Illinois Academy of Science Transactions, Vol. 46, 1953

PROBLEM OF INVERTED ZONES OF METAMORPHISM IN WESTERN NEW HAMPSHIRE

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The general geology of western New Hampshire and neighboring areas (fig. 1) will first be described. Though this description is brief and somewhat incomplete, it will be found adequate for an understanding of the present paper. The stratigraphic details are presented in table 1.

East of the Northey Hill thrust (shown extending N-S across the entire map) is a belt of mantled gneissic domes. These Oliverian domes lie along the crest of a major structural feature, the Bronson Hill anticline. Each dome consists of a central core of feldspar gneiss surrounded or mantled by metamorphosed stratified rocks of Ordovician (?) age or younger. In general the gneissic cores are both conformable and concordant with bedding and schistosity of the adjacent metamorphic rocks (schist, quartzite, and amphibolite), and dips are radially outward.

East of the Bronson Hill anticline are mostly schist and gneiss of Devonian age (Littleton formation). Two large masses of feldspathic gneiss known as the Bethlehem gneiss and Kinsman quartz monzonite are shown on the map. These masses are conformable and in general are concordant with the adjacent rocks. They are considered (Chapman, 1952) to represent highly recrystallized and metasomatized portions of the pelitic Littleton formation.

Between the Northey Hill thrust and the Ammonoosuc thrust (to the west) are highly folded metamorphic rocks derived from volcanics and pelitic and sandy material. Structurally this unit is the southern extension of the Salmon Hole Brook syncline (Billings, 1937), but it is interrupted in a few places by small domical features. The most conspicuous of these is the Lebanon dome (5)* with its core of somewhat foliated granite. In some respects the Lebanon dome resembles the Oliverian domes to the east.

Domal structures are conspicuous features in the middle Ordovician and older rocks of eastern Vermont. At the north of the map is the Strafford dome (3) and immediately south of this is the Pomfret dome (4). In the central part is the Chester dome (8), a large and highly complicated structure, with an apparent core of gneiss. To the south are two smaller domes. The one on the west, showing a gneissic core, is the Sadawga dome (11); the eastern one is the Gilford dome (12).

As can be seen from figure 2, the severity of metamorphism in New Hampshire increases from west to

* Numbers and letters in parenthesis refer to points and quadrangles shown in figure 1.

