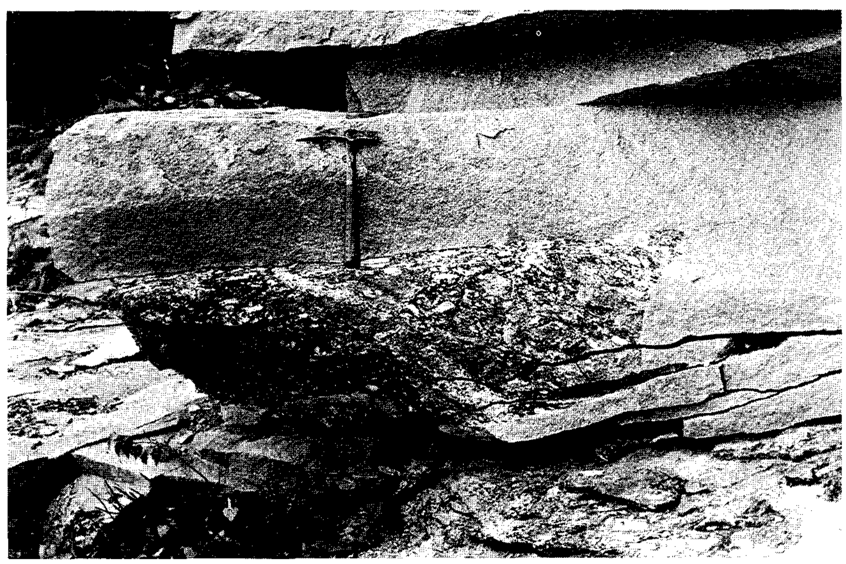

**Plate 8.** Binary quartz monzonite containing an isolated inclusion of Kinsman Quartz Monzonite, from the same outcrop as Plate 5. Note that the foliation of the Kinsman is discordant at the lower contact, but becomes concordant at the upper contact. This suggests that the shear zone may have provided an intrusive conduit for the binary quartz monzonite.

undulating extinction in quartz, kink bands in biotite, and bent, fractured, or tapered feldspar crystals. Tapered albite twins in plagioclase are also common.

The microcline megacrysts are generally surrounded by discontinuous rims of myrmekite, which are sometimes exaggerated by weathering, as in Plate 3. Some of these megacrysts appear to be moderately sheared, and yet are still surrounded by myrmekite rims, indicating that the myrmekite developed after the shearing event. A unique occurrence is shown in Plate 5, where some of the K-feldspar crystals have been completely replaced by quartz, while adjacent crystals are unaltered.

Some degree of foliation is usually present in the Kinsman, and is a result of subparallel alignment of both the tabular feldspar megacrysts and the biotite of the matrix. The foliation is often poorly developed and difficult to measure, particularly in small outcrops or where only a two-dimensional surface is exposed. However, where the foliation is well-developed and exposed, it usually strikes north-northeasterly and dips steeply to the west. The origin of the foliation will be discussed in detail later.

The contact between the Kinsman phase of the pluton and the surrounding Littleton Formation is very complex, and could not be shown in detail on the map of Plate 1. The contact shown on Plate 1 is somewhat subjective, and is intended to separate the area in which the Kinsman definitely predominates or in which any Littleton present may be considered as isolated roof pendants or xenoliths, from the area in which the Littleton may be considered to be predominantly bedrock even though extensively intruded by Kinsman.

The area shown as Dlk on Plate 1 is the contact zone where Kinsman is commonly found injected into the Littleton. Individual contacts within this area are generally concordant, being parallel to both the foliation of the Kinsman and the foliation/bedding of the Littleton. In places the contacts cut the Littleton foliation/bedding at low angles.

The overall nature of this contact zone suggests a lit-par-lit injection of the Kinsman along the foliation planes of the Littleton at scales ranging from centimeters to tens of meters. The contact zone is well exposed in several roadcuts on Interstate 93 north of the village of Ashland.

## Winnipesaukee Quartz Diorite

The Winnipesaukee Quartz Diorite is the major member of the Winnipesaukee Pluton, although it is less abundant and less well exposed than the Kinsman within the Holderness Quadrangle. Because the Winnipesaukee is less resistant to erosion than the other rock units in central New Hampshire, it underlies lowlands containing Lake Winnipesaukee and Squam Lake, as well as a number of smaller lakes.

Hitchcock (1877) considered the "Lake Winnipeseogee Gneiss", as he called it, to be equivalent to the rocks which are now called Bethlehem Gneiss. This view is supported by Billings (1956) although the

Winnipesaukee is generally a quartz diorite in mineralogical composition, whereas the Bethlehem ranges from granodiorite to quartz monzonite. The two are very similar texturally and mineralogically. Chemical analyses of the Winnipesaukee and Bethlehem (Billings and Wilson, 1964) show the Winnipesaukee to be higher in sodium and calcium, which is consistent with its higher plagioclase/K-feldspar ratio.

The mineralogy of the Winnipesaukee Quartz Diorite is very simple, with quartz, plagioclase ($An_{25}$-$An_{40}$) and biotite being the principal components; K/feldspar is minor, and muscovite is rare. Common accessory minerals are apatite, sphene, ilmenite, pyrrhotite, and epidote.

A somewhat unusual variety was found west of White Oak Pond (Plate 1), near the contact with the Kinsman. In hand specimens, it does not appear to differ from the normal Winnipesaukee, but the mafics were found to consist of approximately equal amounts of hornblende and biotite. The remainder of the rock is primarily plagioclase ($An_{33}$), with a small amount of accessory quartz, apatite, ilmenite and epidote. An estimated mode is presented in Table 2 (#8-69).

The Winnipesaukee is generally medium-to-coarse grained and equigranular, although it is locally porphyritic with feldspars as large as 1 cm. in diameter. The rock ranges from nonfoliated to strongly foliated, and measured foliations consistently strike north-northeasterly, and dip about 50°W. Foliations measured by Quinn (1944) in the Winnipesaukee Quadrangle were somewhat more variable, particularly near the contacts, but over most of the pluton, the north-northeasterly strike and westerly dip of the foliation is generally consistent.

The contact between the Kinsman and Winnipesaukee phases of the pluton is best exposed in a roadcut which extends about ¼ mile along Highway 3 and 25 north of Shepard Hill near Holderness. The Winnipesaukee Quartz Diorite is well exposed on the eastern end of the roadcut. Toward the center of the roadcut, concordant layers of Kinsman on the order of several feet thick begin to appear, and these become thicker and more abundant toward the west. Kinsman is the dominant phase at the west end of the roadcut, and outcrops farther west are exclusively Kinsman. The relative ages of the Kinsman and Winnipesaukee are not clearly defined.

Figure 5 schematically illustrates the relationahip between the Winnipesaukee, one of the layers of Kinsman, and some inclusions within the Kinsman, which suggest that the Kinsman has intruded the Winnipesaukee, carrying with it large included blocks which became lodged in the channel. While this configuration could have resulted by coincidence if the Winnipesaukee were intrusive into the Kinsman, it is unlikely.

### Binary Quartz Monzonite

Use of the term "binary quartz monzonite" in this paper follows that

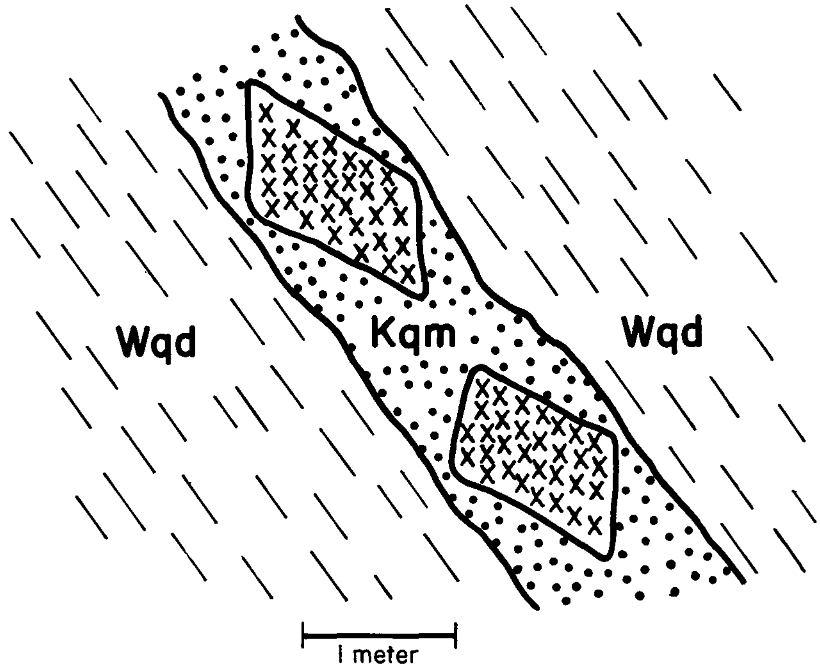


**Figure 5.** Sketch of relationships between the Winnipesaukee Quartz Diorite, the Kinsman Quartz Monzonite, and some inclusions in the Kinsman, suggesting that the Kinsman intruded the Winnipesaukee.

of Greene (1970), and is applied to all of the rocks which have previously been called "binary granite". The rocks in this group may range in composition from granite to granodiorite, but quartz monzonite appears to be the most representative rock type.

The binary quartz monzonites are characteristically medium grained, equigranular, and range from massive to strongly foliated. They tend to be somewhat lighter in color and tend to occur in smaller, more discordant bodies than the other members of the New Hampshire Plutonic Series. As a group, they were formed subsequent to the Kinsman and Winnipesaukee, and the very light-colored finer-grained, and least-foliated variety shown cutting the Kinsman in Plates 6-8 is considered to be the latest phase of the New Hampshire Plutonic Series.

The binary quartz monzonite shown in the northwest corner of the quadrangle on Plate 1 is part of a larger body which has been mapped as Concord Granite in the Cardigan, Rumney, and Plymouth Quadrangles. This intrusion includes both an earlier, coarser-grained, somewhat foliated phase as well as a later phase of the binary quartz monzonite, but both are so closely interrelated that they have been mapped as a single unit (Fowler-Lunn and Kingsley, 1937).

Table 2, Estimated Modes of Some New Hampshire Plutonic Series Rocks

| Sample Number | Rock Type and Location |
|---|---|
| 1-71 | Kqm; Highway 104, north of Wicwas Lake. |
| 3-71 | Sheared Kqm (?); Highway 104, north of Wicwas Lake. |
| 4-71 | Sheared Kqm (?); Highway 104, north of Wicwas Lake. |
| 5-71 | Sheared Kqm (?); see Plate 8; Highway 104, north of Wicwas Lake. |
| 10-69 | Moderately sheared Kqm; Highway 104, north of Wicwas Lake. |
| 8-69 | Hornblende-bearing Wqd; west of White Oak Pond, near the contact with Kqm. |
| 25-71 | Coarse-grained, somewhat foliated bqm; 1/2 mile north of road intersection described for sample #26-71. |
| 26-71 | Coarse-grained, somewhat foliated bqm; road intersection 1/3 mile northwest of Glove Hollow Brook, north of Plymouth Mountain. |

| Sample | 1-71* | 3-71* | 4-71* | 5-71* | 10-69* | 8-69 | 25-71 | 26-71* |
|---|---|---|---|---|---|---|---|---|
| Quartz | 20 | 25 | 15 | 15 | 30 | 5 | 25 | 35 |
| K-feldspar | 30 | 5 | 20 | 5 | 35 | --[1] | 25 | 10 |
| Plagioclase | 35 | 33 | 40 | 55 | 25 | 65 | 35 | 35 |
| Biotite | 10 | 30 | 20 | 24 | 9 | 10 | 10 | 15 |
| Muscovite | 1 | -- | -- | -- | 1 | -- | 2 | 5 |
| Opaque Oxides | 1 | 1 | tr[2] | 1 | tr | tr | tr | tr |
| Garnet | -- | 1 | -- | -- | -- | -- | -- | tr |
| Myrmekite | 3 | 5 | 5 | -- | tr | -- | 3 | -- |
| Hornblende | -- | -- | -- | -- | -- | 15 | -- | -- |
| Epidote | -- | -- | -- | -- | -- | 5 | -- | -- |
| Plagioclase (Anorthite Content ±5%) | 25 | 34 | 28 | 36 | 30 | 33 | 30 | 27 |

* See table 3 for chemical analysis.
1. -- - not present.
2. tr - trace.

**Table 2**

The portion of the binary quartz monzonite stock which lies within the northwestern part of the quadrangle as well as the nearby unit just north of Plymouth Mountain (Plate 1) consists primarily of the earlier phase, which, if it were not for the relative abundance of muscovite, would be difficult to distinguish from the Winnipesaukee Quartz Diorite. Chemical analysis of a sample of this rock from north of Plymouth Mountain (Table 3) shows it to be similar to the Kinsman and Winnipesaukee.

## White Mountain Plutonic Volcanic Series And Related Rocks

Rocks of the White Mountain Plutonic-Volcanic Series are abundant north and east of the Holderness Quadrangle (Fig. 2) and it is somewhat surprising that the series is so poorly represented within the Holderness Quadrangle. The only rocks which probably belong to the White Mountain series are several felsite dikes cutting the outcrop on Highway 3 and 25 in which the Kinsman and Winnipesaukee are in contact. A small White Mountain Series syenite stock, to which these dikes may be related, is located about 4 miles to the north in the Plymouth Quadrangle.

Lamprophyre dikes are common in the Holderness Quadrangle, as in much of central New Hampshire. The type locality of camptonite, a variety of lamprophyre, is located a few miles north, in the Plymouth Quadrangle. The relationship between the lamprophyres and the White Mountain Series is not clear, but the lamprophyres are occasionally observed cutting White Mountain Series rocks in central New Hampshire. Absolute age determinations are not entirely consistent with this view, however. The White Mountain batholith near Conway, New Hampshire has been dated at 180 million years, while recently published K-Ar ages from biotites (Foland, et. al., 1971) give ages of 149-158 million years for the Ossipee complex (Fig. 2). Jack Rice (1970, unpublished ms., Dartmouth College) has made five fission-track age determinations on lamprophyre dikes from the Holderness Quadrangle, and finds an age of 142 ±8 million years, which is about in the middle of the range of White Mountain ages reported by Foland et. al. (1972). An example of a lamprophyre dike is shown in Plate 9.

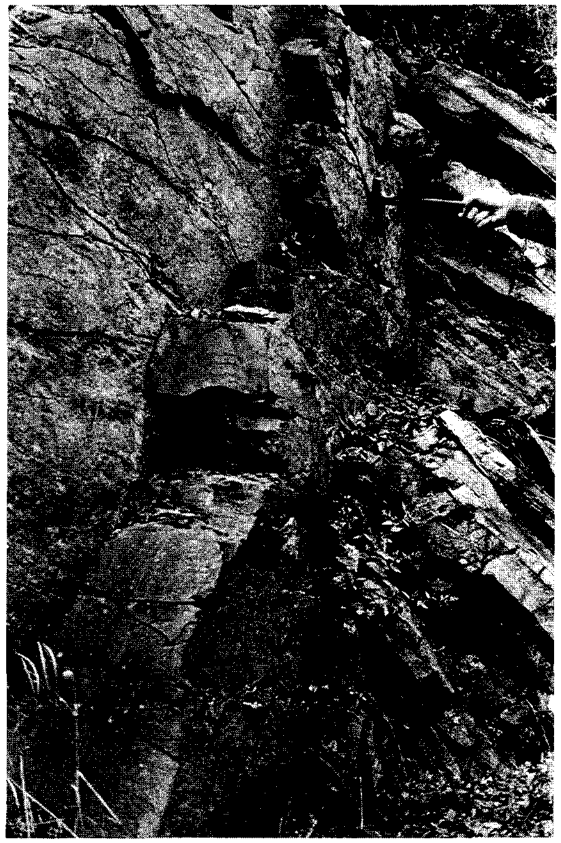

**Plate 9.** Lamprophyre dike from an outcrop on Highway 104 near the village of New Hampton. The rock behind the fracture on which the hammer head rests is Littleton schist. The dike turns sharply to the left at this point. Note the discordance between the dike and the foliation of the Littleton Formation which can be seen below the hammer.

# PETROCHEMISTRY

## Analytical Methods

Chemical analyses of 36 rock samples from the Holderness Quadrangle were made using a combination of X-ray fluorescence and atomic absorption methods. Fe, Ti, Ca, and K were analyzed by X-ray fluorescence, and Si, Al, Na, and Mg were determined by atomic absorption spectroscopy.

After grinding, sample splits for X-ray analysis were weighed, mixed with a heavy absorber ($La_2O_3$), and fluxed with $Li_2B_4O_7$ to form a homogeneous borosilicate glass. The glass was then ground to a fine powder and pressed into a flat cylindrical pellet which could be inserted directly into the sample holder of the X-ray unit. Standards were prepared from samples of known composition (G1 and W1) using the same procedure. Rose and Eggers (1969, unpublished ms., Dartmouth College) used this method to analyze 17 rocks for which previous analyses were available. They found that the accuracy of this method (assuming the earlier analyses were perfectly accurate) for the various oxides as follows: FeO $\pm$2.8% of the amount present; TiO $\pm$3.8%; CaO $\pm$4.7%. $K_2O$ $\pm$4.3%. Replicate analyses showed precision to be much better than accuracy.

Samples for atomic absorption analyses were prepared by fluxing a weighed portion of ground sample with $LiBO_2$. The resulting melt was poured directly into cold, dilute $HNO_3$, in which it dissolved readily upon stirring. The solutions were then diluted and analysed by the standard procedures for each element.

Rose (1969, unpublished ms., Dartmouth College) used this method on 40 previously analysed samples, and found systematic errors in the results for Si and Al. Si was low by an average of 1% of the amount present, while Al was low by an average of 2%. Analyses in Tables 3 and 4 therefore, have arbitrarily been corrected by adding 1% to the measured $SiO_2$ content and 2% to the measured $Al_2O_3$ content. Clark (1972) also noted a similar systematic error in Al. Data were not available for estimation of the accuracy of the Na and Mg determinations.

FeO was determined for some samples using the potassium dichromate method described by Reichen and Fahey (1962). The accuracy of this method was found to be approximately $\pm$10% of the amount of FeO present. Of 11 samples analysed by this method, 5 had positive values for $Fe_2O_3$, 5 had negative values, and 1 had 0% $Fe_2O_3$, which suggests that within the analytical error, the $Fe_2O_3$ content can be considered negligible. The positive values for $Fe_2O_3$ which were obtained are reported, but are not considered significant.

For some of the samples, volatile content was determined by loss on heating at 1400°C, while in others the $H_2O+$ was collected by passing the gasses expelled at 1400°C through a tube containing $CaCl_2$, which was weighed before and after the run.

The accuracy of sampling procedures for coarse-grained porphyritic rocks such as the Kinsman Quartz Monzonite has not been studied in detail by the writer. Baird et. al. (1967) present empirical results showing that for a homogeneous porphyritic quartz monzonite resembling the Kinsman, a 7/8 inch diameter core 12 inches long would provide a representative sample of the outcrop. This is the equivalent to a cubic sample slightly less than 2 inches on a side. All of the samples of porphyritic Kinsman analyzed in this study were at least this large. Although the Kinsman is obviously often inhomogeneous in a single outcrop, samples were collected from within portions which appeared to be homogeneous in a volume much larger than the sample volume. The variations in the analyses of Kinsman sample is therefore considered to be a significant measure of inhomogeneity within the Kinsman on a scale larger than hand specimen size.

## Results

The results of the analyses are presented in Tables 3 and 4, along with CIPW norms calculated for the rocks of the New Hampshire Plutonic Series. Four of 36 original analyses are not inlcuded, because their totals were less than 97%.

Table 3, Chemical Analyses and CIPW Norms of New Hampshire Plutonic Series Rocks from the Holderness Quadrangle

| Sample Number | Rock Type and Location |
|---|---|
| 10-69 | Kqm rich in feldspar and quartz; Highway 104, 1 mile north-east of junction with Highway 3b north. |
| 32-69 | Kqm; 1/4 mile east of Loon Island; Winnisquam Lake. |
| 33-69 | Kqm from contact zone (Dlk); 2 and 1/2 miles north of Highway 104; on Dana Church Road. |
| 7-70 | Biotite segregation in Kqm; Highway 104; north of Wicwas Lake. |
| 8-70 | Kqm; Highway 104; north of Pemigewasset Lake. |
| 10-70 | Moderately sheared Kqm; Highway 104; north of Wicwas Lake. |
| 12-70 | Garnet-bearing Kqm; Highway 104; north of Pemigewasset Lake. |
| 16-70 | Sheared Kqm surrounding fine-grained inclusion (17-70); Highway 104; north of Wicwas Lake. |
| 17-70 | Fine-grained inclusion in sheared Kqm (16-70); Highway 104; north of Wicwas Lake. |
| 1-71 | Kqm; Highway 104; north of Wicwas Lake. |
| 3-71 | Sheared Kqm (?); Highway 104; north of Wicwas Lake. |
| 4-71 | Sheared Kqm (?); Highway 104; north of Wicwas Lake. |
| 5-71 | Sheared Kqm (?) see Plate 8; Highway 104; north of Wicwas Lake. |
| 26-71 | Coarse-grained, somewhat foliated bqm; road intersection 1/3 mile northwest of Glove Hollow Brook; north of Plymouth Mountain. |

**Table 3**

Table 3, Chemical Analyses and CIPW Norms (Continued)

Chemical Analyses

| Sample | 10-69 | 32-69 | 33-69 | 7-70 | 8-70 |
|---|---|---|---|---|---|
| $SiO_2$ | 73.30 | 65.23 | 64.43 | 54.31 | 65.09 |
| $TiO_2$ | 0.28 | 0.94 | 1.29 | 2.73 | 0.76 |
| $Al_2O_3$ | 12.79 | 14.57 | 14.40 | 11.87 | 15.61 |
| $Fe_2O_3$ | 0.00 | 0.00 | 0.00 | 0.00 | nd |
| FeO | 1.49 | 5.78 | 6.69 | 18.04 | 4.35 |
| MgO | 0.37 | 0.74 | 1.59 | 3.26 | 1.87 |
| CaO | 1.33 | 2.24 | 2.71 | 1.79 | 4.53 |
| $Na_2O$ | 2.65 | 2.95 | 2.97 | 0.32 | 3.23 |
| $K_2O$ | 5.58 | 5.40 | 2.79 | 5.91 | 2.21 |
| $H_2O-$ | nd[1] | nd | nd | nd | nd |
| $H_2O+$ | nd | nd | nd | nd | nd |
| Additional[2] | 0.78 | 0.76 | 1.32 | 2.11 | 1.65 |
| Total | 98.57 | 98.61 | 98.19 | 100.34 | 99.30 |

CIPW Norms

| | | | | | |
|---|---|---|---|---|---|
| Qu | 32.13 | 17.54 | 23.69 | 8.08 | 22.51 |
| Or | 33.03 | 31.91 | 16.51 | 34.97 | 13.07 |
| Ab | 22.37 | 24.94 | 25.1 | 2.72 | 27.3 |
| An | 6.48 | 10.51 | 13.46 | 8.9 | 21.52 |
| Di | 0.09 | 0.54 | 0.00 | 0.00 | 0.82 |
| Hy | 3.16 | 10.63 | 14.12 | 36.73 | 10.99 |
| Il | 0.53 | 1.78 | 2.45 | 5.18 | 1.44 |
| Mt | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Cor | 0.00 | 0.00 | 1.55 | 1.66 | 0.00 |
| Others[3] | 0.78 | 0.76 | 1.32 | 2.11 | 1.65 |

1. nd - not determined.
2. Additional - additional loss on heating at 1400°C.
3. Equals sum of $H_2O-$, $H_2O+$, and Additional.

Table 3 (Continued)

Chemical Analyses

| Sample | 10-70 | 12-70 | 13-70 | 16-70 | 17-70 |
|---|---|---|---|---|---|
| $SiO_2$ | 67.82 | 62.31 | 69.90 | 60.54 | 65.35 |
| $TiO_2$ | 0.79 | 0.56 | 0.39 | 1.35 | 0.90 |
| $Al_2O_3$ | 13.12 | 16.42 | 14.65 | 15.61 | 16.09 |
| $Fe_2O_3$ | 0.58 | 0.09 | 0.07 | nd[1] | nd |
| FeO | 4.15 | 4.00 | 1.96 | 8.18 | 4.76 |
| MgO | 0.82 | 0.68 | 0.40 | 1.67 | 1.49 |
| CaO | 1.59 | 1.31 | 1.08 | 3.39 | 3.62 |
| $Na_2O$ | 3.15 | 3.15 | 3.07 | 3.61 | 4.07 |
| $K_2O$ | 4.56 | 6.80 | 5.86 | 2.69 | 1.90 |
| $H_2O-$ | nd | nd | 0.56 | 1.04 | tr[2] |
| $H_2O+$ | nd | nd | 0.30 | 0.67 | 0.60 |
| Additional[3] | 1.92 | 3.28 | 0.36 | 0.47 | 0.17 |
| Total | 98.40 | 98.60 | 98.60 | 99.22 | 98.95 |

CIPW Norms

| Sample | 10-70 | 12-70 | 13-70 | 16-70 | 17-70 |
|---|---|---|---|---|---|
| Qu | 25.18 | 11.29 | 25.42 | 13.72 | 21.19 |
| Or | 26.35 | 40.2 | 34.64 | 15.90 | 11.23 |
| Ab | 26.62 | 26.62 | 25.94 | 30.5 | 34.37 |
| An | 7.9 | 6.51 | 5.37 | 16.82 | 17.96 |
| Di | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Hy | 7.87 | 8.04 | 3.89 | 16.94 | 10.97 |
| Il | 1.5 | 1.06 | 0.74 | 2.57 | 1.7 |
| Mt | 0.84 | 0.14 | 0.09 | 0.00 | 0.00 |
| Cor | 0.2 | 1.48 | 1.28 | 0.58 | 0.74 |
| Others[4] | 1.92 | 3.28 | 1.22 | 2.18 | 0.77 |

1. nd - not determined.
2. tr - trace.
3. Additional - additional loss on heating at 1400°C.
4. Other - equals sum of $H_2O-$, $H_2O+$, and Additional.

Table 3 (Continued)

Chemical Analyses

| Sample | 1-71 | 3-71 | 4-71 | 5-71 | 26-71 |
|---|---|---|---|---|---|
| $SiO_2$ | 66.46 | 56.30 | 63.22 | 60.21 | 63.43 |
| $TiO_2$ | 0.39 | 1.39 | 0.87 | 1.07 | 0.96 |
| $Al_2O_3$ | 16.39 | 17.95 | 17.10 | 17.77 | 16.79 |
| $Fe_2O_3$ | nd[1] | nd | nd | nd | nd |
| FeO | 2.37 | 8.33 | 5.02 | 6.67 | 5.16 |
| MgO | 0.43 | 2.01 | 1.27 | 2.27 | 1.69 |
| CaO | 2.01 | 4.11 | 2.84 | 4.58 | 3.27 |
| $Na_2O$ | 3.52 | 4.02 | 4.69 | 3.70 | 3.41 |
| $K_2O$ | 5.97 | 3.08 | 4.19 | 2.74 | 2.51 |
| $H_2O-$ | nd | 0.23 | tr[2] | tr | tr |
| $H_2O+$ | nd | 1.16 | nd | 0.43 | 0.62 |
| Additional[3] | 2.63 | 0.90 | 0.97 | 0.36 | 0.70 |
| Total | 100.17 | 99.48 | 100.17 | 99.80 | 98.54 |

CIPW Norms

| | | | | | |
|---|---|---|---|---|---|
| Qu | 16.52 | 3.44 | 8.63 | 10.27 | 20.9 |
| Or | 35.31 | 18.24 | 24.8 | 16.18 | 14.85 |
| Ab | 29.76 | 33.96 | 39.61 | 31.28 | 28.82 |
| An | 9.98 | 20.41 | 13.18 | 22.74 | 16.24 |
| Di | 0.00 | 0.00 | 0.78 | 0.00 | 0.00 |
| Hy | 4.77 | 18.00 | 10.53 | 16.13 | 12.1 |
| Il | 0.74 | 2.64 | 1.66 | 2.04 | 1.82 |
| Mt | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Cor | 0.46 | 0.51 | 0.00 | 0.37 | 2.5 |
| Other[4] | 2.63 | 2.29 | 0.97 | 0.79 | 1.32 |

1. nd - not determined.
2. tr - trace.
3. Additional - additional loss on heating at 1400°C.
4. Other - equals sum of $H_2O-$, $H_2O+$, and Additional.

Table 4 Chemical Analyses of Rocks of the Littleton Formation

| Sample Number | Rock Type and Location |
| --- | --- |
| 12-71* | Quartz-muscovite-garnet-chloritoid schist, Dl; near contact with bqm, north flank of Plymouth Mountain. |
| 13-71 | Muscovite-chlorite-garnet-tourmaline (with little quartz), Dl; near contact with bqm, north flank of Plymouth Mountain. |
| 15-71* | Thin-bedded quartz-biotite-garnet-muscovite schist, Dl; Highway 3b, 3 miles north of Highway 104. |
| 21-71* | Fine-grained quartz-biotite-muscovite schist, Dl; Highway 175, 1 mile north-northwest of Church Hill. |
| 23-71 | Fine-grained quartz-biotite-muscovite-garnet schist, Dl; Steele Hill. |
| 29-71 | Quartz-rich, biotite-muscovite-garnet schist, Dl; Interstate 93, 1 mile south of New Hampton-Meredith town line. |
| 31-71 | Quartz-garnet-biotite schist, Dl, surrounding segregation pod (#30-71), Interstate 93, 1 1/2 miles north of Ashland exit (See Table 6). |
| 30-69 | Quartz-biotite-garnet-sillimanite-muscovite schist, Dl, surrounding calc-silicate boudins (#30a-69); Interstate 93 at Meredith-Sanbornton town line. |
| 30a-69 | Calc-silicate boudin, Dl; enclosed in schist (#30-69); Interstate 93 at Meredith-Sanbornton town line. |
| 35-69* | Calc-silicate boudin, Dl, Interstate 93 due west of North Sanbornton. |
| 15-70 | Calc-silicate boudin, Dl; Interstate 93 1 1/2 miles north of Ashland exit. |
| 31-69 | Graphitic quartz-muscovite-biotite-plagioclase schist, Clay Brook Member (Dlc); Highways 3 and 25, 1/2 mile northeast of junction with River Road. |
| 11-71 | Quartz-plagioclase-diopside-actinolite-epidote granofels, Clay Brook Member (Dlc); Highways 3 and 25, 1/2 mile northeast of junction with River Road. |
| 24-71 | Sulfidic, graphitic quartz-feldspar granofels, Clay Brook Member (Dlc?); Interstate 93, 1 mile north of Ashland exit. |

*See table 1 for estimated mode.

**Table 4**

Table 4, Chemical Analyses (Continued)

| Sample | 12-71 | 13-71 | 15-71 | 21-71 | 23-71 | 29-71 | 31-71 |
|---|---|---|---|---|---|---|---|
| $SiO_2$ | 50.48 | 39.95 | 57.21 | 66.21 | 71.51 | 76.08 | 52.09 |
| $TiO_2$ | 1.29 | 1.66 | 1.20 | 0.88 | 0.80 | 0.84 | 1.63 |
| $Al_2O_3$ | 25.62 | 33.07 | 18.69 | 13.49 | 12.17 | 10.34 | 24.41 |
| $Fe_2O_3$ | nd[1] | nd | nd | nd | nd | nd | nd |
| FeO | 10.46 | 9.50 | 7.77 | 5.06 | 4.19 | 4.64 | 9.84 |
| MgO | 2.16 | 1.66 | 2.84 | 3.43 | 1.45 | 1.71 | 3.89 |
| CaO | 0.26 | 0.48 | 0.61 | 2.78 | 0.52 | 0.76 | 0.65 |
| $Na_2O$ | 0.83 | 1.26 | 0.90 | 2.09 | 0.84 | 1.30 | 0.93 |
| $K_2O$ | 3.83 | 6.17 | 3.64 | 2.82 | 3.68 | 1.72 | 4.39 |
| $H_2O$- | 0.08 | tr[2] | tr | nd | 0.26 | tr | nd |
| $H_2O$+ | 3.67 | 4.16 | 3.39 | 0.97 | 0.12 | 0.42 | 1.38 |
| Additional[3] | 1.14 | 1.43 | 0.52 | 0.68 | 2.45 | 0.84 | 0.33 |
| Total | 99.82 | 99.34 | 96.77 | 98.41 | 97.99 | 98.65 | 99.54 |

| Sample | 30-69 | 30a-69 | 35-69 | 15-70 | 31-69 | 11-71 | 24-71 |
|---|---|---|---|---|---|---|---|
| $SiO_2$ | 70.80 | 68.75 | 83.18 | 64.94 | 65.18 | 60.13 | 56.86 |
| $TiO_2$ | 0.91 | 0.76 | 0.34 | 0.64 | 1.11 | 0.85 | 0.72 |
| $Al_2O_3$ | 11.82 | 14.05 | 6.07 | 9.50 | 16.11 | 14.06 | 13.55 |
| $Fe_2O_3$ | nd | nd | 0.23 | nd | 0.00 | nd | nd |
| FeO | 6.94 | 4.98 | 2.45 | 4.40 | 1.62 | 5.80 | 9.17 |
| MgO | 1.73 | 1.15 | 0.66 | 1.61 | 3.02 | 2.26 | 2.26 |
| CaO | 0.88 | 4.72 | 4.62 | 17.41 | 1.90 | 10.05 | 5.46 |
| $Na_2O$ | 1.99 | 0.00 | 0.00 | 0.00 | 2.00 | 0.77 | 0.98 |
| $K_2O$ | 2.34 | 2.29 | 0.14 | 0.01 | 4.27 | 0.64 | 1.71 |
| $H_2O$- | nd | 0.25 | nd | tr | nd | nd | nd |
| $H_2O$+ | nd | 1.17 | nd | 0.24 | nd | 1.22 | 0.72 |
| Additional | 1.50 | 0.56 | 0.51 | tr | 5.23 | 2.2 | 5.12 |
| Total | 98.91 | 98.68 | 98.20 | 98.75 | 100.44 | 98.00 | 96.55 |

1. nd - not determined.
2. tr - trace.
3. Additional - additional loss on heating at 1400°C.

# PETROGENESIS OF THE WINNIPESAUKEE PLUTON

## General Statement

The origin of the Kinsman Quartz Monzonite is a problem of long standing. The Kinsman was originally called the "porphyritic gneiss" by Hitchcock (1877) who believed that the Kinsman consisted of metamorphosed sediments or volcanics. However, Daly (1897) presented a strong case for a magmatic origin, arguing that the contact relations and presence of numerous "horses" or xenoliths of the surrounding rocks within the "porphyritic granite" were characteristic of intrusive rocks. Daly considered the foliation to be a magmatic flow foliation formed subsequent to the deformation of the schists, and suggested that the concordance of foliations between the "porphyritic granite" and rocks at some contacts "is probably owing to an exchanging of the usual batholithic form of intrusion for a sill of sheet-form" (Daly, 1897, p. 787).

Billings (1937, 1956) supported a magmatic origin for the Kinsman Quartz Monzonite, but Chapman (1952) argued that the Kinsman and the Bethlehem Gneiss were formed by granitization of portions of the Littleton Formation. Chapman showed that while the contacts were not always concordant, the foliations of the Littleton Formation were generally consistent with foliations in the Kinsman and the Bethlehem Gneiss. He took these foliations to be relict features preserved during the granitization process. He also cited low sodium content of the K-feldspars in the Kinsman as evidence of metamorphic or metasomatic, rather than magmatic origin. Chemical analyses then available showed the Kinsman and Bethlehem to be higher in $Na_2O$ and CaO than the Littleton, so Chapman assumed a large-scale migration of sodium and calcium through the Littleton Formation to account for its transformation into Kinsman-type rocks. In those portions of the Littleton where the K-feldspar porphyroblasts had already formed, some of the porphyroblasts were replaced by plagioclase to form the Kinsman and Bethlehem Formations.

More recently, Thompson et. al. (1968) have suggested that the Kinsman Quartz Monzonite and Bethlehem Gneiss might have formed from partially mobilized ignimbrite sheets within the Littleton Formation. In Maine, ignimbrite sheets such as the Lower Devonian Kineo Rhyolite are similar to the Kinsman and Bethlehem in size, age, and chemical composition.

Plutonic Series rocks from Table 3, and can best be interpreted as demonstrating differentiation of a calc-alkaline magma. Figure 7 is a plot of $Na_2O$ vs. $K_2O$ for all available analyses of rocks of the New Hampshire Plutonic Series and the Littleton formation (Billings and Wilson, 1964; Clark, 1972; and Tables 3 and 4). A clear distinction can be made between the two groups on the basis of $Na_2O$ content.

Of particular interest here are the three analyses of "Kqm"? (samples 3-71, 4-71, and 5-71 of Table 3) which are taken from the schistose varieties of Kinsman. Sample #5-71 (Table 3) was taken from the outcrop shown in Plate 7. The others are from the same outcrop, but not shown in the photographs. If either granitization or assimilation had partially converted Littleton into Kinsman, one would expect these three analyses to be intermediate in $Na_2O$ content between Littleton and Kinsman. That these samples have higher $Na_2O$ content than the average New Hampshire Plutonic Series rocks indicates that they are not likely to have been derived from Littleton Formation rocks by addition of $Na_2O$. This evidence, together with the observed sharp contacts and unaltered inclusions near the edge of the pluton, appears to eliminate the possibility that any of the Kinsman samples was derived from rocks of the Littleton Formation, either by granitization or by assimilation of inclusions.

From Figure 7, two observations can be made in comparing New Hampshire Series rocks of the Holderness Quadrangle with those elsewhere in the state. First, the average $Na_2O$ content of the Holderness Quadrangle rocks is somewhat higher; and second, the compositional variation particularly in $K_2O$, is much greater. This greater variation in composition is probably due to the fact that samples of the Kinsman analysed during the present study were selected primarily because they showed some evident deviation from the norm, and were not intended to be typical of the bulk composition of the Kinsman. Hopefully, these analyses can be considered a measure of the extreme range of composition within the Kinsman.

If the compositional variation of the Kinsman is not a result of granitization or assimilation, then it must be due to differentiation from a presumably homogeneous original magma. Figure 8 shows a plot of normative An-Ab-Or for analyzed Kinsman samples from the Holderness Quadrangle. The points all lie in the field of initial crystallization of plagioclase, or near the low temperature trough. In the system Q-Ab-Or-An (Fig. 9), the points show a greater scatter, but again fall within the plagioclase field (for [Q+Ab+Or] 95 [An]5) (James and Hamilton, 1969). In a three dimensional plot in the tetrahedron Q-Ab-An-Or, with An at the apex the points would be approximately on a plane including the points $Q_{100}$, $Ab_{55}An_{45}$, and $Ab_{35}Or_{65}$. Hypothetical crystallization paths for liquids of this composition range are shown on Figures 8 and 9. In the system $[Q\text{-}Ab\text{-}Or]_{95}[An]_5$ shown in Figure 9, crystallization would begin with plagioclase, probably andesine, and the composition of the liquid would move along a line including it and the bulk composition. The plagioclase would become more sodic as crystallization proceeded. Orthoclase and plagioclase

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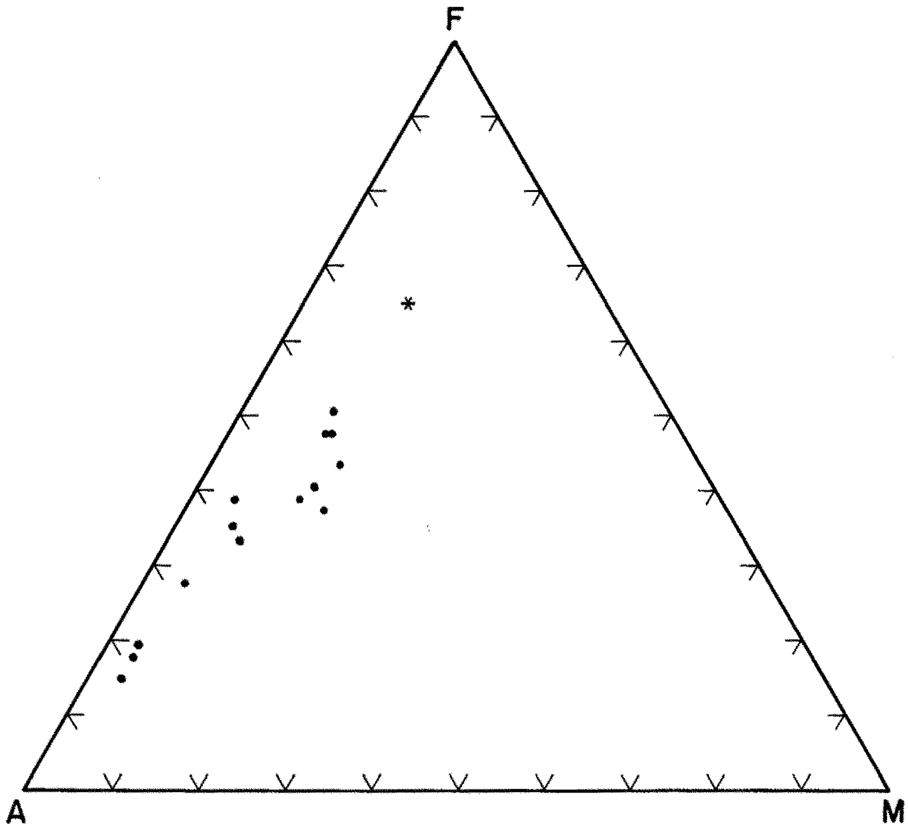



**Figure 6.** AFM diagram of New Hampshire Plutonic Series rocks from Table 3. The solid circles represent New Hampshire Series rocks; the asterisk represents sample #7-70, a biotite segregation in the Kinsman Quartz Monzonite.