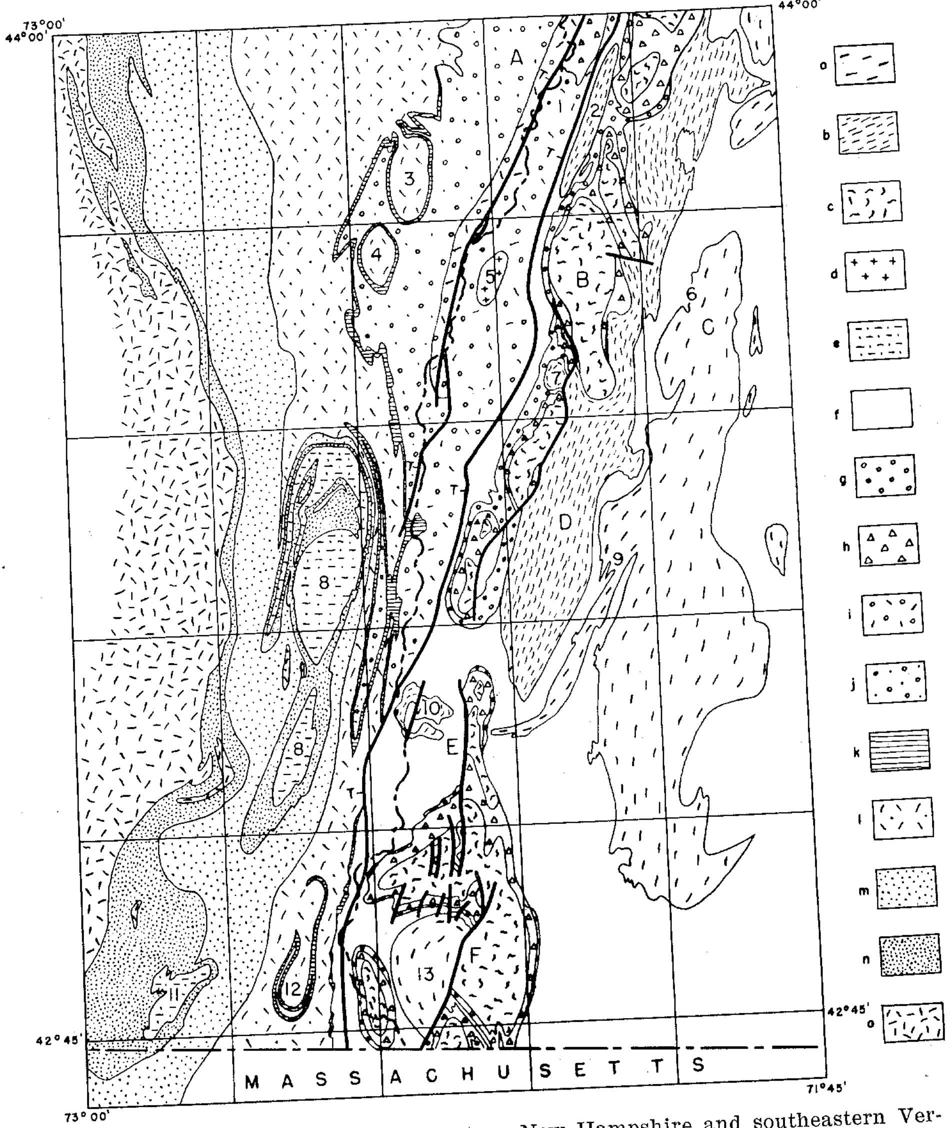

FIG. 1.—Geologic map of southwestern New Hampshire and southeastern Vermont (simplified after Billings, Rodgers, and Thompson).

Key to symbols: a—Kinsman quartz monzonite, b—Bethlehem gneiss, c—Oliverian gneiss, d—Lebanon granite, e—gneiss of Vermont domes, f—Lower Devonian (Littleton formation), g—Silurian (Fitch and Clough formations), h—Upper Ordovician? (Partridge, Ammonoosuc volcanics, and Albee formations), i—Middle or Upper Ordovician? (Orfordville formation), j—Middle Ordovician (Gile Mountain formation), k—Middle Ordovician (Standing Pond volcanics), l—Middle Ordovician (Shaw Mountain, Northfield, and Waits River formations), m—Cambrian or Ordovician (Pinney Hollow, Ottauquechee, Stowe, Moretown, and Cram Hill formations), n—Cambrian or Ordovician (Tyson and Hoosac formations), o—Pre-Cambrian (Green Mountain basement complex).

TABLE 1.—STRATIGRAPHIC DATA FOR WESTERN NEW HAMPSHIRE

Age	Formation	Original composition	Approximate thickness (In feet)
Lower Devonian	LITTLETON	Shales and shaly sands	10,000
Middle Silurian	FITCH	Calcium and magnesium limes shale, and sands	0-700
Middle or Lower Silurian	CLOUGH	Sands and congolomerate	0-1000
UNCONFORMITY			
Upper Ordovician ?	PARTRIDGE	Shale and shaly sands	0-2000
	AMMONOOSUC VOLCANICS	Rhyolitic to basaltic	2000
	ALBEE	Sands and shaly sands	5000
Middle Ordovician ?	ORFORDVILLE	Shales, shaly sands, volcanics	5000

east. In the western part of the state the rocks lie within the biotite zone. To the east the garnet isograd is passed and the rocks may carry garnet. Further east, as we approach the Oliverian domes, staurolite characterizes the next higher grade; and still further east, on the eastern flank of the Bronson Hill anticline (the Oliverian domes), the rocks are in the sillimanite zone.

It has been shown (Chapman, 1952) that in the Sunapee quadrangle (D) the rocks above the sillimanite isograd may be subdivided into three zones: (1) a lower grade zone in which sillimanite and muscovite are stable; (2) an intermediate zone in which microcline develops at the expense of muscovite; and (3) a higher zone in which microcline and garnet develop at the expense of biotite.