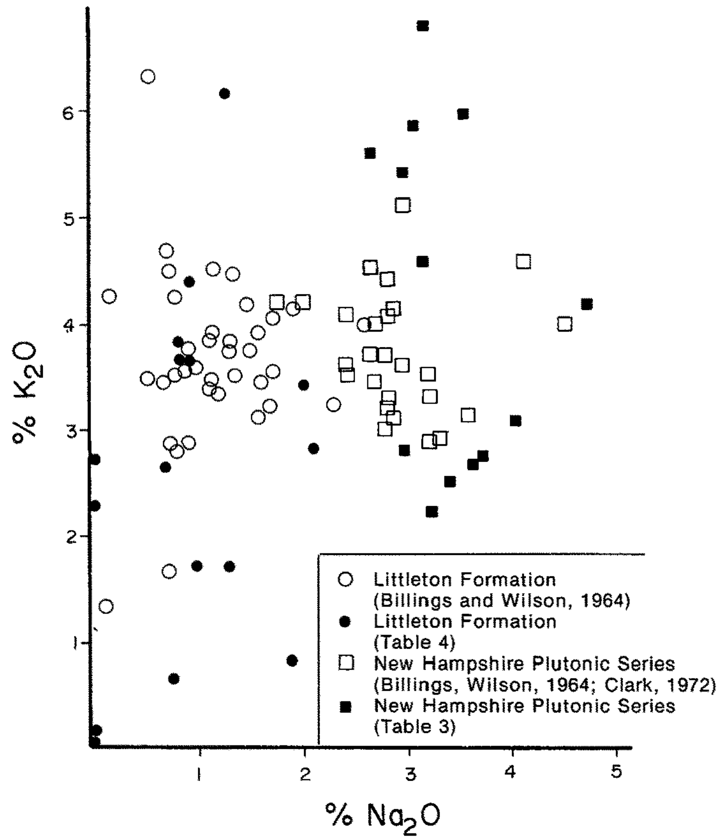



**Figure 7.** $Na_2O$ vs. $K_2O$ for all available analyses of Littleton Formation and New Hampshire Plutonic Series rocks. Analyses from Billings and Wilson, 1964; Clark, 1972; and Tables 3 and 4.

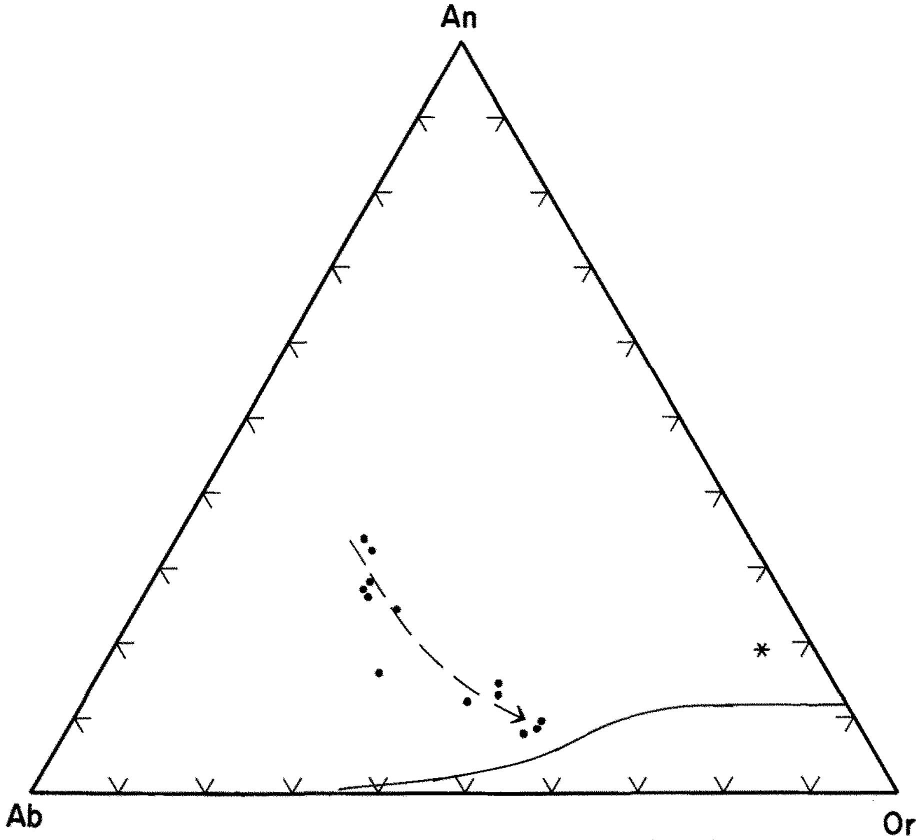



**Figure 8.** Normative An-Ab-Or diagram of New Hampshire Plutonic Series rocks from Table 3. Phase boundary separating plagioclase and alkali feldspar fields at 5000 bars $P_{H_2O}$ after Yoder et. al., 1957). Arrow shows hypothetical crystallization path. Symbols as in Figure 6.

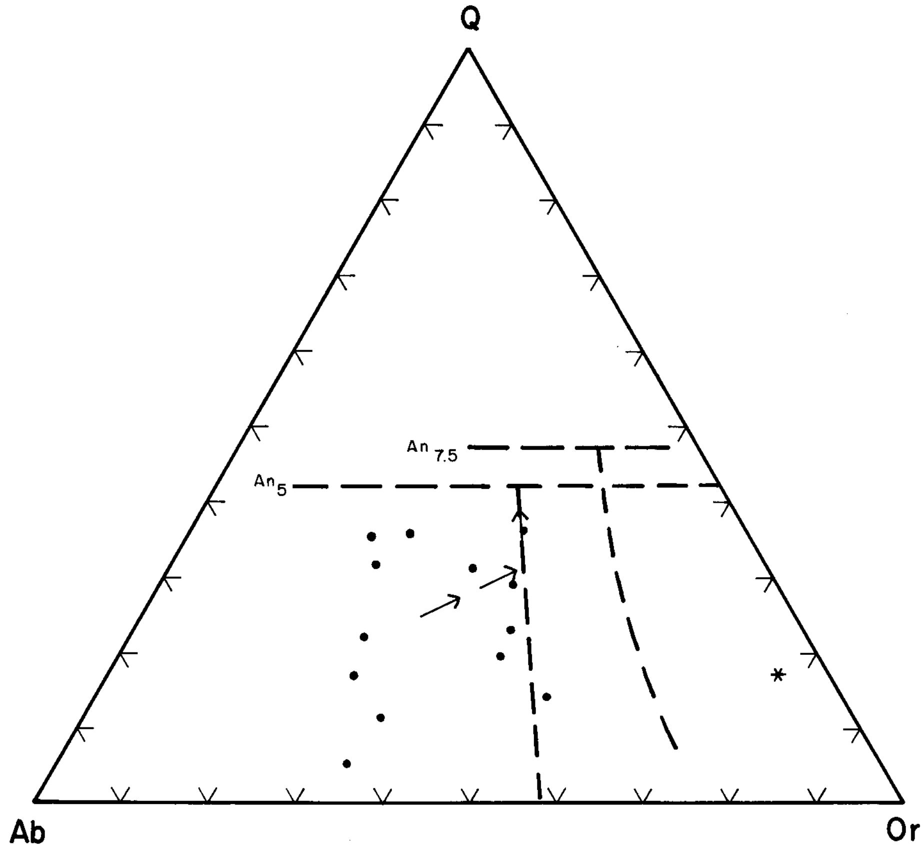



**Figure 9.** Normative Q-Ab-Or diagram of New Hampshire Plutonic Series rocks from Table 3, plotted on the experimental Q-Ab-Or-An system at 1 kilobar $H_2O$ pressure, after James and Hamilton (1968). Minima shown are piercing points, rather than ternary eutectics. Arrows show hypothetical crystallization path at $An_5$. Symbols as in Figure 6.

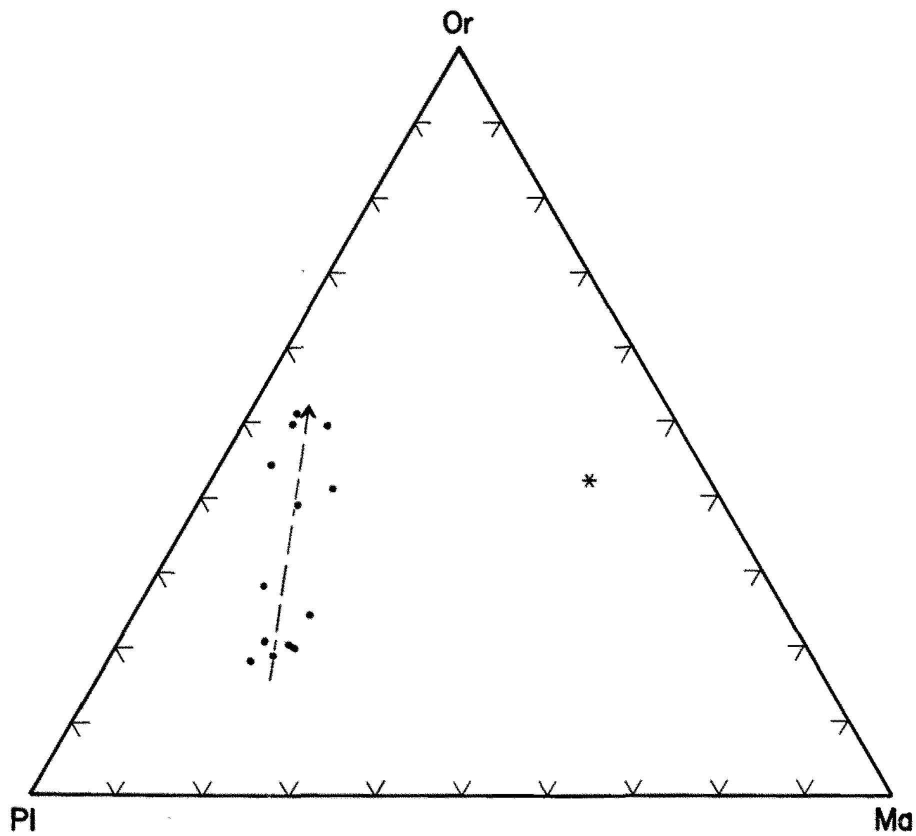



**Figure 10.** Normative Orthoclase-Plagioclase-Mafics diagram of New Hampshire Plutonic Series rocks from Table 3. Arrow shows magmatic differentiation trend. Or-Orthoclase, Pl-Plagioclase, Ma-mafics. Symbols as in Figure 6.

would next crystallize simultaneously, and the liquid composition would move toward the ternary piercing point. The final stage would involve simultaneous crystallization of quartz, orthoclase, and plagioclase.

Most, if not all, of the variation in composition of the Kinsman can be accounted for by crystallization of an initially homogeneous magma along the path described above, affected only by gravitational or mechanical concentration of some of the early-formed plagioclase in parts of the magma, thereby increasing the fraction of orthoclase in the remainder. This interpretation is consistent with relations shown by Figures 8 and 10. Figure 8 shows that the normative An content of the rocks decreases with increasing orthoclase, as would be expected of the earlier more calcic plagioclase were removed, and Figure 10 suggests that the mafic fraction began crystallization early and was concentrated along with plagioclase. Mechanical concentration in the Kinsman is illustrated by a biotite segregation (Table 3, #7-70). This sample is plotted on Figures 8, 9, and 10 as an asterisk, and falls far outisde of the normal range of composition for Kinsman and Littleton rocks.

The presence of K-feldspar megacrysts in the Kinsman is an enigma. Phenocrysts in a magmatic rock are usually thought to indicate an initial slow period of cooling during which a relatively few of the first-formed crystals are nucleated and grow to large size. This is followed by a period of more rapid cooling (usually due to extrusion or to intrusion into cooler rocks) during which the remainder of the liquid is crystallized rapidly to form a finer-grained matrix surrounding the phenocrysts. We have seen, however, that the composition of the Kinsman is such that plagioclase should be the first mineral to crystallize, and should presumably have formed the largest crystals.

Field evidence relating to the time of formation of the K-feldspar megacrysts is ambiguous. The Kinsman contains occasional patches characterized by very high concentrations of K-feldspar; Clark (1972) has observed this feature in the Cardigan Pluton as well. These concentrations could be explained either as the result of filter-pressing a liquid containing early K-feldspar crystals, or by crystallization of residual pockets of potassium-rich liquid after separation of early-formed plagioclase. Evidence for post-magmatic growth of K-feldspars can be seen in the sheared Kinsman rocks shown in Plate 9. Intense shearing has taken place parallel to the foliation, orienting and deforming many of the K-feldspar megacrysts. Some of the megacrysts however, are large, undeformed, and randomly oriented with respect to the foliation, suggesting that they grew *in situ* after the shearing had occurred. Because the rock must have been almost completely crystalline to support shearing stresses capable of deforming the megacrysts, any subsequent crystal growth would have been very late or post-magmatic. Thus, some of the late-formed crystals are metamorphic porphyroblasts.

Four samples of K-feldspar from Kinsman rocks of various textures were all found, using the computer refinement of Wright and Stewart (1968), to be maximum microcline (approximately $Or_{90}Ab_{10}$). A small, but undetermined amount of albite was also observed as an exsolved phase within the megacrysts and/or in the myrmekitic rims. The $Or_{90}Ab_{10}$ composition suggests crystallization in the low-temperature solvus of the AbOr system (Thompson and Waldbaum, 1969). The inversion from monoclinic feldspar to triclinic microcline occurs in the range of 400 to 495°C at high pressures (Tomisaka, 1962). This is considerably below the range of magmatic temperatures. The long period of cooling to be expected in a pluton at depth would probably allow the inversion to take place, regardless of the initial process of formation of the K-feldspar.

Although the chemical composition of the Kinsman indicates that plagioclase should have crystallized first, the generally larger size of the K-feldspars might be attributed to favored growth due to such factors as volatiles, to an abnormal cooling history, or to nucleation processes and rates of crystal growth in the magma. K-feldspar growth subsequent to the shearing process may have been aided by residual solutions squeezed out of the rock during deformation, and does not account for a significant percentage of the K-feldspar in the Kinsman as a whole. It was during a post-intrusive deformational episode that the formation apparently acquired its preferred north-northeast-trending foliation.

The principal difficulties which remain in trying to assign a magmatic origin to the Winnipesaukee pluton are the K-feldspar megacrysts problem, and the problem of explaining the features such as the layered structure in Plate 8. As we have seen, the K-feldspar megacrysts are not consistent with simple crystallization from the melt in the system Q-Or-Ab-An, but this problem is not unique to the Kinsman. For example, some rapakivi granites, although characterized by K-feldspar megacrysts, have compositions such that plagioclase should have crystallized first (Vorma, 1971). It is hoped that further experimental work in multicomponent systems will ultimately determine the factors which control feldspar crystallization and growth.

As we show in the discussion of the structure of the Winnipesaukee pluton, the Kinsman was probably intruded and began crystallizing during an early period of Acadian deformation and was deformed again near the end of the crystallization period. Its megascopic textural and compositional variations can then be explained by a combination of crystal fractionation and shearing process.

# GRAVITY STUDIES

## Previous Gravity Surveys

A gravity survey was made in the Holderness Quadrangle to aid in the interpretation of the subsurface structure of the Winnipesaukee pluton. Two previous regional gravity surveys (Bean, 1953; Joyner, 1963) include the Holderness Quadrangle and surrounding areas in central New Hampshire. Bean surveyed an east-west strip two quadrangles wide from New York to Maine, including the northern three-quarters of the Holderness Quadrangle. Joyner surveyed north and south of Bean's area, from central New Hampshire east to the coast, and after adjusting all of Bean's data by a value of two milligals to correct for a systematic discrepency between the surveys, he produced a composite regional Bouguer anomaly map.

Both Bean and Joyner published Bouguer anomaly maps contoured at 5 milligals. When gravity profiles across the pluton contacts in the Holderness Quadrangle were constructed from their contour maps, they were found to be in sufficient disagreement so as to make attempts at interpretation fruitless, particularly with respect to the structure of the contacts.

It was therefore necessary to make an additional survey across the pluton contacts. This was carried out by the writer in the spring of 1970, using a Worden Gravimeter, Model #308. Stations for this survey (Fig. 11) were concentrated along a line from Meredith south along the shores of Lake Winnisquam, as far as the dam at the southern end of the lake. The two southernmost points (Stations 11 and 12, Table 5) are in the Penacook Quadrangle and not shown in Figure 11. The lake provided a convenient reference level for most of these points. Additional stations were scattered more widely to the west and north, straddling much of the Kinsman-Littleton contact (Plate 1).

## Method of Survey and Data Reduction

Table 5 shows the data collected and the corrections made. The elevation correction is a combination of the free air and the simple Bouguer corrections. The density for the simple Bouguer correction is assumed to be 2.67 gm/cm$^3$, which results in an elevation correction factor for this latitude of .060134 milligals/foot.

Elevations were measured relative to the surface of Lake Winnisquam, which was assumed to be 482 feet as shown on the topographic map. The outlet of the lake is controlled by a dam, so the level is probably quite constant except under unusually dry conditions. For stations along the shore, accuracy is within ±0.5 feet. Stations away from the shore of Winnisquam were located at bench marks or at road intersections whose elevations were indicated on the 1956 United States Geological Survey topographic map of the Holderness Quadrangle.

The latitude correction of —1.3057 milligals/mile N was made with reference to 43°30′ N latitude, the southern boundary of the Holderness Quadrangle.

The instrument correction compensates for daily and day-to-day instrumental drift. One station was occupied twice or preferably three times during the day, and a graph of the variation in gravity reading with time was plotted; from this the correction for daily variation could be determined. At least one station had to have been occupied on a previous occasion, preferably the same one used above, in order to establish the day-to-day instrumental drift.

The corrected values listed in Table 5 are relative to an arbitrary datum. Station #30, located at a U.S.G.S. bench mark on the south side of the cannon (mortar) in the village square in Bristol, has been measured at 980,414.6 milligals absolute gravity by Borns (1972, unpublished ms., Dartmouth College). This value, obtained by comparison with the Farmington, New Hampshire station used by Joyner (1963) may be used to relate this survey to others.

The simple Bouguer anomaly values were contoured at a 2 milligal interval (Fig. 12) and proved to be in good agreement with Joyner (Fig. 13). It was noted from Joyner's (1963) regional map that there is a definite linear trend parallel to the Atlantic coast in addition to an effect of the White Mountain Batholith. A contour map of the regional gravity trend was constructed from Joyner's map by smoothing out his contours, ignoring all local fluctuations. In the Holderness Quadrangle, this regional trend can be approximated by a planar surface striking N 70 E and dipping 0.26 milligals/km to the northwest (Fig. 12). The values for all stations were contoured to remove this trend, and recontoured to produce a residual Bouguer anomaly map (Fig. 14).

Observed gravity profiles (Fig. 15) were constructed from the residual anomaly map along sections A-A′ and B-B′ (Fig. 14). Section B-B′ has been extended to include data taken south of the quadrangle boundary. Figure 15 shows the observed gravity ±1 milligal. The maximum variation between replicate field measurements of gravity was 0.36 milligal. No terrain corrections were made. In the Mt. Kearsarge Quadrangle, an area of greater relief than Holderness, Clark (1972) made 19 terrain corrections whose mean was 0.539 milligal. The maximum deviation from the mean, which is the maximum error introduced by not making terrain corrections, is 0.63 milligal. The profiles of Figure 15 include an error due to contouring which is difficult to estimate because it varies with the spacing of the stations. The estimate of ±1 milligal for the observed profiles, however, appears to be realistic.