Table 2 shows the relation between the zones described above and the facies of the metamorphic rocks in the area. For convenience the zones as shown in the table will be used throughout the rest of this paper.

Relating these metamorphic zones to geologic features we find that the grade of metamorphism increases eastwardly across the Salmon Hole Brook syncline to the Northey Hill thrust. To the east of the thrust is the Bronson Hill anticline with the Oliverian domes almost entirely within the staurolite zone. East of

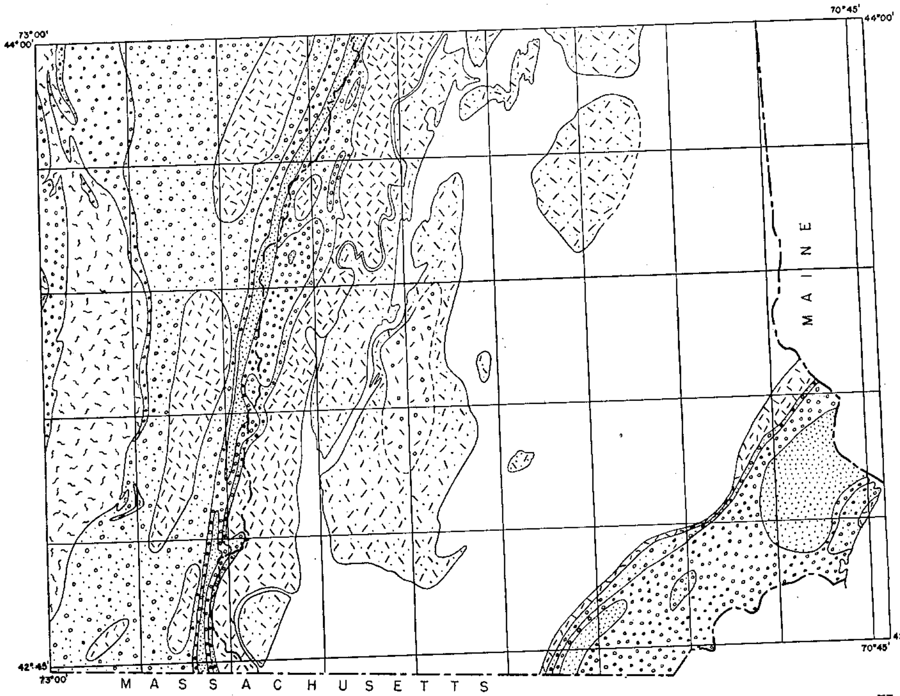

FIG. 2.—Metamorphic zones of southern New Hampshire and southeastern Vermont (modified after Billings, Rodgers, and Thompson).

Key to patterns: vermicular—retrograde zone, stippled—chlorite zone, circles—biotite zone, dots and circles—garnet zone, dashed—staurolite zone, blank—sillimanite zone, dots and dashes—muscovite → microcline zone, circles and dashes—biotite → microcline zone.

the domes is the sillimanite isograd which follows closely the western boundary of the Bethlehem gneiss. This gneiss is considered (Chapman, 1952) to represent a slightly higher grade of metamorphism than the sillimanite rocks to the west. At the east boundary of the Bethlehem gneiss, sillimanite schist and gneiss of the Littleton formation again appear and are of the same metamorphic grade as the sillimanite rocks along the west boundary of the gneiss. In fact we may trace the

TABLE 2.—ZONE AND FACIES CORRELATION FOR WESTERN NEW HAMPSHIRE

Zones	Chlorite	Biotite	Garnet	Staurolite	Sillimanite	Muscovite→ microcline	Biotite→ microcline
Sub-facies	chlorite-muscovite	chlorite-biotite	chloritoid almandine	staurolite-kyanite	sillimanite-almandine		
Facies	greenschist		epidote-amphibolite	amphibolite			Granulite

sillimanite rocks all the way around the south end of the Bethlehem gneiss, more or less isolating this higher grade feldspathic gneiss on the map. Upon passing eastward into the Kinsman quartz monzonite (a gneiss closely resembling the Bethlehem gneiss), the grade again rises to that comparable with the Bethlehem gneiss and then finally to a slightly higher grade in the eastern part of the area.

Considering the relations from a three-dimensional viewpoint, it appears that west of the Oliverian domes (Bronson Hill anticline) the metamorphic zones are superimposed so that the lower grade zones overlie successively higher grade zones, and the zonal boundaries or isograds are inclined to the west. To the east of the Oliverian domes it can be shown in the field that the isograds dip eastwardly at moderate angles, giving a reverse relationship in zone superposition. Evidence for and the significance of this zonal arrangement will now be considered.

At many places in the field the Bethlehem gneiss rests upon rocks of lower grade metamorphism. Particularly striking examples have been described in the Mascoma quadrangle (B) (Chapman, 1939), Bellows Falls quadrangle (E) (Kruger, 1946), and the Sunapee quadrangle (D) (Chapman, 1952). In figure 1 the large masses of Bethlehem gneiss near the towns of Grantham (7) and Bellows Falls (10) are westward extensions of the main mass. They are basin-like structures and overlie slightly lower grade metamorphic rocks. Not only are these structural relations clearly established by geological field studies, but seismological studies (Kruger and Linehan, 1941) show that these basin-like masses have floors at the same depths, as extrapolated from geologic data, and that the underlying rocks may be correlated with a lower metamorphic grade. It seems very reasonable to assume, therefore, that these two masses of Bethlehem gneiss form westward projections from the main body, which extended completely over the Bronson Hill anticline.

Further north in the Mt. Cube quadrangle (A) is an elongate mass of Bethlehem gneiss known as the Indian Pond pluton (2) (Hadley, 1942). This mass may also represent a remnant of a once continuous layer of Bethlehem gneiss. The same is probably true of the Green Mountain pluton of the Moosilauke quadrangle (Billings, 1937) further north.

It seems quite clear that, in the region mapped by the writer, the Bethlehem gneiss represents a zone of higher-grade metamorphism which overlies rocks of lower grade. The zonal boundary which will be referred to as the muscovite→microcline isograd is believed, therefore, to incline eastward at a low to moderate angle.

If we investigate the eastern boundary of the Bethlehem gneiss, we find that it also dips eastward at a moderate angle, but here the metamorphic zone structurally above is of a lower grade (sillimanite zone). The attitude of this dipping contact at numerous localities has been considered in more detail by Fowler-Lunn and Kingsley (1937), Chapman (1939 and 1952), and Heald (1950).

It appears then that the Bethlehem gneiss is a tabular mass, belonging to the muscovite → microcline zone, for the most part underlain and overlain by the sillimanite zone.

In the Cardigan quadrangle (C) (Fowler-Lunn and Kingsley, 1937) the Kinsman quartz monzonite is structurally above rocks of the sillimanite zone. This relationship is conspicuous on Mt. Cardigan (6) where the contact dips east at a low angle. The same relation holds for the Sunapee quadrangle (D) except that the contact is steeper. One striking feature should be noted here, however. The lower boundary of the Kinsman quartz monzonite (which is also the approximate boundary of the sillimanite and muscovite → microcline zones) is very irregular. On Sunapee Mountain (9) the two rock types show interfingering relations so that in places the sillimanite rocks are structurally above and in places structurally below rocks of the muscovite → microcline zone. On the grand scale, however, the muscovite → microcline zone (Kinsman quartz monzonite) rests on the sillimanite zone, and the zonal boundary between the two grades inclines at a high angle to the east.

In the extreme eastern part of the Sunapee quadrangle (D) (Chapman, 1952), Kinsman quartz monzonite of the muscovite → microcline zone passes eastwardly into Kinsman quartz monzonite of the biotite → microcline zone. This highest isograd is only approximately located and information as to its attitude is lacking. It seems likely, however, that the dip of the isograd is very steep.

One further striking example of the muscovite → microcline zone overlying the sillimanite zone is seen in the Keene quadrangle (F) (Moore, 1949) where a mass of Kinsman quartz monzonite (13), about 10 miles long, occupies the center of a large structural basin surrounded by several Oliverian domes.