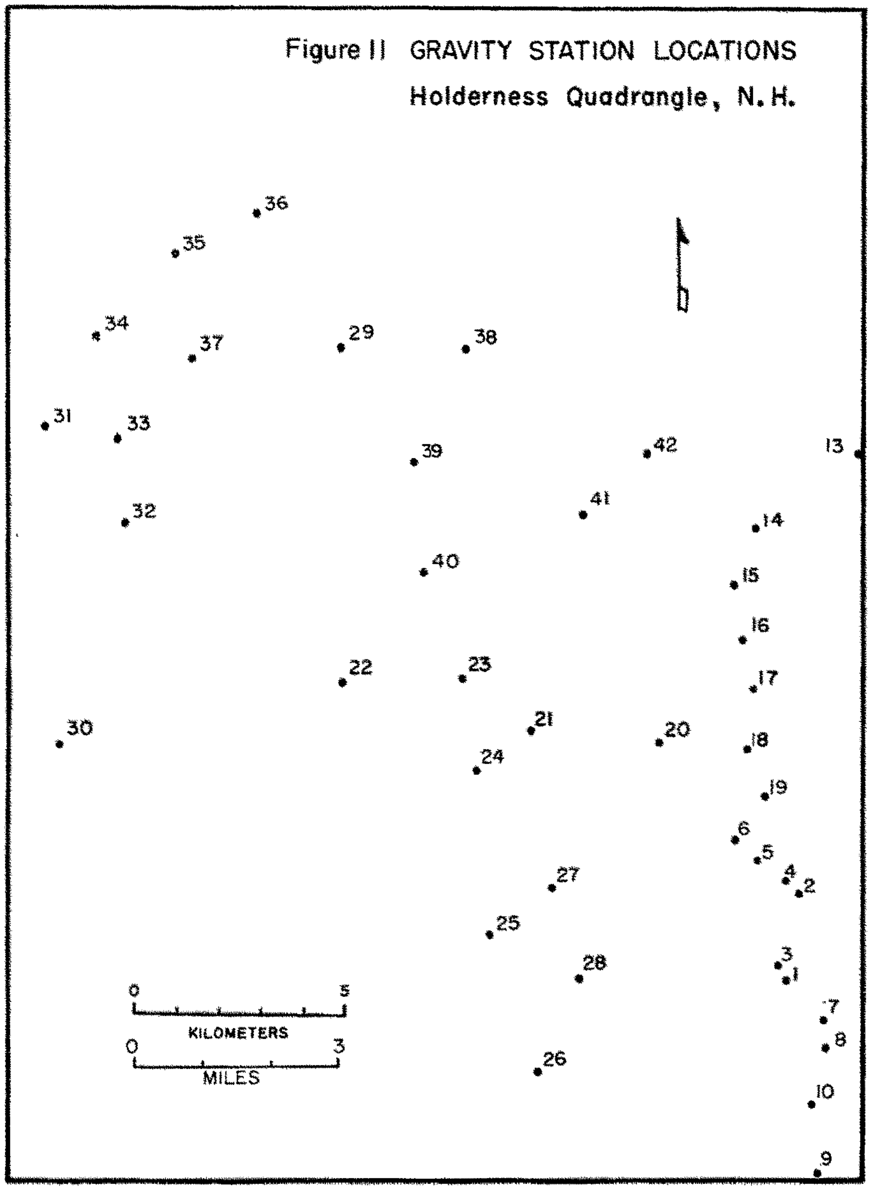



**Figure 11.** Gravity station locations.

Table 5 Gravity Data for Stations Shown in Figure 11*

| Station | Average Instru. Reading (Scale Dw) | X.6455 Mgal per Scale Division (mg) | Elevation Relative to Lake Winnisquam (482 ft.) | Elevation Correction (mg) | Miles N. or S. of 43°30' | Latitude Correction (mg) | Instrument Correction (mg) | Mgals. |
|---|---|---|---|---|---|---|---|---|
| 1 | 810.4 | 523.11 | 4.0 | .24 | 2.9 N | -3.79 | 0 | 519.6 |
| 2 | 811.467 | 523.80 | 0.5 | .03 | 4.21 N | -5.50 | + .02 | 518.4 |
| 3 | 811.95 | 524.11 | 0.0 | --- | 3.2 N | -4.18 | + .01 | 519.9 |
| 4 | 811.1 | 523.57 | 2.3 | .14 | 4.41 N | -5.76 | + .03 | 518.0 |
| 5 | 810.367 | 523.09 | 2.3 | .14 | 4.73 N | -6.18 | + .04 | 517.1 |
| 6 | 809.167 | 522.32 | 2.4 | .14 | 5.01 N | -6.54 | + .05 | 516.0 |
| 7 | 809.967 | 522.83 | 0.7 | .04 | 2.3 N | -3.00 | - .05 | 519.8 |
| 8 | 809.833 | 522.75 | 0.5 | .03 | 1.92 N | -2.51 | - .08 | 520.2 |
| 9 | 810.933 | 523.46 | 0.7 | .04 | 0.15 N | - .20 | -2.46 | 521.0 |
| 10 | 812.9 | 524.73 | 4.0 | .24 | 1.05 N | -1.37 | -2.53 | 521.1 |
| 11 | 811.067 | 523.54 | 1.5 | .09 | 0.6 S | + .78 | -2.67 | 521.7 |
| 12 | 807.1 | 520.98 | 1.3 | .08 | 1.93 S | +2.52 | -2.72 | 520.9 |
| 13 | 811.425 | 523.77 | 1.0 | .06 | 11.0 N | -14.36 | -2.83 | 506.6 |
| 14 | 788.9 | 509.23 | 255.0 | 15.34 | 9.62 N | -12.56 | -2.85 | 509.2 |
| 15 | 797.7 | 514.92 | 150.0 | 9.14 | 8.76 N | -11.44 | -2.86 | 509.8 |
| 16 | 808.767 | 522.06 | 43.1 | 2.59 | 7.94 N | -10.37 | -2.86 | 511.4 |
| 17 | 813.1 | 524.86 | 0.0 | ---- | 7.20 N | -9.40 | -2.87 | 512.6 |
| 18 | 812.4 | 524.40 | 1.0 | .06 | 6.27 N | -8.19 | -2.87 | 513.4 |
| 19 | 812.0 | 524.15 | 2.0 | .12 | 5.65 N | -7.38 | -2.88 | 514.0 |
| 20 | 797.167 | 514.57 | 193.0 | 11.61 | 6.45 N | -8.42 | -4.99 | 512.8 |
| 21 | 769.033 | 496.41 | 526.0 | 31.63 | 6.64 N | -8.67 | -5.14 | 514.2 |
| 22 | 819.833 | 529.20 | 49.0 | 2.95 | 7.31 N | -9.54 | -4.82 | 517.8 |
| 23 | 773.933 | 499.57 | 495.0 | 29.77 | 7.36 N | -9.61 | -4.85 | 514.9 |
| 24 | 773.567 | 499.34 | 547.0 | 32.90 | 5.99 N | -7.82 | -4.89 | 519.5 |
| 25 | 810.4 | 523.11 | 149.7 | 9.00 | 3.62 N | -4.73 | -5.03 | 522.4 |
| 26 | 800.567 | 516.77 | 224.7 | 13.51 | 1.55 N | -2.02 | -5.05 | 523.2 |
| 27 | 800.3 | 516.59 | 268.0 | 16.12 | 4.30 N | -5.61 | -5.09 | 522.0 |
| 28 | 780.167 | 503.597 | 479.0 | 28.81 | 2.95 N | -3.85 | -5.13 | 523.4 |
| 29 | 822.8 | 531.12 | 75.0 | 4.51 | 12.27 N | -16.02 | -4.99 | 514.6 |
| 30 | 826.4 | 533.44 | -16.5 | -0.99 | 6.35 N | -8.29 | -5.01 | 519.2 |
| 31 | 808.867 | 522.12 | 104.0 | 6.25 | 11.10 N | -14.49 | -5.03 | 508.9 |
| 32 | 744.433 | 480.581 | 923.0 | 55.51 | 9.67 N | -12.63 | -5.05 | 518.4 |
| 33 | 783.267 | 505.60 | 478.0 | 28.75 | 10.93 N | -14.27 | -5.07 | 515.0 |
| 34 | 749.267 | 483.95 | 858.0 | 51.60 | 12.48 N | -16.30 | -5.08 | 514.2 |
| 35 | 798.2 | 515.24 | 360.0 | 21.65 | 13.65 N | -17.82 | -5.11 | 514.0 |
| 36 | 829.133 | 535.21 | 58.0 | 3.49 | 14.26 N | -18.62 | -5.11 | 515.0 |
| 37 | 753.833 | 486.60 | 863.0 | 51.90 | 12.14 N | -15.85 | -5.13 | 517.5 |
| 38 | 800.9 | 516.98 | 283.0 | 17.02 | 12.30 N | -16.06 | -5.18 | 512.8 |
| 39 | 779.867 | 503.40 | 494.0 | 29.71 | 10.59 N | -13.83 | -5.21 | 514.1 |
| 40 | 810.2 | 522.98 | 111.0 | 6.68 | 8.94 N | -11.67 | -5.22 | 512.8 |
| 41 | 796.167 | 513.93 | 268.0 | 16.12 | 9.85 N | -12.86 | -5.23 | 511.9 |
| 42 | 789.9 | 509.88 | 337.0 | 20.27 | 10.71 N | -13.98 | -5.24 | 510.9 |

*Stations 11 and 12 are in the Penacook Quadrangle, and not shown in Figure 11.

**Table 5**

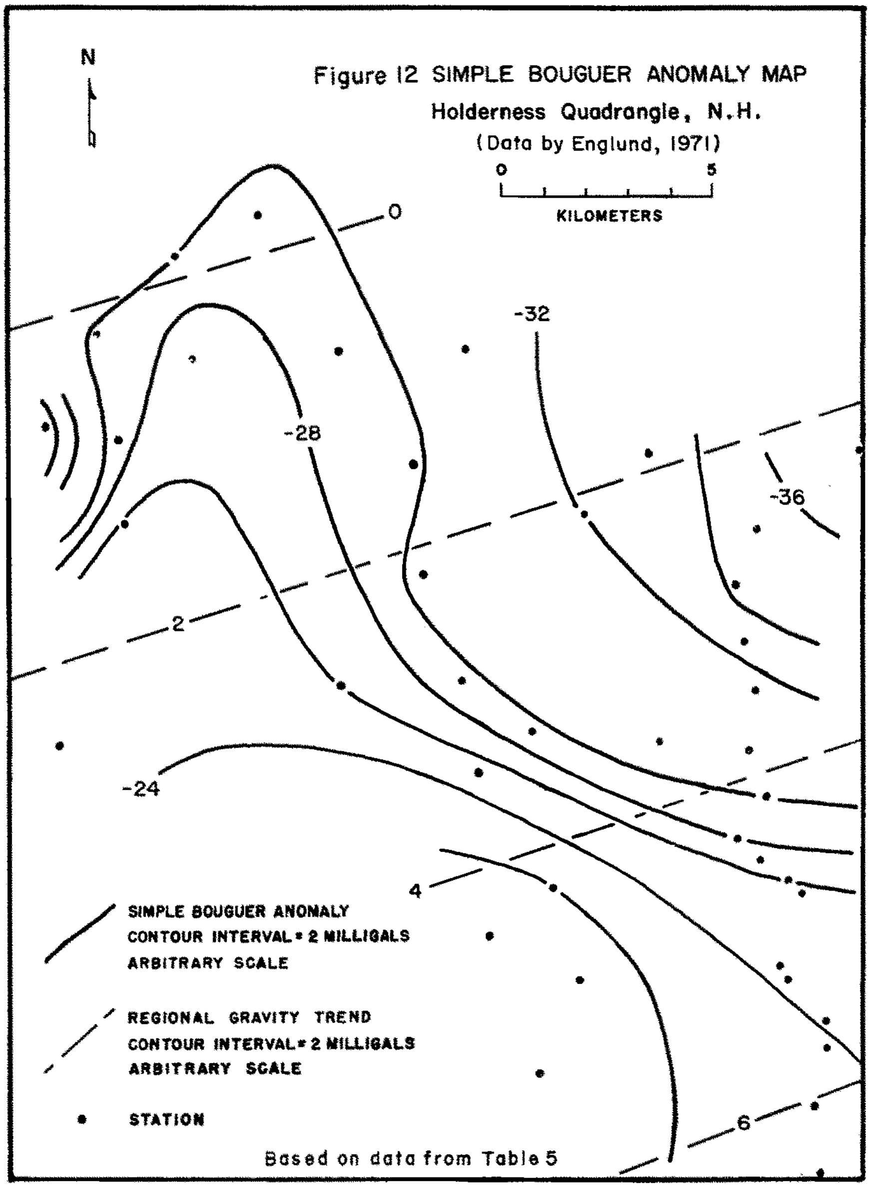



**Figure 12.** Simple Bouguer anomaly map.

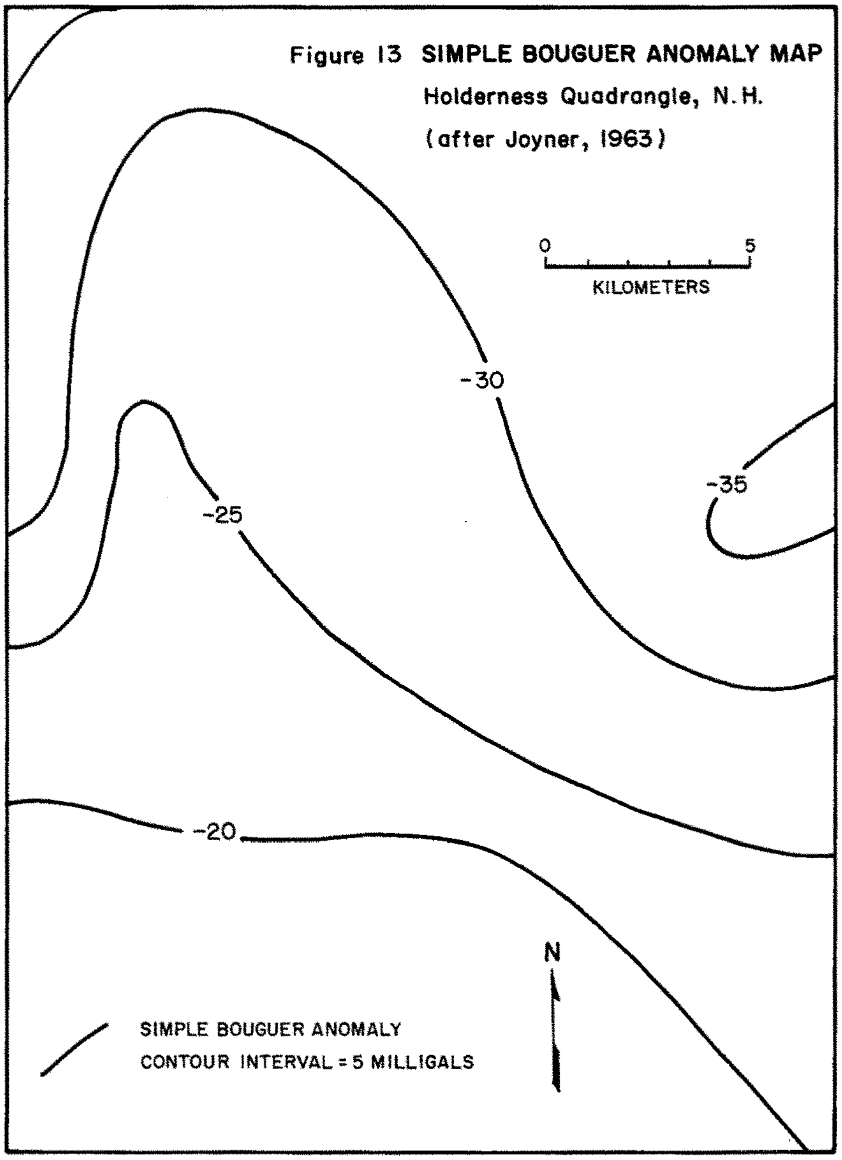



**Figure 13.** Simple Bouguer anomaly map.

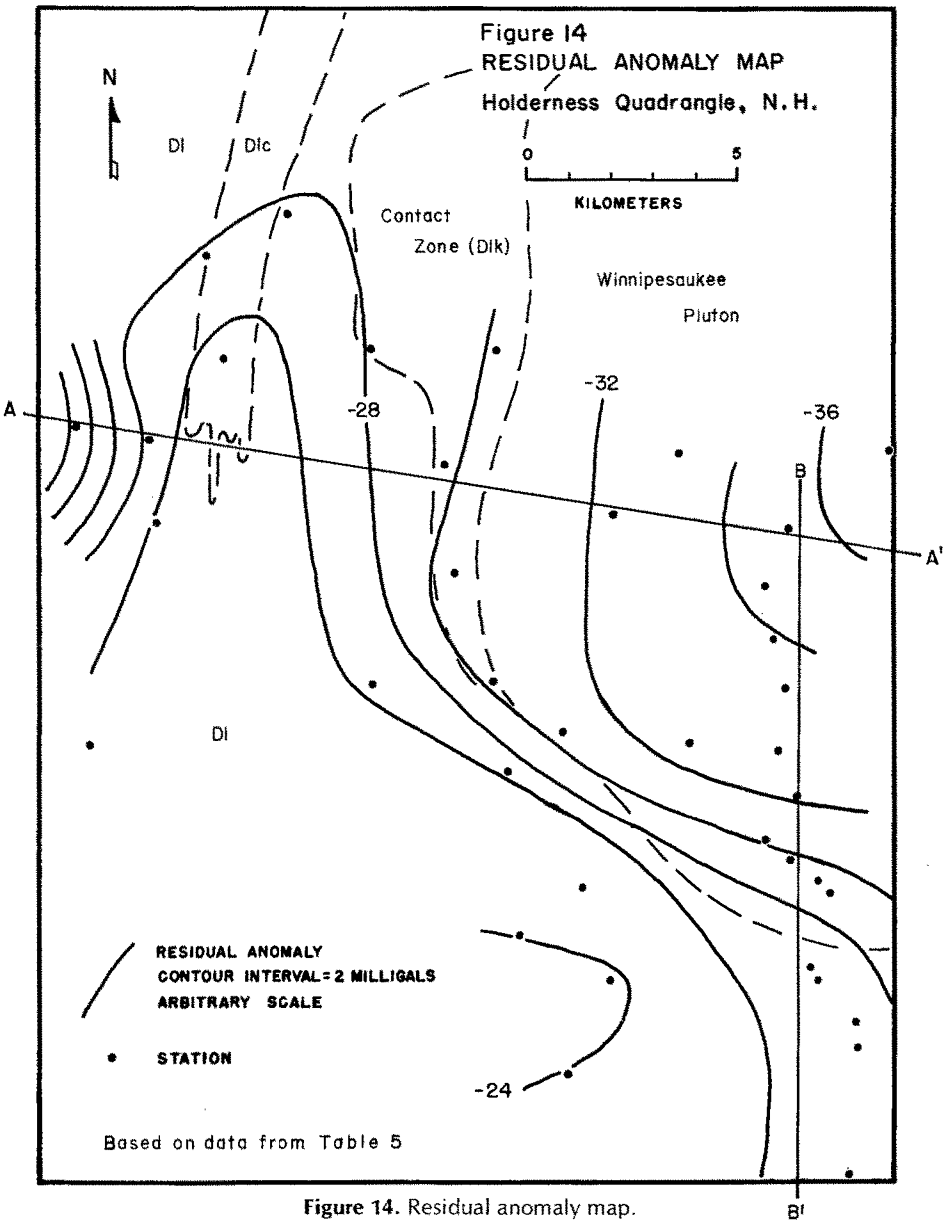



**Figure 14.** Residual anomaly map.

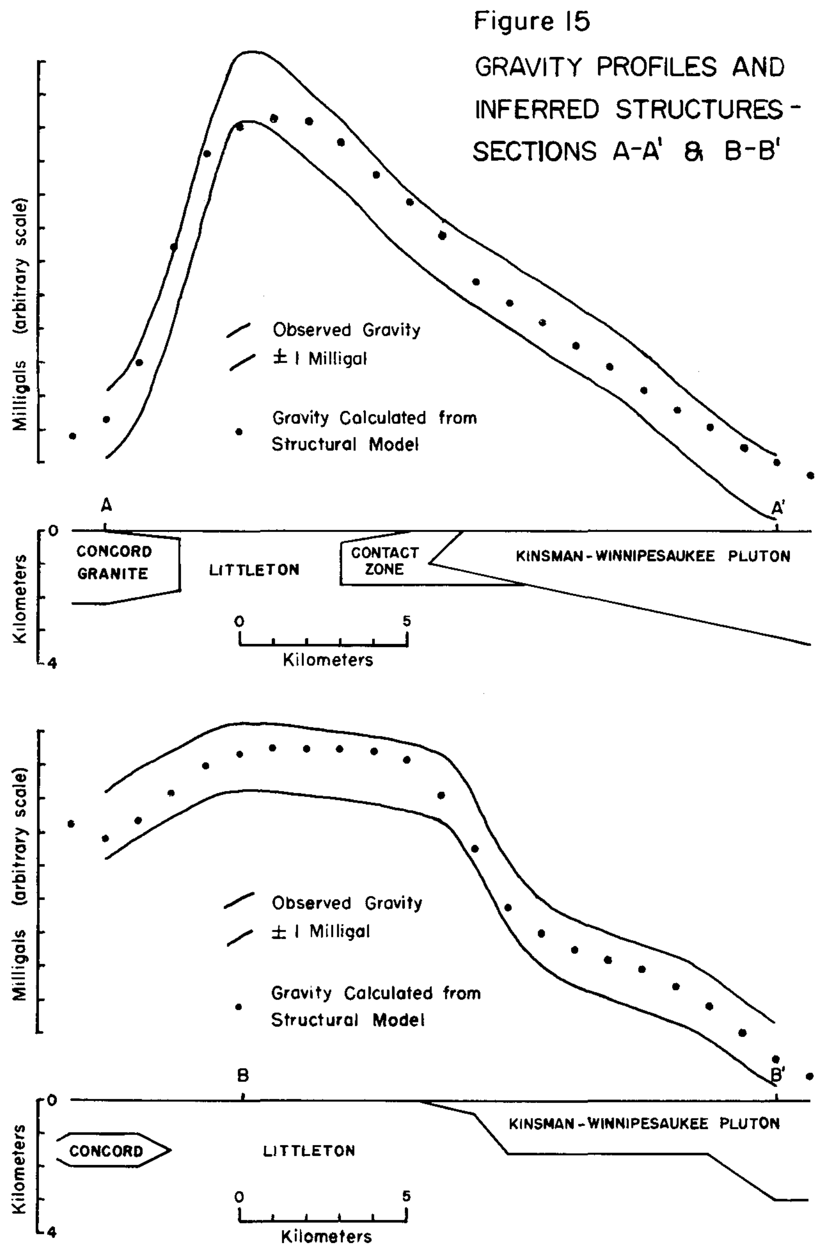



**Figure 15.** Gravity profiles and inferred structures.

## Interpretation of anomalies

A BASIC program (Appendix 1) was written using the equations of Talwani, et. al. (1959) to calculate the cumulative gravitational effect of *n* two-dimensional bodies approximated by irregular polygons. The bodies are assumed to be infinite in the third dimension. This program was used to find structural models whose calculated gravity profiles matched the observed profiles. Because any observed gravity profile can in theory be matched by an infinite number of hypothetical structures, it is necessary to establish some additional constraints on possible models. The principal constraints on constructing such models are the observed contacts at the surface, the dips of the contacts, and the density contrast among the various rock types. Additional aids in model construction are the maximum-depth and minimum-vertical-extent rules (Kane and Bromery, 1968). The maximum depth to the upper surface of a buried source mass is given by the approximate relationship:

$$\text{Dmax} = \text{A}/3.14\ \text{Smax}$$

where Dmax = maximum depth
A = maximum gradient in milligals/km
Smax = anomaly amplitude in milligals.