Theoretical Considerations

The long prevailing idea that regional metamorphism is due to increased temperature and pressure is associated with the concept that grade of metamorphism is in part controlled by depth. Such a line of thinking has led to the idea that in the normal case the grade of metamorphism should increase with depth, and that there is such a thing as a "normal sequence" of metamorphic zones in which a lower grade zone overlies a higher grade zone. Tilley (1925) concluded that the metamorphic zones along the Highland Border in Scotland were not in this "normal" order as Barrow (1912) formerly believed, but that they formed an "inverted sequence." Tilley visualized the zones as being locally overturned mechanically subsequent to the establishment of a "normal sequence." Read (1940), however, pictures the formation of "inverted" sequences not so fortuitously. In fact he raises the question as to whether there is necessarily any normal relationship between metamorphic grade and depth. To him there is little justification for establishing a normal and an inverted sequence. The writer feels, however, that the terms *normal* and *inverted* are helpful when used purely arbitrarily to denote orders of zonal arrangement, rather than to express a standard relation in contrast to an irregular one. In this paper, therefore, the "normal" arrangement of zones is merely one in which the lower grades of metamorphism overlie higher grades, and the "inverted" arrangement expresses the reverse order. Thus no genetic significance is attached to these terms.

The value of determining the attitude of metamorphic zones lies in the light which zonal arrangement throws on the causes of metamorphism.

In the area of western New Hampshire already described, it appears that inverted zonal arrangements are about as common as normal sequences. The Bethlehem gneiss and easterly bordering schists show a normal order, but the same gneiss and the westerly bordering rocks present a clear case of inverted order. The Kinsman quartz monzonite gneiss and the westerly bordering rocks are arranged on a grand scale in the reverse order. In detail, however, several alternations of rock types are noted on Sunapee Mountain (9), with corresponding alternations in the orders of zoning.

West of the Bethlehem gneiss, in the area of the Oliverian domes, it has not been possible to establish with certainty the type of zonal superposition.

If we look further west in New Hampshire and even into western Vermont, we find a very close correlation on a broad scale between the larger structural features and the distribution of metamorphic zones on the map. Thompson (Billings, Rodgers, and Thompson, 1952, p. 15) has remarked on the association of high grade rocks (staurolite zone) and certain domes in eastern Vermont. From figures 1 and 2 it is clear that the rocks of the Strafford-Pomfret dome (3, 4) and the Chester (8), Sadawga (11), and Gilford (12) domes are surrounded by a lower grade of metamorphism. In New Hampshire, furthermore, the Lebanon dome (5) and the structural high (1) in the center of the Mt. Cube quadrangle are also surrounded by a lower grade of metamorphism.

This coincidence of structural and metamorphic highs might be taken to indicate a normal arrangement of metamorphic zones. Where upward displacement of the rock formations is revealed by the development of domes and anticlines, exposing cores of stratigraphically lower formations, isograds have also been uparched, so that more deeply buried metamorphic zones (higher grade) now appear at the surface completely surrounded by more shallow zones of lower grade.

An alternative explanation is that the association of metamorphic and structural highs is not due entirely to domal uplift. It seems quite logical to the writer that the cause of doming may be found in the severity of metamorphism. Metamorphism may have proceeded to such a stage that mobilization of more deeply buried rocks caused doming. Again one may, if one prefers, attribute the rise in metamorphic intensity to domal cores of magma. In any event, it seems clear that the driving force of metamorphism in this area has come from more or less directly below and not from above or laterally.

From his work in the Sunapee quadrangle (D) (Chapman, 1952), the writer concluded that the Kinsman quartz monzonite and Bethlehem gneiss were originally pelitic rocks of the Littleton formation, and not too different in compositional make-up from the rocks of that formation shown separately on the map (fig. 1). Through metamorphic and metasomatic changes, the pelitic sediments were converted to biotite-quartz-feldspar gneiss characterized by large porphyroblasts of potash feldspar. There is little evidence that the rocks of the Littleton formation originally contained abundant feldspar; and it is believed, therefore, that some sodium and perhaps some potassium were added to form sodic plagioclase and microcline, although most of the potash feldspar must have formed at the expense of potential muscovite.