The minimum vertical extent approximation requires an estimate of the density contrast. The formula:

$$\text{Tmin} = \text{A}.031\text{d}$$

where Tmin = minimum thickness in feet
A = anomaly amplitude in milligals
d = density contrast in $\text{gm/cm}^3$

gives the thickness of the infinite horizontal sheet of the specified density contrast which would result in the observed anomaly amplitude.

The relative densities of the rock types in New Hampshire, and presumably in the Holderness Quadrangle as well, are shown in the following table (after Bean, 1953):

Densities of Rock Types

| *Lithologic Unit* | *# of samples* | *Mean Density* | *Range* |
|---|---|---|---|
| Littleton Formation (High Metamorphism) | 76 | 2.82 | 2.70-2.96 |
| Concord Granite | 21 | 2.67 | 2.61-2.75 |
| Winnipesaukee Quartz Diorite | 49 | 2.73 | 2.67-2.83 |
| Kinsman Quartz Monzonite | 11 | 2.70 | 2.68-2.74 |

For the purpose of determining density contrasts for use in constructing models, the Kinsman and Winnipesaukee are considered as a single unit with a mean density of 2.72. Clark (1972) measured the densities of these units in the Mt. Kearsarge Quadrangle, and found somewhat higher values than Bean, but the density contrasts among the units were in good agreement. Because only the density contrast is used in calculating the gravitational effect, the following relative densities are used:

| | |
|---|---|
| Littleton | 0 |
| Contact Zone | −0.05 |
| Winnipesaukee Pluton | −0.10 |
| Concord Granite | −0.15 |

The maximum Bouguer anomaly over the Winnipesaukee pluton is between 10 and 15 milligals. If it is assumed that this anomaly is due entirely to the density contrast between the Winnipesaukee pluton and the Littleton it can be shown by the minimum-vertical-extent formula that the depth at which the density contrast disappears is between 2 and 4 km.

## Structural Model

Figure 16 shows the structural model whose calculated gravity profiles provide the best fit with the observed gravity. It should be noted that the worst discrepancy between the observed and calculated anomalies occurs at the peak of the profile A-A', where the contouring error is likely to be the greatest. Figure 16 is a block diagram combing sections A-A', B-B', and the surface geology. Making the structure sections coincide at their point of intersection provided an additional constraint in the construction of the models.

The most notable feature of the structural models is the shallowness of the plutonic structures. All of the gravity variation in the area can be accounted for by extending the surface rocks down to 3.5 km or less. The rock below 3.5 km in these models is open to any interpretation, subject only to the condition that the density be laterally homogeneous. It is almost certain from the surface geology that this underlying rock must consist of either metasediments like the Littleton, or of Kinsman-Winnipesaukee plutonic rocks. Littleton seems more likely because it is the predominant rock type in the area, and because the structure which would result from making the underlying rock Kinsman would be more difficult to explain. Interpreting the Winnipesaukee pluton as a thin, flat sheet is consistent with the interpretation of Thompson, et. al. (1968) for Winnipesaukee-type rocks in southwestern New Hampshire.

The only feature of the structural model which does not appear at the surface is the small buried pluton of binary quartz monzonite (?) south of section B-B'. Although the present survey did not extend sufficiently far south to show the magnitude of this low, it appears on Joyner's (1963) regional map as a local low of less than 5 milligals magnitude. Binary quartz monzonite was chosen for this body because bqm plutons of similar size are occasionally observed at the surface (Plate 1). Plutons of Kinsman-Winnipesaukee type are possible, however, and could produce the same effect.

On the original gravity map of Joyner (1963), the maximum anomaly

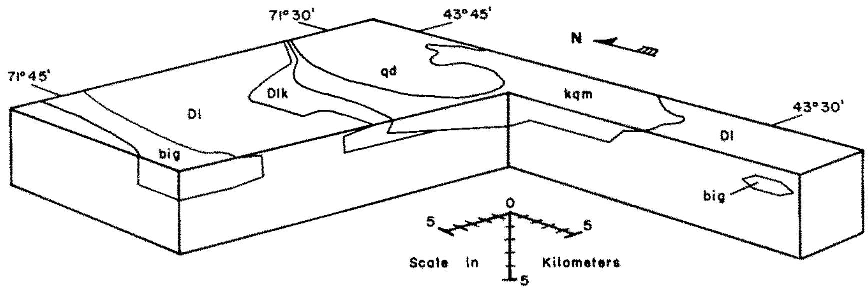



Figure 16

PLUTON STRUCTURES IN THE HOLDERNESS QUADRANGLE — INFERRED FROM GRAVITY ANOMALIES

big – Concord Granite
qd – Winnipesaukee Quartz Diorite
kqm – Kinsman Quartz Monzonite
Dl — Littleton Formation
Dlk – Kinsman-Littleton Contact Zone

**Figure 16.** Pluton structures in the Holderness Quadrangle — inferred from gravity anomalies.

over the entire Winnipesaukee pluton occurs at approximately the intersection of profiles A-A′ and B-B′. The 3.5 km thickness of the pluton at this point is therefore the maximum thickness for the pluton. If this gravity low were interpreted as being due in part to an intrusion of binary quartz monzonite, which is certainly possible, phases of the pluton could be considerably less than 3.5 km.

Clark and Englund (1972) have shown that the Cardigan pluton, which is larger than the Winnipesaukee pluton, has a maximum thickness of about 2 km., and Kane and Bromery (1968) indicate that the Sebago pluton of western Maine may also be shallow. This strongly suggests that all of the New Hampshire series plutons in New England are sheet-like rather than batholithic in nature.

# STRUCTURAL GEOLOGY

## Folding of the Metamorphic Rocks

The metamorphic rocks of the Littleton Formation are generally well foliated, ranging from predominantly schistose to occasionally gneissic or granulose in texture. Where bedding is observed, it is usually found to be parallel or nearly parallel to the schistosity. Because of this parallelism, the term "foliation" will be used loosely to include both schistosity and bedding, unless otherwise specified.

Figure 18a is an equal-area pi diagram showing poles-to-foliation for 192 measurements on the Littleton Formation throughout the Holderness Quadrangle. The pattern is seen to be that of a very broad girdle about a horizontal northeast-southwest axis.

In the field, it was observed that the structural pattern was not constant throughout the quadrangle. Foliations in the northwestern part of the area generally had very steep to vertical dips, whereas in the southern part, the dips were much less steep, and many gentle, open fields were noted.

To illustrate this variation in pattern, the area of the Littleton Formation was divided into three domains, or zones, as shown in Figure 17, and the foliations in each zone were plotted separately and contoured by the Mellis method (Turner and Weiss, 1962). Figures 18b, c, and d show the distinct difference between zones 1 and 3. Zone 2 appears to be intermediate, as the pattern is not significantly different from the overall pattern shown in Figure 18a.

The Littleton Formation has been subjected to a deformational history both intense and complex, as illustrated by the variety of fold styles and orientations observed. Figures 19-24 show some of these folds. It was possible to classify most folds in the field into one of three generations, based on both style and orientation. Deformational events have been labeled $D_1$, $D_2$, and $D_3$; their associated fold generations and axial surfaces are $F_1$, $F_2$, $F_3$, and $S_1$, $S_2$, $S_3$, respectively. $S_0$ is used to designate bedding surfaces.

$F_1$ folds are isoclinal and often recumbent. The dominant foliation in most areas is $S_1$, and is parallel to $S_0$. Thus the pi diagrams of Figure 18 are essentially diagrams of poles to $S_0$ and $S_1$. The plunges of $F_1$ fold axes may vary from horizontal where the foliation is horizontal, to steep where the foliation is near-vertical. $F_1$ folds have been observed on a relatively small mesoscopic scale, one of the larger examples being shown in Fig. 22. The author has been unable to distinguish a major outcrop pattern showing $F_1$ folds on a larger scale, although the pattern

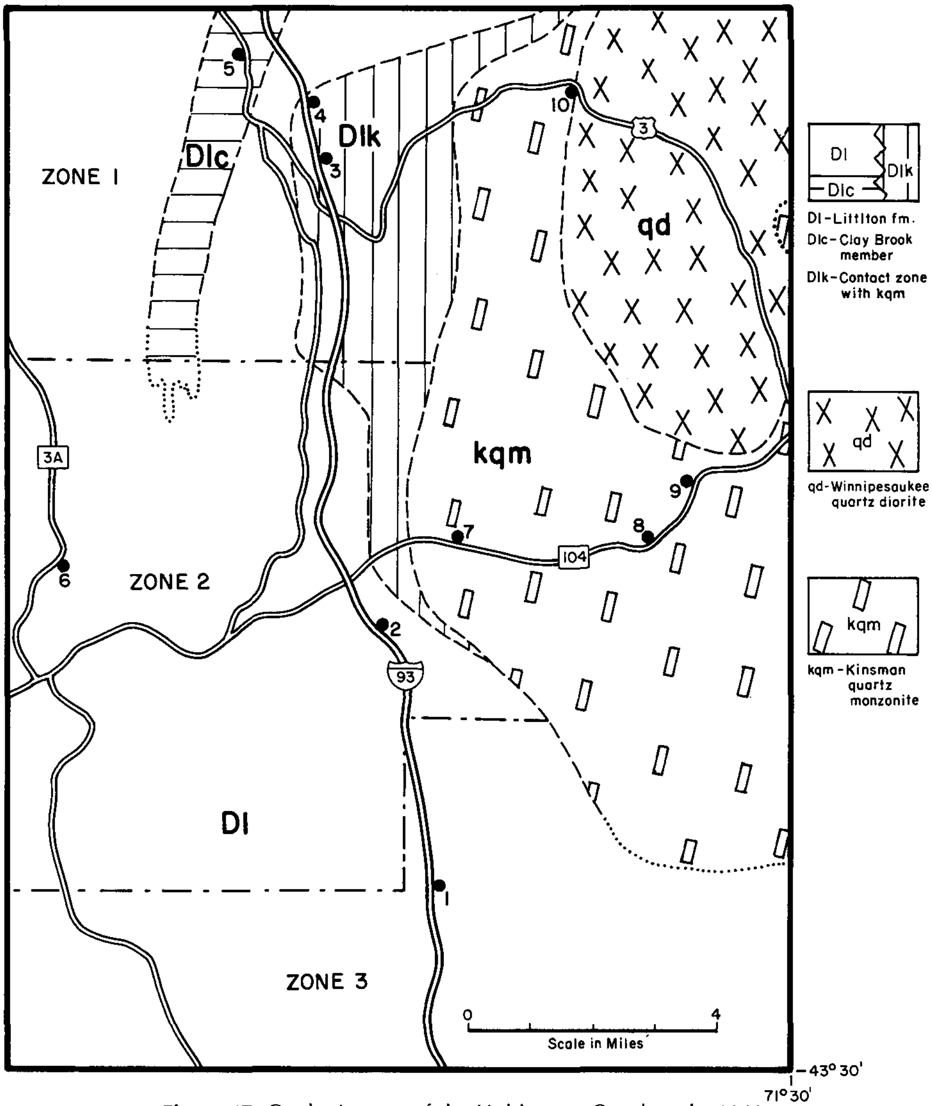
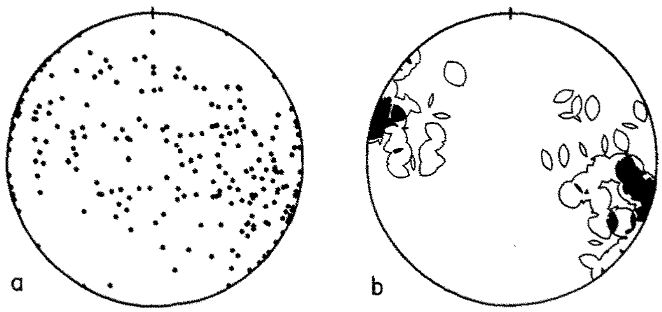
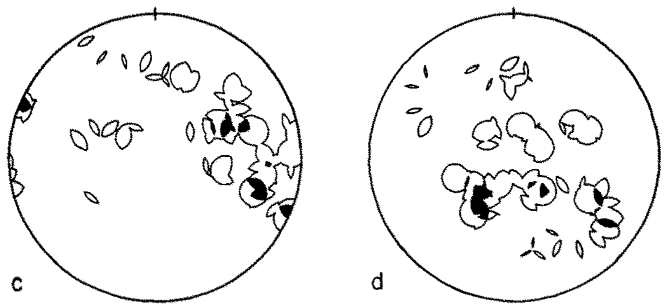



**Figure 17.** Geologic map of the Holderness Quadrangle, N.H.

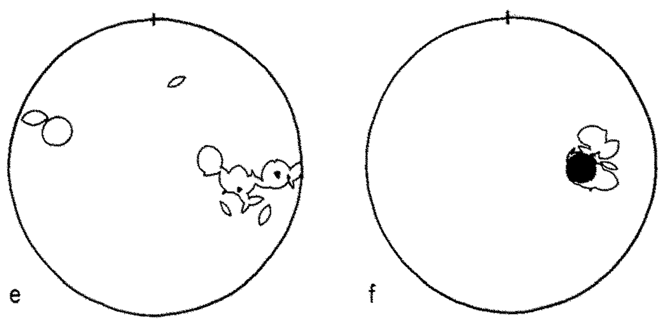
Figure 18
Equal Area Diagrams Showing Poles to Foliation




Total Littleton - 192 pts. Littleton Zone 1 - 72 pts.



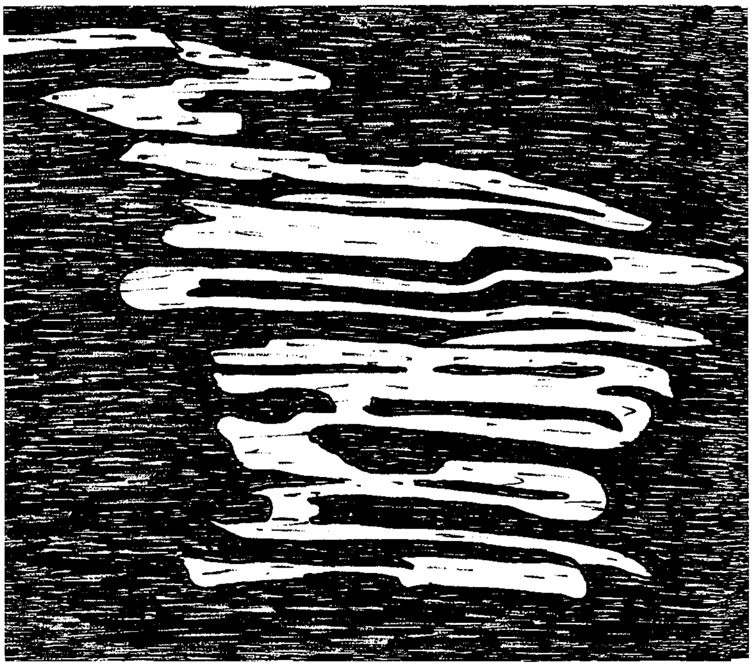

Littleton Zone 2 - 58 pts. Littleton Zone 3 - 62 pts.




Kinsman - 29 pts. Winnipesaukee - 17 pts.

Contour Maxima: b. 4%; c. 7%; d. 7%; e. 12%; f. 20%

**Figure 18.** Equal area diagrams showing poles to foliation.



10 CM

**Figure 19.** Ptygmatically folded quartz vein in the Littleton Formation. Folds of this type generally appear to be associated with $F_1$. The pencil is parallel to the fold axes.

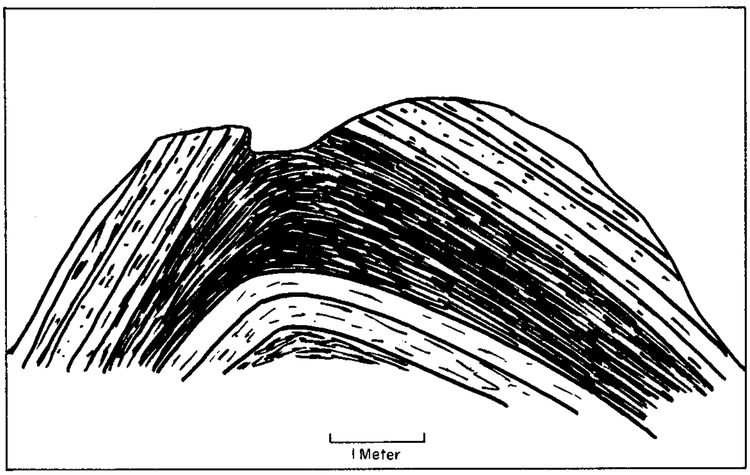



**Figure 20.** Asymmetrical $F_3$ antiform across the road from the ptygmatic folds shown in Plate 11. The sub-horizontal $F_3$ fold axis comes out of the outcrop almost at right angles to the road, slightly to the observer's left, and parallel to a hammer handle in the lower central portion of the photograph. This axis is nearly at right angles to the ptygmatic fold axes across the road.

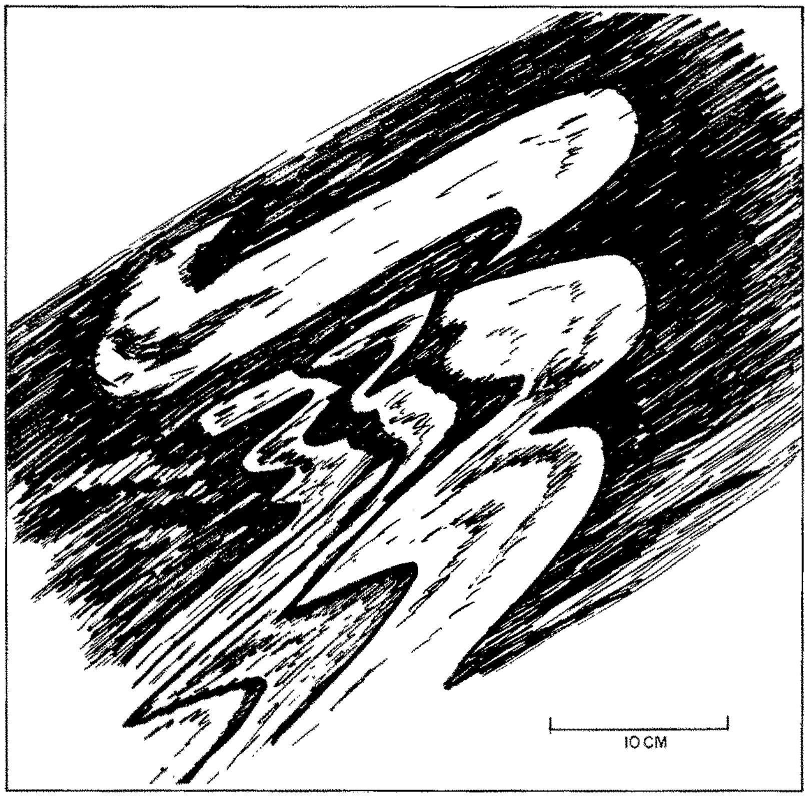



**Figure 21.** Quartzitic layer ($S_0$?) deformed by $F_1$ folding. the $F_1$ axes plunge shallowly to the northwest in the direction pointed out by the pencil. Note the development of an axial plane foliation ($S_1$) above the pencile. The detached, S-shaped fold at the top of the photograph is characteristic of intense shearing parallel to the axial plane.

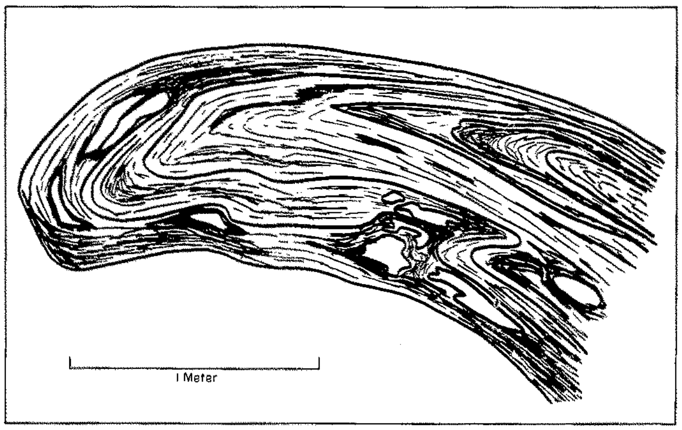



**Figure 22.** Recumbent isoclinal $F_1$ fold, refolded gently about $F_2$. Axes of both $F_1$ and $F_2$ folds strike approximately parallel to the pencil, and plunge shallowly northwest, away from the observer.

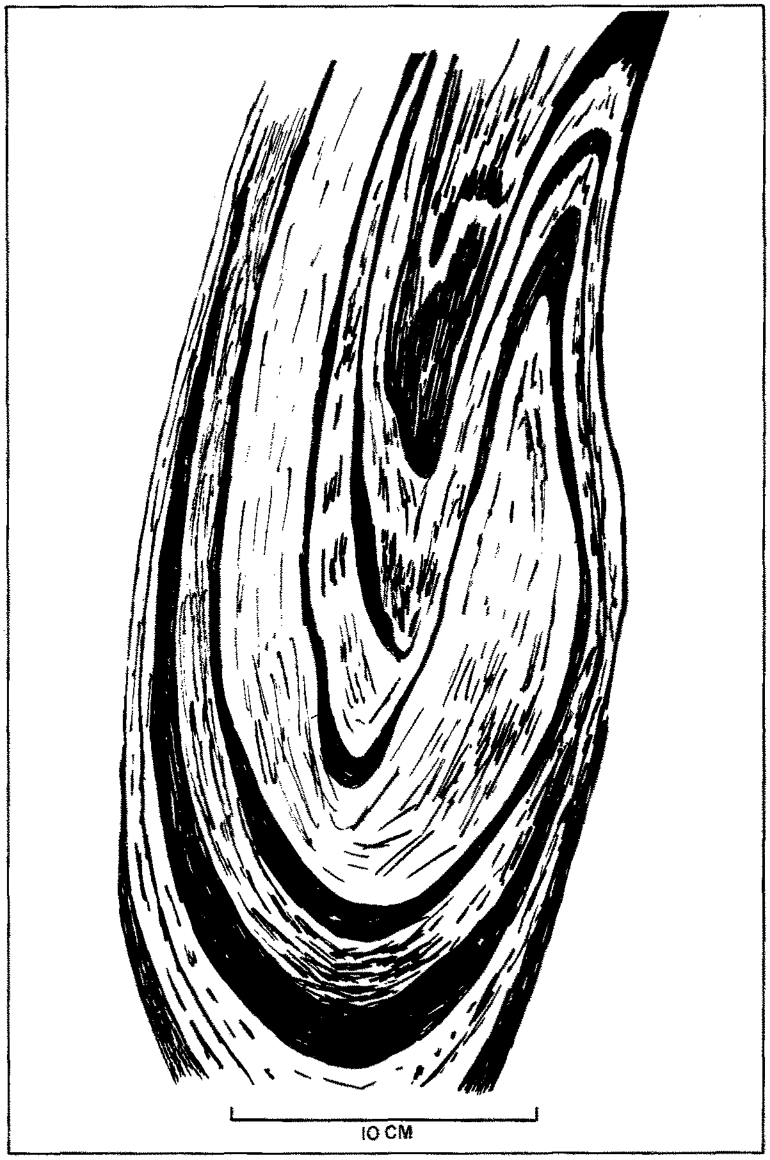



**Figure 23.** Isoclinal $F_1$ fold, refolded isoclinally by $F_3$. From a roadcut on Interstate 93 north of the village of Ashland. The $F_3$ axis is approximately Horizontal, striking northeast. The orientation of the $F_1$ axis could not be determined. Note that away from the hinges, all of the bedding and foliation becomes approximately parallel to $S_3$. Nickel for scale.

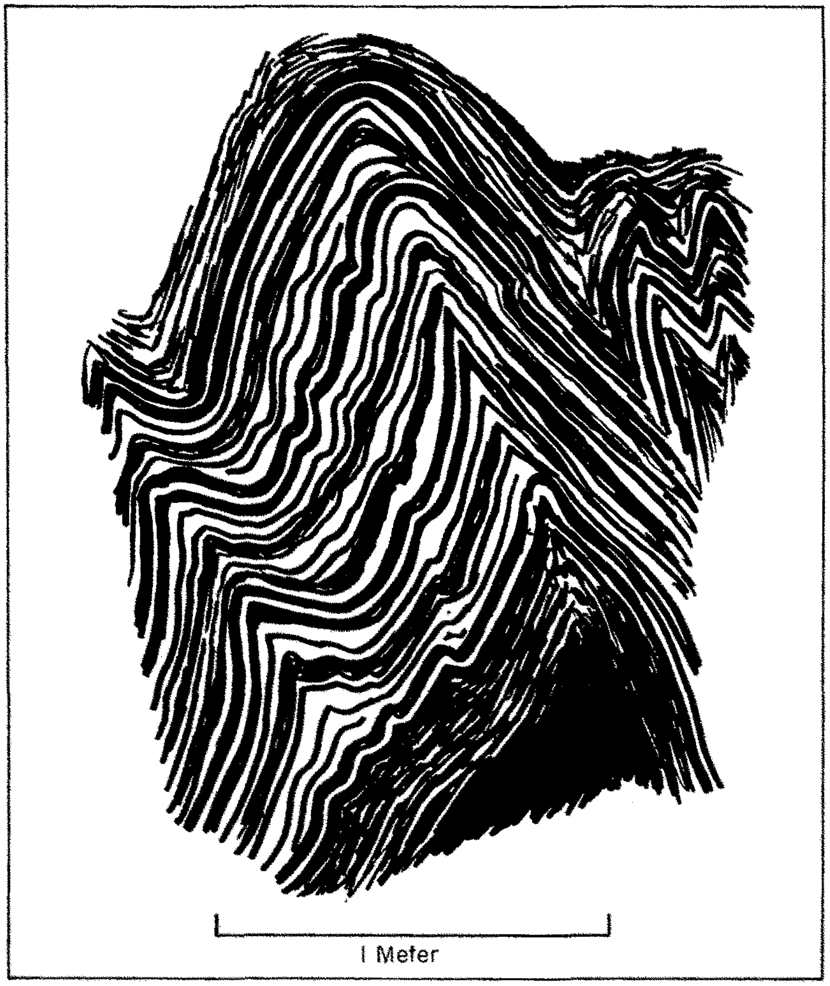
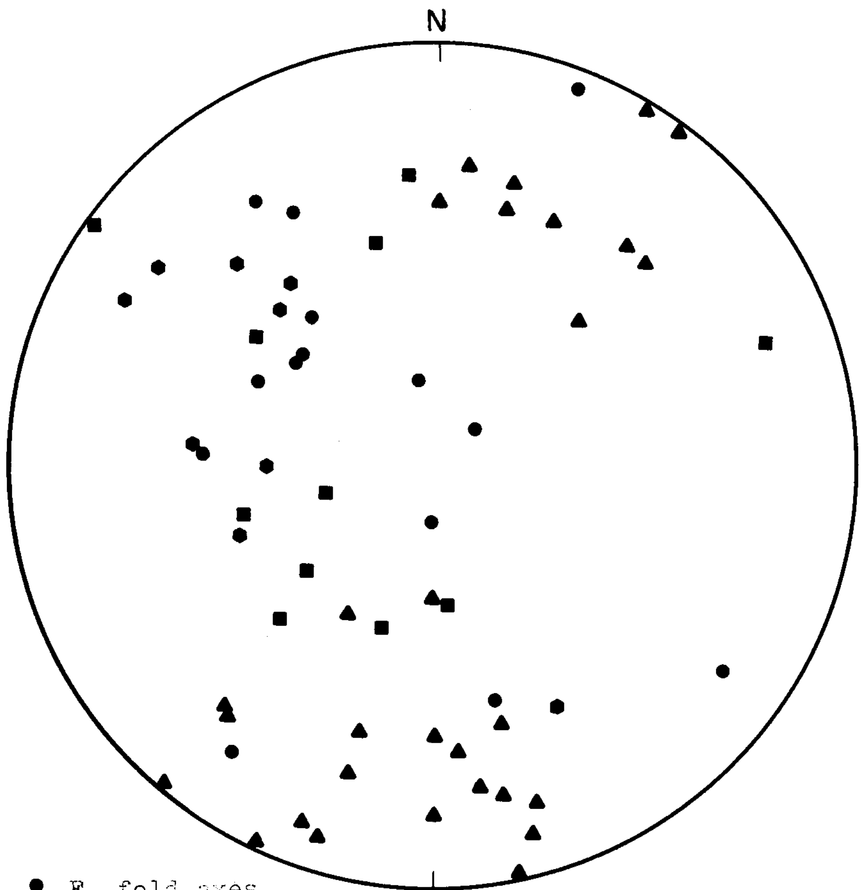



**Figure 24.** $F_1$ (?) folds in thinly bedded Littleton from a roadcut on Interstate 93 north of the village of Ashland. The fold axes plunge steeply parallel to the hammer handle in the lower left of the photograph.




● $F_1$ fold axes
⬢ $F_2$ fold axes
▲ $F_3$ fold axes
■ unclassified axes

**Figure 25.** Point diagram of fold axes and their classification as $F_1$, $F_2$, or $F_3$ based on field criteria.

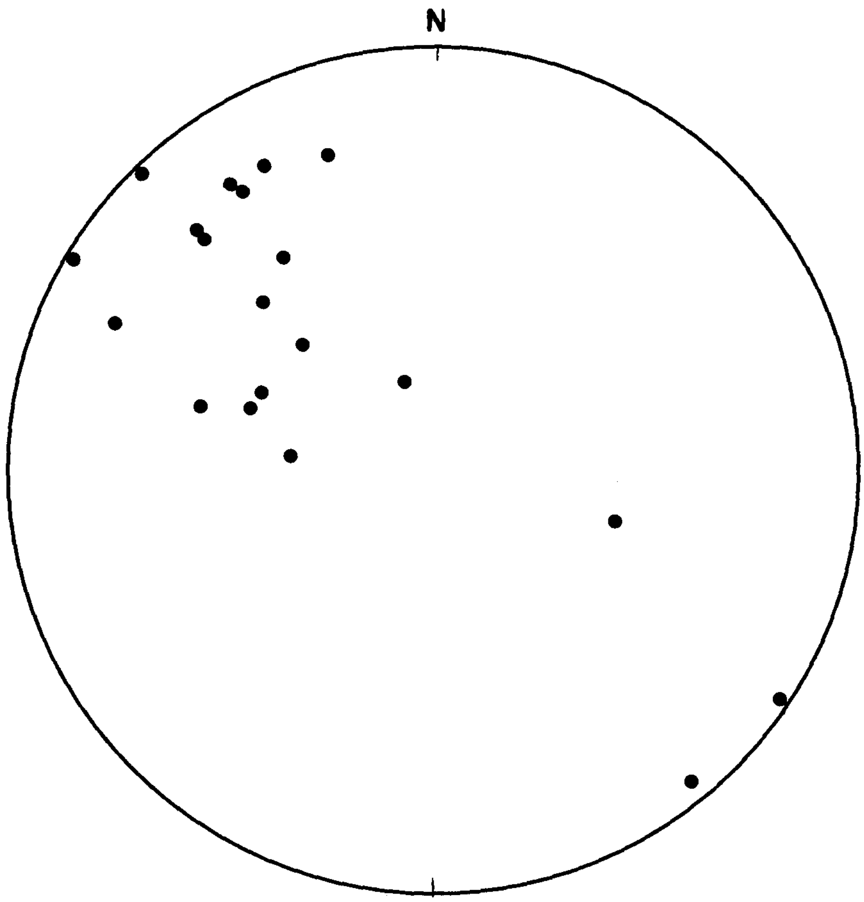



**Figure 26.** Point diagram of mineral lineations.

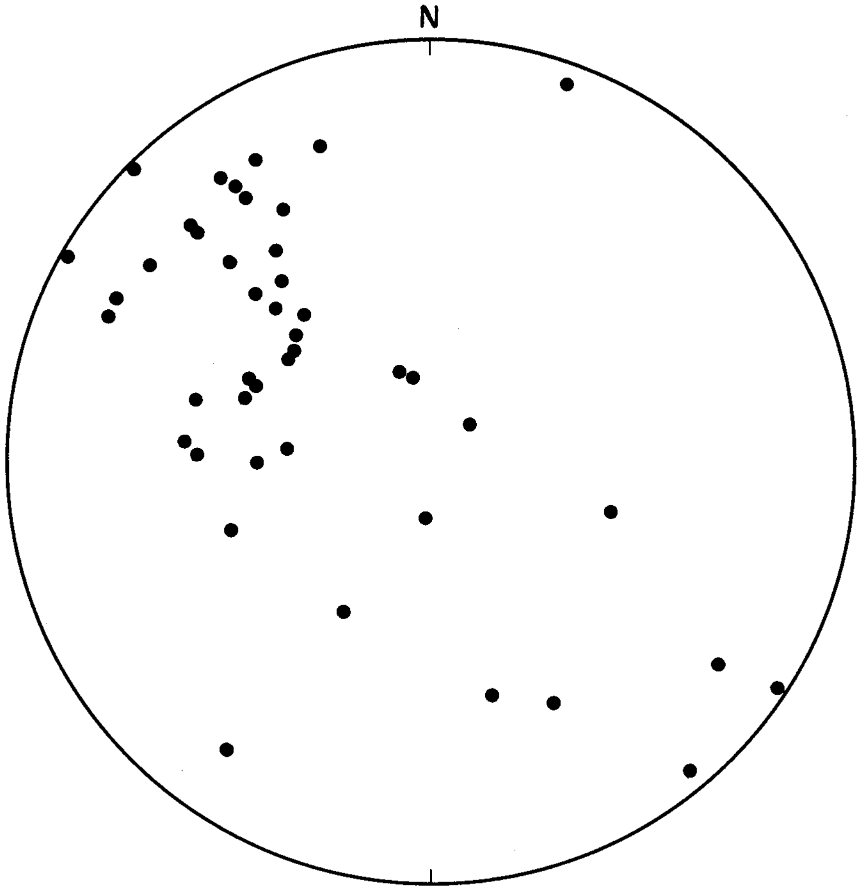



**Figure 27.** Point diagram combining $F_1$ and $F_2$ fold axes and mineral lineations.

of some units (e.g. the Clay Brook Member) may be due to refolding of $F_1$ structures.

$F_2$ folds are gentle, open folds whose limbs rarely dip as steeply as 45°. $S_2$ axial planes are vertical or nearly vertical, and generally strike northwest. Axes of $F_2$ folds follow the same pattern as those of $F_1$. $F_2$ folds have been observed with widths and amplitudes ranging from several feet to several hundred feet. $F_2$ folds apparently represent refolding of $S_1$ schistosity along generally similar tectonic directions.

$F_3$ folds range from open folds resembling $F_2$ to tight, isoclinal folds. $S_3$ axial planes are vertical to nearly vertical, and generally strike between N 10° W to N 40° E. Fold axes rarely plunge more than 30°.

The variation in the intensity of the $F_3$ folds apparently accounts for the variation in the foliation patterns shown between Figure 18b and 18d. In zone 1, the structure is dominated by upright isoclinal $F_3$ folds, whereas in zone 3, the $D_3$ deformation is not more intense than $D_2$. The $F_2$ and $F_3$ folds at approximately right angles in zone 3 resulted in a cross-folding of $S_1$ into an interference structure of domes and basins.

Figure 25 is a point diagram showing the orientations of fold axes and their field classification into fold generations $D_1$, $D_2$ or $D_3$. The diagram shows that $F_3$ fold axes forms a distinct group spatially separated from $F_1$ and $F_2$ fold axes. As noted in the field, the diagram shows that $F_1$ and $F_2$ fold axes are closely related in tectonic origin as well.

A mineral lineation in $S_0$ or $S_1$ was occasionally observed. The lineation was commonly due to orientation of sillimanite, and was consistently found in the same orientation as $F_1$ and $F_2$ fold axes. Measurements of mineral lineations are plotted in Figure 26. In at least one case, a lineation was folded by an $F_3$ fold. No mineral lineations were observed parallel to $F_1$ fold axes. Figure 27 is a point diagram combining $F_1$ and $F_2$ fold axes plus mineral lineations. These points form a broad girdle about an axis striking about N 35° E, plunging about 15° northeast, indicating that $F_3$ has deformed the other structures, and is tectonically younger.

The fold generations have been numbered according to their order of development as deduced from field observations and from the pi and point diagrams. Observations which support this conclusion are as follows:

1) the principal schistosity of foliation appears to be $S_1$. This is generally parallel to $S_0$, and is observed to be deformed by both $F_2$ and $F_3$ folds. $S_2$ and $S_3$ foliations are sometimes observed as crenulations in $S_1$, but are rarely well developed.

2) $F_1$ folds have been observed refolded by $F_2$ folds (Plate 15) and by $F_3$ folds (Figure 23).

3) $F_1$ and $F_2$ fold axes and lineations all have similar orientations, and can be explained as early, subhorizontal, northwest-trending features refolded during $D_3$.

The evidence for considering $D_2$ to be earlier than $D_3$ is admittedly circumstantial, but no evidence to the contrary has been observed. If $D_3$ were assumed to be earlier than $D_2$, it would require $D_3$ folding trending northeasterly, followed by $D_2$ folding in the precise orientation such that $D_2$ axes coincide with $F_1$ axes. This situation would be possible, but not probable.

The foliations in the Winnipesaukee pluton (Fig. 18e, and f) appear to be related to $S_3$. No evidence of $D_1$ or $D_2$ deformation is present in the Winnipesaukee pluton in this area, but Lyons and Clark (1971) find that the Kinsman Quartz Monzonite of the Cardigan pluton was apparently enclosed in recumbent structures, and was probably emplaced synchronously with the folding. The Winnipesaukee pluton was probably emplaced at about this time, during or shortly after the $D_1$ and $D_2$ events.

The inferred sequence of recumbent $D_1$ folding followed by upright $D_2$ folding with approximately parallel axes may be considered as an early and a late stage of the same deformational event. Fyson (1971) indicates that similar time-sequences of recumbent to upright folds are found in many orogenic belts.

The time period during which $D_1$, $D_2$, and $D_3$ deformation took place appears to be restricted to a relatively short interval. The style and orientation of $F_3$ folds indicates that $D_3$ belongs to the main phase of the Acadian orogeny in the Middle Devonian. The conclusion that the Acadian orogeny was a brief but intense event is consistent with recent work by Naylor (1971) in Vermont, where recumbently folded fossiliferous rocks are intruded by an undeformed granitic stock. The fossil and radiometric dates bracket the deformation in a period which Naylor believes may be as short as 10 million years or less.