It is generally agreed by the numerous investigators who have studied the Bethlehem gneiss in detail that it is a tabular body which on the grand scale is concordant as well as conformable with the adjacent non-feldspathic schist and gneiss. The same relations are believed to hold generally for the Kinsman quartz monzonite. If these bodies of feldspar gneiss are the result of recrystallization and metasomatism, it appears that they developed in general along certain beds and preferred layers or zones so as to form an intertonguing relation with non-feldspathic rocks, and that this uneven advance of metamorphism (and metasomatism) has resulted in isograds or zone boundaries which are also irregular and complex. It is to be expected that, in view of this irregularity, pockets of rocks of lower grade metamorphism may be isolated by rocks of higher grade metamorphism and vice versa. Inclusion-like masses of schist in the feldspar gneiss and lens-like bodies of feldspar gneiss in schist may represent such isolated pockets.

In an attempt to restore the isograds in western New Hampshire, the writer has arrived at the following picture. On the broad regional scale the metamorphic zones are arranged in a normal order and incline in a westerly direction. In detail, however, the isograds are highly irregular; and from surface mapping one is impressed by the rapid and frequent changes from normal to inverted order of zones. From figure 3 it is seen that this alternating arrangement is a logical consequence of the tongue-like projections of higher grade zones into lower and vice versa. The metamorphic potential in western New Hampshire apparently decreased in an upward and westerly direction. It seems that the driving force of metamorphism must have had a deep seated origin below the central part of the state.

The irregularity of metamorphic advance is believed to be controlled in large part by rock composition and structure. It is possible, however, that metamorphic zones have been somewhat deformed by the orogeny to the extent that certain irregularities may be due to flowage, folding, and faulting. It seems reasonable that in the higher grades of metamorphism, where metasomatism is more severe, the isograds would be more irregular, whereas in the lower grades, where large scale movement of material is less likely or movement is more diffuse and quantitatively less important, zones of metamorphism would be separated by smoother isograds.

If this line of reasoning is applied to the southeastern part of New Hampshire (fig. 2), one might assume that the decrease in metamorphic potential toward the coast

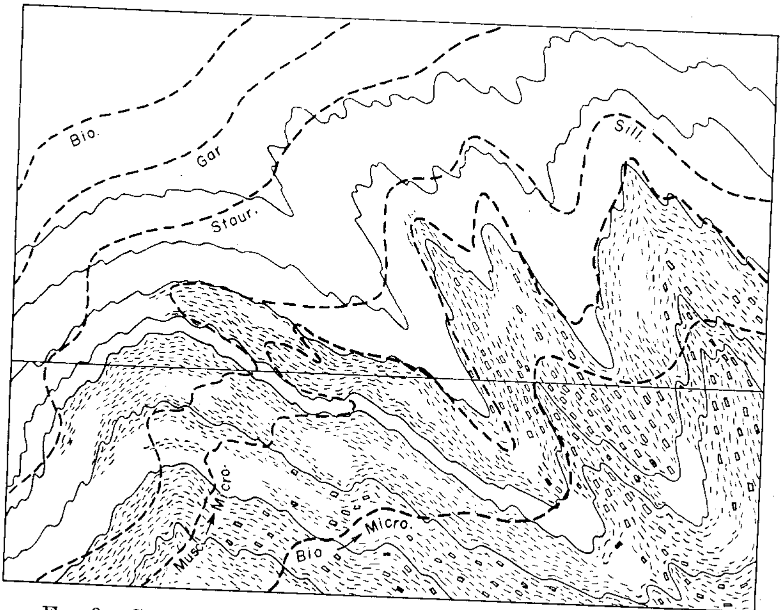

FIG. 3.—Schematic cross section of western New Hampshire, showing folded structure with an Oliverian dome on the west (*left*) and the Kinsman quartz monzonite gneiss on the east (*right*). Horizontal line—present land surface; heavy dashed lines—isograds; dashed pattern—quartzo-feldspathic gneiss.

(southeasterly) is a consequence of the arch of metamorphic zones which culminates in the central part of the state. This gigantic metamorphic anticline, or more properly anticlinorium, may owe part of its arching directly to orogenic movement and isostatic uplift in the central belt.

In conclusion, it appears that the advance of the metamorphic wave into the rocks of western New Hampshire has not come from directly below but obliquely from some depth beneath the central part of the state.

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