## Faulting

No major faults have been recognized in the Holderness Quadrangle. Occasional minor faults have been observed, but none has been traced beyond a single outcrop. It would seem reasonable to associate minor normal faulting with the intrusion of lamprophyre dikes in Jurassic(?) time. The mapping of the metamorphic rocks has not eliminated the possibility that early, large-scale faulting may have occurred during one or more of the Acadian deformations.

## Structural Synthesis

From the evidence presented above the following sequence of events may be deduced for the Holderness Quadrangle.

1) Deposition of the Littleton Formation as turbidites, pelites, and calcareous sands and muds. Continuous deposition probably took place from about the Middle Silurian through Early Devonian, but might have begun much earlier.

2) Isoclinal recumbent folding ($D_1$) about northwest trending axes, resulting in a prominent, nearly horizontal, axial plane schistosity ($S_1$).

3) Gentle, open folding ($D_2$) about northwest trending axes. Little or no development of axial plane foliation ($S_2$).

4) Intrusion of the Winnipesaukee pluton as a large sheet-like body, with lit-par-lit injection of the Kinsman phase along the $S_1$ foliation of

the Littleton. It is probable that the emplacement of the Winnipesaukee pluton occurred concurrently with deformations $D_1$ and $D_2$.
5) Open to isoclinal folding ($D_3$) about northeast trending axes. Originally subhorizontal $F_1$ and $F_2$ axes and lineations were folded into the observed girdle pattern. The foliation and mylonization in the Winnipesaukee pluton imply that prior to $D_3$ it had solidified sufficiently to support shear stresses. The pluton thus may have acted as a buttress during $D_3$, maximizing the deformation of the Littleton in the northwestern part of the quadrangle, while partially shielding the area to the south.
6) Intrusion of the binary quartz monzonites. The emplacement of these bodies probably spanned the period from just prior to just after $D_3$, because both foliated and unfoliated examples are present. Unfoliated binary quartz monzonites have been observed cutting the Kinsman.
7) Intrusion of the White Mountain Series (Jurassic?).
8) Intrusion of lamprophyre dikes (Jurassic?).

Major units belonging to the White Mountain plutonic series have not been found within the Holderness Quadrangle, although the well-known Ossipee and Belknap complexes are nearby. Intrusion of the White Mountain Series in central New Hampshire occurred later than the intrusion of the binary quartz monzonites (Billings, 1956), but prior to the instrusion of the lamprophyre dikes.

The evidence for the deformational event(s) which caused the northwest trending folds ($F_1$ and $F_2$) is of particular interest, because heretofore Acadian folding about northwest-trending axes has not been recognized as a regional feature in New Hampshire. Although most workers in the Merrimack synclinorium in New Hampshire and adjacent quadrangles in Maine have shown at least some evidence of northwest-trending deformation, they have tended to consider the deformation to be isolated and local in nature. When the evidence is considered *in toto*, however, a strong argument can be made for a single, widespread event.

Gilman (1970, unpublished map) has miles-long northwest-trending folds shown by formational outcrop patterns in the Kezar Falls and Newfield Quadrangles, Maine-New Hampshire. Wilson (1969) in the Ossipee Lake Quadrangle, has a very complex pattern of minor structures, which could be explained, at least in part, as northwest-trending folds refolded tightly about northeast-trending axes. In the Gilmanton Quadrangle (Heald, 1955) and the Wolfeboro and Winnipesaukee Quadrangles (Quinn, 1944, 1953), the foliations in the Littleton Formation suggest the same cross-fold pattern of domes and basins as is observed in the Holderness Quadrangle. In the Peterborough Quadrangle (Greene, 1970), northwest-trending folds with associated mineral lineations form a dome and basin interference pattern with northeast-trending folds. Vernon (1971) shows a consistent, northwest-trending mineral lineation in the Concord Quadrangle.

A major northwest-trending deformation, as suggested by the above evidence, could also explain some structures observed further west, such as the northwest-trending hinge of the nappe in the Claremont Quadrangle (Thompson, et. al., 1968), and the steeply plunging folds of the Hardy Hill (Clough) quartzite in the Mascoma Quadrangle (Chapman, 1939).

The lack of abundant map patterns showing northwest-trending folds in New Hampshire is perhaps one of the major arguments against their interpretation as a regional phenomenon. A possible explanation lies in the fact that if recumbent, isoclinal folds such as $F_1$ are folded tightly by upright folds at nearly right angles, such as $F_3$, the outcrop pattern of an $F_1$ fold as seen on a geologic map would not differ significantly from that of an $F_3$ fold. The distinguishing criteria, of course, would be the plunge of the fold, as determined from the dips of the beds at the hinge. However, outcrops may not be available exactly on the hinges of tight folds, and plunges would then be inferred from associated minor folds. If the dominant minor folds were related to $F_3$, and the earlier deformation were not suspected, the actual structure might not be recognized.

The structure of the Clay Brook Member is an excellent example to illustrate the problems involved. The outcrop pattern on Plate 1 suggests the possibility of an isoclinal fold with its hinge at the southern end of the unit. The southern contact of the unit is complex, however, and no outcrops were identified as being directly on the hinge of the fold. Minor fold axes in the area have both shallow northeast plunges ($F_3$) and steep north plunges ($F_1$).

Several interpretations are possible which are consistent with this evidence. If the major fold is an $F_3$, then the earlier $F_1$ folds may be either on such a small scale that they do not affect the outcrop pattern, or on such a large scale that only a portion of one limb of an $F_1$ fold is seen here refolded. On the other hand, if the Clay Brook Member is the oldest unit, it must have been initially recumbently folded about $F_3$ fold axes into its present position, and in the Holderness Quadrangle would be a synform plunging steeply to the north, with the oldest rocks in the core.

An adequate understanding of the structures in central New Hampshire and particularly the importance of the $D_1$ deformation, will depend on much better stratigraphic control within the metasediments. This has been notably lacking, perhaps precisely because of unsuspected complexities in the structure. The recent discovery of the Roundtop Quartzite Member as a mappable unit provides hope that it may be useable as a marker horizon, although its outcrop pattern, as mapped to date, neither supports nor refutes any of the hypotheses mentioned above.

It is clear that if the $D_1$ deformation was a major event which produced large-scale nappe structures corresponding to the minor folds, then the commonly accepted interpretation that stratigraphic subdivisions of the Littleton Formation get younger toward the axis of the Merrimack synclinorium cannot be correct. Indeed, the existence of any structure resembling a synclinorium in the classical sense would be questionable.

Although the existence of a regional northwest-trending deformation has not been definitely established, the evidence is sufficiently strong so as to warrant a regional structural study with this hypothesis in mind.

# METAMORPHISM

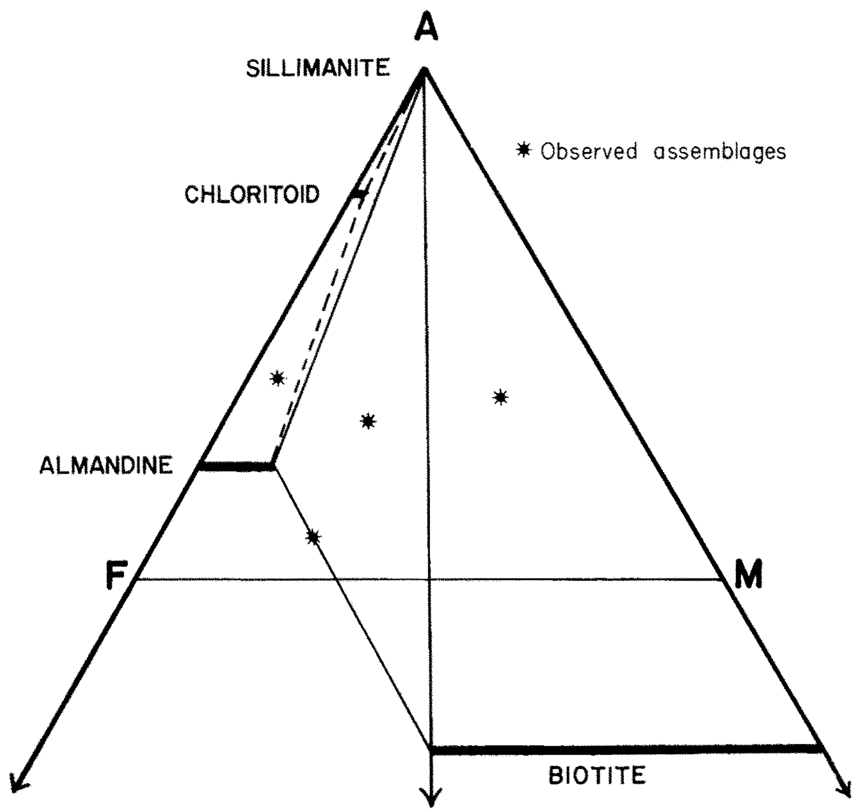
## Grade of Metamorphism

The mineral assemblages of the metasedimentary rocks of the Holderness Quadrangle are typical of sillimanite grade metamorphism. Sillimanite is the only $Al_2SiO_5$ polymorph observed, and is usually present as fibrolite, indicating that it was not formed from earlier kyanite or andalusite. Occasionally, larger crystals of sillimanite are observed. These are generally oriented parallel to $F_1$ or $F_2$ fold axes. Muscovite, rather than K-feldspar, is the normal potassium-aluminum silicate, indicating that the rocks in this area have not attained the K-feldspar-sillimanite zone. One sample of sulfidic schist containing K-feldspar was observed within the Kinsman-Littleton contact zone north of the village of Ashland. The mineral assemblages in the Littleton Formation include, in addition to quartz: biotite-sillimanite-(muscovite)-(plagioclase); biotite-garnet-(muscovite)-(plagioclase); and biotite-sillimanite-garnet-(muscovite)-(plagioclase). A chloritoid-quartz assemblage was also observed. Biotite and garnet are often partially altered to chlorite. Phase relations in the pelitic rocks are shown in Figure 28.

The mineral assemblages present in the calc-silicate rocks are represented in Figure 29. These are: Quartz-diopside-grossularite; quartz-diopside-grossularite-clinozoisite; quartz-plagioclase-actinolite; and plagioclase clinozoisite-diopside. The common occurrence of diopside and grossularite in most of these rocks is consistent with the amphibolite facies sillimanite-grade metamorphism observed in the pelitic schist.

The temperature and pressure of the rocks during metamorphism can be estimated from experimentally determined univariant equilibrium curves as shown in Figure 30. Unfortunately, there is still considerable disagreement as to the location of the kyanite-andalusite-sillimanite triple point, as is evident in comparing two of the most recent determinations by Richardson et. al. (1969), and Holdaway (1971). We can say, however, that the temperature of regional metamorphism in this area must have been at least 500°C and less than about 750°C, and the results of Richardson, et. al., suggest that the minimum may have been as high as 622°C.

The experimentally determined stability curves of Figure 30 cannot necessarily be considered to accurately represent natural systems because the experimental curves are produced under relatively restricted conditions and refer to pure chemical species, and to oxygen




**To K-feldspar**

**Figure 28.** Phase relations of amphibole facies pelitic rocks of the Holderness Quadrangle. Projection of AFMK Tetrahedron is from muscovite onto AFM surface (Thompson, 1957). Quartz and muscovite are present. Plagioclase may be present.

fugacities different than those in the natural rocks. The muscovite and quartz to form K-feldspar and sillimanite involves the release of $H_2O$ from the muscovite. The curve in Figure 30 is for a closed system in which $P_{H_2O} = P$ total. If the system is open, or if other gases are present, $P_{H_2O} < P$ total, and the curve will shift to the left. In a system completely open to $H_2O$, the univariant curves for muscovite will have a negative slope. French (1965) has shown that in systems in equilibrium with graphite at about 600°C, $P_{H_2O}$ is not likely to exceed 50% of P total. Thus, the muscovite of the muscovite-graphite assemblage observed in the Clay Brook Member would be expected to go to K-feldspar and sillimanite at considerably lower temperatures than suggested by Figure 30.

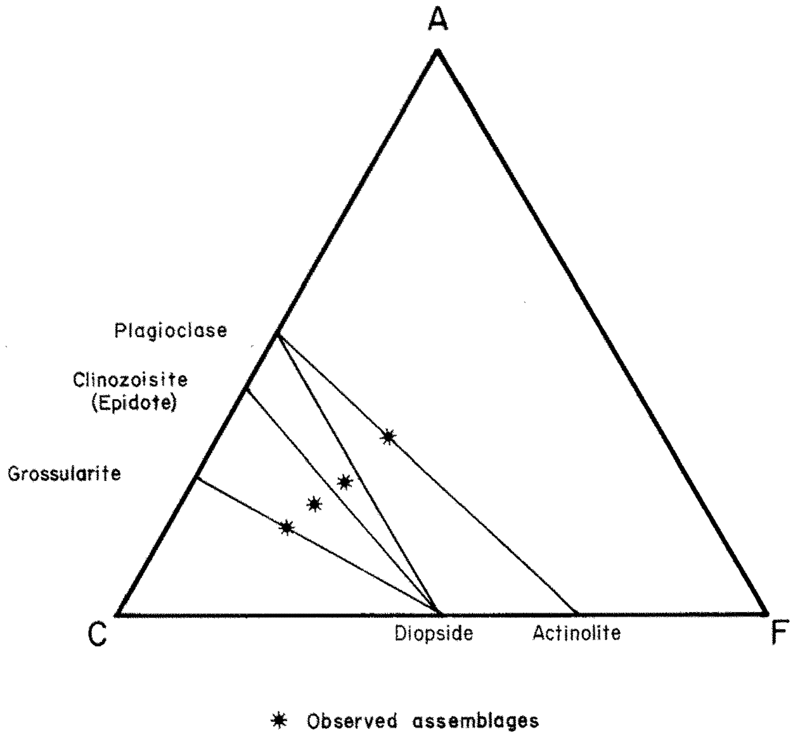



**Figure 29.** ACF diagram for amphibolite facies showing mineral assemblages in calc-silicate boudins and calc-silicate rocks of the Holderness Quadrangle.

Most of the rocks of the Littleton Formation show some degree of retrograde metamorphism, usually as alterations of biotite and garnet to chlorite. Prograde metamorphism reaction usually involves the loss of water, and retrograde reactions are inhibited unless the water is retained in the rock, or unless water is added during cooling. An unusual assemblage containing chloritoid-quartz from the north slope of Plymouth Mountain (Plate 1) may be an example of retrograde metamorphism. The sample (#12-71, Table 4) was collected within a few hundred feet of the contact with the binary quartz monzonite, and while the intrusion may have locally increased the temperature of the schist, it may also have been the source of the water needed for the formation of chloritoid.

This chloritoid has only 1.4% MgO by microprobe analysis (J. B. Lyons, verbal communication), and because MgO does not have a strong effect on the univariant curve for the breakdown of chloritoid (Ganguly, 1968) pp. 286-289), the curve reproduced in Figure 30 may be a fair index of its upper stability field. Other stability curves deduced by

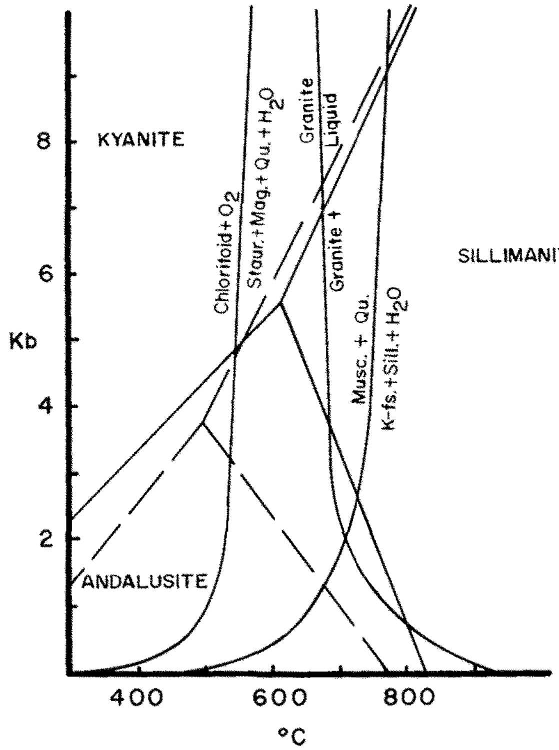



**Figure 30.** Some experimentally determined stability relations for silicates. Kyanite-andalusite-sillimanite fields from: Richardson, et. al. (1969), solid lines: Holdaway (1971), dashed lines. Chloritoid curve from Ganguly and Newton (1968); granite curve from Tuttle and Bowen (1958); muscovite curve from Segnit and Kennedy (1961).

Ganguly (1968) show that in any case chloritiod is not stable above 550°C to 600°C, and that under appropriate conditions it supplants the mineral staurolite.

If the lower temperature and pressure $Al_2SiO_5$ triple point show in Figure 30 (Holdaway, 1971) is accurate, it is then possible that chloritoid could be a stable prograde phase within the sillimanite zone. The stability fields of chloritoid (Ganguly and Newton, 1968) and sillimanite intersect in a small area (Fig. 30), and according to Ganguly, chloritoid may be stable at even higher temperatures when oxygen fugacity is very low. If the chloritoid of Plymouth Mountain is not of retrograde origin, the mineralogy there specifies a very narrow P-T range of approximately 525°C, at $P_{total}$ of close to 4 kilobars (assuming that Holdaway's (1971) triple point is correct).

## Anatexis

The granite melting curve of Figure 30 indicates that partial melting is likely to accompany sillimanite-grade metamorphism, if $P_{H_2O} = P_{total}$, and that the sillimanite K-feldspar zone is not likely to be attained without partial melting occurring first. As we have seen, however $P_{H_2O} < P_{total}$ could move the muscovite stability curve to the left of the melting curve, which would then mean that partial melting would not occur until after the sillimanite-K-feldspar zone had been reached.

Numerous light-colored pods and lenses occur in the Littleton Formation, and could be interpreted either as metamorphic or anatectic segregations. Chemical analyses and CIPW norms of three such segregations are presented in Table 6, and their normative Q-Ab-Or compositions are plotted on Figure 31, along with the ternary minimum of the granite system (Tuttle and Bowen, 1958). The segregations do not fall near the minimum, and therefore are considered to be not of anatectic origin, but to be metamorphic segregations. This is also to be inferred from the high normative corundum in the three rocks.

Because anatexis has not taken place, we can conclude that the regional metamorphism in the Holderness Quadrangle never attained temperatures sufficiently high (approximately 675°C) for anatexis to occur, or that $P_{H_2O}$ in the rocks was low at the time of metamorphism, or that there was a conjunction of both factors.

## Relative Age of Metamorphism

The occurrence of large sillimanite crystals oriented parallel to $F_1$-$F_2$ fold axes suggests that the metamorphism may have occurred prior to or during the period of $D_1$-$D_2$ deformation. An alternate explanation is that subsequent to deformation, $F_1$ and/or $F_2$ fold axes provided lines of weakness in the rock along which the sillimanite needles grew preferentially. The concentric mineral zones around isolated calc-silicate boudins indicate that metamorphism also took place after the $D_1$-$D_2$ deformation. Also, the garnets of the Littleton schists are free from the spiral trains of inclusions often found when garnets are growing during deformations.

A consistent picture develops if we infer that rapid tectonic burial and deformation of sediments occurred during $D_1$-$D_2$. Simultaneous intrusions of the earlier phases of the New Hampshire Plutonic Series caused rapid initial heating and a longer period during which high temperatures were maintained and the rocks gradually lost $H_2O$ and $CO_2$.

Metamorphic reactions appear to have been essentially complete prior to $D_3$ deformations, because the effects of this deformation are primarily mechanical. For example, biotites parallel to $S_1$, are folded and sometimes crenulated by $D_3$, but have not recrystallized into a strong $S_3$ foliation.

Table 6, Chemical Analyses and CIPW Norms of Three Segregation Pods

Chemical Analyses

| Sample | 30-71 | 6-71 | 18-71 |
|---|---|---|---|
| $SiO_2$ | 69.36 | 75.28 | 62.08 |
| $TiO_2$ | 0.12 | 0.31 | 0.85 |
| $Al_2O_3$ | 15.73 | 13.97 | 17.89 |
| $Fe_2O_3$ | nd[1] | nd | nd |
| FeO | 2.02 | 1.36 | 7.79 |
| MgO | 0.76 | 0.40 | 2.29 |
| CaO | 0.89 | 1.05 | 1.36 |
| $Na_2O$ | 2.21 | 1.95 | 1.74 |
| $K_2O$ | 6.63 | 2.58 | 2.49 |
| $H_2O-$ | 0.15 | tr[2] | tr |
| $H_2O+$ | 0.28 | 0.99 | 0.57 |
| Additional[3] | 0.83 | 0.91 | 0.29 |
| Total | 98.98 | 98.80 | 97.35 |

CIPW Norms

| | | | |
|---|---|---|---|
| Qu | 30.22 | 26.52 | 50.33 |
| Or | 14.73 | 39.2 | 15.23 |
| Ab | 14.72 | 18.65 | 16.51 |
| An | 6.76 | 4.42 | 5.2 |
| Di | 0.00 | 0.00 | 0.00 |
| Hy | 18.6 | 5.41 | 2.98 |
| Il | 1.61 | 0.23 | 0.59 |
| Mt | 0.00 | 0.00 | 0.00 |
| Cor | 9.84 | 3.28 | 6.06 |
| Other[4] | 0.86 | 1.26 | 1.9 |

1. nd - not determined
2. tr - trace
3. Additional - additional loss on heating at 1400°C.
4. Other - Equals sum of $H_2O-$, $H_2O+$, and Additional.

Table 6

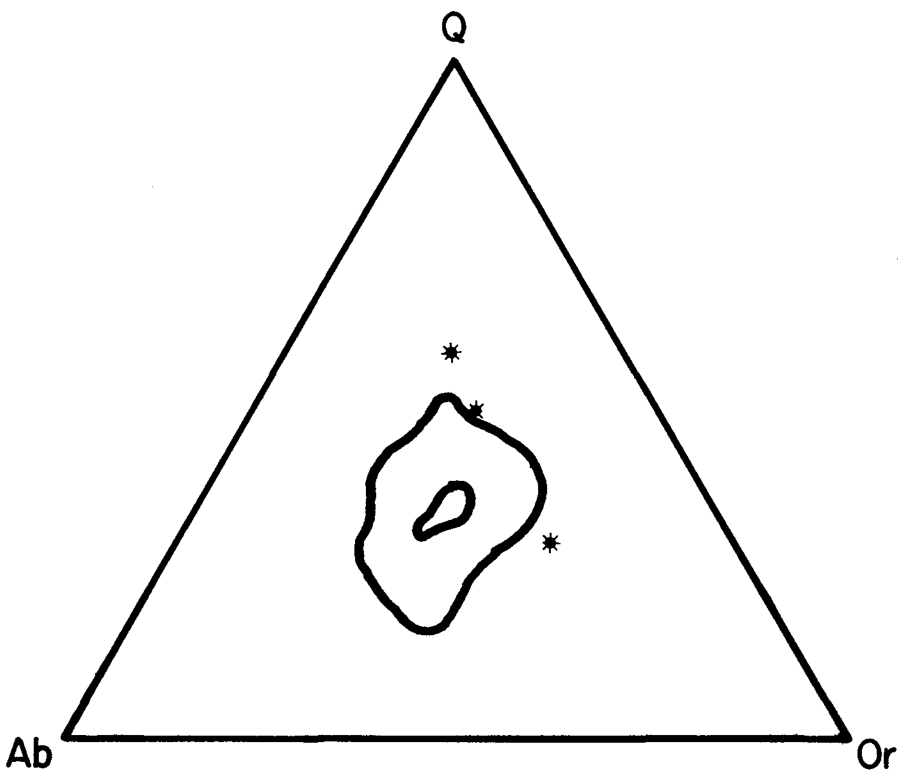



**Figure 31.** Plot of normative Q-Qb-Or of three segregation pods (Table 6). Contours show the ternary minimum for plutonic rocks after Tuttle and Bowen (1958). Contours are 1% and 6% per .25% area contour.

## TECTONIC SYNTHESIS

The recent development of the concept of plate tectonics has provided a remarkably consistent model to explain the distribution and nature of the currently active tectonic areas of the world (Le Pichon, 1968). By using the magnetic anomaly patterns of the sea floor to determine the velocity of plate movements, the separation of Europe and America can be extrapolated back in time to a period about 180 million years ago, when the present Atlantic Ocean began to open. Although direct evidence for plate movements in earlier periods is not available, it may be inferred that old orogenic belts such as the Appalachian-Caledomide belt represent the sites of former convergent plate junctions.

Several authors (Wilson, 1966; Dewey, 1969; Bird and Dewey, 1970) have applied the plate tectonics model to the origin of the Appalachian-Caledonian belt. In its most general form, this model involves the opening and closing of a "Proto-Atlantic Ocean", beginning in the late Precambrian and culminating with the collision of two continental masses in the Late Silurian-Early Devonian.

Figure 32 shows the model proposed by Bird and Dewey (1970) for the tectonic history of the Appalachians from late Precambrian through the Devonian. In terms of the present geography, the boundary between Logan's zone and the piedmont of zone A lies along the Green Mountains — Berkshire Highlands, and the boundary between zones A and B lies somewhat to the east of the Bronson Hill anticlinorium. Thus, the Holderness Quadrangle is probably on or just to the east of the zone A — zone B boundary.

Figure 32 shows the Late Ordovician Taconic orogeny as uplift and compression resulting from the onset of plate convergence. No evidence of the Taconic orogeny is present in the Holderness Quadrangle, except insofar as some of the metasediments may have had their source area in the Taconic highlands to the west. Continuous depositions from the Ordovician through Early Devonian in the Merrimack synclinorium would be consistent with a similar depositional history along strike, in the Aroostook-Matapedia belt of New Brunswick and Maine (Pavlides, Boucot, and Skidmore, 1968).

The Acadian orogeny, which is well represented in the Holderness Quadrangle, is generally consistent with the plate tectonic model. The rapid and deep burial with associated metamorphism and plutonism, (Naylor, 1971) and the complex, intense deformation culminated by compressional folding with a consistent regional trend, can be readily visualized as the result of continental rise sediments caught between two colliding continental plates.

In the Holderness Quadrangle, the most difficult tectonic problem in this or any other model is to explain the northwest-trending folding. The recumbent $F_1$ folds and the upright $F_2$ folds, both having northwest-trending axes, were formed after the deposition of the Littleton Formation. The $F_1$ and $F_2$ folds were formed prior to the upright, northeast-trending $F_3$ folds. As discussed earlier, the entire sequence of events, including the deposition of the Littleton Formation deep burial and metamorphism, and the deformation cycles $D_1$, $D_2$ and $D_3$ must have taken place during a relatively short period of time (Naylor, 1971). While the $D_3$ deformation can be assigned to compression during the final closure of the Proto-Atlantic, clearly the $F_1$ and $F_2$ folds cannot have been formed directly by this same mechanism. It is unlikely that two colliding plates would give rise to horizontal compressive forces acting at right angles to the direction of plate movement.

The recumbent style of the $F_1$ folds suggests a gravitational mechanism for their formation. In order to develop folds at right angles to those formed by the major compression, one can visualize a large mass of sedimentary rocks being uplifted at the point where two continental plates make initial contact. Because real continental outlines are irregular, it is almost certain that collision would not take place simultaneously along an entire continental margin. The uplifted rocks would tend to move downslope and outward in all directions from the point of maximum uplift, but movement would probably be greatest to the sides, into the remaining trough between the two continental plates, where the gravitational potential would be greatest. Movement in the upper levels of the pile would probably take the form of low-angle gravity faulting similar to that described in Maine by Moench (1970), and grading downwards into predominantly recumbent folding as is observed in the Holderness Quadrangle.

The regional pattern to be expected from this model would be a series of roughly concentric arcs, delineated by the axes of $F_1$ and $F_2$ folds, about the area of maximum uplift. Although the available evidence suggests that the $D_1$-$D_2$ deformation was quite widespread throughout much of central New Hampshire and adjacent areas, no systematic study of these structures has yet been made, and the data are not adequate to test the hypothesis.

It is tempting to consider the $D_1$ deformation to be equivalent to the emplacement of the nappes on the Bronson Hill anticlinorium. Both are recumbent, and occurred after the deposition of the Littleton Formation and prior to the regional northeast-trending $D_3$ deformation. If the hinges of the nappes generally trend northeasterly, however, as indicated by Thompson et. al. (1968) a genetic relationship between the nappe formation and $F_1$ folding in the Holderness Quadrangle is not evident. A possible solution is suggested by the folding model proposed by Fyson (1971), in which early folding of sedimentary rocks at depth is primarily recumbent. During recumbent folding, the rocks become

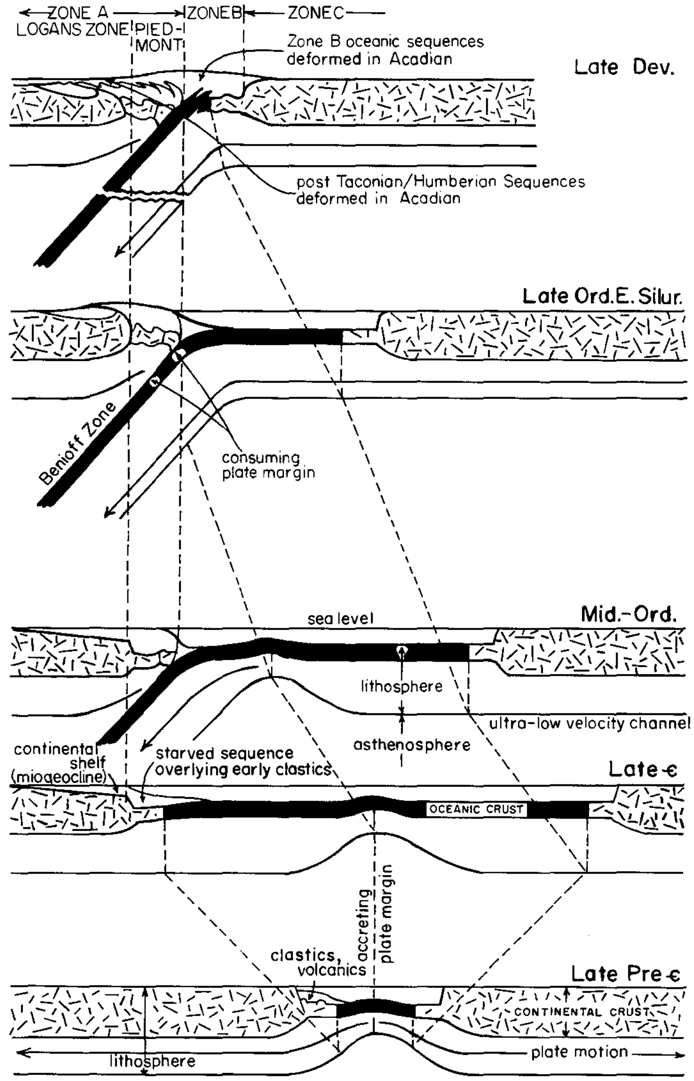



**Figure 32.** Suggested plate evolution of the Appalachian Atlantic and relationships of plate consumption to the Taconic and Acadian orogenies. The positions of plate margins are relative. The position of the Holderness Quadrangle relative to these sections would be on or near the Zone A — Zone B boundary (after Bird and Dewey, 1970).

"welded" into a more homogeneous mass, and subsequent deformation will result in primarily upright folding. Using this model, the following sequence of Acadian deformation can be proposed for central and western New Hampshire:

1) A large mass of material was uplifted at the initial point of contact between two colliding continental plates. This point was either northeast or southwest of the present location of the Holderness Quadrangle.

2) Gravitational forces caused the uplifted material to flow into the gap remaining between the two plates. At the depths which are now exposed in the Holderness Quadrangle, the deformation occurred primarily as recumbent folds ($F_1$). Recumbent folding continued until "welding" was completed and/or until horizontal movement was stopped by a buttress, at which point the folding became dominantly upright ($F_2$).

3) Continued closure of the continental plates moved the "welded" mass of recumbently folded rocks toward the west-northwest, into their present location, and at the same time greatly shortened the previously undeformed section. This shortening was accomplished at first by the emplacement of large-scale nappes, while the final stages of compression formed upright folds ($F_3$) over the entire region. The remobilization and emplacement of the gneiss domes of the Bronson Hill anticlinorium also occurred during or after the $F_3$ folding (Thompson, et. al. 1968).

The junction of the two continental plates in the New England region was completed with the $D_3$ deformation. The Appalachian-Caledonian Mountains had attained their maximum elevation; perhaps at the same stage in their history as the Himalayas are today. Their subsequent Paleozoic history was dominated by erosion and corresponding isostatic uplift.

In the Mesozoic, the opening of the present Atlantic Ocean began. It seems likely that most of the post-Acadian activity, such as the Triassic normal faulting and volcanism, the plutonism and volcanism of the White Mountain Plutonic-Volcanic series, and the Late Jurassic intrusion of lamprophyre dikes, was a direct or indirect result of the tensional forces associated with the opening of the Atlantic.

## SUMMARY and CONCLUSIONS

Sediments of the Littleton Formation were subjected to a complex process of burial, deformation, metamorphism and intrusion during the Acadian orogeny, which is currently considered to be the result of the collision of two continental plates. The initial deformation involved recumbent isoclinal folding with northwesterly trending axes, producing an approximately horizontal axial plane foliation parallel to bedding. This foliation and the bedding were subsequently gently folded about northwesterly trending axes and vertical axial planes, ranging from gentle, open folds in the southern part of the Holderness Quadrangle to isoclinal folds in the northwest.

The Winnipesaukee pluton of the New Hampshire Plutonic Series is of magmatic origin, emplaced as a large, relatively thin sill or sheet into the Littleton Formation. Emplacement occurred during the initial recumbent folding, and the contacts are approximately concordant with the foliation of the schists. A series of superimposed sills of this type could have provided the immediate source of heat required for the sillimanite grade metamorphism in central New Hampshire, although the source of the magma itself is unknown. The rocks of the Winnipesaukee pluton were themselves deformed by the late stage of folding, resulting in the development of axial plane foliation within the Kinsman Quartz Monzonite and Winnipesaukee Quartz Diorite. Intrusion of binary quartz monzonites, also of the New Hampshire Plutonic-Volcanic series, occurred in the area during and after the final deformation. The White Mountain Plutonic-Volcanic Series is represented in the Holderness Quadrangle only by several small lamprophyre dikes of Late Jurassic age.

The recognition of the early period of northwesterly trending recumbent folding is potentially of great significance in interpreting the geologic history of New Hampshire. If this deformation is found to be regional in extent, it will require a major reexamination of the structural interpretation of the Merrimack synclinorium. A regional northwesterly trending deformation would also considerably restrict the number of possible tectonic models for the Acadian orogeny.

# APPENDIX

## Computer Program for Calculating Gravity Profiles

```
90   COMPUTES GRAVITATIONAL EFFECT OF UP TO 10 2-DIMEN-
     SIONAL BODIES,
91   EACH APPROXIMATED BY A POLYGON OF UP TO 99 VERTI-
     CES.
92   COORDINATES MUST BE GIVEN IN KILOMETERS.
100  DATA START IN LINE 1000 — GIVE NUMBER OF STATIONS,
110  X COORDINATE FOR EACH STATION; NUMBER OF BODIES.
120  FOR EACH BODY — GIVE DENSITY CONTRAST;
130  NUMBER OF VERTICES; X & Z COORDINATES FOR EACH
     VERTEX,
140  CLOCKWISE, LISTING FIRST VERTEX AGAIN AS LAST VERTEX.
141  FULL SCALE FOR PLOT IS 35 MGALS. IF ABS (MAX ANOMALY)
     IS GREATER
142  OR MUCH LESS THAN 35 MGALS, ADJUST SCALE IN LINE 905
     (E.G.,
143  IF ABS (MAX ANOMALY) = 66 MGALS, LINE 905 SHOULD
     READ "LET V = V/2")
144
145
146
150  PRINT "DISTANCE"
160  PRINT "FROM ORIGIN", "ANOMALY"
170  PRINT "IN KM", "IN MGAL"
180  PRINT
181  PRINT
184  PRINT
185  PRINT
190  LET K = 6.67
200  DIMX(100), T(100), P(100), R(100)
205  DIM G(10,100)
210  READ M'    # OF STATIONS
220  FOR Q = 1 TO M
230  READ P(Q) ' X COORDINATE OF STATION Q
240  NEXT Q
245  READ S'    # OF BODIES
250  FOR S1 = 1 to S
```

```
260 READ D
270 READ N '      # OF VERTICES
280 FOR I = 1 TO N+1
290 READ X(I),Z(I) '      COORDINATES OF VERTEX I
300 NEXT I
310 FOR Q = 1 TO M
320 FOR I = 1 TO N+1
330 LET X(I) = X(I) - P(Q)
340 LET R(I) = X(I)*X(I) + Z(I)*Z(I)
350 IF X(I) = 0 THEN 390
360 IF X(I)<0 THEN 470
370 LET T(I) = ATN (Z(I)/X(I) )
380 GO TO 510
390 IF Z(I)< 0 THEN 430
400 IF Z(I)> 0 THEN 450
410 LET T(I) = 0
420 GO TO 510
430 LET T(I) = -1.5708
440 GO TO 510
450 LET T(I) = 1.5708
460 GO TO 510
470 IF Z(I)<0 THEN 500
480 LET T(I) = ATN(Z(I)/X(I) + 3.1415927
490 GO TO 510
500 LET T(I) = ATN (Z(I)/X(I) ) - 3.1415927
510 LET X(I) = X(I) + P(Q)
520 NEXT I
530 FOR J = 1 TO N
540 LET T = T(J+1)
550 IF T< -3.14159 THEN 580
560 IF T> 3.14159 THEN 600
570 GO TO 610
580 LET T = T +6.28318
590 GO TO 610
600 LET T = T -6.28318
610 LET X(J) = X(J) - P(Q)
620 LET X(J+1) - P(Q)
630 LET B=X(J+1)-X(J)
640 LET C=Z(J+1)-Z(J)
650 LET H = X(J)*Z(J+1) - Z(J)*X(J+1)
660 LET E = H/(B*B+C*C)
670 LET E1=B*T
680 IF R(J+1)=0 THEN 700
690 GO TO 720
700 LET R = 10E-25
710 GO TO 760
720 IF R(J)=0 THEN 750
730 LET R = R(J+1)/R(J)
740 GO TO 760
```

```
750  LET R=10 E25
760  LET E2 = .5 * C * LOG(R)
770  LET F = E*(E1+E2)
780  LET G=G+F
790  LET X(J) = X(J) + P(Q)
800  LET X(J+1) = X(J+1) + P(Q)
810  NEXT J
820  LET G(S1,Q) = G*2*K*D
830  LET G = 0
840  NEXT Q
850  NEXT S1
860  FOR Q=1 TO M
870  FOR S1 = 1 to S
880  LET V = V + G(S1,Q)
890  NEXT S1
900  PRINT P(Q), INT(10*V+.5)/10
905  LET V=V
910  PRINT TAB (INT(V+35.5);"*"
915  LET V=0
920  NEXT Q
1000 DATA 13, -2, 0, 4, 6, 8, 10, 12, 14, 16, 18, 20, 22
1100 DATA 1
1200 DATA .2,4
1300 DATA 0, 0, 5, 0, 5, 2, 0, 2, 0, 0
1400 DATA -5,0,1,0,1,2,-5,2,-5,0
1500 DATA -.05
1600 DATA 6
1700 DATA 9.3,0,10.6,1,14.6,2,8,2,8,.4,9.3,0
1800 DATA -1,5
1900 DATA 10.6,0,40,0,40,3,20,3,10.6,1,10.6,0
9000 END
```

## Field Methods

Geological mapping of the Holderness Quadrangle was carried out during the summers of 1969-70-71. Initial reconnaissance mapping in 1969 was done mainly along the extensive network of roads in the area. Subsequently, pace and compass traverses were made through the heavily wooded areas, as well as traverses along ridges and stream beds. Short, closely spaced traverses were generally made to define contacts, as attempts to follow contacts resulted in considerable difficulty in locating positions accurately on the map. Average spacing between adjacent traverses was approximately ½ mile or less. The traverse density was variable, concentrating along contacts or other areas of interest noted during the initial reconnaissance.

Mapping was done on a 15′ topographic map enlarged to 1:20,000 scale. Location of outcrop positions was generally done by a combination of pace and compass, and topography. Aerial photographs were available, but were not effective due to the extensive forest cover.

Contacts shown on the map as "accurate" were located in the field to within about 20 feet; while "approximate" contacts were located to within 300 feet.

Exposed bedrock comprises approximately 1% of the map area. Outcrop areas are generally small and well scattered; it is rare to walk a mile without encountering any outcrops. Sampling for petrographic and chemical analysis was done on a selective, rather than systematic basis. Because of this, atypical rocks were more heavily represented in the sample suites. However, attempts were made to make each sample as representative of its outcrop or rock type as possible.

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